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Groundwater Recharge Estimation in Southern Africa

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Preface

Water is life! This has been a slogan of many water awareness campaigns. But the deep truth in these words only emerges as water stress and water scarcity increase. No wonder that the World Water Forum has listed the big water issues in our time as:

- · Meeting basic needs,
- · Securing the food supply, and
- · Protecting ecosystems.

And there has been recognition of another international priority issue:

- · Ensuring the knowledge base.

No sustainable development of a scarce natural resource, and thus of life, is possible without understanding the resource and managing it wisely according to this growing understanding. Therefore I am thrilled that our region has dug deep to put its wisdom on groundwater recharge together to help secure a healthy people and a healthy environment in this water scarce part of the world. In South Africa we like to refer to groundwater as the hidden treasure. It is vital to understand its value and put it to use.

I am pleased that the authors have not stopped at quantification of recharge which merely refers to how much can be abstracted from the resource. If we truly want to manage our scarce water resources in a sustainable way we need to start understanding the surface catchment and the underground aquifer as an interlinked hydrological system and these in turn linked to the water-dependent ecosystems. Science should be ahead of policy to be able to help shape policy and to support its practical implementation. Modern water legislation, like the National Water Act of 1998 in South Africa, strives towards Integrated Water Resources Management. Two key principles underpinning this Act are " that all water, wherever it occurs in the water cycle, is a resource common to all, which shall be subject to national control and "it is necessary to recognise the unity of the water cycle". Without such understanding of the hydrological system we would not be able to move forward into practical implementation from such noble principles.

I realise that behind this summary and integration of knowledge are many years of research effort in several countries in our region. I can visualise much of this being done in the Kalahari under extreme working conditions. There has also been exemplary sustained international support for several of the studies. I would like to salute All these participants and supporters.

This brings me to a concluding remark. No country will in future be able to manage its water resources in isolation, not only because hydrological systems do not follow political boundaries, but also because of the vast international pool of knowledge and technology that will be required to achieve sustainable development. I therefore welcome the UNESCO International Hydrological Programme, the publisher of this book, as an essential mobilizer of good water science across national boundaries into our region.

All this is captured in the isiZulu word

Masibambane! meaning "lets work together"

Ronnie Kasrils.

Ronnie Kasrils, MP
Minister of Water Affairs and Forestry
Republic of South Africa

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This book contains a selection of papers initially presented at various courses and workshops in Southern Africa since 1996, and several papers on new research findings. The following organisers and facilitators are thanked for their input during and after the various courses and workshops held in the past few years, which has formed a basis for this book:

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PART I

Review and Challenges of Recharge Estimation in Southern Africa

1. Review of Groundwater Recharge Estimation in Arid and Semi-Arid Southern Africa

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ABSTRACT A review is presented of groundwater recharge estimation in arid and semi-arid Southern Africa based on three decades of recharge investigations in the region. Principles of methods currently in use are described and references from the Southern African region are given when possible. The methods are reviewed in terms of limitations, applicability at different fluxes, temporal and spatial scales and are rated on accuracy, ease of application and costs. Promising methods for recharge estimation are the Chloride Mass Balance (CMB), Cumulative Rainfall Departure (CRD), Extended model for Aquifer Recharge and moisture Transport through unsaturated Hardrock (EARTH), Water Table Fluctuation (WTF), Groundwater Modelling (GM) and Saturated Volume Fluctuation (SVF) methods. Particularly those methods that are based on relationships between rainfall, abstraction and water level fluctuations such as the CRD and EARTH have the potential to forecast groundwater recharge. The CMB, CRD and GM methods are discussed in more detail.

1.1 Introduction

In arid and semi-arid areas, assessment of groundwater recharge is one of the key challenges in determining the sustainable yield of aquifers as recharge rates are generally low in comparison with average annual rainfall or evapotranspiration, and thus difficult to determine precisely. For this paper we limit ourselves to groundwater recharge estimation in arid and semi-arid Southern Africa with aridity as defined by Lloyd (1986) on the basis of average annual rainfall: hyper-arid: 0-50 mm/yr; arid: 50-200 mm/yr and semi-arid: 200-500 mm/yr. About 22% of Southern Africa (SADC) falls within the boundaries of aridity as shown in Figure 1.1.

Figure 1.1 Aridity in Southern Africa.

1.1.1 Three decades of recharge studies in Southern Africa

In Southern Africa, most regional and local recharge studies (including groundwater exploration projects) have been carried out in semi-arid Botswana, Namibia and South Africa over the last three decades.

Botswana
The first systematic study of groundwater recharge in Botswana was carried out in 1974 by Jennings in the eastern part of the country in collaboration with researchers of the South African University of Witwatersrand. During the 1970s and 1980s studies were also carried out in the Kalahari (Verhagen et al., 1974; Mazor et al., 1977; Foster et al., 1982; De Vries and Von Hoyer, 1988). In 1987 the co-operation programme Groundwater Resources Monitoring and Recharge Study (GRES) was jointly launched by the Botswana and Netherlands governments aiming at a better understanding of recharge processes in the country. The first phase of GRES concentrated on Precambrian aquifers in south-eastern Botswana and was completed in 1991 (Gieske, 1992). The second phase (GRES II) expanded into the Kalahari Basin and was completed in 1997 (Selaolo, 1998; Beekman et al., 1996; 1999; De Vries et al., 2000). Methods used included analysis of precipitation and evapotranspiration, study of environmental isotopes and rainfall chemistry, and analysis of transport processes in both saturated and unsaturated zones. GRES investigations revealed that considerable recharge of the order of 10 to 50 mm/yr takes place under favourable conditions in the eastern part of Botswana. A decreasing recharge trend was observed from 5 mm/yr in the fringe of the Kalahari to less than 1 mm/yr towards the central part. Lower recharge rates, or even hardly any recharge at all, may be expected where rainfall drops below 400 mm/yr.

Namibia
Despite Namibia being the driest country in Southern Africa, large-scale recharge studies have only been conducted since the nineties. The Namibian and German governments jointly launched a co-operation program in 1992 for recharge studies in the north-eastern part of the country focusing on the karst areas of Otavi Mountain Land (Schmidt and Ploethner, 2000). For the past two decades annual rainfall has been below the long-term mean annual rainfall of 550 mm, resulting in recharge being less than 9 mm/year. For the adjacent Kalahari Catchment to the east, Klock (2001) determined recharge at 1 mm/year. This figure was based on regionalised site-specific hydrochemical data using satellite imagery and was verified by a groundwater model. Recharge in the area may range from 0.2 to more than 100 mm/year. Central Namibia will face an urgent need for additional secure water resources within the next decade and therefore a groundwater investigation was initiated in 1999 in northern Otavi Mountain Land to determine the long-term sustainable abstraction and short-term emergency bulk groundwater abstraction from the promising Tsumeb aquifers (Bufler et al., 2000). Recharge is also being investigated for the Stampriet Artesian Basin in the south-eastern part of Namibia through a joint co-operation program between the Namibian and Japanese governments and the International Atomic Energy Agency (IAEA).

South Africa
First systematic recharge studies carried out in South Africa date back to the early 1970s in the western Transvaal (Bredenkamp and Vogel, 1970; Bredenkamp et al., 1974) and in the northern Cape (Smit, 1978). Recharge studies were mostly carried out at a local scale and as part of a larger groundwater resources assessment project. It was during the international groundwater recharge workshop in Turkey in 1987 that an urgent need was expressed for developing new and improving existing practical methods for recharge estimation in arid and semi-arid areas

(Simmers, I. (ed.), 1988). In South Africa in particular, the growing need for reliable recharge estimation originated from a desire to better (sustainably) manage limited water resources. The Water Research Commission of South Africa therefore initiated the project "Preparation of a Manual on Quantitative Estimation of Groundwater Recharge and Aquifer Storativity". The manual, published in 1995 (Bredenkamp et al., 1995), presents a great variety of well-tested (semi-empirical) methods that are widely employed in South Africa and contains a wealth of recharge case studies and data covering the past 30 years. Of particular interest is that in the manual a first attempt was made at rating different estimation techniques in terms of ease of application, reliability and availability of data.

1.1.2 Overview of results

Figure 1.2 shows all reported recharge values up to 1997, including those from more humid Southern African regions, and mostly from Botswana (Beekman et al., 1996), South Africa (Bredenkamp et al., 1995) and Zimbabwe (Houston, 1988) as a function of annual rainfall. The diagram shows up to a factor of 100 difference in recharge values at the same annual rainfall. The method which has most consistently been applied over the range of annual rainfall illustrated here is the Equal Volume (spring flow) method (modified water balance). The results of other methods, such as the Saturated Volume Fluctuation modelling, mostly fall within this band. The Botswana results fall within the elongated shaded ellipse, clearly showing the trend of decreasing recharge with decreasing rainfall from south-eastern Botswana to the Central Kalahari. The area below the band of the spring flow values indicates mostly results obtained with the river baseflow (hydrograph separation) method. It seems that this method consistently underestimates recharge in this range of rainfall values. The area above the spring flow band indicates some anomalous recharge values determined through chloride profiling in St. Lucia, South Africa (Bredenkamp et al., 1995).

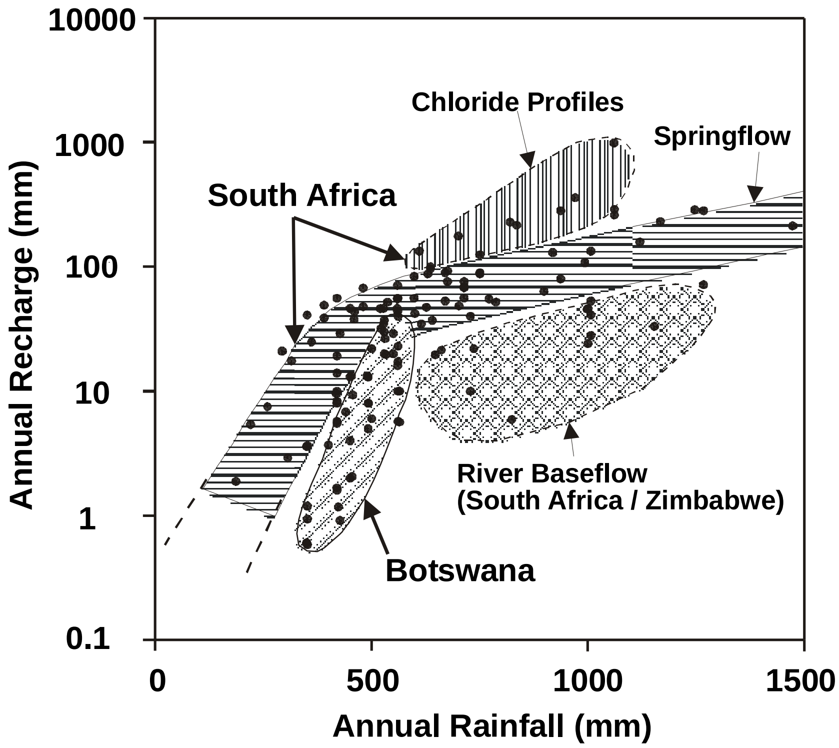


Figure 1.2 Results of recharge studies in Southern Africa (modified after Beekman et al., 1996).

For annual rainfall less than 500 mm/yr large differences exist between values found. Since a wide range of environmental factors, such as vegetation, geology and geomorphology, may have contributed to this discrepancy, a satisfactory explanation cannot be given here. Further work is needed to update the diagrams for studies carried out since 1997.

1.2 Recharge - Definition, concepts and variability

1.2.1 Recharge – Definition and concepts

Groundwater recharge can be defined in a broad sense as “an addition of water to a groundwater reservoir”. Four main modes of recharge can be distinguished:

A. “Downward flow of water through the unsaturated zone reaching the water table”
B. “Lateral and/or vertical inter-aquifer flow”
C. “Induced recharge from nearby surface water bodies resulting from groundwater abstraction” and
D. “Artificial recharge such as from borehole injection or man-made infiltration ponds”

In this paper we focus on the first mode: natural recharge by downward flow of water through the unsaturated zone, which is generally the most important mode of recharge in arid and semi-arid areas. Mechanisms of infiltration and moisture transport that are likely to occur for this mode are illustrated in Figure 1.3. Main sources of recharge are rainfall, surface water bodies (ephemeral or seasonal rivers, lakes, estuaries) and irrigation losses.

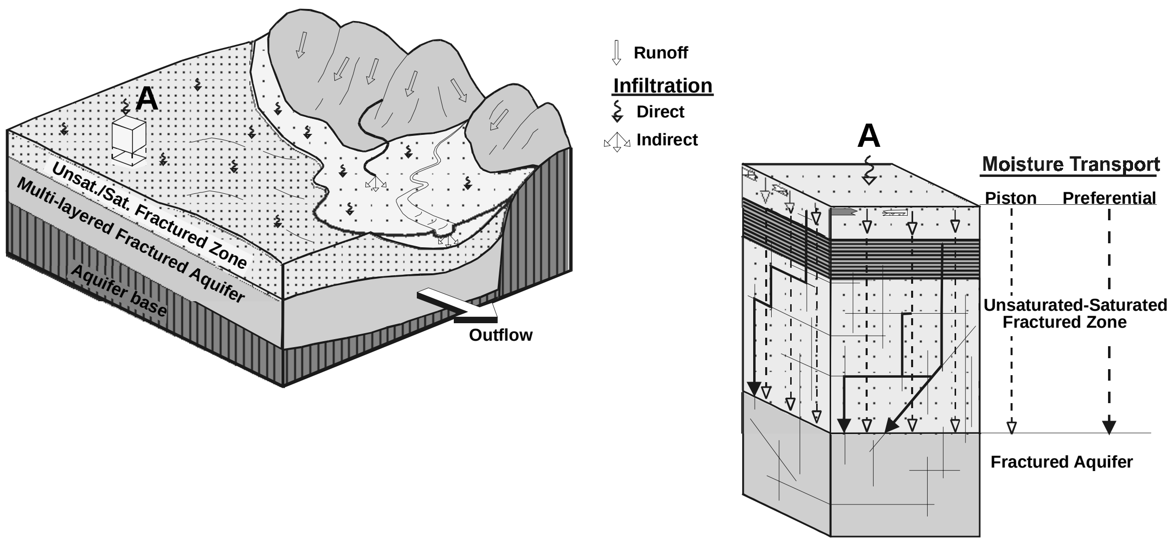


Figure 1.3 Mechanisms of infiltration and moisture transport (after Beekman et al., 1996).

Recharge can be expressed in various forms, e.g. as a percentage of annual rainfall, or in mm/year. It can be classified according to Beekman et al. (1999):

I. origin of water (Lloyd, 1986; Lerner et al., 1990; De Vries and Simmers, 2002):

a. direct / diffuse recharge: direct infiltration of precipitation and subsequent percolation through the unsaturated zone to a groundwater body, i.e. water added to the groundwater reservoir in excess of soil-moisture deficits and evapotranspiration,

b. indirect / non-diffuse recharge: percolation to the water table through riverbeds,
c. localised recharge: accumulation of precipitation in surface water bodies, and subsequently concentrated infiltration and percolation through the unsaturated zone to a groundwater body.

II. flow mechanism through the unsaturated zone:
a. piston / translatory flow: precipitation which is stored in the unsaturated zone, is displaced downwards by the next infiltration / percolation event without disturbance of the moisture distribution,
b. preferential flow: flow via preferred pathways / macro-pores, which are sites (e.g. abandoned root channels, burrows, fissures) or zones (e.g. stream beds) in the unsaturated zone with a relatively high infiltration and / or percolation capacity.

III. area on which it acts:
a. point recharge: recharge at a site, with no areal extent,
b. line recharge: recharge from a line source, such as a drainage feature or river,
c. areal recharge: recharge over an area.

IV. time scale during which it occurs (for both episodic and perennial recharge):
a. present-day recharge: recharge occurring within a time frame of days / months,
b. short-term recharge: recharge covering a short period, in the past or predicted for the near future within a time frame of months / years,
c. long-term recharge: recharge over a longer period, in the past (palaeo-recharge) or predicted for the future (accounting for climate change) within a time frame of tens up to thousands of years.

1.2.2 Recharge - Spatial and temporal variability

Determination of groundwater recharge in arid and semi-arid areas is neither straightforward nor easy. This is a consequence of the time variability of precipitation in arid and semi-arid climates, and spatial variability in soil characteristics, topography, vegetation and land use (Lerner et al., 1990). Moreover, recharge amounts are normally small in comparison with the resolution of the investigation methods. The greater the aridity of the climate, the smaller and potentially more variable is the recharge flux (Allison et al., 1994).

1.3 Guidelines for Recharge Estimation

There are as many methods available for quantifying groundwater recharge as there are different sources and processes of recharge. Each of the methods has its own limitations in terms of applicability and reliability. The objective of the recharge study should be known prior to selection of the appropriate method for quantifying groundwater recharge as this may dictate the required space and time scales of the recharge estimates (Scanlon et al., 2002). Water resource evaluations for instance would require information on recharge at large spatial and temporal scales whereas assessments of aquifer vulnerability to pollution would require more detailed information at local and shorter time scales.

Development of a conceptual model of recharge in the study area should also precede selection of the appropriate recharge estimation method in order to reduce both uncertainty as well as costs of quantifying recharge. Such a model should describe the location, timing and probable mechanisms of recharge and provide initial estimates of recharge rates based on climatic, topographic, land use and land cover, soil and vegetation types, geomorphologic and (hydro-) geologic data (including recharge sources, flow mechanisms, piezometry, groundwater exploitation, etc.). Some guidelines for recharge estimation are given in Lerner et al. (1990) and Scanlon et al. (2002), but a user-friendly framework for recharge estimation does not yet exist.

1.4 Overview of Recharge Estimation Methods

1.4.1 Classification of recharge estimation methods

Recharge estimation methods can be classified according to:

- *Hydrogeological provinces*: regions of similar climate and geology with similar geomorphologic history (e.g. alluvial fans and riverbeds, sand and sandstone, volcanic, etc.; Lerner et al., 1990),
- *Hydrologic zones*: atmosphere, surface water, unsaturated and saturated zones (Bredenkamp et al., 1995; Beekman et al., 1999; Scanlon et al., 2002), or
- *Physical and Tracer approaches*: direct versus indirect, water balance and Darcyan physical methods and chemical, isotopic and gaseous tracer methods (Lerner et al., 1990; Kinzelbach et al., 2002).

The ideal classification accommodates for all above-mentioned criteria.

1.4.2 Commonly used methods

An overview of commonly used recharge estimation methods in Southern Africa is given in Table 1.1. The methods are grouped according to hydrologic zones and further sub-divided into physical and tracer approaches. A brief description of the principle and references is given for each method. Methods referring to surface water and unsaturated zones estimate potential recharge whereas methods referring to the saturated zone estimate actual recharge. A review of commonly used methods is given in the section 1.5. Methods excluded from the overview and review due to either a too qualitative nature, large inaccuracy or a too complicated nature for application in the (semi-)arid environment are the rainfall-recharge relationships, soil-moisture / water budgets (Schulze, 1995), seepage meter, applied and heat tracers and (semi-)quantitative methods which involve ^{2}H, ^{18}O (Beekman et al., 1996) and ^{4}He (Selaolo, 1998). To our knowledge ^{36}Cl has not yet been applied in the region for recharge estimation.

Examples of integrated approaches, i.e. combining various methods, are the:

- “Combined chemical and isotope mass balance approach” (Beekman et al., 1999), and
- “Recharge” Excel spreadsheet model (Van Tonder and Xu, 2000)

The combined chemical and isotope mass balance approach is based on dating moisture and groundwater using the Chloride Mass Balance and ^{14}C groundwater dating methods.
The “Recharge” spreadsheet model enables analysis of hydrogeological data by commonly used estimation methods from Table 1.1 and gives an opportunity to calculate a weighted average recharge rate after having assigned weighting factors to each of the methods used.

A semi-quantitative approach is currently being applied to crystalline basement aquifers of Central Namaqualand in South Africa to define the recharge potential (Adams, pers. comm.). The approach is based on integrating spatial climatic and (hydro-)geologic datasets in a GIS environment and can be considered a derivative of the DRASTIC approach (Aller et al., 1987), which is used for aquifer vulnerability mapping. The approach has the potential to become quantitative once it is combined with recharge estimation methods of Table 1.1.

Table 1.1 Recharge estimation methods applied in (semi-)arid Southern Africa.

Zone	Approach	Method	Principle	References
Surface Water	Physical	**HS**	Stream hydrograph separation: outflow, evapotranspiration and abstraction balances recharge	10
Surface Water	Physical	**CWB**	Recharge derived from difference in flow upstream and downstream accounting for evapotranspiration, in- and outflow and channel storage change	4
Surface Water	Physical	**WM**	Numerical rainfall-runoff modelling; recharge estimated as a residual term	5
Unsaturated	Physical	**Lysimeter**	Drainage proportional to moisture flux / recharge	2
Unsaturated	Physical	**UFM**	Unsaturated flow simulation e.g. by using numerical solutions to Richards equation	2, 4
Unsaturated	Physical	**ZFP**	Soil moisture storage changes below ZFP (zero vertical hydraulic gradient) proportional to moisture flux / recharge	2, 3, 6
Unsaturated	Tracer	**CMB**	Chloride Mass Balance – Profiling: drainage inversely proportional to Cl in pore water	1, 2, 3, 6
Unsaturated	Tracer	**Historical**	Vertical distribution of tracer as a result of activities in the past (^3H)	1, 2, 3, 6
Saturated – Unsaturated	Physical	**CRD**	Water level response from recharge proportional to cumulative rainfall departure	2, 9
Saturated – Unsaturated	Physical	**EARTH**	Lumped distributed model simulating water level fluctuations by coupling climatic, soil moisture and groundwater level data	3, 7
Saturated – Unsaturated	Physical	**WTF**	Water level response proportional to recharge / discharge	2
Saturated – Unsaturated	Tracer	**CMB**	Amount of Cl into the system balanced by amount of Cl out of the system for negligible surface runoff / runon	1, 2, 3, 6
Saturated	Physical	**GM**	Recharge inversely derived from numerical modeling groundwater flow and calibrating on hydraulic heads / groundwater ages	2, 3
Saturated	Physical	**SVF**	Water balance over time based on averaged groundwater levels from monitoring boreholes	2
Saturated	Physical	**EV-SF**	Water balance at catchment scale	2
Saturated	Tracer	**GD**	Age gradient derived from tracers, inversely proportional to recharge; Recharge unconfined aquifer based on vertical age gradient (^3H, CFCs, $^3H/^3He$); Recharge confined aquifer based on horizontal age gradient (^{14}C)	1, 6, 8

HS: Hydrograph Separation – Baseflow
CWB: Channel Water Budget
WM: Watershed Modelling
UFM: Unsaturated Flow Modelling
ZFP: Zero Flux Plane
CMB: Chloride Mass Balance
CRD: Cumulative Rainfall Departure
EARTH: Extended model for Aquifer Recharge and Moisture Transport through Unsaturated Hardrock
WTF: Water Table Fluctuation
GM: Groundwater modelling
SVF: Saturated Volume Fluctuation
EV-SF: Equal Volume - Spring Flow
GD: Groundwater Dating

[1]Beekman et al., 1996
[2]Bredenkamp et al., 1995
[3]Gieske, 1992
[4]Lerner et al., 1990
[5]Sami and Hughes, 1996
[6]Selaolo, 1998
[7]Van der Lee and Gehrels, 1997
[8]Weaver and Talma, 1999
[9]Xu and Van Tonder, 2001
[10]Xu et al., 2002

1.4.3 Recharge forecasting

Forecasting groundwater recharge has become increasingly important, particularly with regard to the envisaged climate change impacts on Southern Africa's limited water resources (Kirchner, 2003; Cavé et al., 2003). Methods that have great potential to forecast recharge are those that have established relationships between rainfall, abstraction and water level fluctuations, such as the CRD, EARTH, Auto Regression Moving Averages and empirical methods. Critical in reliable forecasting of recharge is the accuracy of forecasting rainfall in terms of frequency of events, quantity and intensity. In Southern Africa there is a wealth of rainfall records, often dating back to the beginning of the previous century and this should form a sound basis for future predictions. Note that the accuracy of forecasting recharge is further complicated by the non-linearity of groundwater resources in their response to rainfall. Forecasting should accommodate for the propagation of uncertainty in input parameters.

1.5 Review of Recharge Estimation Methods

1.5.1 Commonly used methods

A review of commonly used recharge estimation methods in (semi-)arid Southern Africa is presented in Table 1.2. Methods are evaluated in terms of limitations, applicability (range of fluxes, spatial and temporal scales) and ratings (accuracy, ease of application, cost).

Table 1.2 Review of commonly used recharge methods for (semi-)arid Southern Africa.

Zone	Method	Limitations	Applicability[2]			Rating[3]		
			Flux (mm/yr)	Area (km^2)	Time (yrs)	Acc.	Ease	Cost
SW	HS	Ephemeral rivers	400-4000 (0.1-1000)	10^{-4}-1300 (10-1000)	0.3-50 (1-100)	2-3	1-2	1-2
	CWB	Inaccurate flow measurements	100-5000	10^{-3}-10	1d-1yr	2-3	2	3
	WM	Ephemeral rivers	1-400	10^{-1}-$5*10^{5}$	1d-10yr	2	2-3	3
Unsaturated [1]	Lysimeter	Surface runoff	1-500 (0-200)	0.1-30m^2	0.1-6	2	3	3
	UFM	Poorly known relationship hydraulic conductivity - moisture content	20-500	0.1-1m^2	0.1-400	3	2	2
	ZFP	Subsurface heterogeneity; periods of high infiltration	30-500	0.1-1m^2	0.1-6	3	2	2
	CMB	Long-term atmospheric deposition unknown	0.1-300 (0.6-300)	0.1-1m^2	5-10000	2	1	1
	Historical	Poorly known porosity; present ^3H levels almost undetectable	10-50 (10-80)	0.1-1m^2	1.5-50	2-3	2-3	3
Sat. - Unsat.	CRD	Deep (multi-layer) aquifer; sensitive to specific yield (S_y)	(0.1-1000)	(1-1000)	(0.1-20)	1-2	1-2	2
	EARTH	Poorly known S_y	(1-80)	(1-10m^2)	(1-5)	1-2	2	1
	WTF	In/ouflow and S_y usually unknown	5-500	$5*10^{-5}$->10^{-3}	0.1-5	2	1	1
	CMB	Long-term atmospheric deposition unknown	0.1-500	$2*10^{-6}$ – >10^{-2}	5- >10000	2	1	1
Saturated	GM	Time consuming; poorly known transmissivity; sensitive to boundary conditions	(0.1-1000)	(10^{-6}-10^{6})	(1d–20yr)	1-2	3	3
	SVF	Flow-through region; multi-layered aquifers	(0.1-1000)	(1-1000)	(0.1-20)	1-2	1-2	2
	EV-SF	Confined aquifer	(0.1-1000)	(1-100)	(1-100)	1-2	1-2	1-2
	GD	^{14}C, $^3H/^3He$, CFC: poorly known porosity / correction for dead carbon contribution	^{14}C: 1-100 $^3H/^3He$, CFC: 30-1000	^{14}C,$^3H/He^3$, CFC: $2*10^{-6}$->10^{-3}	^{14}C: 200-200000 $^3H/He^3$, CFC: 2-40	3	2-3	3

[1] All methods for estimating fluxes through the unsaturated zone assume diffuse vertical flow whereas in reality flow along preferred pathways is the rule rather than the exception. These methods therefore tend to overestimate the diffuse flux.
[2] Data in brackets are estimates from Southern Africa; Rainfall may be up to 2000 mm/ in a year; other data represent global values and are taken from Scanlon et al. (2002).
[3] Ratings for methods applied to semi-arid Southern Africa.

The aim of rating is to advance an on-going discussion among a wide range of stakeholders on the selection of appropriate methods for recharge estimation. The ratings are based on the authors experience and on ratings given by Bredenkamp et al. (1995), van Tonder and Xu (2000), Kinzelbach et al. (2002) and a recent workshop on the "Framework for recharge estimation in Southern Africa" project (Beekman et al., 2003).

With regard to the applicability of methods, data has been adopted from Scanlon et al. (2002). Regarding ratings, the approach of *accuracy* rating is adopted from Kinzelbach et al. (2002): Class 1: difference from true value within a factor of 2, Class 2: within a factor of 5 and Class 3: within a factor of 10 or more. *Ease of application* is related to data requirements and data availability and is rated from 1: easy to use to 3: difficult to use. *Cost* is rated from 1: inexpensive to 3: expensive.

1.5.2 Promising methods

The following methods can be applied with greater certainty in arid and semi-arid Southern Africa: CMB, CRD, EARTH, WTF, GM and SVF. These methods have in common that they estimate recharge based on linking specific information from the atmosphere, unsaturated and saturated zones. Greater certainty in the results from the GM method may be obtained if groundwater levels and ages are linked. Three of these methods: CMB, CRD and GM are widely applied and will be discussed in more detail. They represent an increasing complexity in their use and data requirements.

Chloride mass balance (CMB)
This method is based on the assumption of conservation of mass between the input of atmospheric chloride and the chloride flux in the subsurface. It can be used for both estimating a moisture flux in the unsaturated zone by means of a profiling technique when diffuse (piston) flow is assumed and for recharge. Comparison of moisture flux and recharge provides insight into the mechanism of recharge. Note that in a wider context, mechanisms of recharge and recharge rates can be considered crucial in the assessment of vulnerability of groundwater resources to pollution.

For a steady state between the chloride flux at the surface and the chloride flux beneath an upper zone where evapotranspiration and mixing of rainfall and pore water takes place and excluding runoff and run-on, a site specific moisture flux can be calculated for the unsaturated zone by (Eriksson and Khunakasem, 1969):

$$R_{sm} = \frac{P * Cl_p + D}{Cl_{sm}} = \frac{TD}{Cl_{sm}} \tag{1}$$

where R_{sm} is the moisture flux (diffuse or slow flow component; mm/yr), P is rainfall (mm/yr), Cl_p and Cl_{sm} are chloride concentrations in rainfall and soil moisture (mg/l), and D is dry chloride deposition ($mgm^{-2}yr^{-1}$). The sum of $P*Cl_p$ and D is also referred to as "Total atmospheric chloride Deposition" (TD) and originates from both precipitation and dry fall out. A better estimate of the moisture flux is obtained from a mass balance which integrates chloride and moisture contents cumulatively (c) over a specific depth interval (Gieske, 1992):

$$R_{sm(c)} = \frac{TD * TM}{TC} \tag{2}$$

where TM is total moisture content (mmm^{-2}) and TC is total chloride content (mgm^{-2}). Simply substituting Cl_{sm} in Eq. (1) for the chloride concentration in groundwater at the water table (Cl_{gw}) gives a total recharge rate R_T:

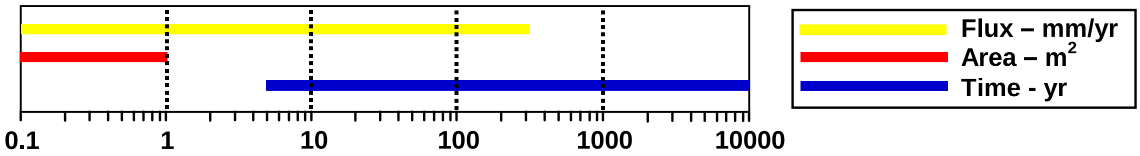
$$R_T = \frac{TD}{Cl_{gw}} \tag{3}$$

Cl_{gw} originates from different flow components in the unsaturated zone. For an areal R_T, Cl_{gw} represents the harmonic mean of chloride concentrations in groundwater. If it is assumed that R_T originates from only two flow components in the unsaturated zone which are fully mixed in the groundwater: a (slow) diffuse (R_{sm}) and a (quick) preferential (R_{pr}) flow component and $Cl_{pr} \ll Cl_{gw} < Cl_{sm}$, the relative contribution of either R_{sm} or R_{pr} to R_T expressed as a fraction is calculated as (Sharma and Hughes, 1985):

$$f(R_{pr}) = 1 - f(R_{sm}) \approx \left(\frac{Cl_{sm} - Cl_{gw}}{Cl_{sm}} \right) \tag{4}$$

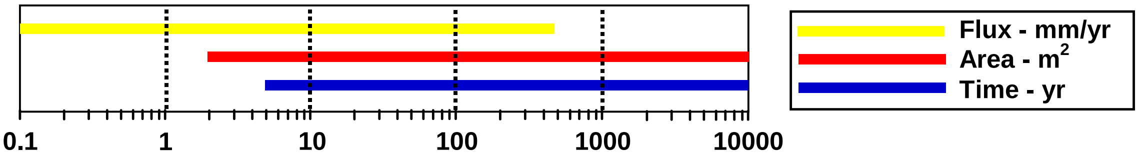
Applicability

Unsaturated Zone: Moisture Flux



Most reliable estimates of site-specific moisture fluxes may be obtained through a multiple tracer profiling approach (Simmers et al., 1997). This approach aims at deducing and quantifying where possible relevant transport processes occurring in the unsaturated zone. For example, the CMB method may reveal the thickness of the evapotranspiration zone and moisture fluxes, ^{18}O and ^{2}H profiling may provide insight into the evaporation process and moisture fluxes (Beekman et al., 1996) and ^{3}H profiling may highlight zones of preferred pathways (thereby (in)validating the use of the various methods; see Selaolo et al., 2003).

Saturated Zone: Recharge



The CMB method for the saturated zone may be especially useful in areas where groundwater levels do not fluctuate or data on groundwater levels are lacking.

Limitations

For the unsaturated zone preferential flow seems to be more the rule than the exception. Moisture fluxes may therefore be overestimated. The CMB method should not be applied in areas underlain by evaporates or areas where upconing or mixing of saline (ground) waters occurs. The method should be applied with great caution in areas close to the sea where rainfall chloride contents are highly variable.

In fractured rock systems, the applicability of the CMB method is complicated if (1) additional chloride is produced through weathering of the rock matrix and when (2) time is needed to develop a new equilibrium between groundwater chloride concentrations in the rock matrix and fractures following a change in environmental conditions (Cook, 2003). If additional chloride is being produced, a recharge rate derived from a CMB should be considered a minimum. In the case of a larger fracture spacing it takes longer to develop a new equilibrium in chloride concentrations. The estimated recharge may therefore not represent changed environmental conditions (e.g. change in recharge due to climate or land-use change).

Data requirements
Long-term averages of P, Cl_p and D; Moisture flux: Cl_{sm} and volumetric moisture content; Cl_{gw}

Ratings
Accuracy: 2; Ease of application: 1; Cost: 1
Although this method may not be as accurate as other methods, differences in recharge estimation are still within a factor of five. Measured atmospheric input of chloride (often only short term records are available) is assumed to be representative for a long period and is thus an area of concern as rainfall and chloride deposition during the past may be different from today. Other areas of concern include the uncertainty in the measured chloride content of rainfall and rainfall amount, depending on the type of rain gauge used, pollution and analytical errors when measuring relatively low chloride contents (Beekman and Sunguro, 2002; Adams, 2002). Despite these shortcomings, the CMB method is highly recommended, also for fractured rock systems (see Cook, 2003); it is relatively simple and it is the least expensive method.

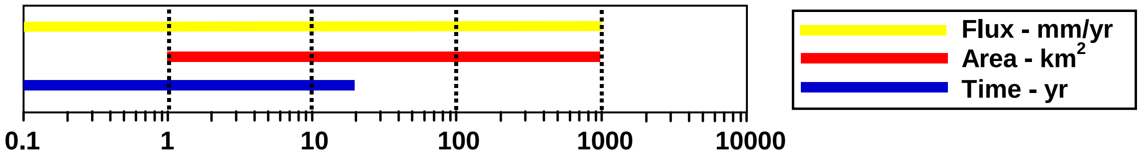
References
Eriksson and Khunakasem, 1969; Gieske, 1992; Bredenkamp et al., 1995; Beekman et al., 1996; Simmers et al., 1997; Cook, 2003.

Cumulative Rainfall Departure (CRD)
The CRD method is based on the premise that water level fluctuations are caused by rainfall events. Bredenkamp et al. (1995) applied the method extensively with success in South Africa. Recently, the method was revised to accommodate for trends in rainfall time series (Xu and Van Tonder, 2001). Recharge is calculated as (Xu and Van Tonder, 2001):

$$R_T = rCRD_i = S_y \left[\Delta h_i + \left(Q_{pi} + Q_{outi} \right) / \left(AS_y \right) \right] \qquad with$$
$$CRD_i = \sum_{i=1}^{N} P_i - \left(2 - \frac{1}{P_{av}\, i} \sum_{i=1}^{N} P_i \right) iP_t \tag{5}$$

where r is that fraction of a CRD which contributes to recharge, S_y is specific yield, Δh_i is water level change during month i (L), Q_p is groundwater abstraction (L^3/T), Q_{out} is natural outflow, A is recharge area (L^2), P_i is rainfall for month i (L/T) and P_t is a threshold value representing aquifer boundary conditions. P_t may range from 0 to P_{av}, with 0 representing a closed aquifer (no outflow), and P_{av} representing an open aquifer system (for instance controlled by spring flow). The ratio r/S_y can be estimated based on Eq. (5) through an optimisation process, which minimises the difference between calculated and observed water level fluctuations over a specific time interval. The CRD method and estimation of the r/S_y ratio has been built into a user-friendly Excel program for recharge estimation called REME (Xu and Van Tonder, 2001).

Applicability



Limitations
The method cannot be applied in areas where there are no groundwater level fluctuations. In the above form it should only be applied to unconfined aquifers.

Data requirements
Monthly rainfall records, water levels, borehole abstractions and aquifer properties including storativity and size of recharge area.

Ratings
Accuracy: 1-2; Ease of application: 1-2; Cost: 2
Groundwater levels of fractured aquifers with small storativity are particularly sensitive to rainfall recharge. Simulation of water levels based on the CRD method and hence recharge estimation is fairly accurate in these cases, provided that storativity can be determined. The uncertainty in recharge estimation increases with increasing depth to the water table. Rainfall, water levels and abstraction rates must be representative for the recharge area of the aquifer. By taking into account different ranges of rainfall, the CRD method will give reasonable estimates of recharge rates. Accuracy of estimation increases with better spreading of boreholes over the recharge area of the aquifer and increased frequency of monitoring data.

References
Bredenkamp et al., 1995; Xu and Van Tonder, 2001.

Groundwater Modelling (GM)
The aim of modelling groundwater flow is usually to predict the aquifer piezometry under various groundwater stress situations. The general three-dimensional groundwater flow equation assuming uniform fluid density and viscosity is formulated as (Bear, 1972):

$$\frac{\partial}{\partial x_i}\left(K_{ij}\frac{\partial h}{\partial x_j}\right) + q_s = S_s\frac{\partial h}{\partial t} \qquad (6)$$

where i,j represent principal coordinate directions, K is hydraulic conductivity tensor (L/T), h is hydraulic head (L), S_s is specific storage (1/L), x is a space coordinate (L), t is time (T) and q_s represents fluid sources (such as recharge) or sinks (such as abstraction) per unit volume (1/T). Eq. (6) can be solved for various complex flow configurations by means of numerical modelling techniques. Recharge can be calculated based on known piezometry, K and S parameter values, and other inflow (e.g. riverbank infiltration, inter-aquifer flow, etc.) into and outflow (natural drainage, abstraction, evapotranspiration) from the aquifer. Confidence in calculated recharge will improve when the velocity distribution of groundwater or groundwater ages based on the hydraulic model match groundwater ages derived from radio-nuclide transport modelling. Three dimensional solute transport in a groundwater system, accounting for fluid sources and sinks and radio-nuclide transport (radioactive decay as a first-order irreversible process), excluding sorption, can be described by the following advection-dispersion equation (Spitz and Moreno, 1996):

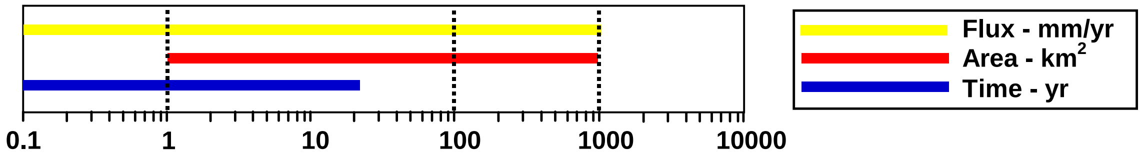
$$\frac{\partial c}{\partial t} = \frac{\partial}{\partial x_i}\left(D_{ij}\frac{\partial c}{\partial x_j}\right) - \frac{\partial(v_i c)}{\partial x_i} + \frac{q_s}{\theta}c_s - S \qquad with \tag{7}$$

$$S = \lambda c \qquad and \qquad \lambda = \frac{\ln 2}{T_{1/2}}$$

where c is dissolved concentration (M/L^3), c_s is concentration of fluid source or sink, t is time (T), v is advective transport velocity (L/T), x is direction of flow, D is dispersion coefficient, θ is the porosity of the medium, S represents a sink due to decay of the radio-nuclide (e.g. ^{14}C), λ is a rate constant that characterizes decay (1/T) and $T_{1/2}$ is half-life of the radio-nuclide (T). The transport equation is linked to the flow equation through:

$$v_i = -\frac{K_{ij}}{\theta}\frac{\partial h}{\partial x_j} \tag{8}$$

Applicability



Limitations
Groundwater modelling is time consuming, sensitive to boundary conditions and difficult to calibrate.

Data requirements
Conceptual hydrogeological model, daily/monthly rainfall records, water levels, borehole abstractions, aquifer characteristics including storativity, hydraulic conductivity, porosity, dispersion characteristics, radio-nuclide concentrations (e.g. ^{14}C), etc.

Ratings
Accuracy: 1-2; Ease of application: 3; Cost: 3
The accuracy of recharge estimation relates directly to the degree of discretization of the groundwater system and to the accuracy of the parameter values. Once the age or velocity distribution in an aquifer based on the flow model matches the age distribution of groundwater, a higher degree of confidence is gained in the recharge estimate. With regard to ^{14}C dating of groundwater, correction models may have to be constructed to account for sources or sinks of carbon. These correction models require a proper insight into the hydrochemistry of water-rock interactions operating in the aquifer, hence ^{14}C dating and thus recharge estimation is becoming a challenging task. Both flow and transport modelling require advanced hydrogeological and hydrochemical skills and costs involved are usually high due to the vast amount of hydrogeological and hydrochemical data required.

References
Bredenkamp et al., 1995; Gieske, 1992; Beekman et al., 1999; Kinzelbach et al., 2002.

1.6 Summary and Future Outlook

A wealth of recharge estimation methods for (semi-)arid areas is currently available with each method having its own limitations. Whereas one method can be applied in site specific studies, the other can better be used in regional studies; whereas one method represents a short time scale, e.g. from event based recharge to daily/monthly/yearly recharge, the other represents a much longer time scale, ranging from decades to thousands of years. Clarity on the aim of the recharge study is crucial in choosing appropriate methods for recharge estimation. Confidence in recharge estimates improves when applying a multitude of methods (Beekman et al, 1996; De Vries and Simmers, 2002; Scanlon et al., 2002).

In Southern Africa, experience in recharge estimation covers a time span of at least three decades. This experience formed the basis for this review of recharge estimation. We conclude at this stage that the following methods can be applied with greater certainty in the arid and semi-arid parts of the region: the CMB, CRD, EARTH, GM, SVF and WTF methods. From these methods the CMB is the easiest to apply and the least expensive whereas GM is the most difficult and expensive method.

Future work should focus on quantifying the time lag between rainfall and water level response, on episodic recharge and on forecasting within the context of climate change. Three decades of work on recharge assessment in the region should be collated and synthesized and translated into user-friendly products (such as manuals, databases, decision support systems and analysis programs) to better serve the groundwater practitioner in estimating recharge and the water manager in properly and effectively using the results. This would pave the way also for dealing with urban recharge, an issue, which has not yet received sufficient attention in Southern Africa.

1.7 Acknowledgements

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2. Challenges in Estimating Groundwater Recharge

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ABSTRACT In the preface of the proceedings of a conference on recharge estimation in Turkey in 1987, Ian Simmers noted:

"No single comprehensive estimation technique can yet be identified from the spectrum of methods available; all are reported to give suspect results."

16 years later, this statement still holds, although in the Southern African context, recharge processes, particularly in fractured rock environments, are better understood due to findings made during structured, well funded, and long term research programmes (Bredenkamp et al., 1995; Beekman et al., 1996). The focus of this paper is to briefly outline the limitations of some of the more common recharge estimation methods, and offer suggestions for improving the overall reliability of recharge estimations in the region.

2.1 Some Commonly Identified Method Limitations

2.1.1 Unsaturated zone methods

While essentially using either a physical or chemical mass balance to obtain an estimate, the underlying assumption made when using unsaturated zone methods is that recharge occurs via diffusive processes (i.e. piston flow occurs). Thus, at sites where preferred pathways influence the recharge flux, as has been shown to occur at sites throughout sub-Saharan Africa, the reliability of resulting estimates is questionable. For example, upon receiving 450 mm of rainfall over a three-day period, only 2 of 18 soil moisture tubes taken near the Free State town of Dewetsdorp during the 1988 wet season showed an increase in moisture content below a depth of 1m (Kirchner et al., 1991). Water levels in boreholes increased significantly during this period, however, thereby confirming recharge via preferred pathways.

The use of unsaturated zone techniques is further complicated when the potential for both diffusive and preferred pathway flow to occur at a given site during different seasons is considered, such as would be expected where surficial highly plastic clays occur. At these sites, rapid recharge via shrinkage cracks could be expected at the start of the wet season, with diffusive processes becoming more important as the clay swells in response to moisture content increases with continuing rainfall. The prevailing climate would also be of significance at these sites, the depth of seasonal moisture variation (and thus the depth of shrinkage crack development), greater in semi-arid to arid areas, as opposed to sub-tropical zones.

The main limitation of physical mass balance models is the uncertainty associated with determining the parameters necessary for recharge estimation, such as soil moisture and its relationship with hydraulic conductivity. For example, the practicing geohydrologist must often make significant assumptions on the basis of unsaturated zone data obtained from small, supposedly representative, sample sizes when applying these methods. If, say, a sandy aquifer is pump tested, the calculated hydraulic conductivity represents an average for profiles influenced by water table draw down, both spatially and with depth. In comparison, a sand permeability can be determined in the laboratory on a sample the size of a beer coaster, the result no doubt correct for that particular sample, but in all likelihood much less representative of the aquifer as a whole. Further, it is assumed that, in the case of:

- In situ samples, they have not been disturbed during sampling;
- Disturbed samples, they can be remoulded in the laboratory to represent site conditions;
- In situ measuring devices, such as lysimeters, they have been sited at a place that is representative of the unsaturated zone as a whole, their installation not having impacted upon site hydraulic behaviour.

2.1.2 Saturated zone methods

Chloride Mass Balance (CMB)

Since being initially proposed by Eriksson and Khunakasem (1969), the Chloride Mass Balance (CMB) method has been applied during recharge investigations worldwide in recent time (Edmunds and Gaye, 1994; Wood and Sanford, 1995; Bazuhair and Wood, 1996). When using the chloride mass balance method, input parameters (i.e. rain) are generally difficult to measure, highly variable (i.e. seasonal effects), prone to pollution, and poorly understood, while unsaturated and saturated zone processes can be easily measured and understood at sites where diffuse recharge is occurring. In the Southern African context, the absence of long-term rainfall quality data for sites across the region is one of the main factors limiting its application. For example, significant seasonal variations are apparent in the chloride concentration of monthly composite rainfall samples taken in Bloemfontein since February 2002 (see Figure 2.1). The cause of the observed variations has not yet been determined, although contamination of rainfall samples with airborne dust is suspected (Bean, 2003). If this is the case, significant annual variations in rainfall chloride concentration could be expected, the potential for dust contamination being greater in drier years when there is less vegetative cover.

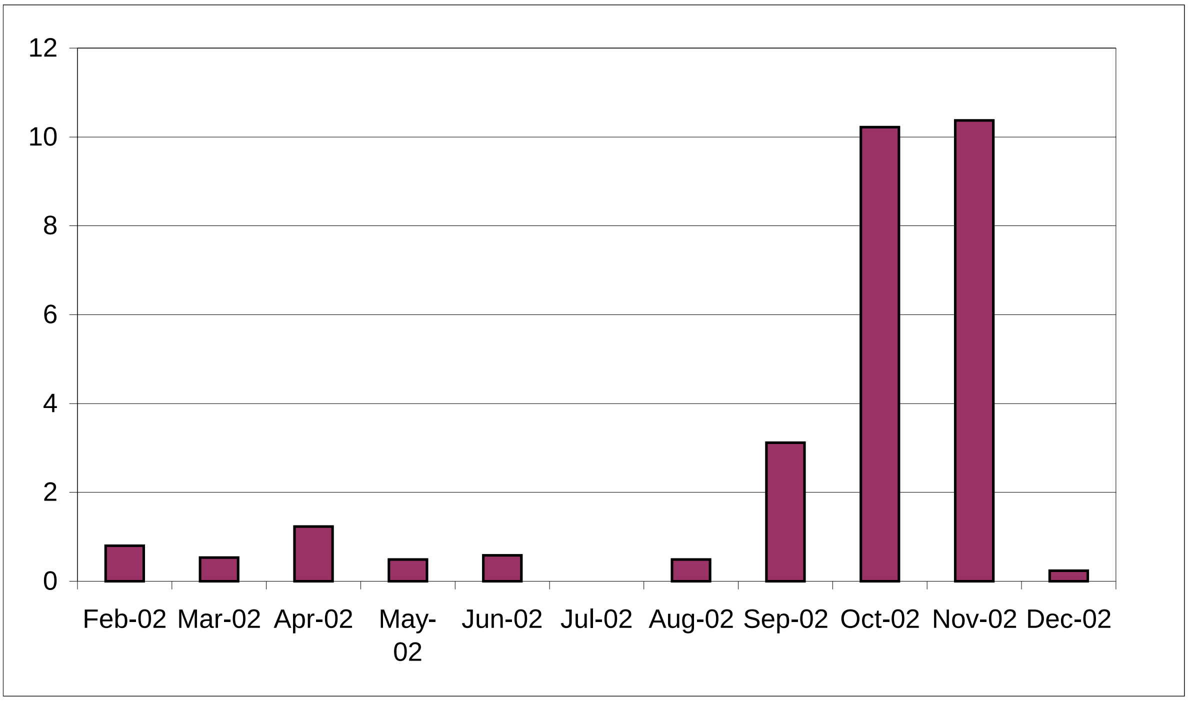

Figure 2.1 Chloride concentration in monthly composite samples taken in Bloemfontein in 2002.

With respect to measured chloride in groundwater, an error of 10% is considered acceptable at several laboratories, while others can only accurately measure dissolved chloride concentrations greater than a given amount (say 5 mg/l). For recharge determination, rainfall chloride concentrations must be accurate and precise to at least 0.1 mg/l (i.e. if the chloride concentration of rainfall at a given site is assumed to be 1 mg/l, but it is actually 0.5 mg/L, calculated recharge will be twice what it actually is).

The groundwater chloride concentration is often assumed to stay constant during recharge studies, although this cannot be stated with any certainty within any investigated aquifer system unless long-term monitoring data is available. Realistically, however, this is only likely to be a problem within aquifers that could potentially contribute to the salt load (i.e. marine sediments), or sites where pollution is occurring.

Another assumption made by practitioners applying the CMB is that the recharge pulse only moves vertically through the unsaturated zone. Within the Karoo Basin, where shallow, sandy soils can overlie less permeable sedimentary sequences on gently sloping ground, this probably rarely occurs (Sami and Hughes, 1996). Under these conditions, interflow/through flow is encouraged, resulting in higher soil water chloride concentrations than would be expected if only vertical flow through the unsaturated zone occurred (i.e. a horizontal distance path is longer, allowing more chloride from the zone of evapo-transpiration to be dissolved from the root zone). Thus, it is the chloride concentration of the standing surface water derived from throughflow, and not rainfall, that must be determined (Wood, 1999). This is often problematic because these samples must be taken as soon as possible after rainfall, often from remote or poorly accessible areas.

Given that input chloride concentrations can vary significantly from site to site within a region of investigation, it is unsurprising that CMB estimations are site specific. Work undertaken at Hotazel (Northern Cape) and Petrusburg (Free State) by Bean (2003), confirms that potential geomorphological controls on groundwater chloride concentration must be considered when applying the method. Indeed, these geomorphological controls are often integral to any conceptual geohydrological model developed for the site, particularly in Southern Africa where most exploitable aquifers are unconfined or semi-confined in character.

Isotope methods

Allison et al. (1983) developed a semi-empirical method of estimating recharge using the stable isotopes ^{2}H and ^{18}O. According to Amore et al. (2000), the method has never been shown to be applicable under field conditions, perhaps not surprising given that, while the relative abundance of both isotopes is influenced by evaporation, transpiration effects are not represented on infiltrating soil water (Zimmerman et al., 1967). Thus, they are more for qualitative, as opposed to quantitative use during recharge studies, particularly given that both can be regarded as conservative tracers.

Regional-scale estimations have been undertaken by using groundwater age data to estimate the aquifer replenishment rate (Kotze et al., 2000). While several isotopes can be used to date groundwater, perhaps the most commonly used is ^{14}C, a radioactive isotope that can be used determine groundwater ages up to 60000 years. Age determination using the method can be a daunting task, however, due to difficulties associated with accounting for ^{14}C dilution by alternative carbon sources. In Southern Africa, this is a particular problem given the high carbonate content of many aquifers in semi-arid and arid areas.

As an example of the problems associated with dating using ^{14}C, consider groundwater sampled from Borehole 5.1, Florida, U.S.A. ($^{14}C_{DIC}$ = 37 pmC and ^{13}C soil gas = -12 $^{o}/_{oo}$). This sample was assessed using adjustment models suggested by several different researchers (see Figure 2.2), the resulting estimates varying between 8000 years and 400 years. The interpretation of ^{14}C results can therefore be very subjective, a condition that is not conducive to obtaining reliable estimates of recharge.

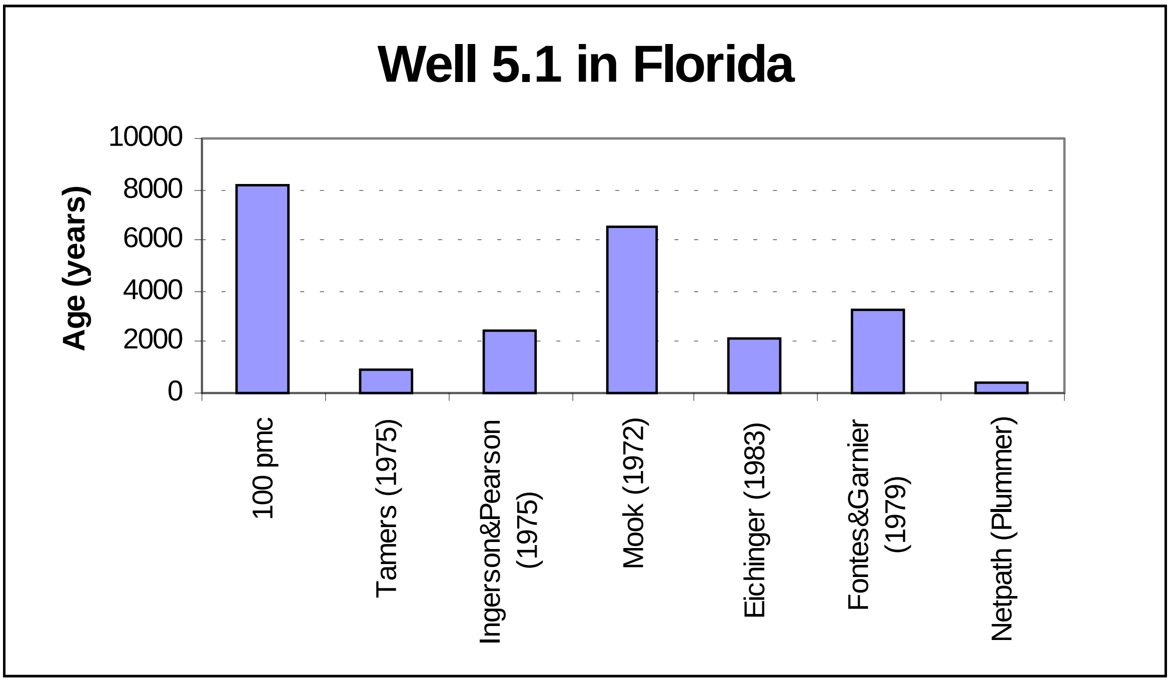

Figure 2.2 Range of ages obtained by using different adjustment models to account for carbon dilution during ^{14}C dating (Plummer and Sprinkle, 2001).

Water balances (SVF method, CRD method, Groundwater models)

The accuracy of water balance methods is largely dependent on the quantity and quality of data available for interpretation (e.g. spreading of the boreholes over the aquifer, frequency of water level and abstraction data, correctness of the conceptual model and boundary conditions). A problem with the SVF-method is that the measured water levels must be representative for the aquifer as a whole, with lateral inflow and outflow along the boundaries known. Even the steady state calibration of recharge in a groundwater model could be problematic in a case where the outflow flux (or inflow flux for that matter) is not known with a high degree of certainty. Nevertheless, providing the boundary conditions are well known, a steady state model can be used to obtain an average recharge for the different zones in the aquifer with good effect.

The major advantage of SVF-type estimations is that they allow recharge estimations to be made from current data. While this probably isn't a cause for concern in wetter areas where aquifers are recharged annually and are regularly "flushed", in semi-arid to arid areas comprising thick unconsolidated aquifers it becomes a problem because groundwater residence times are much longer. At these sites (e.g. the Kalahari), it is possible that the background Cl concentration has been influenced by recharge that occurred during a previous wet period, say, 4000 years ago, rather than recharge processes today. It should be appreciated, however, that SVF techniques are most applicable to unconfined to semi-confined aquifer systems, as the water level that corresponds for the aquifer under investigation is known, which is often not the case in layered aquifer systems.

A shortcoming with most models is that uniform recharge over the model area is assumed, which is quite clearly not the case in most instances. However, unless detailed and often expensive investigations are undertaken initially, it is almost impossible to identify areas of preferred recharge. For example, at Meadhurst near Bloemfontein, recharge is a factor of two higher in lower lying areas as compared to the surrounding dolerite hills, the implications being that run-off from the relatively bare hills accumulates as surface water in depressions prior to recharge. Similar recharge behaviour has also been observed elsewhere, particularly in semi-arid to arid areas (Vogel and van Urk, 1975; Wood and Sanford, 1995; Bazuhair and Wood, 1996; Bean, 2003). Indeed, given current industry competitiveness, most consultants would not be prepared to take the risk of budgeting for such a detailed investigation at the time of tender submission, their site conceptualisation and resulting model unrepresentative of site conditions as a result.

Recharge estimation is also possible using another water balance type method, the EARTH model (Beekman et al., 1996), monthly water level and rainfall data for a given borehole being a minimum requirement. Again, for the saturated part of the model, the storage co-efficient must be known for a unique recharge estimate to be obtained.

2.2 The Balance between Recharge and Sustainability – A Namibian example

In terms of sustainable management, it is important that the water level in a given borehole be maintained (i.e. the static water level after two years of pumping is the same as was initially), which by inference requires that extraction be less than recharge. In practice, however, groundwater resource management is more complicated than this given the spatial variations in recharge. Thus, while the total extraction from a given borehole is much less than total recharge to the aquifer system as a whole, where recharge is exceeded locally, water levels will still decline over time. It is therefore important that management approaches not be over-generalized, site specificity instead considered within the framework of a comprehensive holistic management strategy for respective aquifer systems.

Long-term trends in monitoring data collected from well fields in the Windhoek Aquifer provide an example of unsustainable management practices in a semi-arid area (see Figure 2.3). Abstraction here totalled 103 Mm^3 for the period 1950 to 2000 (i.e. about 2 Mm^3/a), resulting in an average annual water table decline of 0.8 m/a (see Figure 2.4), with Murray (2002) estimating that this must decrease to about 1.7 Mm^3/a (i.e. 2.2% of the average annual rainfall over an area of 213 km^2) to ensure long term sustainability. However, given the total area of the Windhoek aquifer is larger than the 213 km^2, the potential exists for sustainability to be restored if water from new boreholes drilled outside of the currently exploited catchment area is used to complement the well field supply.

The low chloride concentration (approaching 5 mg/L in some instances) of groundwater extracted from the Windhoek Aquifer is also of interest. Given that the weighted average chloride concentration of site rainfall was about 1 mg/l, recharge estimated using the CMB method was 20%, a significant over-estimation. When site geology was also considered, however, it was apparent that the low chloride groundwater was associated with fault zones that occur within the quartzites of the Aus Mountains. Thus, the observed low groundwater chloride concentrations can be attributed to rapid recharge along the trend of these structures.

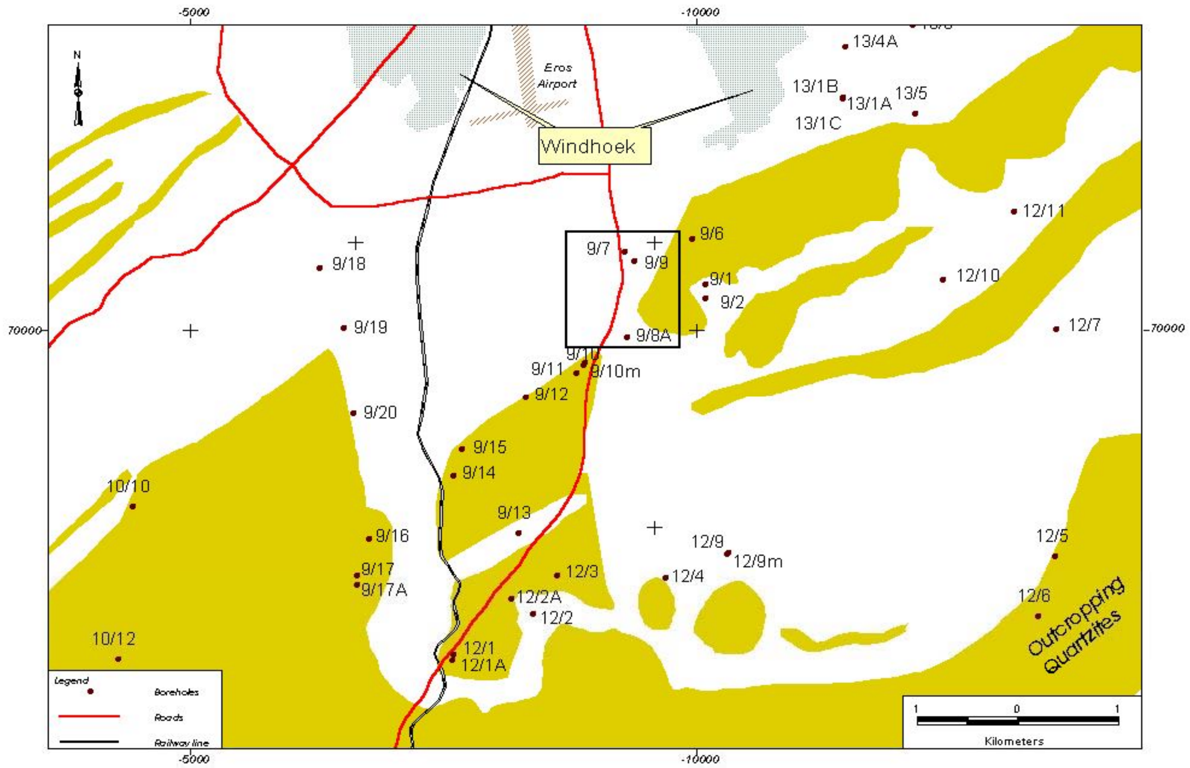

Figure 2.3 Layout of abstraction boreholes in the Windhoek Aquifer.

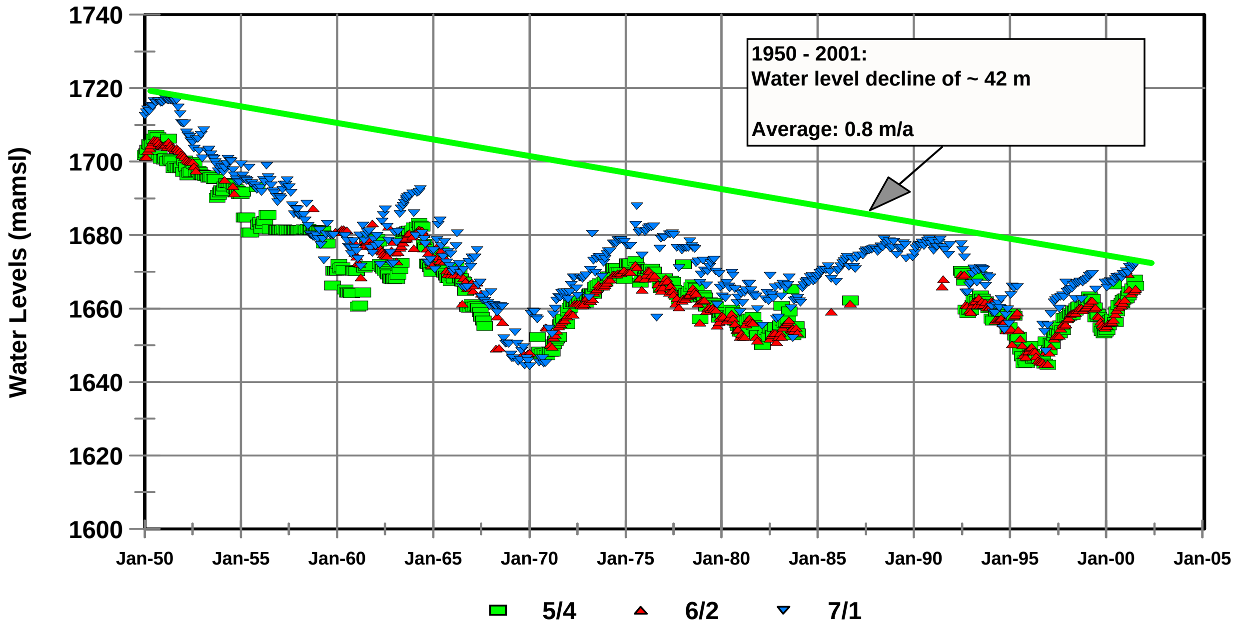

Figure 2.4 Water level data of three boreholes in Windhoek, Namibia (after Murray, 2002).

Note that the groundwater table in Figure 2.4 shows an average annual decline of 0.8m over a 50-year period.

For the Windhoek Aquifer, unsaturated zone storage is not always available. In such a case, the component of rainfall that is not run-off rapidly recharges the aquifer via preferred pathways. Indeed, at sites where groundwater chloride concentrations are higher than those in rainfall, observed increases could be just as easily be attributed to the dissolution of airborne dust derived chloride that has been deposited on site surfaces as opposed to evaporation.

The absence of unsaturated zone storage could also explain the high recharge (50%) calculated using the CMB method for fractured Table Mountain Group aquifers near Cape Town (Weaver and Talma, 2000). Visual inspection of the outcropping mountain peaks in the area confirms that there is little, and in many places, no unsaturated zone. Thus, given that a) recharge water is only resident in the unsaturated zone for a limited time due to the lack of storage; b) transpiration cannot take place because vegetation has difficulty becoming established at such sites; and c) evaporation cannot take place because rapid recharge occurs via preferred pathways, then recharge can be defined as that component of rainfall that does not leave the recharge area as run-off.

2.3 Approach to Recharge Estimation in Some Difficult Terrains

2.3.1 Episodic recharge

Providing the conceptual recharge model and method limitations are understood, the CMB method is a practical, low cost method of estimating site recharge. This is significant in Southern Africa because, while considerable funding has traditionally been made available to undertake research into water quality issues, limited funding appears to have been set aside for ensuring that water levels are measured at monthly intervals in aquifers of strategic importance. Other techniques, which provide better estimates because they model current aquifer response to recharge as opposed to the long-term averages that geochemical and isotopic methods provide, cannot be utilized. This is perhaps of most concern in areas where episodic recharge, a feature of semi-arid and arid landscapes, occurs.

Moisture retention within the unsaturated zone, particularly in semi-arid to arid areas draped with thick sand sediments, can have a significant influence on recharge during a given rainfall event (Foster et al. 1982; de Vries et al., 2000). While recharge is diffusive in these environments, the recharge pulse is only mobilized by extraordinary rainfall, and is thus episodic in character. In terms of effectively managing groundwater resources in these areas, it is not only essential that the amount of recharge be quantified, but also the amount of rainfall that is required to initiate recharge, as this will dictate the amount of groundwater that must be held in reserve. Recharge in these areas occurs maybe once in every 10 years, over which time resource requirements could change drastically (i.e. a new groundwater supply for a town or mine is required). Given the time span involved, greater flexibility in the use of the resource is required, such that allocations in these areas be based on the total amount of water that can be used between recharge return periods rather than an annual average.

Given the lack of water level data for much of semi-arid and arid Southern Africa, alternative methods of episodic recharge prediction are required. One technique successfully applied in the Northern Cape Province of South Africa was to relate amount effects observed in regional rainfall data to groundwater isotope values. Dansgaard (1964) initially made reference to the amount effect, whereby the isotopic composition of rainwater is related to the depth of precipitation received at a given site. Work undertaken by Bean (2003) confirms that amount effects can be identified in International Atomic Energy Agency and World Meteorological Organization (IAEA/WMO; 2003) datasets for Johannesburg and Windhoek, but not for Cape Town, possibly a consequence of the latter's Mediterranean climate. Thus, using an approach similar to that taken by Harrington et al. (2002), the minimum amount of rainfall required for recharge to occur can be predicted (Figure 2.5).

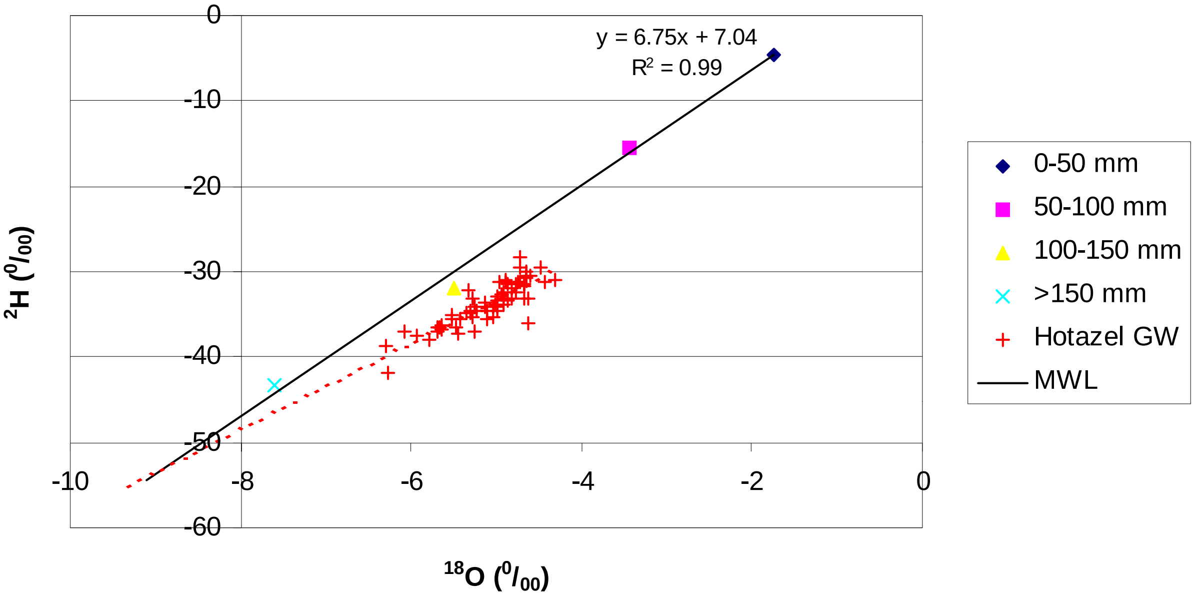

Figure 2.5 Amount effect for Windhoek rainfall determined using IAEA/WMO (2001) data.

Note that in Figure 2.5 the weighted average isotopic composition for monthly rainfalls of 0 to 50, 50 to 100, 100 to 150, and >150 mm intervals become successively more depleted in ^{18}O. When Hotazel groundwater data (used with permission of BHP Billion) for those sites recharged directly from rainfall (^{18}O<-4.3 $^{o}/_{oo}$) is included, it is apparent that a linear trend line can be extended to intercept the Windhoek MWL, the point of interception indicating the amount of monthly rainfall required before recharge occurs. Results indicate that monthly rainfall must exceed at least 150 mm before recharge occurs, which when considered against 40 years of monthly rainfall data for Hotazel, suggests that recharge has occurred no more than 13 out of a possible 504 months during that period, and is thus episodic in character.

2.3.2 Recharge in fractured rock terrains

Given the limitations of the CMB method, particularly at sites where recharge occurs via preferred pathways, it may seem paradoxical that its use be recommended within some fractured rock terrains. Nevertheless, its application is justified when the variable hydraulic properties of fractured rock aquifers are considered.

On a regional scale, fractured rock aquifers are commonly regarded as equivalent porous mediums for modelling purposes, a necessity given the significant variations in porosity, hydraulic conductivity, and storage that occur between adjacent areas. Thus, even where long-term water level data is available, the hydraulic conditions that contribute to the observed water table response at a given site following recharge represent an average for the area surrounding a given borehole. The CMB method negates the need for measuring or estimating these hydraulic parameters, as it already represents a long-term average of recharge. This is not to say that water levels should not be taken at fractured rock terrains, but rather that recharge calculated using water balance methods be checked using CMB methods in those areas completely overlain by a porous unsaturated zone of significant thickness. Indeed, the comparison of results obtained using multiple estimation techniques is recommended during all recharge investigations, whether conducted in fractured rock or porous environments.

2.4 Future Challenges

The impact of climate change on recharge estimation has the potential to be significant, the main reason being that the underlying assumption of the CMB method, that present recharge is a reflection of what occurred in the past, may no longer hold true. It is not just the potential for some areas of Southern Africa to become wetter and others drier in response to global warming that is of concern, but also the potential for rainfall chloride concentrations to change over time. Given that monthly variations in rainfall chloride concentrations at two inland areas have been attributed to the influence of airborne dust (Bean, 2003), it is reasonable to assume that conditions that affect the generation of dust are of importance regionally. Thus, if the Kalahari were to become significantly wetter, one would assume the amount of airborne dust, and hence rainfall chloride concentration to decrease. However, the re-equilibration of groundwater chloride concentrations necessary before the CMB method could be again reliably applied in the area would not be apparent for several (most likely thousands) of years afterwards.

One of the major future challenges is the identification of recharge areas, and unless detailed and expensive investigations are undertaken initially, it will be impossible to identify these zones.

2.5 Conclusions

Given the inherent inhomogeneity of catchment and aquifer parameters, there is no set procedure that can be followed to enable site recharge to be estimated. What is most important is that, wherever possible, multiple estimation techniques be applied, the shortcomings of each clearly understood by the investigator in question. However, the absence of long-term monthly water data at most sites will result in a continued dependence on the combined application of the CMB method with stable isotope techniques for some time yet.

In order to obtain more reliable estimates, a concerted effort must be given to obtaining input parameter data, including monthly water levels, seasonal rain, surface, and groundwater quality (chemical and isotopic), and the storage and management of this data in centralized databases. This can be best achieved with government funds, although given the recent changes in legislation requiring industry to ensure monitoring is undertaken in some countries in Southern Africa, there is considerable scope for private money to contribute to data collection. This can perhaps best be achieved if a standardized monitoring code of practice is developed for industries operating in the region, which outlines minimum monitoring frequencies for input parameters necessary for recharge estimation.

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PART II

Recharge Estimation Related to the Unsaturated Zone

3. Multiple Tracer Profiling in Botswana – GRES Findings

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ABSTRACT Moisture transport through the unsaturated zone in semi-arid Botswana has been studied extensively through the "Groundwater Resources Monitoring and Recharge Study" project (GRES: 1987-1997; Gieske, 1992; Selaolo, 1998; Beekman et al., 1999). Various methods were applied to estimate moisture fluxes and to elucidate recharge mechanisms. Most successful were the profiling methods that used a multitude of the tracers Chloride, ^{2}H, ^{18}O and ^{3}H. Moisture fluxes were found to decrease sharply with decreasing annual rainfall from around 11 mm/year in shallow soils and weathered rock overlying Precambrian basement in south-eastern Botswana to less than 1 mm/year in Kalahari sediments of Central Botswana. Most importantly, the studies revealed evidence for preferential flow occurring at great depths. This means that the use of methods that assume diffuse moisture transport require validation in order to derive meaningful results.

3.1 Introduction

Moisture transport studies in Botswana in the vadose zone date back to the early 1970s and were mainly carried out in the Kalahari. Early judgments suggested 6 m Kalahari sand cover as the limit of recharge, with vegetation effectively taking out all available moisture. However, subsequent studies, using ^{13}C, ^{14}C, ^{3}H, ^{2}H and ^{18}O, carried out in both the Central Kalahari (Verhagen et al., 1974, Mazor et al., 1977) and in the Gordonia region of the southern Kalahari (Verhagen, 1984), suggested that recharge is common. Foster et al. (1982) carried out isotope studies in the fringe of the Kalahari Basin and pointed out that tritium levels in the unsaturated zone were very low while chloride levels were found to be very high. From their study, it was concluded that recharge under present climatic conditions was unlikely. De Vries (1984) corroborated this view through a numerical study of receding regional groundwater levels since the last pluvial period (12000 BP). The conclusion from this study was that present-day overall recharge is probably less than 1 mm/yr and that part of present groundwater flow could be a residual from a previous wetter climatic period.

Since the late 1980s, systematic research was carried out for a period of ten years within the framework of the GRES project. The investigations centred around moisture transport through the vadose zone and recharge to the underlying aquifers in crystalline basement in the south-eastern part of the country (see Figure 3.1 - GRES-I: 1987-1992; Gieske, 1992) and in Karoo aquifers in the Kalahari Basin towards the central part of the country (GRES-II: 1992-1997; Selaolo, 1998; Beekman et al., 1999). In this paper, we present some results of multiple tracer profiling studies along a transect from south-eastern to central Botswana.

In section 3.2, the basic theory of methods used relating to multiple-tracer profiling is given. Section 3.3 discusses the long-term annual rainfall, total chloride deposition patterns and meteoric water lines (MWL's) for Botswana. These constitute the input function for quantifying moisture fluxes and for elucidating recharge mechanisms. The fourth section deals with shallow and deep multiple tracer profiling in the unsaturated zone in the GRES study areas of Nnywane-

Pitsanyane, Letlhakeng-Botlhapatlou and Central Kalahari. A summary of the findings is given in section 3.5.

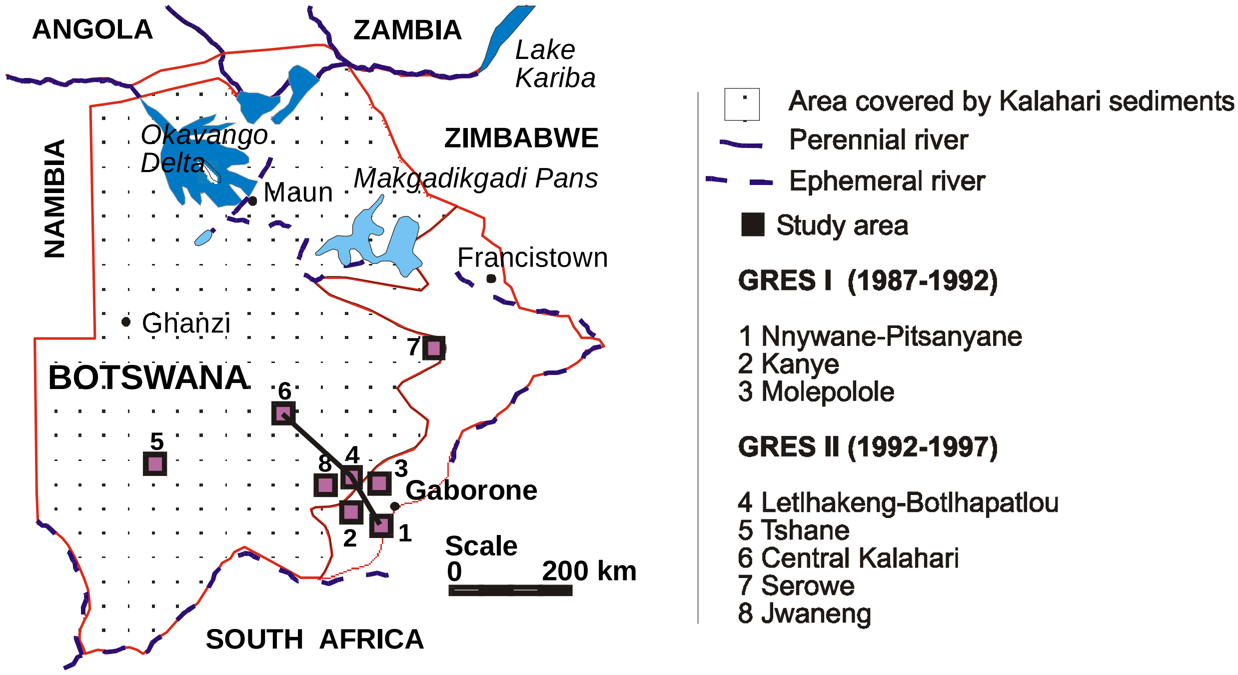

Figure 3.1 Map of Botswana showing GRES study areas (Beekman et al., 1996).

3.2 Theoretical Aspects of Multiple Tracer Profiling

The following tracers were used in the multiple tracer profiling studies along the southeast – central Botswana transect to determine moisture fluxes and to elucidate recharge mechanisms: Chloride, the stable isotopes ^{2}H and ^{18}O and the radioactive isotope ^{3}H. Each tracer provides insight into certain aspects of transport processes operating within the vadose zone and the aquifer. Chloride has been used in a wide range of environments for the determination of both moisture fluxes and recharge rates and has also been used successfully to unravel different recharge regimes in the past. It should be noted that moisture fluxes only represent potential recharge while actual recharge can only be determined when also the saturated zone or the water table is included in the analysis. ^{2}H and ^{18}O are valuable for estimation of evaporation and for delineation of the origin of groundwater. They may also be used in the estimation of moisture fluxes and recharge rates. The presence of elevated ^{3}H activity in moisture and groundwater is an indication of recently infiltrated water and may also be used for deciphering recharge mechanisms in terms of multi-model flow. The strength of multiple tracer profiling (see e.g. Simmers et al., 1997) is that various information obtained from each tracer, including the lithological and other information about the groundwater system, can be combined so that a more complete picture emerges of both fluxes and transport mechanisms.

In this paper, moisture fluxes have been calculated based on the Chloride Mass Balance (CMB), $^{2}\delta$ Isotope Displacement (ID) and the Tritium (^{3}H) methods. Detailed information on these methods, including limitations, underlying assumptions and applicability can be found in Allison et al. (1984), Lerner et al. (1990); Gieske (1992), Bredenkamp et al. (1995); Beekman et al. (1997) and Selaolo (1998).

3.2.1 *Chloride Mass Balance (CMB)*

This method is based on the assumption of conservation of mass between the input of atmospheric chloride and the chloride flux in the subsurface. For a steady state between the chloride flux at the surface and the chloride flux beneath an upper zone where evapotranspiration and mixing of rainfall and pore water takes place, the moisture flux can be calculated as (Eriksson and Khunakasem, 1969):

$$R_{sm} = \frac{P * Cl_p + D}{Cl_{sm}} = \frac{TD}{Cl_{sm}} \tag{1}$$

where R_{sm} is the moisture flux (diffuse or slow flow component; mm/yr), P is rainfall (mm/yr), Cl_p and Cl_{sm} are chloride concentrations in rainfall and soil moisture (mg/l), and D is the dry chloride deposition ($mgm^{-2}yr^{-1}$). The sum of $P*Cl_p$ and D is also referred to as "Total atmospheric chloride Deposition" (TD) and originates from both precipitation and dry fall out. Comparison of moisture flux and recharge provides insight into the mechanism of recharge.

3.2.2 *Isotope Displacement - $^{2}\delta$ (ID)*

Allison et al. (1984) proposed a conceptual model for calculation of moisture fluxes and recharge rates in (semi-) arid areas using the stable isotopes ^{2}H and ^{18}O. They proposed the following simple relationship between displacement of δ-values of soil water from the local meteoric water line ($^{2}\delta = a^{18}\delta + b$; with a=8 and b=10 for the Global MWL) and moisture flux:

$$\Delta\delta = \frac{C}{\sqrt{R_{sm}}} \tag{2}$$

with Δδ as the displacement of either ^{2}H or ^{18}O from the local MWL (in $^{o}/_{oo}$) and C representing the slope of a line through the inverse of the square root of moisture fluxes obtained from other methods (e.g. from the CMB) and δ displacements from the local MWL observed in a $^{2}\delta$ - $^{18}\delta$ diagram for different sites. The method assumes that rainfall events are evenly distributed over a year and that moisture is transported through the vadose zone by means of piston flow. Neither one of the assumptions seem to always hold in semi-arid environments, but for a wide range of climatic regimes in Australia, Allison et al. (1984) did find a valid relationship for C=20. The method is more sensitive to moisture fluxes of less than 10 mm/yr.

3.2.3 *Tritium (^{3}H)*

A downward moisture flux can be calculated from (Allison et al., 1994):

$$R_{sm} = \frac{10\sum_{i=1}^{z} T_i \theta_i h_i}{\sum_{i=1}^{t} wT_{pi} \exp(-\lambda i)} \tag{3}$$

where T_i is the concentration of ^{3}H (TU) for a specific depth interval h_i; θ is volumetric moisture content; T_{pi} is the mean ^{3}H concentration in rain for year i before present; λ is the tracer decay constant (0.05576 yr^{-1}) and w is a weighting factor which takes into account year to year variation in recharge (w~1 in general). This method also assumes vertical piston flow. If vapour transport develops, fluxes would be overestimated, especially where these are known to be low.

3.3 Rainfall, Chloride Deposition and Meteoric Water Lines

3.3.1 Rainfall

Rainfall monitoring in Botswana has been a continuous process since the early 1900s (Bhalotra, 1987). Expansion of this network has mainly been a concerted effort between the Department of Meteorological Services and the Department of Geological Survey.

Rainfall in Botswana is restricted to the summer season and most rainfall events occur as convective thunderstorms of limited areal extent and relatively short duration. Daily amounts of 10 to 30 mm are common. Occasionally, however, large-scale frontal systems bring heavy, widespread rains for periods ranging from one or two days to several weeks. Figure 3.2a illustrates the long-term average rainfall in Botswana. Tyson (1986) has shown that rainfall is modulated slightly by a periodicity of between 18 and 20 years, which has clearly manifested itself in the wet period of the 1970s and the drought of the 1980s. Although there seems to be a general trend, strong local and regional deviations occur due to the different geological and geomorphological conditions.

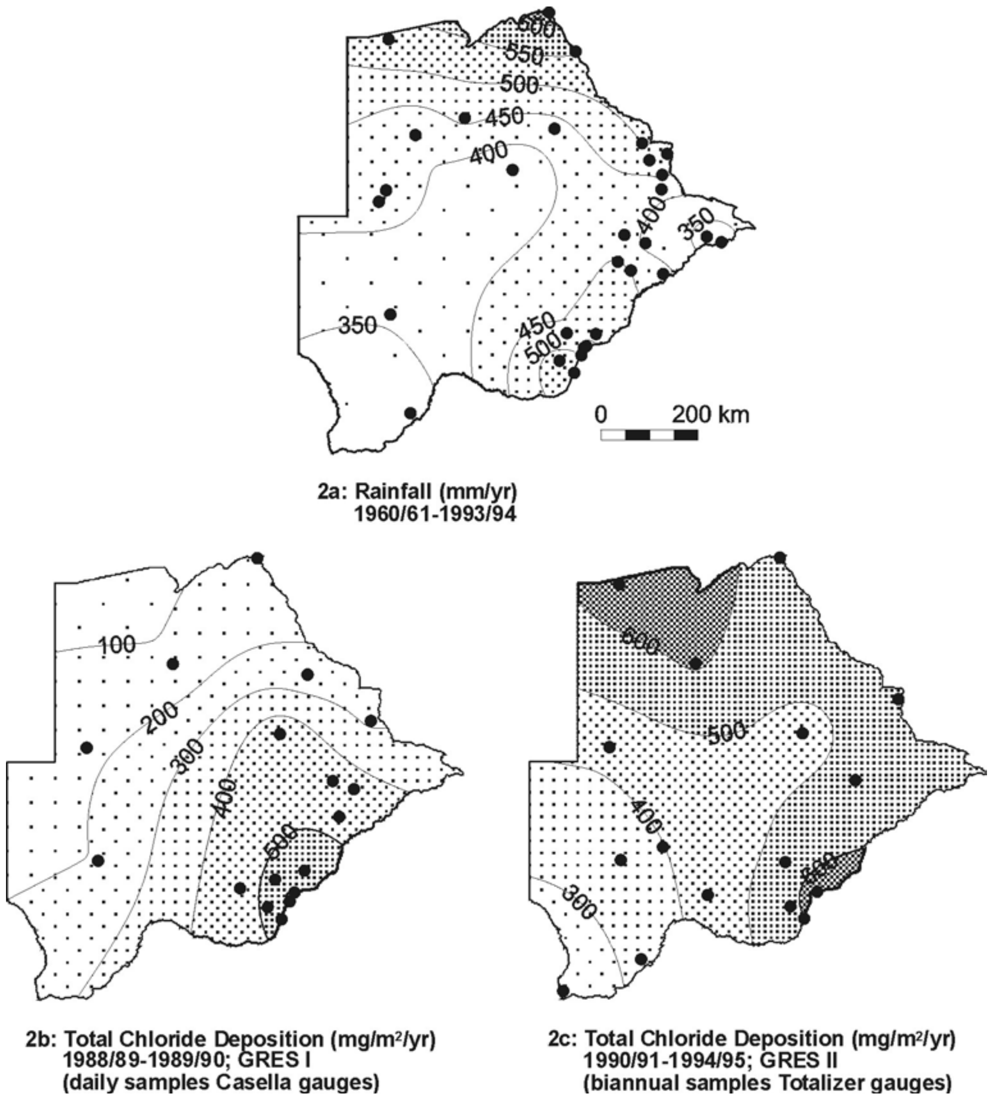

Figure 3.2 Rainfall and Total Chloride Deposition Patterns (Beekman et al., 1996).

3.3.2 Chloride Deposition

Rainfall sampling for the determination of chemical constituents has been carried out systematically by the Botswana Geological Survey since 1983 mainly to determine the total chloride deposition (TD) for groundwater recharge estimation. Within the GRES project, various sampling methods and different types of rain gauges were used to evaluate the validity of the results (Gieske, 1992; Selaolo et al., 1994; Beekman et al., 1996; Selaolo, 1998). Factors that influence the accuracy and validity of TD determination are the design of the rain gauge; the material of the funnel and reservoir; the preparation, sampling and conservation procedures; contamination and instrumental accuracy of Cl analysis.

Large variations in TD (inter-annual, between different types of rain gauges and between rain gauges of the same type) were interpreted from an experimental station at the fringe of the Kalahari (Selaolo, 1998; Beekman et al., 1999). Despite these variations, an increase in TD at this station was observed with increasing rainfall. Figures 3.2b and 3.2c illustrate the difficulty in interpretation of TD-results from the countrywide network for two different periods of measurement corresponding to two different types of rain gauges. For mean annual rainfall increasing in a northeasterly direction from 300 to 650 mm/yr (Figure 3.2a) different trends in TD are observed. The differences between the two series of TD values can neither be explained by differences in rainfall amounts nor by expected differences in TD due to the different types of rain gauges. Clearly, further experimentation and long-term monitoring is needed to establish the causes of such variations.

3.3.3 Meteoric Water Lines (MWL's)

Systematic sampling of rainfall for ^{18}O and ^{2}H started at the Lobatse Geological Survey rainfall station in 1991 and was extended to the fringe of the Kalahari as well as to the Central Kalahari environment. Figure 3.3 shows MWL's based on individual rainfall events.

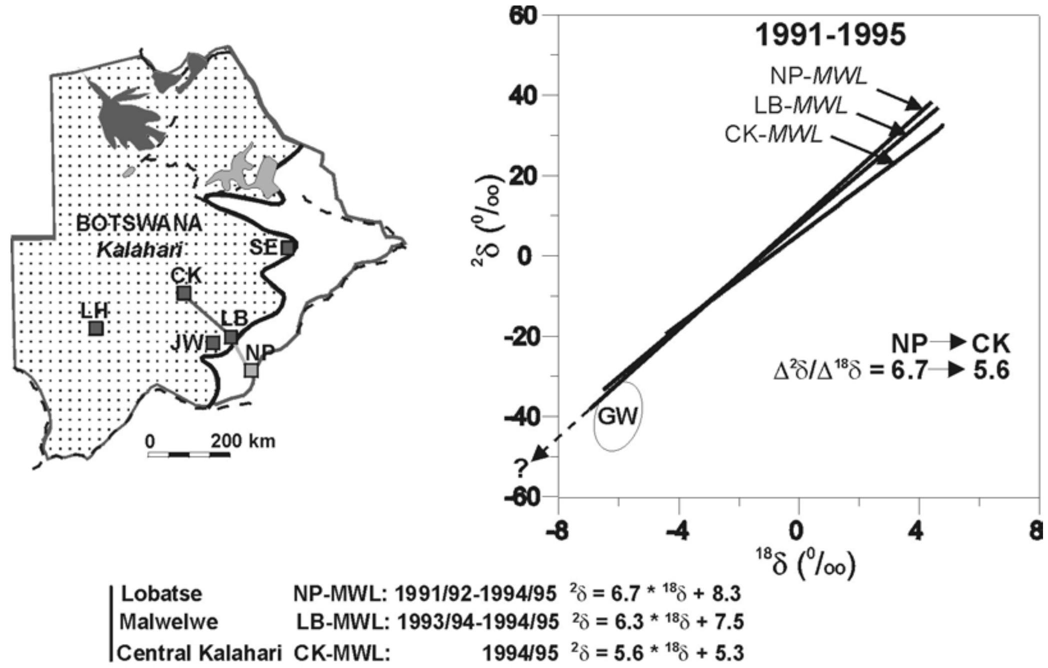

Figure 3.3 Botswana MWL's (Beekman et al., 1999).

A decrease in slope and deuterium excess value can be seen from the south-eastern part of the country (Lobatse: NP-MWL) via the fringe of the Kalahari (Malwelwe: LB-MWL) towards the Central Kalahari (CK-MWL). This decrease may be attributed to an increasing evaporative enrichment of raindrops below the cloud base (Rozanski et al., 1993). The Botswana MWL's do not differ much from the Pretoria MWL. For the Letlhakeng-Botlhapatlou (LB) area at the fringe of the Kalahari, it was found that over the whole range of rainfall amounts the most depleted ^{18}O and ^{2}H contents closely resemble the groundwater composition (Beekman and Selaolo, 1994). Variation in isotopic composition was largest for rainfall amounts of events less than 30 mm suggesting that events in excess of 30 mm significantly contribute to groundwater recharge.

Events with relatively large amounts of rainfall were also analysed for ^{3}H since 1992. Activities for stations from the south-eastern part of the country towards the Central Kalahari (Lobatse, Malwelwe and Central Kalahari) ranged from 1.5 to 8.2 TU with a weighted mean of 4 TU for the 1994/95 rainy season.

3.4 Multiple Tracer Profiling in the Unsaturated Zone

It is generally recognized that groundwater recharge occurs along various routes such as direct infiltration through outcrop areas, hill slope runoff percolating through colluvial and alluvial fans, river bed infiltration and through soils of varying thickness (Allison et al., 1994). The nature of the zone above the water table in these geomorphologic settings is highly complex. Figure 3.4 illustrates some of the geomorphologic controls on moisture transport through the vadose zone that can be expected in a typical semi-arid environment, like the Botswana Kalahari. Both unsaturated and saturated conditions exist in a heterogeneous medium characterized by primary and secondary fracture porosity. Strong variations in hydraulic behaviour occur through small differences in moisture retention. In semi-arid areas, in particular, where rainwater after infiltration is often subjected to high evapotranspiration only a few percent of the annual rainfall normally percolates down to the water table.

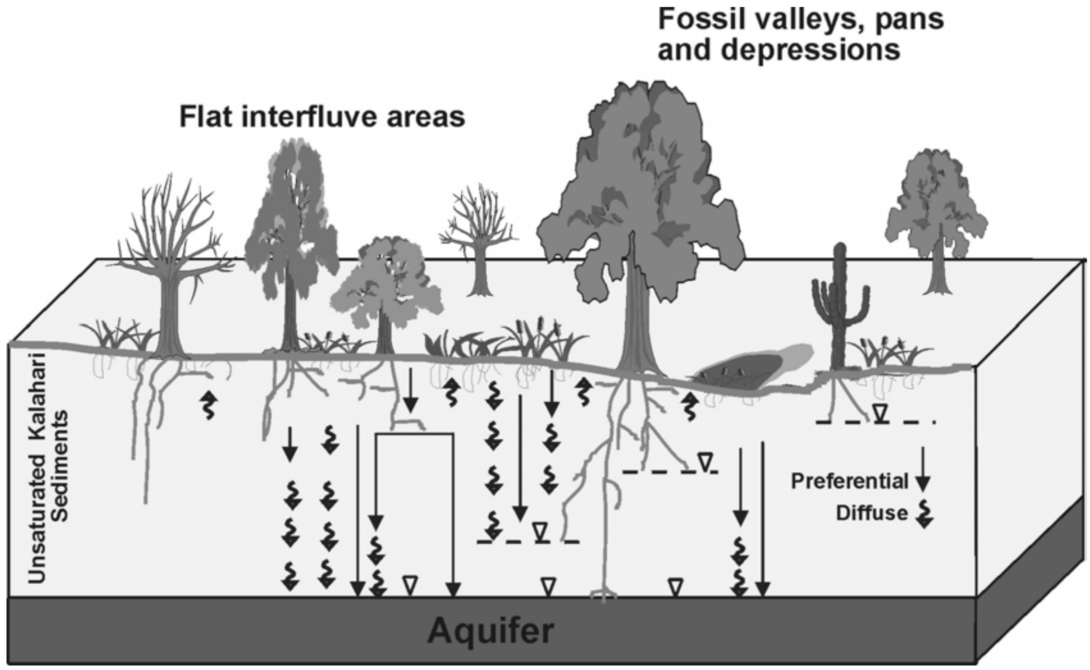

Figure 3.4 Mechanisms of infiltration and moisture transport in the Kalahari (Selaolo, 1998).

In view of the heterogeneous nature of the vadose zone, transport is expected to be multi-modal, with slow diffuse percolation through the topsoils and relatively fast preferential flow through cracks, root channels, and fractures. An understanding of moisture transport through this heterogeneous zone is therefore essential for an appraisal of the relative importance of contributing recharge components.

In the following sections, an overview of results of the GRES multiple tracer profiling studies along a transect from south-eastern to central Botswana is given. More detailed information is found in Gieske (1992), Selaolo (1998) and Beekman et al. (1999).

3.4.1 Nnywane-Pitsanyane (NP - average rainfall 500 mm/yr)

Several small dolomite aquifers lie about 10 km north of Lobatse in a north-south trending valley along a major geological fault. Soil thickness is about 10 m along the valley axis, but decreases to less than 1 m towards the slopes. Highest aquifer transmissivities and storativities occur in the valley where the water table is about 40 m below the surface. Because the aquifers form an important water resource for the Lobatse water supply, the hydrogeology and hydrochemistry of the area is well known (see e.g. Jennings, 1974 and Gieske, 1992).

Shallow profiles

A small study area of 200 by 200m was selected on the main valley axis and twenty shallow profiles varying in depth from 1 to 10 m were drilled with a hand auger system. Samples, taken at regular intervals were analysed for Cl, tritium and stable isotopes. Detailed analytical procedures and results were reported by Gieske et al. (1990) and Gieske (1992). Figure 3.5 illustrates some of these results.

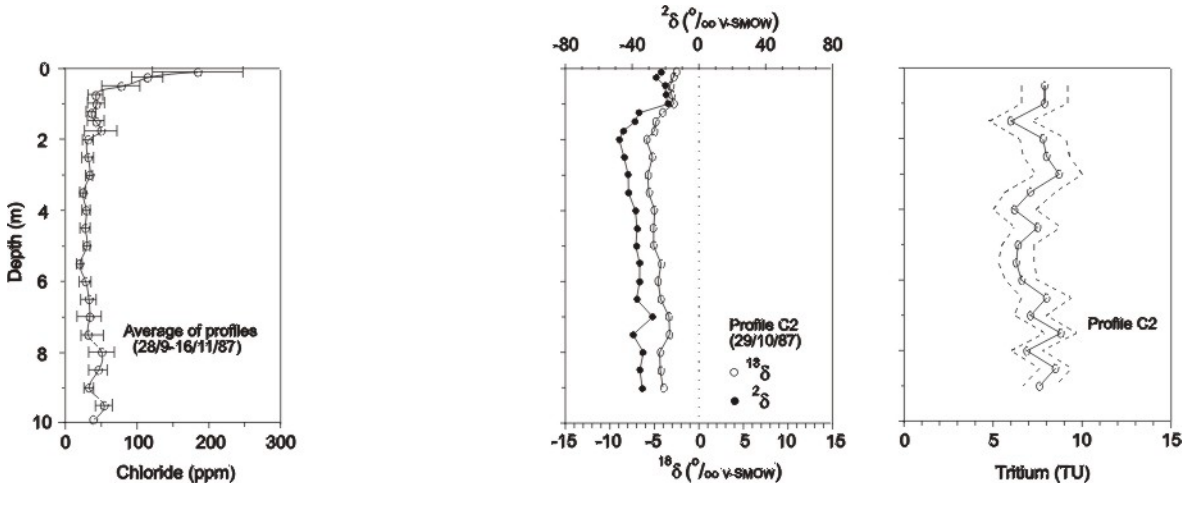

Figure 3.5 Nnywane-Pitsanyane shallow profiles (Gieske, 1992).

The averaged chloride profile illustrates that the root zone in the area does not in general lead to a maximum in the chloride concentration, as is the case in e.g. many Australian examples (Allison et al., 1994). It was concluded from the chloride mass balance method, that the average moisture flux through the soils in the area amounted to about 11 mm/yr. It was also observed that, even on the small scale of 200 by 200 m, variability in moisture transport was high. Surface drainage lines were found, below which infiltration was much enhanced. Regionalization and averaging of recharge fluxes obtained through vertical profiles is therefore not straightforward, even when drilled on such a dense and regular grid.

Stable isotope profiles in the area generally do not show evaporative profiles, characterized by increasing enrichment towards the surface (Allison et al., 1994). The profiles show that pure evaporation from depths below 2 m is very low. Therefore the stable isotope composition of soil moisture below this depth is characterized by the depleted isotope content of above normal rainfall events.

Results obtained through analysis of tritium activities indicate an average moisture flux of 16 mm/yr, slightly higher than the average value obtained from the chloride mass balance method. Discrepancies between the two methods seem to be due to the fact that tritium can be transported in both liquid and vapour phases while the chloride ions are only transported in the liquid phase. The difference between the two methods increases with decreasing recharge rates (Gieske et al., 1995). At higher moisture flux rates and higher saturation of sediments, anion exclusion may become important, leading to apparently higher recharge values obtained through the use of the chloride mass balance methods.

The generally low flux rates through the valley soils contrast with the generally fast reaction of the groundwater table to seasons with above average rainfall. Most recharge in the area therefore seems to occur along other avenues. Multi-modal recharge has been taken into account through groundwater flow modelling and through so-called mixing cell models (Adar et al., 1988, Gieske and De Vries, 1990).

3.4.2 Letlhakeng-Botlhapatlou (LB - average rainfall 420 mm/yr)

The area is located in the fringe of the Kalahari Basin about 100 km to the northwest of Gaborone (Figure 3.1) and is characterized by a flat topography intersected by several fossil valleys and small pans below which a major aquifer has formed in Karoo sandstones. Average depth to the water table is 50 m with groundwater flow towards the northwest.

Thirteen shallow profiles (1.5 to 11.5 m depth) representing various geomorphologic and vegetation settings were drilled with hand augers. Moisture was extracted from sediments in the laboratory using thoroughly tested techniques (vacuum distillation) and procedures to minimise inaccuracies that result from incomplete recovery of moisture (Obakeng et al., 1997). Samples were analysed for Cl, ^{2}H, ^{18}O and ^{3}H (Gieske et al., 1995; Beekman et al., 1997; Selaolo, 1998). One deep profile of 41.7 m was drilled with a multipurpose drilling rig deploying an advanced sampling technique (Beekman et al., 1997; Selaolo, 1998).

Shallow profiles

Results of some shallow profiles sampled in the LB area are summarised in Figure 3.6. Moisture fluxes in the area were calculated between 1 and 10 mm/yr, with an average of about 4 mm/yr, based on the chloride mass balance method. For the deepest profile LM, however, a higher flux of 18 mm/yr was calculated (Gieske et al., 1995). The absence of chloride peaks in this profile and the sudden decrease in moisture content at 9m depth suggest that flushing has occurred following heavy rainfall.

Large changes in chloride contents with depth, as illustrated by profiles LBA-7 and Matlolakgang, are attributed to both evapotranspiration and multi-modal moisture transport (diffuse and preferential flow). Assuming that downward moisture transport is adequately described with a bimodal flow model (Sharma and Hughes, 1985) the contribution of preferential flow to the total moisture flux in the area amounts to about 50% for the upper 10m of the heterogeneous zone above the water table.

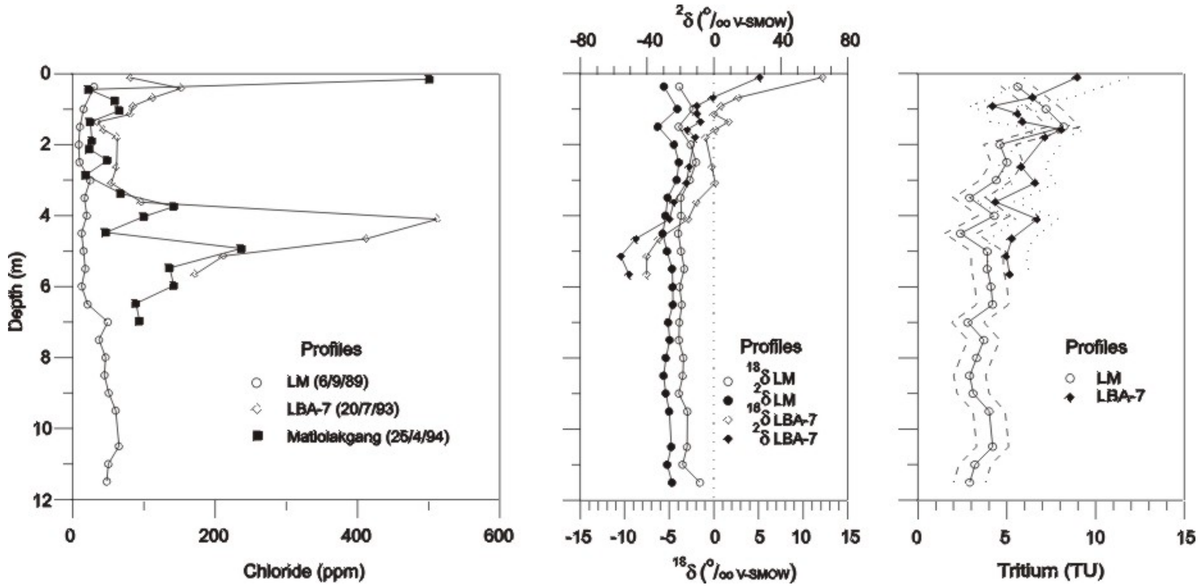

Figure 3.6 Letlhakeng-Botlhapatlou shallow profiles (Beekman et al, 1996; Selaolo, 1998).

Stable isotope profiles in the area generally show increasing enrichment towards the surface as a result of evaporation (see e.g. profile LBA-7; Beekman et al., 1997; Selaolo, 1998). The absence of such an isotopic enrichment for profile LM and the nearly identical stable isotopic composition of moisture over the whole profile support the interpretation of flushing after heavy rainfall. Displacement of the isotopic compositions of soil moisture from the LB-MWL below the zone of pure evaporation (~2 m) is generally constant and is proportional to the inverse of the square root of moisture fluxes (derived from chloride mass balances). Thus, it appears that the isotope displacement method proposed by Allison et al. (1984) to estimate moisture fluxes from stable isotope profiles can also be used for the Kalahari environment.

Analysis of tritium activities of moisture from profile LM reveals a moisture flux of 9.8 ± 2.2 mm/yr, which is lower than the flux determined through the chloride mass balance method (18 mm/yr) and is different from the results obtained for the NP area. As earlier indicated, flushing of the profile by heavy rainfall may explain the anomalous character of this profile. The tritium flux calculated for LBA-7 (5.5 ± 1.2 mm/yr) is higher than the chloride flux (2.5 mm/yr), which is in line with expected increased tritium transport in the vapour phase at recharge rates below 10 mm/yr (Cook and Walker, 1995; Gieske et al., 1995).

Long term monitoring (up to five years) of moisture contents by neutron probe measurements at several sites in the area has confirmed the occurrence of flushing at depths of up to 4 m following heavy, above average, rainfall (Beekman et al., 1997; Selaolo, 1998). From these observations, a general trend can be seen showing a threshold value of about 400 mm of rainfall below which hardly any soil moisture seems to penetrate below the main zone of evapotranspiration (see Figure 3.7).

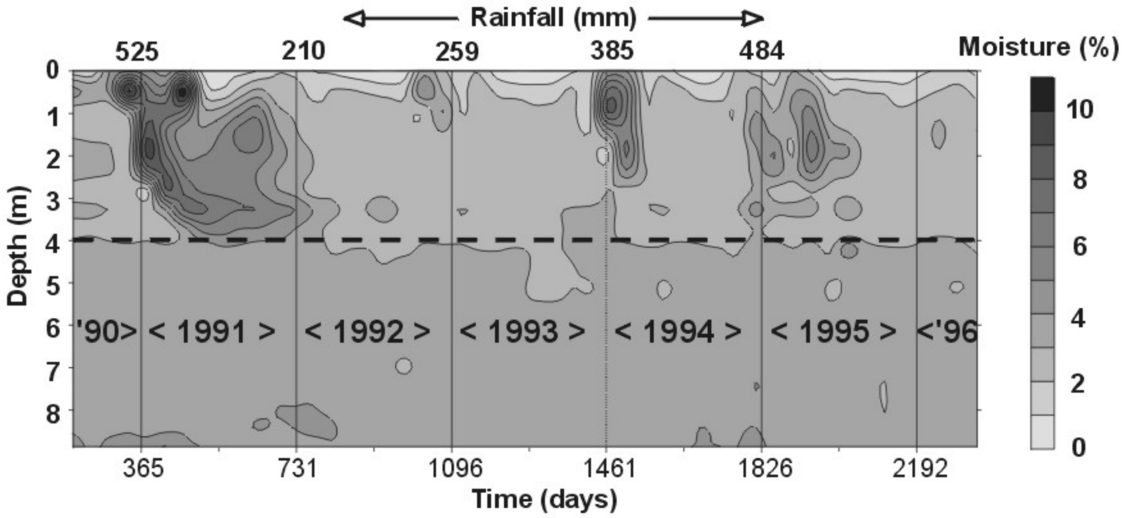

Figure 3.7 Volumetric moisture contents Maipatlelo profile (Beekman et al, 1997; Selaolo, 1998).

Deep profile

The deepest unsaturated zone profile (LB-3B), which so far has been augered with the GRES rig in the Botswana Kalahari, reached a depth of 41.7 m. The profile was augered in the middle of a palaeo-valley. Depth to the water table at the site is 54 m. The upper 7.5 m of the sequence consists of Kalahari deposits (mainly fine clayey and silty sediments and a 1 m thick calcrete layer near the surface) and is underlain by weathered and fractured Karoo sediments. Results of the multiple tracer profiling are shown in Figure 3.8. Below a depth of 6 m, chloride concentrations reduce drastically and indicate preferential flow of 'diluted' moisture bypassing the main zone of evapotranspiration. Lower chloride content in soil moisture below this depth compared with the chloride content in groundwater (100-125 mg/l) suggests that groundwater at this site is derived from a larger area than only the small valley course. Considering the entire profile, a moisture flux of 2.9 ± 1.2 mm/yr was computed based on the chloride mass balance method. Excluding the evapotranspiration zone, with its seasonal characteristics, a moisture flux of 8.6 ± 3.4 mm/yr was calculated.

The effects of evaporation are clearly restricted to the top 1 m of the profile as shown by the plot for stable isotopes. Displacement of the isotopic compositions of soil moisture from the LB-MWL below this zone is rather uniform: a linear fit through the stable isotopic compositions shows a nearly identical slope as the slope of the LB-MWL. A moisture flux of 2.0 mm/yr was found using the method of Allison et al. (1984).

The moisture flux derived from the tritium data, for the upper 6m interval, is 13.6 mm/yr, much higher than the flux determined by the chloride mass balance method for the same interval of 0.6 mm/yr. This supports again the observation that tritium transport in the vapour phase may become dominant at low recharge rates (<10 mm/yr). The 23 to 1 ratio between the tritium and chloride flux is the highest ratio thus far found in Botswana. It clearly demonstrates the inaccuracy of moisture flux calculation based on (uncorrected) tritium activities of moisture from the Kalahari semi-desert. Tritium activities of soil moisture, however, proved to be extremely valuable in this environment to elucidate transport mechanisms. Relatively high tritium activities, ranging from 3 to 6 TU, were found below the evapotranspiration zone at depths of 7 m, 17 m, 28-30 m and 38 to 41 m and indicate that preferential flow does occur to greater depth in the semi-arid environment of the Kalahari.

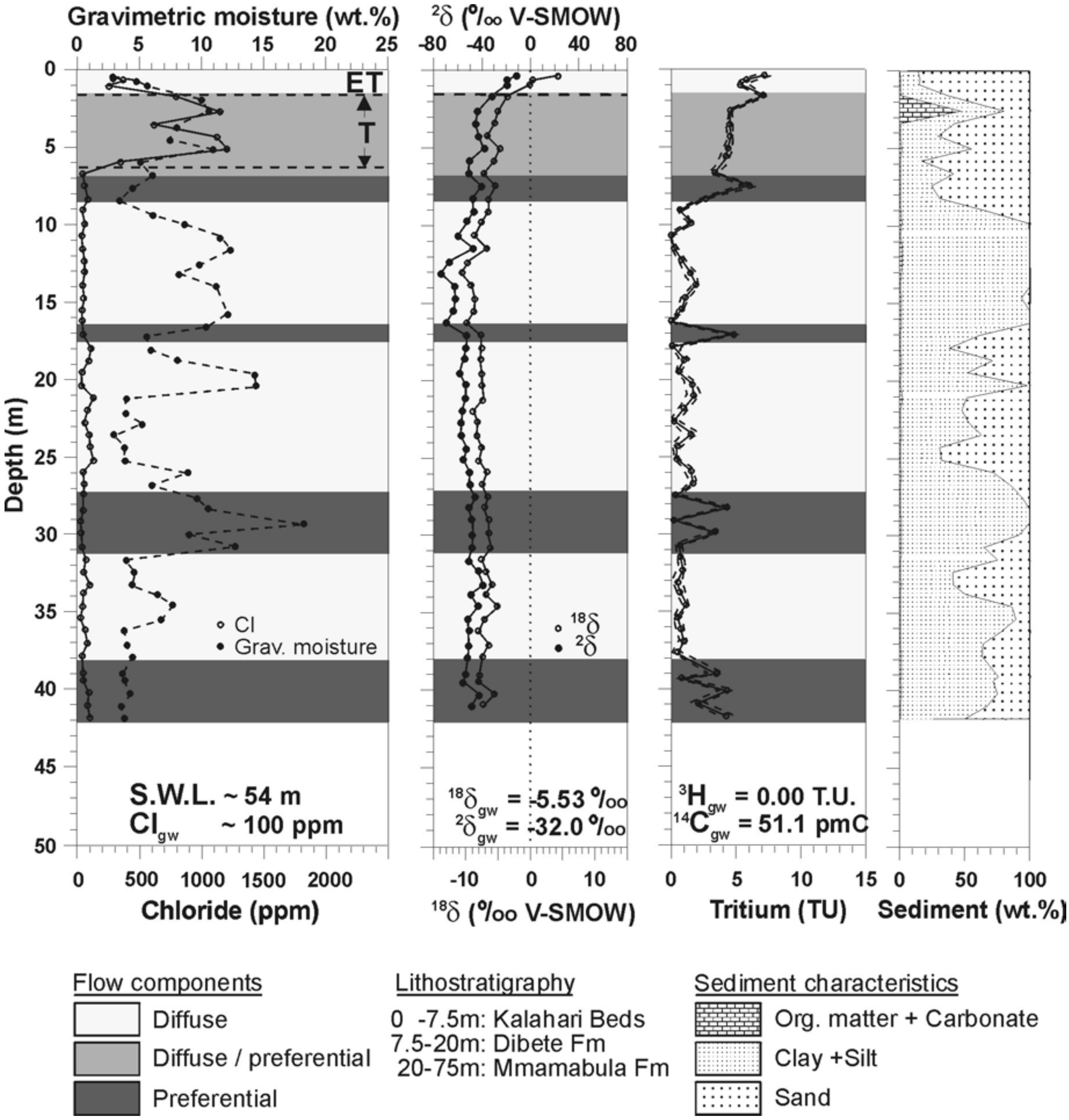

Figure 3.8 Multiple tracer profile LB-3B (after Selaolo, 1998).

3.4.3 Central Kalahari (CK - average rainfall <400 mm/yr)

A small area in the Central Kalahari, just north of the Khutse Game reserve, some 250 km northwest of Gaborone, was selected for a detailed moisture flux and recharge study (Figure 3.1). Pans and small linear dunes are the most prominent features in this part of the Kalahari Basin.

Deep profile

Two 28 m deep-cored holes, a few metres apart from each other, were drilled in 1995. Depth to the water table at the site is about 70 m. The upper 23 m of the sequence consists of Kalahari deposits (sands and calcretes) and is underlain by weathered and fractured Karoo sediments. Figure 3.9 shows the results of studies related to the CK-2B and CK-3B sediments and cores.

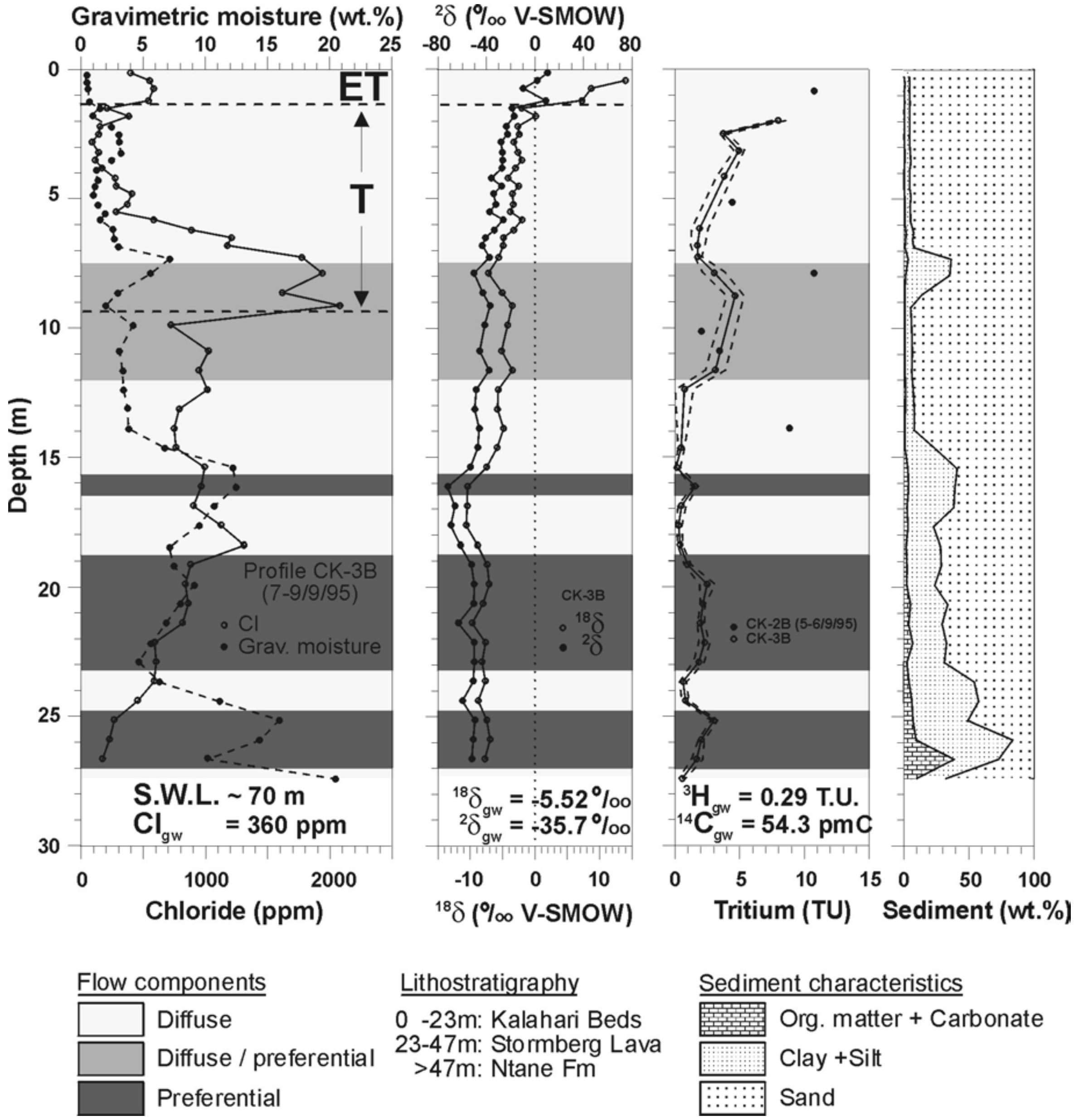

Figure 3.9 Multiple tracer profiles CK-2B and CK-3B (after Beekman et al., 1996).

High chloride concentrations in the upper 2 m of the profile indicate the main evapotranspiration zone. The pronounced peak between 6 and 9 m depth, corresponding to a layer of relatively finer grained sediment with higher moisture content, probably originates from transpiration. The chloride concentration decreases drastically below 9 m suggesting significant preferential flow. Further changes in chloride content with increasing depth may either be explained by changes in climatic conditions or by changes in relative contributions of different flow components. The existence of preferential flow was also deduced from macro-pores in the diamond-drilled core between 15 and 20 m depth. A moisture flux of 0.5 ± 0.1 mm/yr was calculated based on the chloride mass balance method for the upper 7.5 m of the profile. This figure is much lower than fluxes obtained in the NP and LB areas, and reflects probably the decreasing rainfall towards the Central Kalahari to below a critical seasonal precipitation threshold of 400 mm/yr.

The stable isotopes profile shows an increasing enrichment in the upper part and confirms that the thickness of the zone of evaporation is not more than 2 m. Displacement of the isotopic compositions of soil moisture from the CK-MWL (Figure 3.3) below this zone down to 7.5 m depth is rather uniform: a linear fit through the isotopic compositions resembles the slope of the CK-MWL. A moisture flux of 1.1 ± 0.1 mm/yr was obtained using the isotope displacement method.

The moisture flux derived from tritium data, for the same 7.5 m interval, is 3.8 ± 0.5 mm/yr, higher than the flux determined with the chloride mass balance and the stable isotope methods and again supports the observation that tritium transport in the vapour phase may become dominant at low recharge rates. Tritium activities of soil moisture proved again to be extremely valuable to elucidate transport mechanisms. High tritium activities of up to 4.6 TU for CK-3B relative to the surrounding levels for the following depth intervals: 7.5 to 12 m, 15.75 to 16 m, 18.75 to 23.25 m and 24.75 to 27 m clearly indicate a significant contribution of preferential flow to moisture fluxes and groundwater recharge. Note that changes in tritium activities in the lower part of the profile correlate well with changes in chloride concentration. Differences in tritium activities of profiles CK-2B and CK-3B suggest that preferential flow does not necessarily occur at the same depth level, despite the small distance between the profiles of only a few metres.

3.5 Summary and Concluding Remarks

Recharge processes were extensively studied in Botswana during the period 1987-1997 within the framework of the GRES project. Methods used included analysis of precipitation and evapotranspiration, study of environmental isotopes and rainfall chemistry, and analysis of transport processes in both the unsaturated and saturated zones.

Input functions for vadose zone studies

Through long-term monitoring of rainfall chloride content, total chloride deposition (TD) maps of Botswana were produced for the application of the chloride mass balance method. Although there seems to be general agreement on TD values in southeastern Botswana, it was shown during the second phase of the project that values for the northern part of the country may have to be adjusted upwards. In general, TD values not only seem highly variable in time but also appear to depend on sampling techniques and on the type of rain gauge, hence the need for continued monitoring and experimentation. We also recommend that the monitoring network be expanded to the entire Southern African sub-continent.

A decrease was found in the slope of local MWLs from the southeast of Botswana towards the Kalahari Basin, which may be explained by an increase in evaporation. Good agreement was also found with data from the Pretoria station in South Africa. Measured rainfall tritium activities were in accordance with model predictions (Doney, 1992).

Multiple tracer profiling

Extensive studies of moisture transport through the unsaturated zone were made through the analysis of a large number of shallow and deep profiles, sampled for chloride, stable isotopes and tritium. Results from the studies indicate that moisture fluxes generally decrease with decreasing rainfall from southeastern Botswana: 11 mm/yr on the average with 500 mm/yr rainfall to the Central Kalahari: 0.5 mm/yr with less than 400 mm/yr rainfall. Moisture fluxes were shown to be highly variable spatially and of an essentially multimode nature with slow diffuse flow through the top soils, and relatively fast preferential flow through soil cracks, root channels and fractures. Evidence for multimode moisture transport is given indirectly by the higher recharge rates derived from saturated zone recharge estimation (Gieske, 1992; Selaolo, 1998; Beekman et al., 1999), and directly by elevated tritium activities at great depths. Significant amounts of post-bomb tritium were found in two deep profiles from the LB and CK areas, proving for the first time, that preferential flow does occur up to great depths (41 m) even under the climatic and environmental conditions of the Kalahari. Considering the uncertainty in (a) depths of the root zone with its upward as well as downward flow, (b) the occurrence of multimode flow and (c) the total chloride deposition, neither the relatively shallow profiles (<10 m) nor the deep 27.8 m CK-3B and 41.7 m LB-3B profiles, warrant a detailed reconstruction of palaeo-recharge. Note that the contribution of preferential flow to the total moisture flux in the LB area for instance, assuming a bimodal flow model, was estimated at over 50% for the upper 10m of the heterogeneous zone above the water table (section 3.4.2).

Results of the chloride mass balance method indicate that moisture fluxes in fossil valley systems are lower than fluxes in topographically higher interfluve areas. Regionalization of moisture fluxes and recharge rates, however, remains a problem. Furthermore, uncertainty with respect to the long-term chloride deposition values and exceptional climatic events causing sudden flushing, complicate comparison between various profiling methods.

The displacement of the stable isotope composition of moisture from the local MWL is inversely proportional to the square root of the moisture flux in accordance with Allison et al. (1984). This seems to be only valid in the drier parts of the country.

Tritium transport appears to become dominated by vapour transport when fluxes decrease below 10 mm/yr. Therefore, systematically higher flux rates are found when applying the tritium method for uncorrected tritium activities than when the chloride mass balance method is used. Figure 3.10 illustrates this for a large number of studies in Australia, USA, Israel and Botswana (Gieske et al., 1995; Beekman et al., 1996).

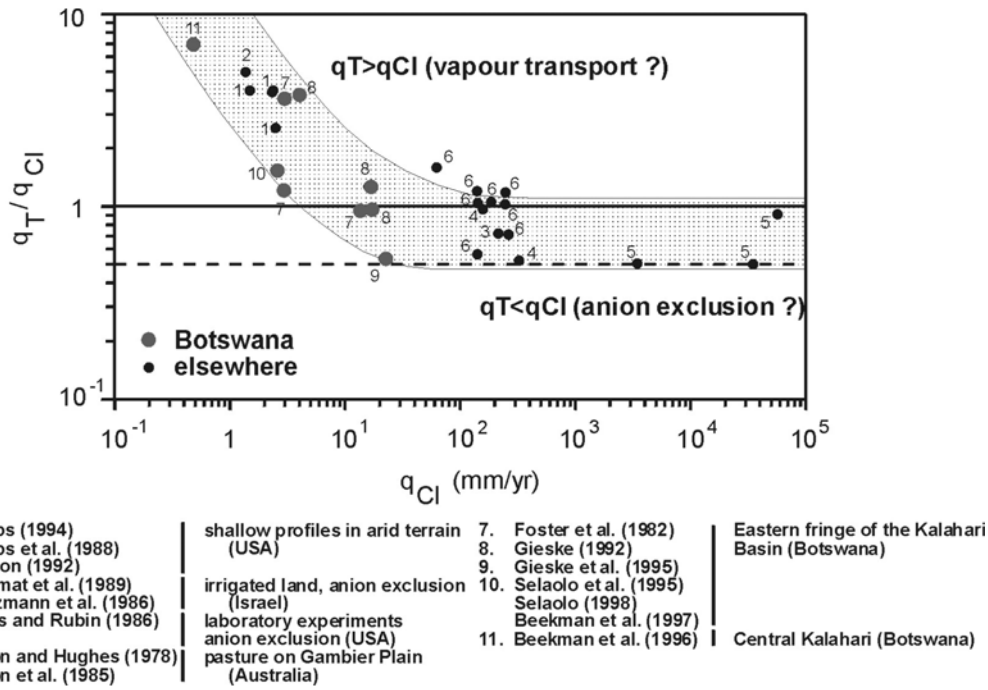

Figure 3.10 Comparison between moisture fluxes determined by the Chloride Mass Balance and Tritium Profiling methods (Beekman et al., 1996).

In conclusion, the tritium method cannot be used straightforward for determination of slow diffuse infiltration when flux rates are below 10 mm/yr. The use of tritium as a tracer in elucidating recharge mechanisms through preferential pathways remains, however, extremely valuable.

3.6 Acknowledgements

This paper is based on the results of a long-term co-operation programme on groundwater recharge and water resource evaluation in Botswana (GRES: 1987-1997) between the Botswana Geological Survey Department (DGS), the University of Botswana (UB-Geology Department) and the Vrije Universiteit Amsterdam, The Netherlands (VUA - Centre for Development Cooperation Services and Faculty of Earth Sciences). The Botswana and Netherlands Governments jointly funded the GRES project.

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4. Recharge Quantified with Radiocarbon in Three Studies of Karoo Aquifers in the Kalahari and Independent Corroboration

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ABSTRACT Environmental isotope data from a "snapshot" sampling hold out the promise of producing acceptable estimates of ground water recharge for resource management purposes. In three major groundwater development projects in Botswana, estimates of recharge to the Karoo aquifers in the Kalahari were based on residence times derived from radiocarbon (^{14}C), supported by other isotope data. In the assessment, three factors needed to be considered: 1) the model leading to acceptable values of residence times 2) the initial, or recharge, radiocarbon value and 3) appropriate values of aquifer porosity.

In each of the three studies, porosity had been measured on numerous drill cores obtained from the principal fractured sandstone aquifers. The resulting isotope-based recharge values correspond reasonably with independent recharge assessments using the equal volume method to analyse long-term rest level observations in two cases; in the third, recharge was independently assessed on the basis of the chloride mass balance in both unsaturated and saturated zones. It is concluded that a) the isotope snapshot approach can give acceptable values for recharge in the development of groundwater resources and provides rational management information early in the life of a groundwater supply scheme; b) the exponential model and an initial radiocarbon value of 85 % atmospheric are realistic in this environment and c) the effective porosity appears to be the appropriate parameter in the calculation of recharge. This also provides an insight into the behaviour of the aquifers.

4.1 Introduction

The quantification of recharge is increasingly given attention in the management of groundwater as the ultimate measure of the long-term sustainability of the resource. Traditionally, recharge can be assessed on the basis of long-term accumulation of water level and abstraction data in e.g. the equal volume method of interpreting hydrographs. Since their discovery in the environment some 50 years ago, cosmogenic radioisotopes in groundwater provided the possibility of assessing the residence time or the rate of recharge to aquifers (e.g. Libby, 1955). The assessment of recharge using the "isotope snapshot" approach is attractive in principle as it can provide up-front information on recharge that can be built into the management plan at the inception of exploitation.

In the isotope approach a mean residence time has to be calculated, based on some concept of transit time distribution. Amongst the available models: piston flow, dispersion and exponential (e.g. Zuber, 1986; Verhagen et al., 1991) the latter has been generally employed as a suitable "lumped parameter" model for phreatic aquifers. This approach is regarded as particularly appropriate where the borehole mixes all the water it intersects. Otherwise, little control can be exercised over sampling conditions, as existing boreholes have to be employed, often of differing and poorly known construction. The only control which can be exercised is to sample from boreholes in constant production or during extended pumping tests.

Tritium (^{3}H) is a conservative tracer, simple to interpret, although useful only in shallower and more actively turned-over groundwater, particularly in the southern Hemisphere. In sedimentary, mainly silicious, terrain such as the deeper multi-layer Karoo aquifers of the Kalahari, with mean residence times typically in excess of 1000 years, radiocarbon is useful as

1) initial (recharge) concentrations can be estimated with some confidence and
2) its subsequent behaviour in the saturated zone can be regarded as effectively conservative.

The well-mixed model assumes that the pumped well produces a mixture of different ages or transit times from the different lithological units penetrated. In a simple approach, recharge R can be estimated:

$$R = \frac{\sum_i n_i H_i}{T} \tag{1}$$

where T is the mean residence time (derived by the exponential model), n_i the porosity and H_i the thickness of the individual saturated units (i) penetrated.

The question arises as to which value of porosity should be employed. Most sedimentary aquifers in southern Africa have both primary and fracture porosity. In pump test analysis, the derived storage coefficient tends to increase with increasing pumping time (months to years; Verhagen et al., 1999), as deeper seated water is gradually released from the pores. It is argued here that, at an appropriate time scale, an environmental tracer effectively labels all the water in the aquifer by advection or diffusion. The approach taken is to use the effective porosity. Porosity is not always measured directly on aquifer material in groundwater investigations, but clearly is of fundamental importance in the meaningful isotope assessment of recharge.

The problem of the behaviour of a tracer in a fractured porous medium has been discussed extensively by Zuber (1994). When water is transported mainly through the fractures, the tracer will diffuse into the more or less stagnant water in (micro) pores. In this way, the transport of tracer is delayed, the relevant retardation factor R_p being given by:

$$R_p = \frac{t_t}{t_w} \equiv \frac{n_p + n_f}{n_f} \tag{2}$$

where t_t and t_w are the transit times of tracer and water and n_p and n_f the matrix and fracture porosities respectively. As usually $n_p >> n_f$, high retardation factors can be attained. This may lead to tracer-based mean residence times of an order of magnitude - or more - too high.

Isotope data was collected in three groundwater investigations in Karoo aquifers of the KaIahari over the past 20 years, allowing recharge estimates to be made. Long-term conventional geohydrological observations are now available in two cases, which allow recharge to be calculated from well field hydrographs. In the third case, recharge was calculated using the chloride balance method. In the three investigations reviewed below, acceptable correspondences were found between the isotope-based recharge figures, using the criteria mentioned above, and those obtained from other techniques, based on different criteria.

The three investigations were conducted in Botswana (Figure 4.l), respectively at Jwaneng diamond mine (Verhagen, 1987; 1993), Orapa diamond mine (Mazor et al., 1977; Verhagen and Morton, 1989) and the PaIla Road wellfield extension (Wellfield Consulting Services, l994).

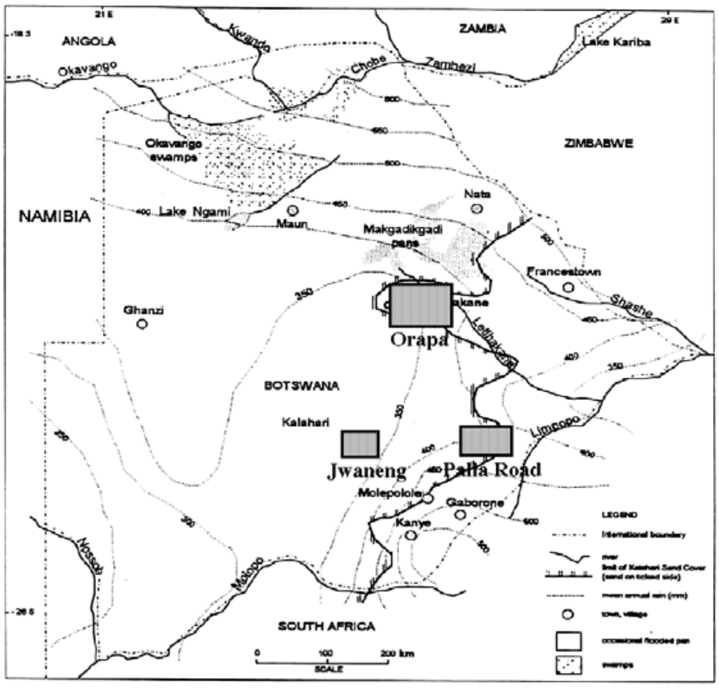

Figure 4.1 Map of Botswana, showing isohyets and the three wellfield study areas.

4.2 Hydrogeology

The generalised geological section of the Karoo aquifers of the Kalahari basin is shown in Figure 4.2. The principal aquifers are the more ubiquitous Ntane (or Clarens) fine grained aeolian largely non-calcareous sandstone, which can be locally indurated and cemented, overlying the Ecca (carboniferous) sequence of sandstones, siltstones and mudstones with locally better developed coarser grained sandstone horizons. Overlying the Ntane, except at the basin margins, is the Stormberg basalt. This unit confines the underlying sandstone aquifers and is aquiferous mainly due to jointing and fracturing. Along the fault structures in the south and north, major fracturing has produced considerable secondary porosity in the different aquiferous units, which are often juxtaposed through extensive block faulting, resulting in local hydraulic discontinuities. The entire basin has a cover of semi-consolidated Kalahari Beds, mainly sand and calcretes, which constitute minor, often perched, local aquifers.

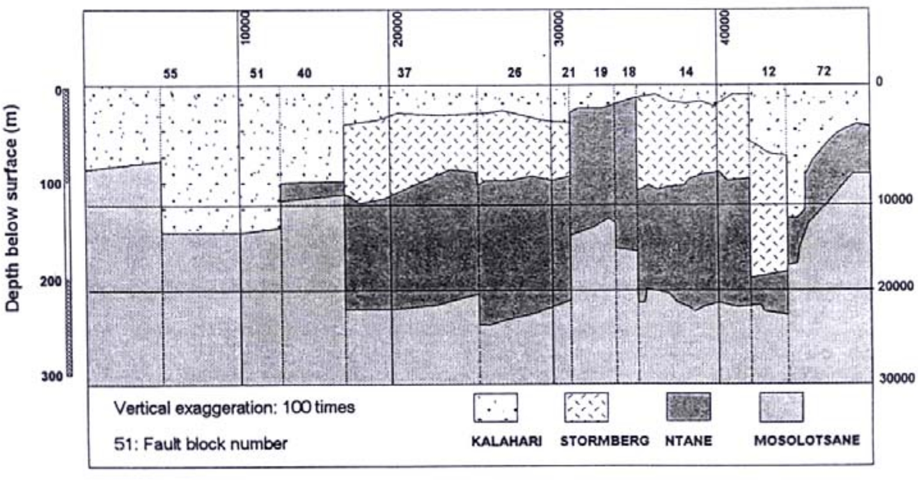

Figure 4.2 Generalised Karoo lithology showing typical block faulting, such as found at Orapa and Palla Road (after Blecher and Bush, 1993).

4.3 Initial – or Recharge – Radiocarbon Value

On account of the heterogeneous and complex recharge pathways (Beekman et al., 1996) in the Kalahari cover, it is not readily apparent which of the available hydrochemical/isotopic models to employ in order to "correct" radiocarbon values for residence time calculations. A survey of the Kalahari conducted in 1971/72 when thermonuclear tritium was still evident in groundwater systems in the southern Hemisphere (Verhagen et al., 1974) provided empirical evidence of 85 % of atmospheric for the radiocarbon content of recently recharged water. Such values have been used by other workers in similar environments (Vogel, 1967) and have been employed in calculations of mean residence times in this study.

4.4 Jwaneng Mine Wellfield

The well field supplying Jwaneng mine in southern Botswana taps a fluviatile sandstone aquifer. The structure, interpreted as a fault-controlled delta or alluvial fan in the Ecca facies, dips northwestwards below an increasing thickness of a mudstone/siltstone aquitard (Figure 4.3). The land surface is flat and underlain by some 20 metres of Kalahari deposits.

Exploitation of the well field commenced in 1980 from 14 wells and has been increased to some 107 m^3a^{-1} from 28 high-yielding wells at present. Radiocarbon measurements on water from production wells in the early 80's in the range of 55 pMC to 75 pMC showed ongoing recharge, clearly disproving the belief that no significant modern recharge occurs in the Kalahari (Foster et al., 1982). The geographical distribution of radiocarbon values appeared paradoxical, as they increase in the direction of increasing confinement of the sandstone. A few kilometres further north, deeper village boreholes intersecting the same sandstone unit give vanishing radiocarbon (Figure 4.3).

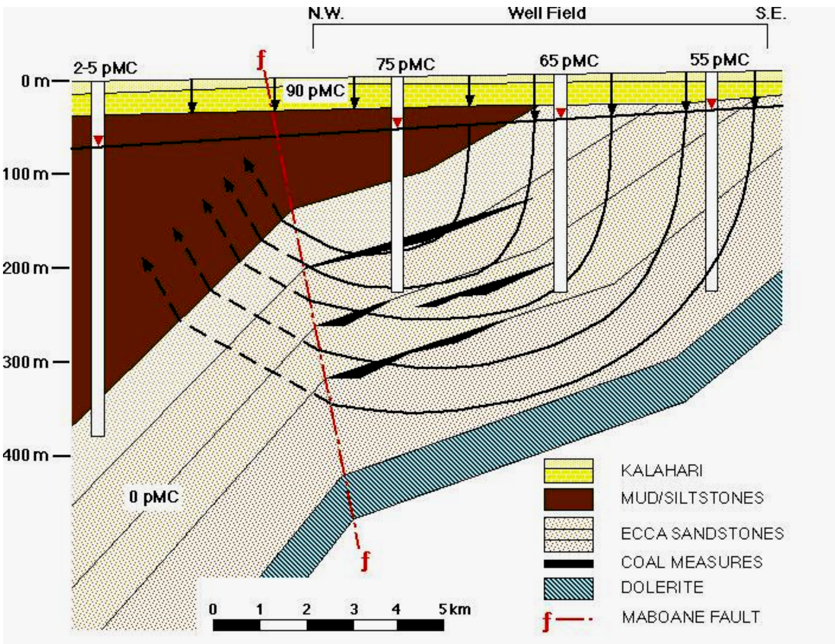

Figure 4.3 Schematic section of aquifer tapped by the Jwaneng well field, with well field and deep village boreholes, mean radiocarbon values and inferred flow lines.

These observations led to a first conceptual model of the aquifer's hydrology shown as schematic flow lines in Figure 4.3. Probably infrequent recharge to the sandstone sub-crop beneath the Kalahari Beds cover would produce mounding which drives groundwater down-dip in a northerly direction. The well field boreholes, all drilled to about 230 m, intersect the entire aquifer and full range of flow lines in the SE. NW-wards, boreholes only partially penetrate the aquifer and intersect only the more recent flow lines, which results in increasing ^{14}C values. The flow is clearly arrested between the well field and the deep village boreholes further northwards at ~0 pMC. The input from recharge should therefore be balanced by leakage into the aquitard over some 100 km^2 of contact with the sandstone.

Time series (Verhagen, 1993) of radiocarbon for several production boreholes over the period of full exploitation (1983-1993) show no significant changes (Figure 4.4), which suggest that flow relationships have not been modified significantly over that period.

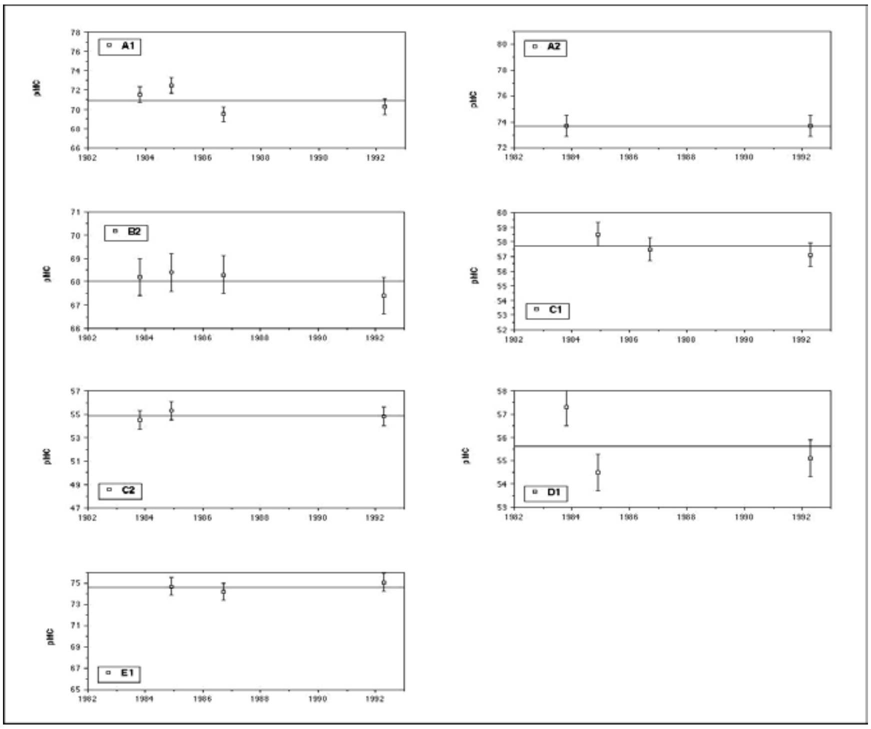

Figure. 4.4 Radiocarbon time series for production wells 1983 to 1992.

Recharge values were calculated using exponential model mean residence times based on radiocarbon values, giving 3.7 mm a^{-1} in the SE (at 55 pMC) to 4.7 mm a^{-1} in the NW (75 pMC). A mean porosity value of 15% was used, based on the effective porosity measured on several drill cores obtained from the sandstone aquifer during original well field development.

An analysis of a 12-year hydrograph of average water levels for the well field (Figure 4.5) revealed a number of periods of equal water level (van Rensburg and Bush, 1994). Recharge was calculated on the basis of the equal volume method, which assumes that the abstraction over the period equals the recharge. A best fit of the data for the relationship between inferred recharge and the rainfall over the same period was obtained for a storativity of 0.0008, giving a recharge of 1.4% of the mean annual rainfall of some 350 mm. In the modified Hill method, assuming a 200 mm threshold of rainfall before recharge can occur, a recharge of 1.98% of rainfall and a storativity of 0.005 is obtained. The long-term apparent S-value, given by the ratio (cumulative abstraction)/(cumulative dewatered volume) over 12 years, approaches 0.02. The range of resulting recharge values of 3.0 – 4.9 mm a^{-1} overlaps the range determined from isotope observations.

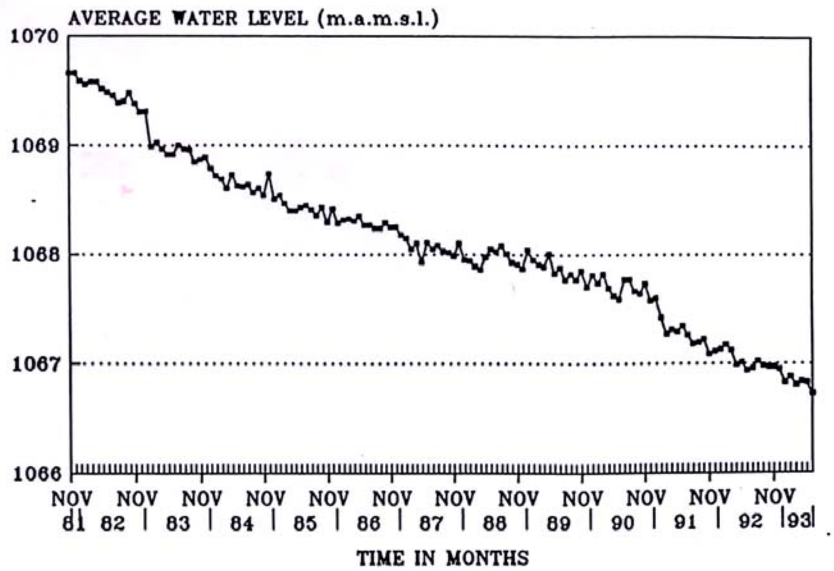

Figure 4.5 Mean hydrograph for the Jwaneng well field (after van Rensburg and Bush, 1995).

It is clear therefore that the figure for storage increases with increasing time scale. In the short-term (months) responses, the (confined) storativity applies. In the (partial) dewatering of the aquifer on the time scale of years an "effective" porosity or specific yield applies. For the recharge estimates using radiocarbon, the more appropriate factor appears to be storage represented by the effective porosity. Implicitly included in the storage relevant to the isotope approach could well be the inferred reverse leakage from the aquitard.

4.5 Orapa Mine Wellfields

Various well fields were established to supply the Orapa diamond mine in northern Botswana since the latter sixties (Figure 4.6). The fractured and jointed basalt cover (up to 80 m) confines the underlying Ntane sandstone aquifer locally, but allows widespread hydraulic connection between it and the shallow overlying Kalahari Beds aquifer, as seen in hydrographs (Mazor et al., 1977). During exploitation, the sandstone aquifer shows confined behaviour. In the isotope approach, the system was treated as a multi-layer phreatic (i.e. diffusely recharged) aquifer, with rest levels in the basalt. Slow lateral groundwater movement, under a regional piezometric gradient of 0.001 in the sandstone aquifer is further restricted on account of extensive block faulting and intrusion by diabase dykes. Radiocarbon values are found to lie in the range 1.3 - 45 pMC, with one outlier at 75 pMC. Only protracted pump tests were sampled in an attempt to reduce the localising effect of individual boreholes and transient mixtures (Verhagen et al., 1999).

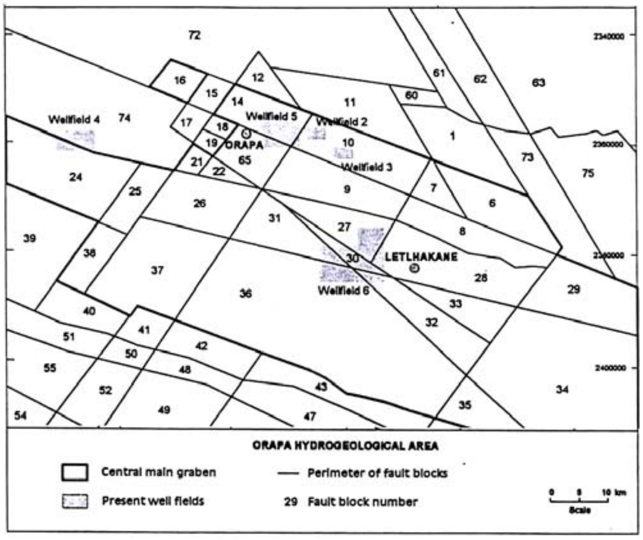

Figure 4.6 Map showing positions of Orapa mine well fields and major block faulting (after Blecher and Bush, 1993).

Porosities ranging from 17 % to 30 % were measured on numerous borehole cores, with a mean of 22 %. The bulk fracture porosity of the basalt is unknown, but conservatively was taken as 2%. Recharge was calculated using radiocarbon data and the exponential model. Even in the absence of significant regional flow, age stratification can be produced by areally restricted intense rain events (Figure 4.7), which are likely to be the major recharge mechanism in the area (Verhagen et al., 1999). Recharge values range from 0.6 to 2.4 mm a^{-1} with a single value of 21.5 mm a^{-1} for a well that intersected a major basalt fracture. Including this value a mean of 3.7 mm a^{-1} is obtained; eliminating the presumed outlier gives a mean of 1.1 mm a^{-1}.

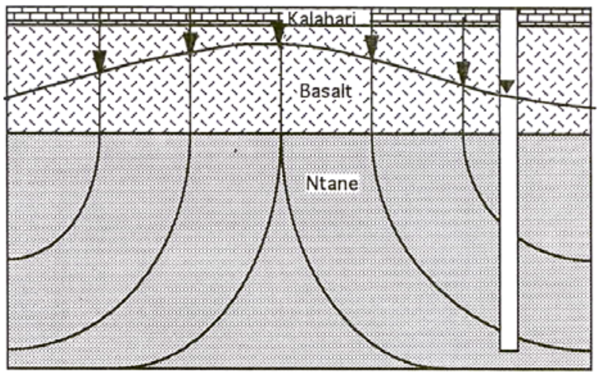

Figure 4.7 Schematic model showing mounding from recharge from a local rainfall event which produces age stratification in the underlying sandstone aquifer.

Analysing more than 20 years of rest level and rainfall data, good simulations of piezometric level fluctuations could be achieved with cumulative rainfall departure methods, taking into account the abstractions from the well fields (van Rensburg and Bush 1995). With the equal volume method the saturated volume fluctuation data for one of the well fields provided recharge values, which are plotted against corresponding integrated rainfall in Figure 4.8. This gave (confined) storativities of 0.00022 to 0.00126 and recharge values of 2.7 mm a^{-1} to 5.8 mm a^{-1}. Considering the uncertainties in parameters used in both the isotope and rest level fluctuation approaches, and the limited number (9) of documented isotope sampling points, a reasonable correspondence in the recharge figures was obtained.

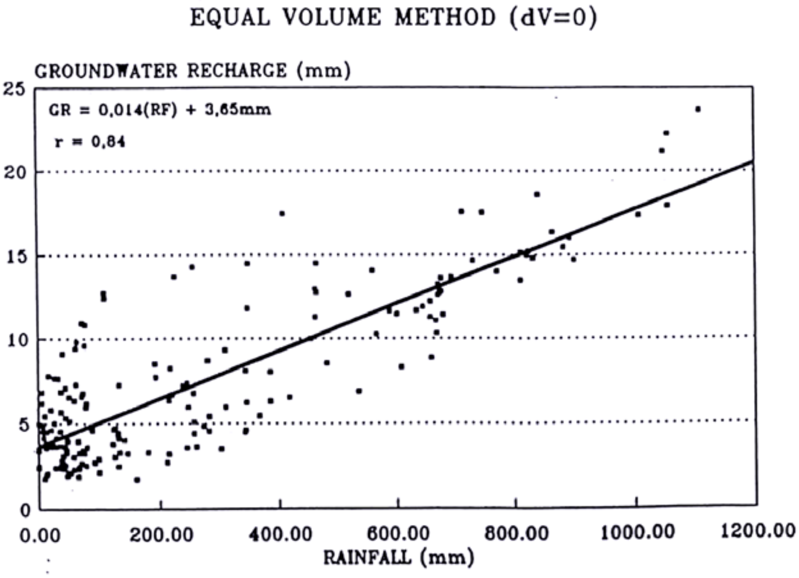

Figure 4.8 Recharge calculated for equal volume periods read off a four year hydrograph for Orapa well field 4, plotted against rainfall for the same period (after van Rensburg and Bush, 1995).

Close on 30 years ago, when water management was still predicated on the principle of groundwater mining, isotopic data had shown that significant recharge is occurring to the confined sandstone at Orapa (Verhagen, 1974; Mazor et al., 1977). Recharge quantification on the basis of radiocarbon has now produced figures for recharge, which are roughly compatible with, but lower than, those obtained by the equal volume method. All the model parameters used effectively maximise the former. An explanation may be found in the low and probably heterogeneous flow in the sandstone aquifer as seen in near-vanishing radiocarbon values obtained for some first strikes at the basalt-sandstone contact. On the other hand, one borehole, which intersected a basalt fissure, gave a value of 75 pMC in the underlying sandstone.

Although the exercise has underlined the problem of an isotope assessment of recharge to a confined aquifer, the radiocarbon snapshot approach did produce realistic, if possibly conservative, recharge estimates.

4.6 Palla Road Wellfields

As part of an investigation for the extension of an existing well field in eastern Botswana, isotope data was obtained from a number of existing and newly drilled boreholes. The main aquifer is the aeolian Ntane sandstone, a fractured unit usually covered by basalt, and extensively block-faulted. In contrast with the previous two cases, the extensive study area of some 1650 km^2 has both a topographic and piezometric gradient of some 3 ‰. Groundwater radiocarbon values, in many cases with measurable tritium, cover a wide range, with a preponderance of values between 65 pMC and 110 pMC (Figure 4.9). Groundwater flow is quite heterogeneous. High radiocarbon values at depth in major fractures suggest substantial (natural) fracture flow. There is evidence also of substantial intergranular flow in major sandstone blocks.

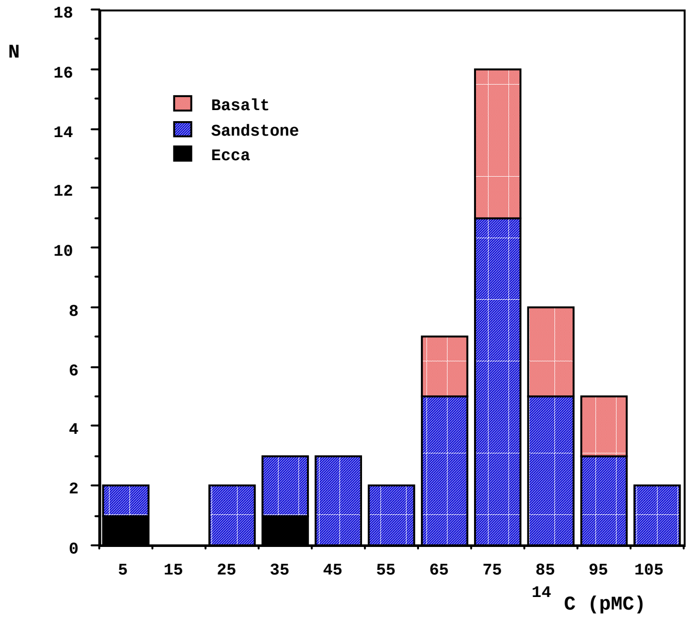

Figure 4.9 Frequency histogram of radiocarbon values for Palla Road boreholes and their distribution in the three main lithological units.

Some 52 measurements of porosity were performed on a 142 m length of core of Ntane sandstone obtained from a single borehole. The mean value of porosity obtained is 0.012 with a standard deviation of 0.004. As at Orapa, the fracture porosity of the basalt is taken to be 2 %. Recharge calculated only for those 24 boreholes with adequate lithological data gave a mean value of 6.6 mm a^{-1}. This value, higher than found at Jwaneng and Orapa, is ascribed to the more active topography-driven, lateral groundwater transport and higher rainfall (Figure 4.1).

Two independent recharge estimates were made (Wellfield Consulting Services, 1994). The first was based on the chloride balance of groundwater, which is justified in view of ongoing, regional flushing of the aquifers. Sampling sites were selected to ensure minimal contamination by intrusion of saline water from the deeper aquifers. With figures for rainfall and dry deposition of chloride (Beekman et al., 1996), chloride-based recharge values range from 4.5 mm a^{-1} to 8.5 mm a^{-1} with a mean of 5.9 mm a^{-1}. The second was made assessing the unsaturated zone chloride balance in a limited number of shallow (2 metre) soil profiles, from which recharge values in the range of 2 mm a^{-1} to 5 mm a^{-1} were determined. These values approximate, but are lower than, those obtained from radiocarbon measurements from the saturated zone. This could be further evidence that diffuse recharge tends to follow preferential flow paths.

4.7 Discussion and Conclusions

The problem of fractured porous aquifers addressed in this paper was theoretically treated by Zuber (1994) in terms of micropore diffusion, which would give unrealistically high tracer-based mean residence times. However, satisfactory correspondence was achieved between recharge figures obtained using a) the exponential model b) 85 % atmospheric for the initial or recharge value of radiocarbon c) the effective porosity on the one hand and independent techniques in each of the three case studies on the other. These results, summarised in Table 4.1, lend weight to the approach taken in this study employing isotope-based recharge calculations. This approach seems to apply to the relatively higher residence times addressed with radiocarbon, where the assumption of the uniform labelling of all the water in the aquifer by the tracer appears to be justified. Whether this assumption would be valid for the much shorter time scales addressed with tritium is open to investigation.

Table 4.1 Comparison of isotope-based and independent assessments of recharge.

Site	Isotope-based recharge assessment (mm/a)	Independent recharge assessment (mm/a)	Method of independent recharge assessment
Jwaneng well field	3.7 (SE) – 4.7 (NW)	3.0 - 4.9	Equal volume; Modified Hill
Orapa well field	3.7 (high) ; 1.1 (low)	2.7 - 5.8	Equal volume
Palla Road well field	6.6 (mean)	5.9 (mean) 2.0 - 5.0	Chloride balance, saturated Chloride balance, unsaturated

The implication of these results for the aquifer behaviour of the fractured sandstones of the Kalahari is that on the time scales of radiocarbon, the transport through the matrix of the sandstone is sufficiently active to account for effectively all the tracer in the output of wells which often are located on fractures to enhance yield.

Classical hydrograph methods of recharge assessment require many years of rest level observations, along with an appropriate choice of the storage factor. Unsaturated zone methods tend to underestimate recharge and evidence mounts of the importance of preferential flow. Restricted lateral groundwater mobility, with resulting predominance of vertical water losses – common in this environment - may generate chloride values which again would provide unrealistically low recharge estimates.

This study has demonstrated that the concept of the “isotopic snapshot” can give a meaningful first estimate of recharge. Even where there are no direct measurements of aquifer porosity, an estimate based on general knowledge of the aquifer material can still provide a useful recharge value for management purposes.

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PART III

Recharge Estimation Related to the Saturated Zone

5. Perspectives on Recharge Estimation in Dolomitic Aquifers in South Africa

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ABSTRACT Groundwater has a history of neglect and unsustainable utilization in South Africa. Despite recent developments in policy, and the importance of groundwater as a strategically important resource on which many rural communities depend, water management in South Africa remains largely determined by surface water systems. An exception is the dolomitic aquifers of South Africa, which for many towns, rural areas and farming are the only water resource. This paper reveals some of the results and characteristics of dolomitic aquifers that have been obtained using a variety of methods and ancillary techniques.

5.1 Introduction

Assessment of groundwater recharge and aquifer storage is the key element in the evaluation and management of such resources. Research projects on recharge estimation undertaken in the Republic of South Africa (RSA) have been supported by the Water Research Commission and the Department of Water Affairs and Forestry, and dolomitic aquifers have been the focus of many of these studies (Figure 5.1).

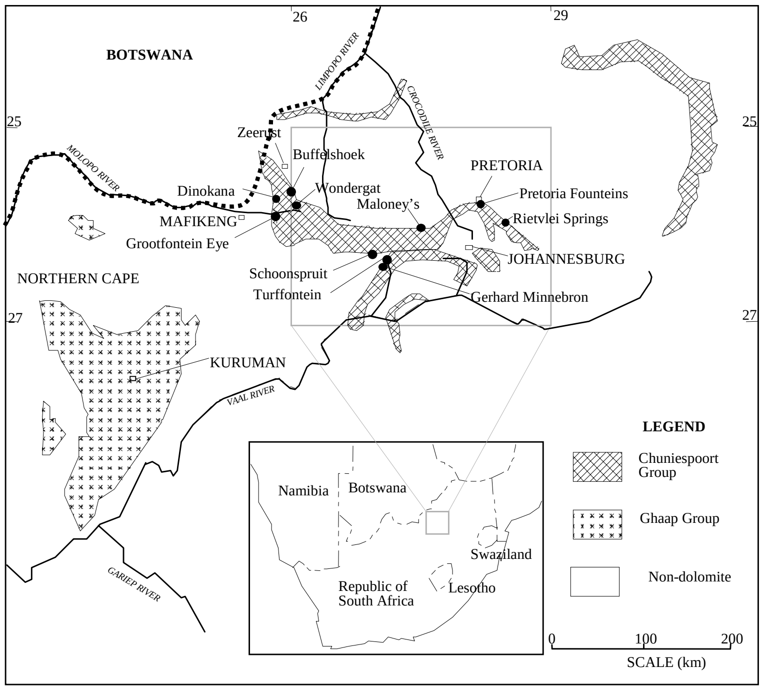

Figure 5.1 Locality map showing the dolomitic aquifers of South Africa and some of the important springs and towns that are partially or fully dependent on dolomitic groundwater.

The importance of groundwater recharge estimation in groundwater resources evaluation and in environmental impact assessments is reflected in Table 5.1.

Table 5.1 Degree of relevance of recharge estimation to groundwater resource evaluation and environmental impact assessment.

Groundwater Resources Evaluation	Environmental Impact Assessment
. allocation of abstraction permits (high) . aquifer utilisation (high) . risk assessment (high) . boreholes yields (high) . harvest of springs (high) . aquifer storage (medium) . conjunctive use (medium) . borehole protection (low)	. aquifer vulnerability to pollution (high) . dewatering engineering (high) . base flow (high) . interaction with wetlands/lakes (high) . groundwater contamination (medium) . instream flow requirements (low) . vegetation (low) . rehabilitation (low)

Strengths, weaknesses, opportunities and threats regarding assessment and utilization of the groundwater exploitation potential, are the following:

Strengths: Significant progress in quantifying recharge using different methods

- publications, manuals and reports on the estimation of recharge and storativity;
- broadened perspectives on hydrological interactions providing a greater understanding of how groundwater systems function;
- progress in estimating aquifer storativity more reliably, especially in dolomitic aquifers that cover the greater part of the RSA.

Weaknesses: Lack of a multi-disciplinary approach to groundwater management

- viewing the components of the hydrological cycle in isolation and not holistically in an integrated way;
- interdependence between recharge and aquifer storativity, which complicates the resolution of water level responses;
- pursuing complex methods rather than focussing on the solution of practical problems and aquifer management;
- devoting too much time to refining recharge estimations over short-term time intervals;
- insufficient integration of physical and tracer techniques for recharge estimation.

Opportunities: Collation of knowledge and its application in groundwater management

- collating information which has been gathered in the RSA over many years and from studies such as GRES in neighbouring Botswana (Gieske, 1992; Selaolo, 1998; Beekman et al., 1996; 1999);
- extending studies of recharge as part of new projects on groundwater monitoring;
- more extensive application of hydrodynamic simulation of dolomitic aquifer systems on a regional scale despite inadequacies in the resolution of parameters such as recharge and storativity;
- more effective characterization of aquifer systems according to borehole water level response using simplistic approaches (such as the cumulative rainfall departure method);
- enhance the application of mass transport models as better estimates of groundwater recharge are becoming available;
- incorporating knowledge acquired into hydro(geo)logical education and training.

Threats: Poor management of aquifers

- difficulty in implementing the National Water Act leading to uncontrolled abstraction, over-exploitation and deterioration of groundwater resources;
- aggravation of pollution problems and effects of contamination on aquifer systems;
- neglecting the role of groundwater in sustaining the ecological balance;
- losing perspective on groundwater systems by using highly sophisticated modelling packages and methods.

5.2 Perspectives on Groundwater Recharge Estimation

Recharge is governed by the intricate balance between several components of the hydrologic cycle, each of which is a function of several controlling factors:

- Rainfall: f(intensity, frequency, variability, spatial distribution)
- Evapotranspirative losses: f(temperature, wind, humidity and phreatophites)
- Discharge losses: f(interflow, springs, base flow, lateral flow and artificial discharge)
- Catchment: f(soil type, thickness, spatial distribution, topographical feature, vegetation)
- Geology: f(rock types, characteristics of fracture networks, occurrence of dykes)

This has resulted in the development of complex hydrological models to accommodate for the interrelationships between groundwater, surface water and pollution. The complexity of the models often limits their practical value as does a lack of real data, which often have to be approximated by regional parameters or estimates by an expert. Studies in the RSA have shown that a simple approach can reveal much about groundwater recharge, and that sparse data should not necessarily be a constraint.

The need to simulate the high variability of recharge over short intervals of time is not as critical as is generally perceived, because annual values of recharge can be converted into monthly values by applying annual average recharge coefficients to monthly rainfall. This is a valid approach because groundwater stored underground behaves in a similar fashion to a surface water reservoir in smoothing the variability of monthly run-off. This is illustrated by Figure 5.2, which shows the correspondence between the flow of the Pretoria Grootfontein springs fed from dolomite, and the surface inflow into the Rietvlei Dam averaged over several years, notwithstanding the high variability of monthly runoff. Linear correspondence between groundwater levels and spring flow was observed and indicates that the groundwater hydrographs also reveal smoothing of short-term variable recharge.

Analysis of the water balance of a surface or groundwater reservoir often does not require refined assessment of the short-term variable inflow because the integrated inflow/recharge in relation to rainfall over longer time intervals generally conforms to a linear relationship. The averaging period producing a linear response is characteristic of the catchment, regardless of whether or not it yields surface flow or groundwater recharge. Studies of the response of springs to rainfall have revealed much about the recharge characteristics of aquifers as has been demonstrated in the case of the Dinokana spring (Xu et al., 1993). The flows of the dolomitic springs in a specific month corresponded remarkably well to the average antecedent rainfall over several months. Similarly, monthly groundwater levels conform to the average rainfall over a number of preceding months.

The dependence of both groundwater levels and springflow on the average antecedent rainfall requires that estimates on water balances must also incorporate this average rainfall, for example when using the equal volume method for recharge estimation.

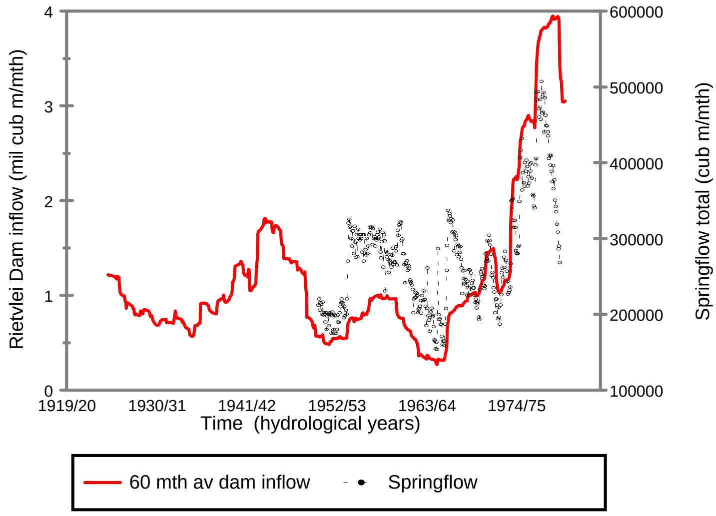

Figure 5.2 Comparison between the moving average inflow over 60 months to the Rietvlei Dam and the flow of the Grootfontein eye near Pretoria.

5.3 Methods for Recharge Estimation in Dolomitic Areas

Recharge is one of the most difficult parameters to determine reliably by water balance methods, because of the dependence of groundwater levels on both recharge and aquifer storativity. The interdependence of these parameters makes it difficult to find a unique solution for either recharge or storativity without having reliably determined the other parameter by an independent method. When the drawdown-response to abstraction from pumping tests is used for determining storativity in dolomitic aquifers, the concept of effective radius must be taken into account, otherwise it can lead to unreliable results. Much progress, however, has been made in deriving better estimates of S values (Verwey et al., 1995; Botha et al., 1998), by using empirical approaches (Bredenkamp et al., 1995) and hydrodynamic modelling.

Several methods can be used to quantify groundwater recharge but the reliability of the estimates is often questionable because of simplifying assumptions and uncertainties of some key parameters that are required. A wide range of methods has been incorporated as part of a WRC project which culminated in the compilation of a manual (Bredenkamp et al., 1995). New perspectives have since been gained in a follow up study on groundwater monitoring, which has focussed on the cumulative rainfall departure and moving average methods (WRC-K5/838).

In the following sections we will highlight experiences gained in the application of those methods that have proven to be practical and reliable in recharge estimation in dolomitic areas despite their sometimes simplistic approach.

5.3.1 Hydrochemical and isotope tracer methods

Chloride Mass Balance (CMB)

This method entails determining the recharge coefficient from the ratio of chloride in rainfall to chloride concentrations at the deepest point in the soil profile. It can be represented by the equation:

$$RE_{av} = a \times \frac{CL_{rf}}{CL_{soil\ moisture}} \qquad (1)$$

where RE_{av} is average annual recharge; a is recharge coefficient; and Cl is chloride concentration. Critical to its application is the chloride concentration of the rainfall input.

A simplified approach was adopted using the data collected from rainfall in Botswana. The average annual chloride input of that rainfall appears to match that of a few analyses of inland rainfall in the RSA. Results show a reasonably consistent relationship with annual rainfall. The chloride input, usually expressed as total deposition (mg/m^2), has been converted to mg/l of rainfall. A similar relationship to that of inland rainfall also appears to hold for rainfall sampled along the Zululand coastal region (Figure 5.3). Chloride concentrations of rainfall in coastal regions are significantly higher than those of rainfall from the inland.

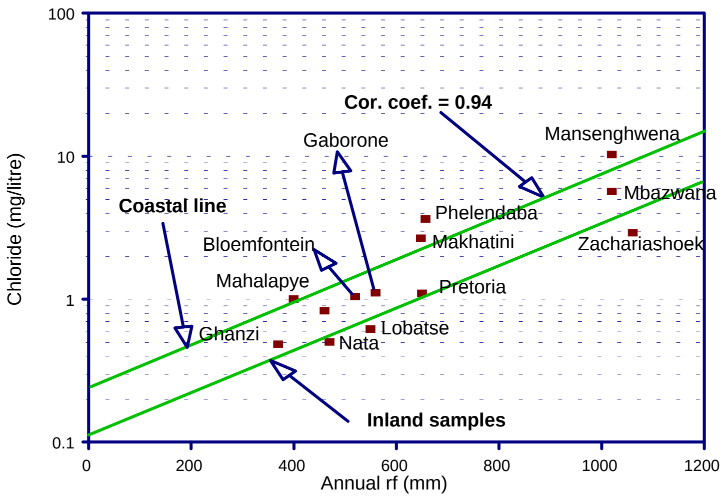

Figure 5.3 Average annual concentrations of chloride derived from rainfall collected in Botswana and elsewhere in the RSA.

The chloride method has been applied in different regions in the RSA and results from chloride profiles in the unsaturated zone portray a surprisingly consistent relationship between rainfall and potential recharge (Figure 5.4), which could be used as a first approximation of the average recharge in a region. Note that Kuruman, Grootfontein and Zuurbekom represent dolomitic aquifers.

The variability of chloride concentrations in relation to fluctuations of annual rainfall, based on measurements at Lobatse of Botswana, shows a decline in chloride concentration as the annual rainfall increases. This conforms to a reduction in chloride concentrations as more prolonged rainfall showers occur as well as higher rainfall, which produces higher percentages of recharge. However, a sample of mixed rainfalls would still produce an acceptable average value of the recharge. Further measurements of chloride in rainfall would therefore help to refine estimates of recharge by means of the chloride method.

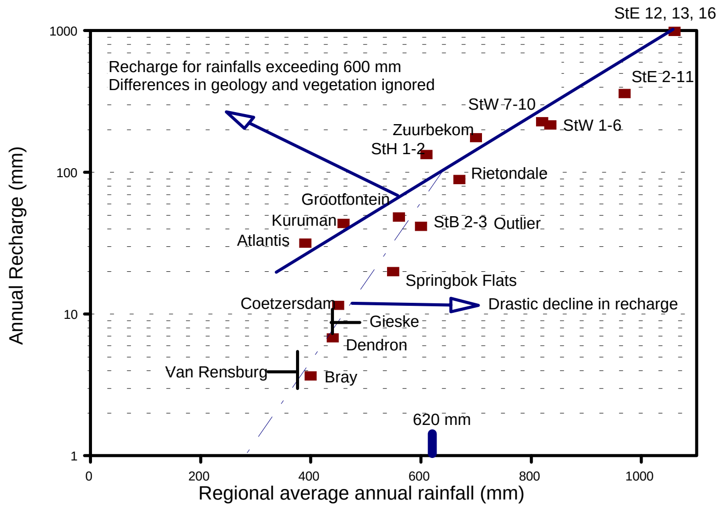

Figure 5.4 Regional rainfall-recharge relationships derived from chloride profiles in the unsaturated zone.

The CMB method can also be applied to the saturated zone to estimate a 'true' total recharge originating from both diffuse and preferential flow components through the unsaturated zone. Recharge over an entire drainage area can be determined by integrating ratios of chloride contents in rainfall to that of groundwater over the whole area. This requires that both spatial chloride input of rainfall and the chloride in groundwater be determined reliably. This method has been applied by Gieske (1992) and Sami and Hughes (1996) but has not been used extensively in the RSA although its wider application to estimate spatial variability of recharge is promising.

An elegant application of the CMB method is to make an estimate of recharge from the ratio of chloride concentration of rainfall to that of water discharging from a spring. Recharge rates would then be integrated both spatially and temporally. The reliability of the assessments based on the chloride concentrations of springs may be checked by means of cumulative plots of the temporal variations of chloride (Figure 5.5). Chloride concentrations for Maloney's eye (or spring) for instance show that little contamination has yet occurred, whereas the Upper Turffontein eye, suggests that significant contamination has occurred, thereby invalidating application of the chloride mass balance method.

Tritium Profiling (TP)

Assumptions that apply for this method are basically the same as for the chloride profiling method, i.e. infiltrating water displacing water in the unsaturated zone in a piston-like manner and that no preferential bypassing of flow to greater depths occurs. Rainwater that infiltrated in 1962/63 may be distinguished by its higher tritium concentrations. Comparison of this water with moisture from the post 1962/63 period may reveal information on moisture fluxes or potential recharge. This method has been successfully applied in the Bo Molopo area where average recharge was estimated between 5% and 8% of the average annual rainfall (Bredenkamp et al., 1974; Bredenkamp et al., 1995). In the Rietondale area of Pretoria, tritium profiles have yielded a potential recharge of about 8% and in the Atlantis area about 10% of the average annual rainfall. Verhagen et al. (1974; 1984) and Vogel and Van Dijken (1974) have applied the method on a limited scale.

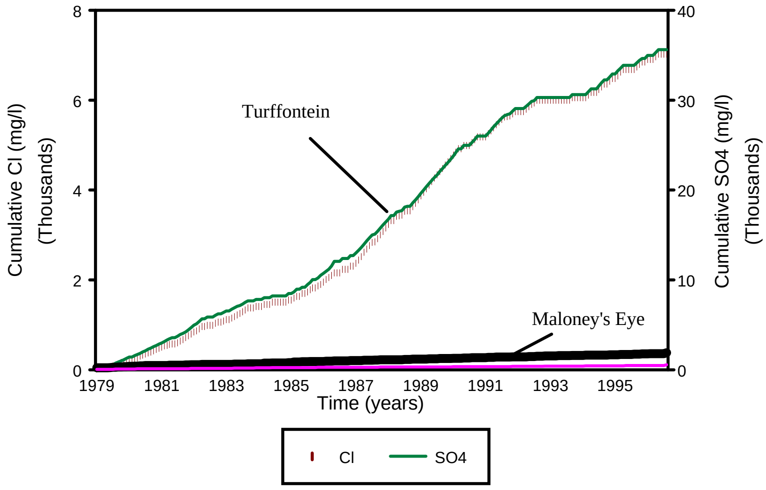

Figure 5.5 Cumulative plots of chloride and sulphate concentrations vs. time for the Upper Turffontein and Maloneys eyes.

In Botswana, the tritium profile method yielded recharge estimates which agree well with chloride profiles for moisture fluxes in excess of 10mm/year (Gieske, 1992). Applicability of the tritium method is presently limited by inadequate thickness of soil, causing the 1962/63 spike to be no longer detectable. However tritium concentrations, which have also been drastically decayed due to the short half-life time, have become difficult to determine accurately.

Characterising recharge by means of hydrochemistry and isotope physics

The chemistry of water discharging from a spring provides valuable information on the nature of the recharge areas. Hydro(geo)chemistry can be used as a vital supplement to enlightening the groundwater balances of spring catchments and the reliability of estimates of recharge. Sulphate concentrations generally indicate the extent of pollution from mining activities and a comparison of sulphate and chloride concentrations may reveal imbalances in the relative contributions of contaminants. On a grand scale, the ratio of rainfall to evaporation determines the concentration of dissolved constituents and of chloride, thus providing a relative index of recharge.

Springs exhibiting lower TDS and bicarbonate concentrations are partially recharged by non-dolomitic formations, which may indicate higher rates of recharge. The Pretoria Fountains dolomitic aquifer receives part of its recharge from the weathered granites of the large granitic dome (see Figure 5.1). The Rietvlei springs are also fed by recharge from both dolomite and the Pretoria shale formations and quartzite. The Maloneys dolomitic spring has the lowest bicarbonate concentration of all the dolomitic springs of which Barnard (1997) has shown that part of the recharge is derived from the Transvaal Formations. The relative contribution of non-dolomitic recharge has been estimated at about 16% of the total. In the case of Kuruman eye, the variations of the chemical composition of spring water clearly indicates that about 40% of the recharge is derived from the Asbestos Mountains although this recharge area is about one-third of the total recharge area. This has been derived from a simple two-box mixing model, which has provided similar results to those obtained by a complex mixing model (Adar and Kotze, 1997).

Recharge, chemistry and ^{14}C concentrations

The chemistry of dolomitic waters may assist in explaining anomalous ^{14}C concentrations in the following puzzling cases:

- ^{14}C concentrations of several springs are relatively low, despite frequently occurring recharge. ^{14}C values also remained relatively constant over many years in spite of higher bomb ^{14}C inputs (see Figure 5.6). The presence of tritium in some springs of which flow responds to rainfall, clearly suggests shorter turnover times than what is interpreted from ^{14}C concentrations. Note that the plotted ^{14}C concentrations of the Grootfontein eye in Figure 5.6 were shifted by 48 months as these lagged in comparison to the springflow.

- In springs such as Turffontein and Gerhard Minnebron, contamination of the aquifer is evident, in spite of the ^{14}C concentrations of the spring water having remained fairly constant. In a few dolomitic springs, ^{14}C contribution from bomb origin has been found. In the case of Kuruman eye clear inputs of higher tritium and ^{14}C have occurred although the higher concentrations were not sustained.

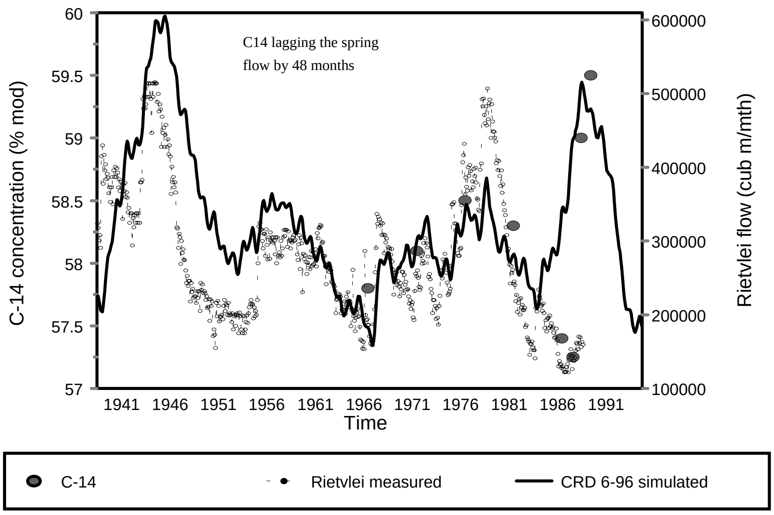

Figure 5.6 ^{14}C concentrations of the Grootfontein eye in relation to the measured and simulated flows.

An alternative explanation for the observed ^{14}C concentrations can be derived from examining the hydrochemical evolution of groundwaters. In spite of most dolomitic springs in the RSA having reached saturation in bicarbonate, no carbonate precipitates are found at the outlets of springs, as is the case with most carbonate aquifers elsewhere in the world. It appears that rapidly recharging water in dolomitic terrain partially bypasses the biological soil zone where equilibrium of ^{14}C and carbonate solution is normally established. The slightly acidic water dissolves dolomite rock with zero ^{14}C concentration, as the water moves through karstic channels and fractures. This results in lower than "normal" ^{14}C concentrations, which erroneously may be ascribed to radioactive decay of ^{14}C. Bicarbonate and chloride concentrations of springs show correspondence, thus further supporting this hypothesis and suggesting that bicarbonate concentrations also provide a measure of recharge in these cases (see Figure 5.7). Low concentrations of bicarbonate in dolomitic aquifers manifest that rapid recharge occurs which bypasses the zone of biological activity yet, upon entering cavernous dolomite, dissolution takes place, increasing the carbon content in groundwater.

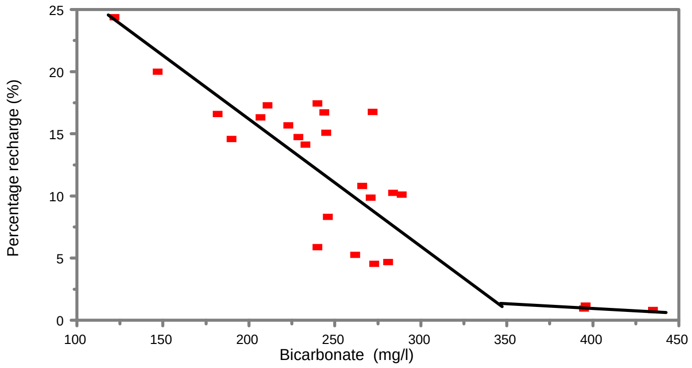

Figure 5.7 Recharge (determined by CMB method and expressed as % of average annual rainfall) for dolomitic aquifers in relation to bicarbonate concentrations.

The preceding provides an explanation as to why contamination is observed in springs such as the Turffontein and Gerhard Minnebron whilst ^{14}C concentrations have remained fairly constant. In the case of Kuruman eye, the discharging water is a mixture of recharge from two catchments having different chemical signatures. During normal rainfall conditions, the relative contributions of recharge from the two catchments are balanced, but during periods of high rainfall, a greater proportion of recharge is contributed from the surrounding mountainous area, which has a different chemical and isotopic signature. This results in an increase in ^{14}C and tritium in the spring water. This temporary imbalance is gradually restored to the original ratio as normal flow is restored. Even if the characteristics of the recharge from different sub-catchments are different, the isotopic and chemical signature of the water would not change much.

Plotting the ^{14}C concentrations against the alkalinity of spring water (Figure 5.8) shows that ^{14}C concentrations increase as the alkalinity increases. This is in agreement with the previous finding that the lower the alkalinity, the more the ^{14}C concentrations have been 'diluted'. Apart from this effect, the admixing of bomb ^{14}C will also be a function of the turnover time of the aquifer, which further complicates interpretations based on ^{14}C concentrations of spring water.

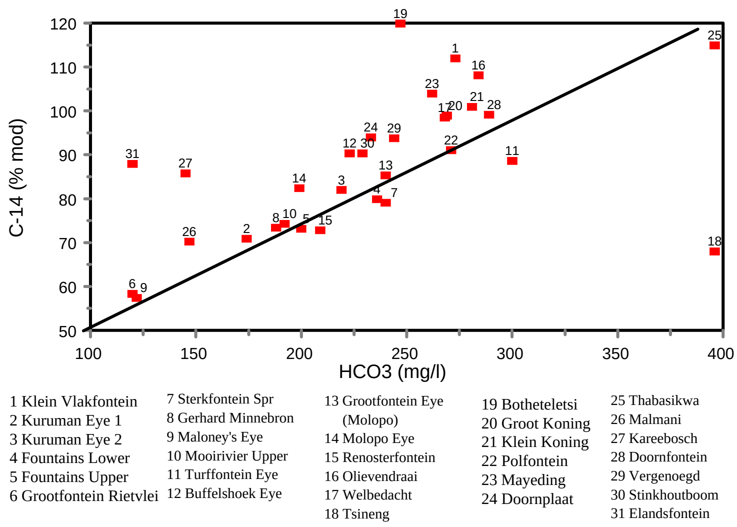

Figure 5.8 Concentrations of ^{14}C plotted against alkalinity (HCO_3).

Outliers in Figure 5.8 can be explained as follows:

- Elandsfontein: part of its recharge is derived from shale and quartzite formations;
- Tsineng: this eye appears to have a component of old water;
- Botheteletse: emanating from an aquifer with a small turn-over time - ^{14}C probably increased by admixture of bomb ^{14}C and 'biological' contamination at the outlet.

Dating of groundwater from dolomitic aquifers by ^{14}C should thus be treated with great caution. In view of the general support of the use of natural isotopes in groundwater studies, this has been examined more extensively in a WRC project (Contract No. K5/838) on groundwater monitoring (Bredenkamp, 1999). The value of ^{14}C measurements in dolomitic areas appears to be more in the bomb-increase being used as a marker to identify the cumulative recharge that have accrued since the introduction of bomb ^{14}C to the atmosphere. ^{14}C concentrations of springs that have been monitored only conform to the input of ^{14}C if the values are plotted incorporating a lag time. In the case of the Grootfontein spring at Rietvlei in Pretoria, the lag time was estimated at 48 months.

5.3.2 Physical methods

Equal volume (EV)

An estimate of the average recharge for a delimited aquifer can be obtained by means of the equal volume method whereby the aquifer storativity is eliminated from the water balance equation (dV=0):

$$\text{Inflow} - \text{Outflow} + \text{RE} - \text{Q} = \text{S} \times \frac{\text{dV}}{\text{dt}} \qquad (2)$$

where Inflow is the average lateral inflow; Outflow is the average lateral outflow; RE is groundwater recharge; Q is the net discharge from the groundwater system which could either be pumping or the discharge from a spring or both; dV is change in saturated volume of the aquifer; dt is the time increment over which the water balance is calculated and S is aquifer storativity.

The recharge is equal to the abstraction during periods when the saturated volume of the aquifer at the beginning and end of the period are the same (dV = 0). These periods can be identified in two ways, namely:

- for the status of equal-saturated-volumes derived from a composite hydrograph of the integrated groundwater level fluctuations, as measured at different monitoring boreholes spread over the aquifer;
- from periods of equal-status of spring flows.

Because the water level fluctuations in some cases respond to rainfall over long periods, the recharge derived in this way will not reveal the true variability of recharge, unless it is related to the correct causative rainfall. The longer the equal volume period, the more consistent the recharge estimates will be as it approaches the long-term average recharge, which conforms to a linear relationship.

In the case of the Rietpoort dolomitic aquifer near Zeerust different rainfall/recharge relationships for periods before and after 1983 have been found (Figure 5.9). The discrepancy most likely originates from the diversion of natural flow of the Molopo spring into a pipeline as from 1983.

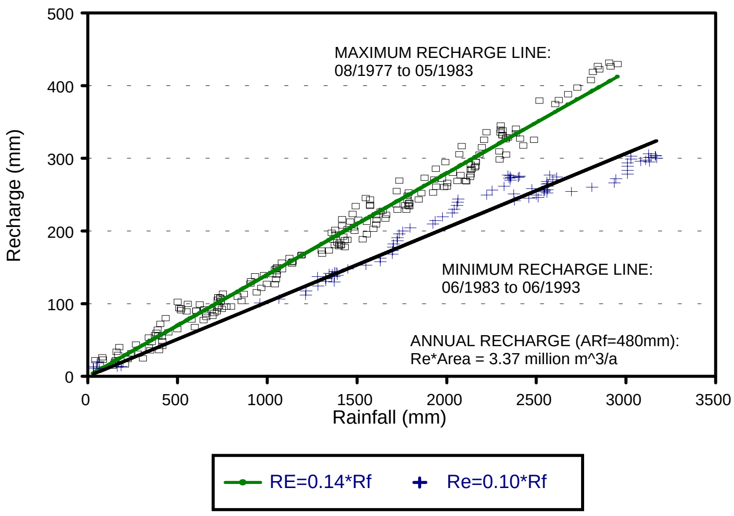

Figure 5.9 Groundwater recharge of the Rietpoort compartment near Zeerust based on the equal volume method.

In the case of the Buffelshoek eye, the EV method revealed clear evidence for an exponential rainfall-recharge relationship. Note that in case of discrepancies, e.g. due to abstraction or poor measurements, the rainfall-recharge relationship shows a clear shift (Figure 5.10), which provides a means of checking the reliability of the data.

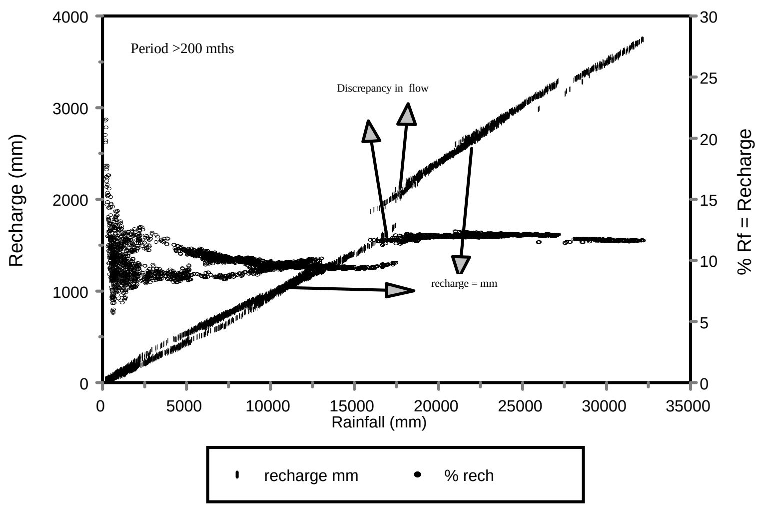

Figure 5.10 Anomalies in rainfall-recharge relationships of springs.

Other physical methods

Methods that also have been applied to dolomitic aquifers include the Modified Hill (MH), Cumulative Rainfall Departure (CRD) and the Moving Average Rainfall (MA) methods. The Modified Hill method represents a simplified version of the groundwater balance and aims at estimating storativity; the CRD method provides a means to mimic groundwater levels; and the Moving Average Rainfall method relates groundwater levels to moving average rainfall. Further details of these methods are given in Bredenkamp et al. (1995).

The CRD method has been developed into the most advanced one as compared to the other two methods. It not only gives the temporal response of groundwater levels to recharge, but also enables the derivation of aquifer characteristics. The method incorporates a short-term and long-term memory of the aquifer, which better conforms to the natural hydrological balance of an aquifer than the moving average method (MA). The coefficient of recharge, which is either used by or derived from the CRD method, represents an integrated average value. It can be derived when the storativity of the aquifer has been determined reliably. An examination of storativity values and inferred aquifer depth, derived from an analysis of groundwater hydrographs in the Zeerust, Grootfontein, dolomitic area in the Northwestern Province, has revealed an exponential decrease in storativity with depth (Figure 5.11). Alternatively, the storativity could be ascertained if the recharge would have been estimated, e.g. by means of the chloride method.

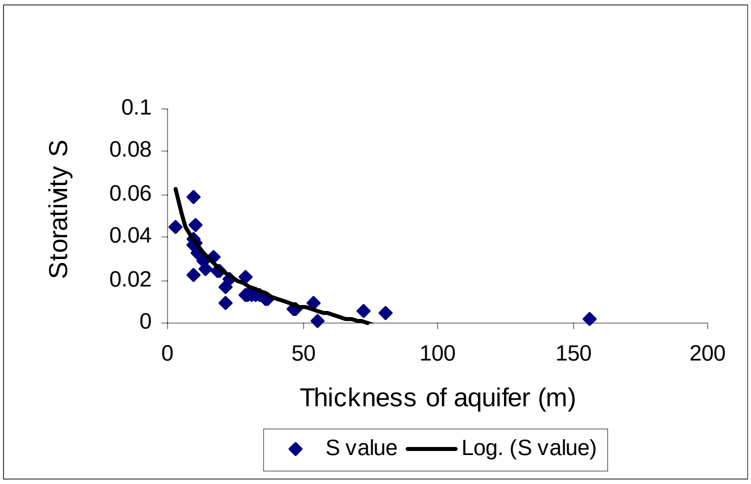

Figure 5.11 Change in storativity with depth for a dolomitic aquifer in the Northwest Province of RSA.

Recently, the CRD method was revised to accommodate for short-term rainfall records (Xu and Van Tonder, 2001). As shown in Figure 5.12, simulation using this new CRD method demonstrates that the responding water levels lag the rainfall events by many months. In the case of Wondergat, the best simulation encompasses a lag of 23 months as is also shown by the simulation of groundwater levels according to the moving average method.

Although the CRD and MA methods yield simulations that correspond closely, the CRD relationship represents a more acceptable hydrological model of the groundwater balance. Both methods, however, help to acquire a better understanding of the groundwater balance apart from determining critical parameters of the aquifer. The MA method could also be used to simulate the response of the aquifer to different scenarios of recharge and abstraction.

The CRD and MA methods provide a useful check of the reliability of water level and spring flow records, provided that reliable rainfall records are available. The flow of springs could best be simulated by the MA estimate of recharge and the area of the aquifer. Missing data of both water levels and spring flow can therefore be simulated.

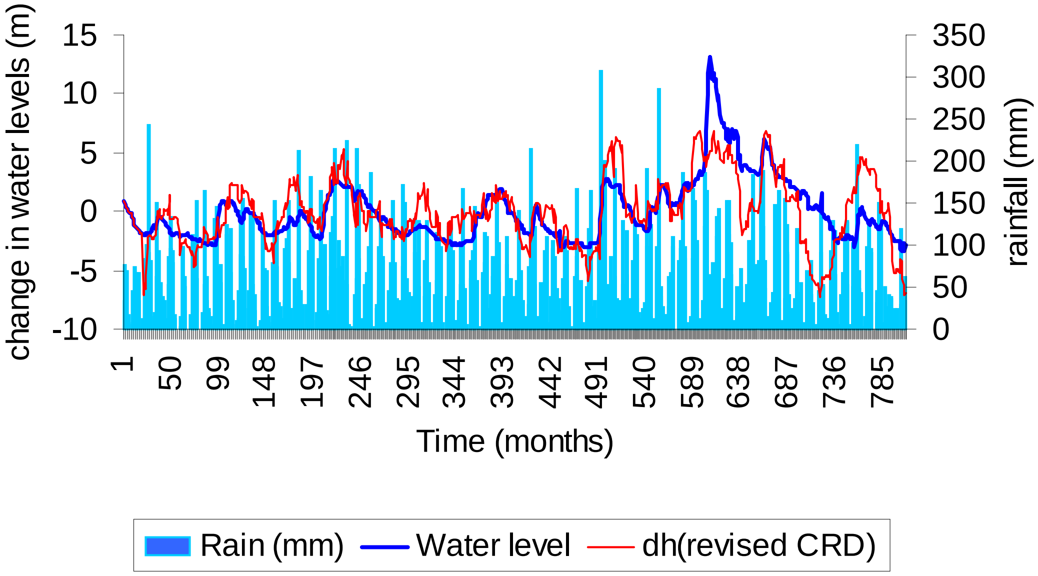

Figure 5.12 Simulation of water level fluctuation in Wondergat for rainfall averaged over the preceding 23 months.

Furthermore, application of the CRD and MA methods to different series of monitored groundwater levels has significantly contributed towards a better understanding of the spatial variability of recharge. Basic requirements are reliable rainfall records and groundwater levels (WRC contract K5/83). The methods work particularly well in the case of dolomitic aquifers, as they are highly permeable aquifers equalizing spatial differences in recharge. The integrated response of an aquifer could be represented by the saturated volume fluctuation that is derived from the summation of water level responses of different boreholes in the aquifer as illustrated in Figure 5.13 for the Lichtenburg aquifer. The impacts of different rates of abstraction could be inferred by increasing the abstraction from which the probabilities of exceedance of specified limits of groundwater levels could be assessed.

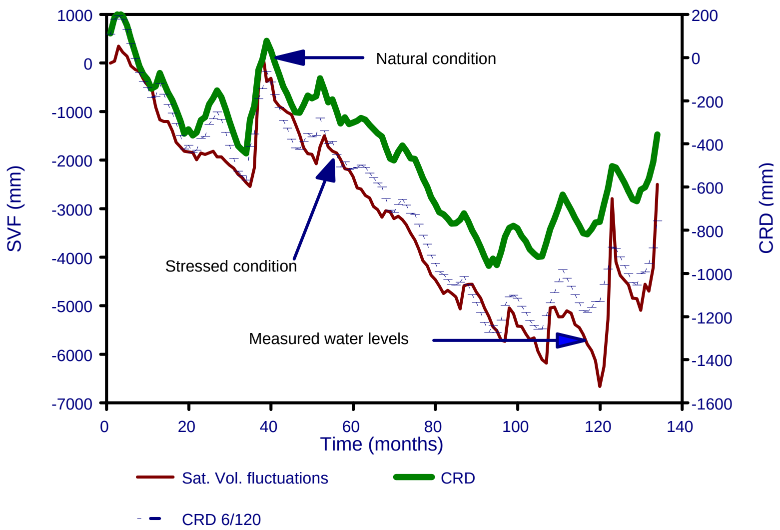

Figure 5.13 Groundwater levels in the Lichtenburg area (expressed as fluctuations of Saturated volumes) in relation to the CRD series.

5.4 Conclusions

Significant progress has been made in the RSA in studies of groundwater recharge. These have complemented the research efforts of the GRES project in Botswana. Software has now been developed to apply the CRD/MA methods. The significance and value of improved quantitative estimation of recharge using simplistic methods can be summarized as follows:

- improved assessments of the sustainable yield of groundwater resources of dolomitic and other aquifers;
- improved estimates of aquifer storativity and aquifer depth;
- better insight into recharge variability for groundwater and pollution studies;
- enhanced understanding and interpretation of natural isotopes in groundwater studies;
- improved hydrodynamic aquifer simulations based on better recharge estimates;
- improved management of groundwater systems based on simple assessment tools.

With regard to further studies on groundwater recharge the emphasis should be on:

- practical evaluations and applications, e.g. aquifer management based on monthly recharge estimates;
- storativity estimates of fractured aquifers should be validated by analysis of pumping tests and inverse modelling using models similar to AQUAMOD codes (Van Tonder et al., 1997);
- combined chemical and isotope approaches to estimation of recharge;
- re-examining the value of (anomalous) ^{14}C concentrations in dolomitic aquifers for quantitative/qualitative assessments of recharge;
- further improvement and testing of CRD method for simulation of recharge and for evaluating the performance of aquifers for different scenarios of exploitation.

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6. A Box Model for Estimating Recharge – The RIB Method

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ABSTRACT A new method called Rainfall Infiltration Breakthrough (RIB) is proposed for estimating groundwater recharge of aquifers (accommodating for pores, fractures and their combination) where groundwater level fluctuations occur resulting from rainfall recharge. The method shows resemblance to the cumulative rainfall departure (CRD) method in that it also simulates groundwater levels, but differs from the CRD method in that it accounts for the manner in which recharge occurs. It therefore provides a better opportunity for explaining the occurrence of recharge under different conditions. The method is user-friendly and requires only few spatial data. The RIB method can be used for estimating either recharge or aquifer storativity. Validation of the method under typical South African conditions is discussed based on model-generated and known cases.

6.1 Introduction

Infiltration of rainfall is influenced by many factors such as rainfall intensity and duration, surface topography, geology and factors related to moisture transport (moisture content, hydraulic conductivity, etc.). Estimation of recharge from infiltration and unsaturated zone studies, however, is complex and often inaccurate. Therefore, hydrogeologists generally resort to comparing rainfall events directly with groundwater levels, that is if water level fluctuations do occur (Wenzel, 1936; Sophocleous, 1991; Bredenkamp et al., 1995; Wu et al., 1996). In South Africa, such comparisons were often made for recharge estimation of fractured rock aquifers such as the Karoo (Kirchner et al., 1991) and dolomite aquifers (Bredenkamp et al., 1995).

Xu and Van Tonder (2001) demonstrated that even though the departure of rainfall from the mean rainfall is negative, natural water levels may continue to rise as long as there is a surplus of recharge as opposed to discharge. It is often difficult, however, to relate water level changes directly to rainfall events. Particularly when water level fluctuations are out of phase with rainfall, the absence of recharge may be interpreted incorrectly. A filter or transfer function can be introduced to accommodate for the delayed transfer of moisture through the unsaturated zone to the water table. If it is assumed that rainfall events are linearly related to the change in water level, time series of recharge can be derived. Analytical relationships between water levels and recharge have been formulated in the past (De Vries, 1974).

In this paper we propose a new filter relating water level fluctuations to rainfall aiming at better understanding and estimating groundwater recharge.

6.2 Conceptual Framework

6.2.1 Recharge process

Only part of infiltrated rainfall breaks through the zone where evapotranspiration occurs and percolates to the groundwater table. This part is here referred to as recharge. The arrival time of this water at the water table is delayed due to the 3-dimensional spreading of moisture. The duration of the recharge event is prolonged with increasing thickness of the unsaturated zone (see Figure 6.1). The breakthrough water not necessarily results from a single rainfall event, but may represent a series of preceding rainfall events. Depending on the characteristics of the aquifer, that portion of the rainfall that breaks through may cause a rise in water level, and subsequently an increase in discharge down-gradient.

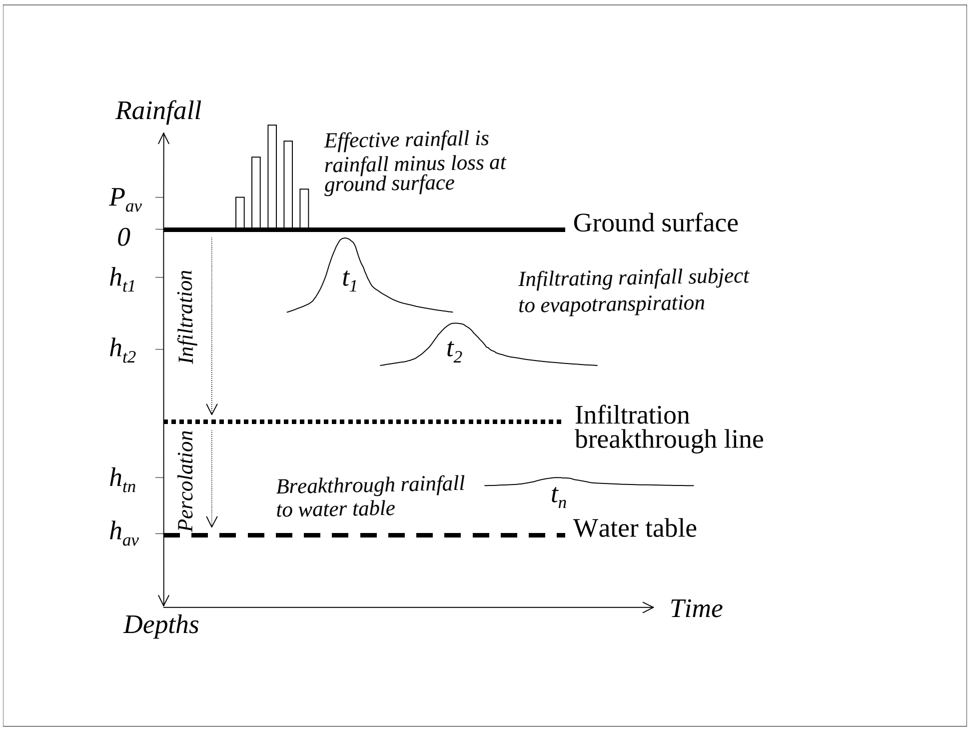

Figure 6.1 Sketch of the Rainfall Infiltration Breakthrough process.

6.2.2 Water level response

The water level response to rainfall recharge is a function of many hydrogeological factors. Most critical are depth to water level, structure and texture of the unsaturated zone, and characteristics of the rainfall. The following types of responses can be distinguished:

- Rapid response: within hours, days or a month of intensive rainfall. In this case, recharge normally occurs via preferential flowpaths like extensional fault zones;
- Intermediate response: over a time span of a year or two. In this case, recharge normally occurs through direct and indirect flowpaths; and
- Slow response: taking longer. In this case, recharge normally occurs through diffuse flowpaths.

6.3 Theory

6.3.1 Rainfall Infiltration Breakthrough (RIB)

If records of rainfall events are expressed as a time series: P_1, P_2, P_3, …, P_n, a RIB_i (or recharge) can be defined as:

$$RIB(i)_m^n = r\left(\sum_{i=m}^{n} P_i - \left(2 - \frac{1}{P_{av}(n-m)} \sum_{i=m}^{n} P_i\right)\sum_{i=m}^{n} P_t\right)$$

$$\begin{aligned}&(i = 1, 2, 3, \ldots I)\\ &(n = i, i-1, i-2, \ldots N)\\ &(m = i, i-1, i-2, \ldots M)\\ &\mathrm{M} < \mathrm{N} < \mathrm{I}\end{aligned} \qquad (1)$$

where r is that fraction of the above mentioned cumulative rainfall departure (CRD) which contributes to the RIB (is recharge percentage); P_{av} is average rainfall over the entire rainfall time series; P_t is a threshold value representing aquifer boundary conditions, which can be determined during the simulation process (P_t may range from 0 to P_{av}, with 0 representing a closed aquifer, and P_{av} representing an open aquifer system); the symbol i represents a sequential number of a rainfall record, while parameters m and n, introduced as memory markers, represent the start and end of a time series length, during which period rainfall events contribute to the breakthrough RIB(i). Note that Eq. (1) reduces to the CRD as defined by Bredenkamp et al. (1995) if rainfall events from P_m to P_n do not show any trend and thus cumulative rainfall averages equate to P_{av}.

Eq. (1) indicates that the i-th month rainfall infiltration breakthrough is attributed to the cumulative effect of rainfall series P_m, P_{m+1}, P_{m+2}, …, P_n with a weighting factor that is a function of the moving average of a rainfall time series. It shows that RIB(i) is the weighted average of the rainfall series P_m, P_{m+1}, P_{m+2}, …, P_n. The weighting factor is not necessarily constant and may be positive or negative depending on whether or not the amount of rainfall during the period of interest exceeds the moving average rainfall.

Three Rainfall Infiltration Breakthrough scenarios can be distinguished based on different time scales as illustrated in Figure 6.2:

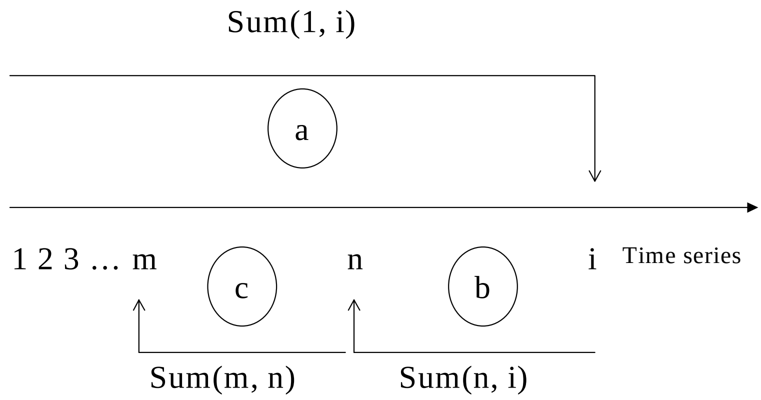

Figure 6.2 Time lag scenarios a-c.

Scenario a: RIB_1^i is the cumulative result of all previous rainfall. This often represents a combination of point and diffuse recharge mechanisms.

Scenario b: RIB_n^i results from the previous N rainfall events immediately before the current rainfall event. This can be observed in fractured rock aquifers where the infiltration process takes place relatively quickly.

Scenario c: RIB_m^n results from a limited rainfall series between m and n. This often occurs in unconsolidated aquifers where the infiltration process may behaviour like a piston flow.

6.3.2 Relationship between RIB and groundwater level fluctuation

Time series of recharge events based on Eq. (1), i.e. R_1, R_2, R_3, …, R_n, are shown in Figure 6.3.

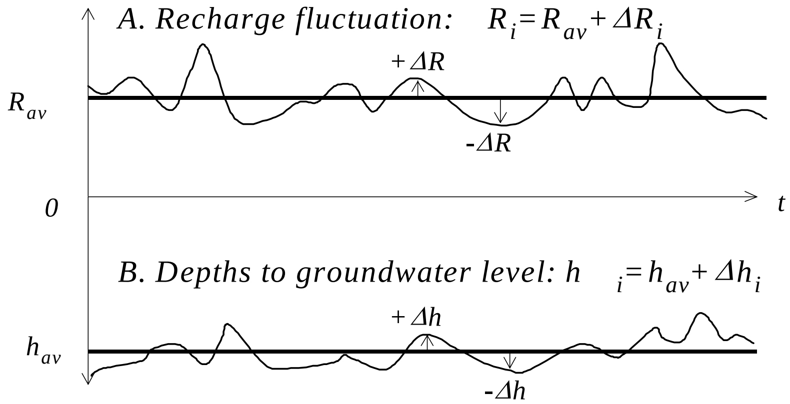

Figure 6.3 Recharge series and groundwater level fluctuations.

Recharge R can be replaced by a RIB which in turn comprises the following two terms:

$$RIB(i)_m^n = \left(RIB(i)_m^n\right)_{av} + \Delta\left(RIB(i)_m^n\right) \tag{2}$$

Average recharge RIB_{av} is a function of the average saturated aquifer thickness h_{av}, aquifer transmissivity T and average distance of groundwater flow L away from the groundwater mound, whereas recharge fluctuation ΔRIB is a function of aquifer storativity S and the groundwater table fluctuation Δh (Haijtema, 1995; Xu et al., 2001):

$$\begin{aligned} \left(RIB(i)_m^n\right)_{av} &= 2Th_{av}/L^2 \\ \Delta\left(RIB(i)_m^n\right) &= S\Delta h_i \end{aligned} \tag{3}$$

Combining Eqns. (2) and (3) result in the following expression for water level fluctuation:

$$\Delta h_i = (1/S)\cdot\left(RIB(i)_m^n - 2Th_{av}/L^2\right) \tag{4}$$

It is assumed that RIB is the driving force behind a monthly water level change, provided that the other stresses are relatively constant. As defined in Eq. (3), ΔRIB can be positive and negative, depending on rainfall intensity and frequency. The groundwater level will rise if the ΔRIB is positive, and decline if it is negative.

Eq. (4) describes a linear relationship between RIB and water level change. Accommodating for other stresses, which may be exerted on an aquifer, Eq. (4) is modified as follows:

$$\begin{aligned} &\Delta h_i = (1/S)\cdot\left(RIB_m^n\right) - \left(2Th_{av}/SL^2\right) - \left(Q_{pi} + Q_{outi}\right)/(AS) \\ &\qquad (i = 0, 1, 2, 3, \ldots I) \end{aligned} \tag{5}$$

Term $(Q_{pi} + Q_{outi})/(AS)$ in Eq. (5), represents abstraction from the aquifer and outflow at the i-th month, and is only required if a pumping borehole affects that part of the study area where water level readings are taken.

Eq. (5) can also be used to estimate the ratio of recharge to aquifer storativity by minimising the difference between the calculated and measured Δh_i series.

6.3.3 *Critical remarks*

The RIB concept is based on a box model method, which makes use of the convolution function. The input function is rainfall, the output is the infiltration breakthrough and the transfer function is a dynamic weighting factor. The numeric form of the convolution can be written as:

$$RIB(i)_m^n = \sum_{i=m}^{n} P_i \cdot \left(r - 2rP_t(n-m) \Big/ \sum\nolimits_m^n P_i + rP_t / P_{av} \right) \qquad (6)$$

where the expression between brackets represents the transfer function or weighting factor.

Parameters m and n in Eq. (6) are determined through simulations using a Solver that is available in Excel spreadsheets. Note that the weighting factor is a function of the parameters r, P_{av}, P_t, n and m. Generally, rainfall time series are composed of random and deterministic components, the latter in the form of trends and periodicities. To a certain degree, a short series of data often displays a trend, which is reflected in Eq. (1). The RIB method has therefore been formulated to account for such a trend.

A close inspection of Eqns. (1) through (5) reveals the following:

- If rainfall P_i is constant over time, $P_i = P_{av}$ and groundwater levels do not fluctuate naturally. Steady state conditions prevail.
- Only the ratio r/S can be determined through water level simulation.
- Eq. (4) is similar to that used in the EARTH Model (Van der Lee and Gehrels, 1990). The difference is that h_{av} is replaced by h_{i-1} in the SATFLOW module of the Earth Model.
- Since Eq. (4) reflects a natural situation, it cannot accommodate for variable pumping rates. Eq. (5) does account for changes in pumping and outflow rates $(Q_{pi} + Q_{outi})$.
- Term $(Q_{pi} + Q_{outi})/(AS)$ of Eq. (5) is only required if the influence of pumping and/or outflow on water level changes is evident. This may be true in cases of highly fractured dolomite aquifers, where high values of transmissivity are encountered.
- Eq. (5) implicitly assumes that there are no long-term trends in the rainfall pattern.

The RIB method is a lumped parameter method. It does not address parameter variations in space, and should thus be applied with caution.

6.4 Case Studies

In the following sections the RIB method is compared with the CRD method for closed and open aquifer systems.

6.4.1 Closed aquifer system

A hypothetical aquifer with recharge of 2% of the rainfall over a closed area of 5x5 km^2 has a borehole at the centre pumping at a rate of 15000 m^3 per month. The aquifer has a storativity of $1x10^{-3}$. Water levels over 24 months are generated using Modflow-based software.

The water level series is simulated using a computer program named RIB. Simulated water levels are compared with generated levels in Figure 6.4, where dh(crd) refers to water levels calculated using Bredenkamp et al. (1995) CRD method, while dh(rib) represent water levels calculated using Eq. (5). The average modelled recharge is 1.79% of the rainfall.

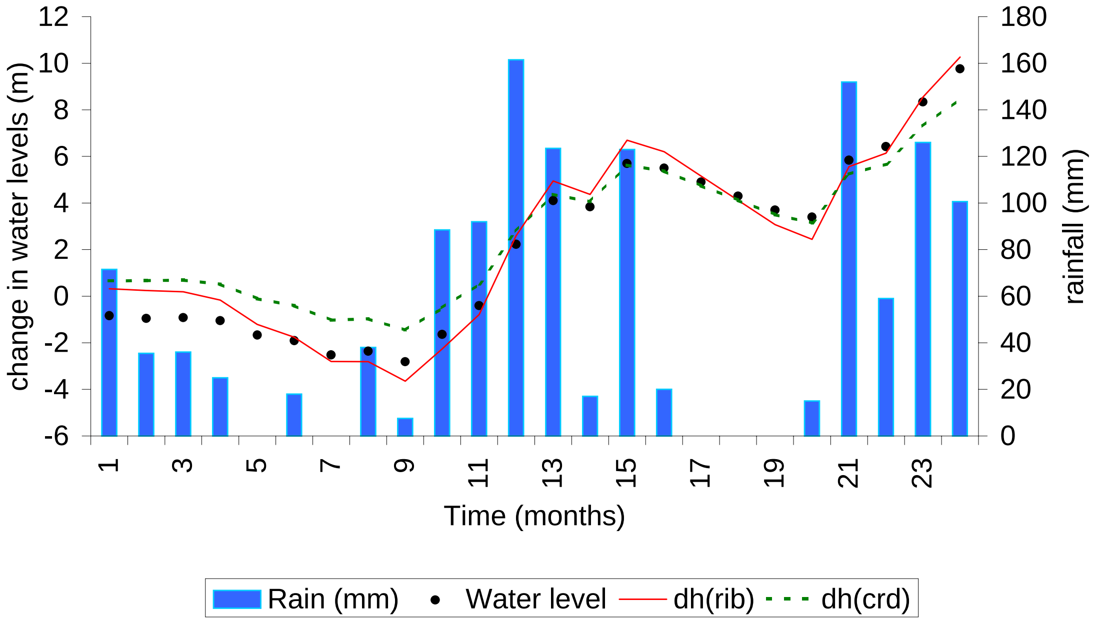

Figure 6.4 Simulation of groundwater fluctuation using the RIB and CRD methods based on model generated data.

6.4.2 Dolomite aquifer

The Grootfontein aquifer in South Africa is compartmentalized by dolerite dykes. The compartment situated in the recharge zone covers an area of $1.25x10^3$ km^2. Aquifer storativity has been estimated at 2.39% (Bredenkamp et al. 1995). Both the CRD and the RIB methods were applied to this case. Results are shown in Figure 6.5. Based on the RIB method, a recharge value of 5.71% of the rainfall was calculated, whereas the CRD method revealed a value of 11%.

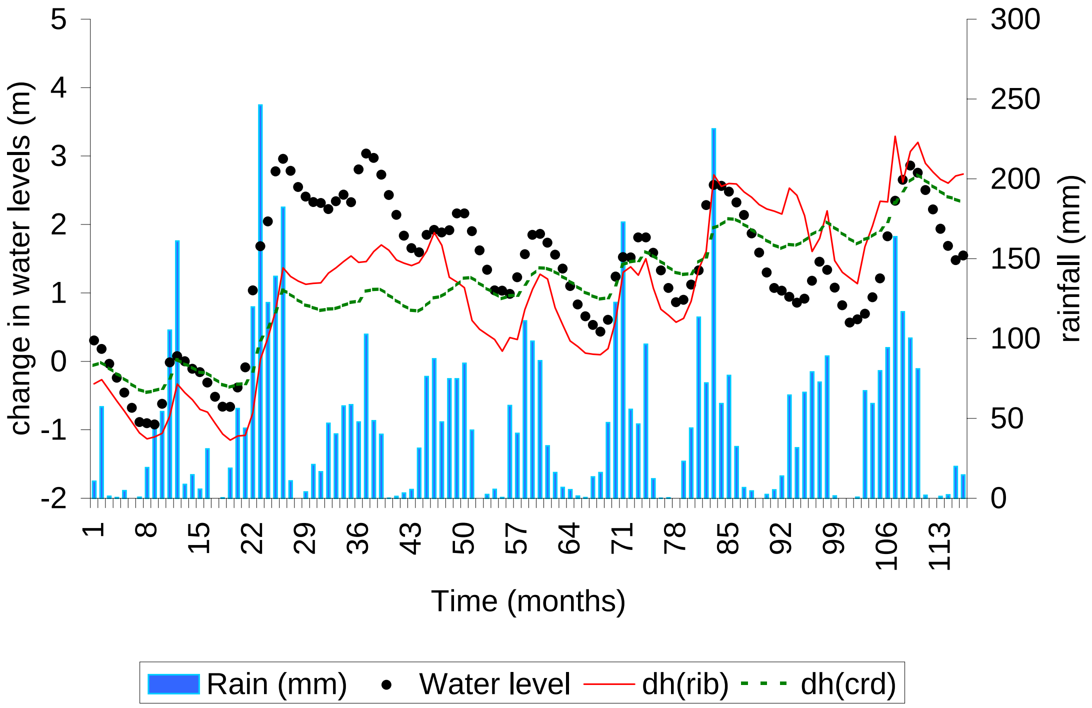

Figure 6.5 Simulation of groundwater fluctuation using the RIB and CRD methods based on data from Grootfontein compartment.

6.4.3 Karoo aquifer

The Karoo aquifer in Dewetsdorp was investigated by Kirchner et al. (1991). It covers an area of 21 km^2 with aquifer storativity estimated at 0.19%. In this case, both CRD and RIB methods were applied and yielded an average recharge of 1.67% of the rainfall as shown in Figure 6.6. The application of Eq. (5) produced a better fit using Scenario b "RIB_n^i" (see Section 3.6.1) with incorporation of 10 months of memory (n=10). This indicated that the unsaturated zone possesses preferential flowpaths. If the memory was not considered, i.e. Scenario a "RIB_1^i" in Section 3.6.1, it would give a recharge value of 1.45% as reported previously (Xu and Van Tonder, 2001).

6.5 Conclusions and Recommendations

The RIB method is a powerful tool for groundwater recharge estimation. It accommodates for rainfall series with trends and is capable of simulating different scenarios of aquifer recharge. The ratio r/S can be estimated for shallow aquifers and does not require large amounts of spatial data. The estimation can be optimized through a Solver built in Excel. The applicability of the RIB method for deep-seated aquifers still remains to be verified.

Although the RIB method is sensitive to the depth to the groundwater table, it can be applied to most boreholes in South Africa as these are drilled within a depth range of 100 m. Prior to the application of the method one must be certain that rainfall data are directly responsible for water level fluctuations. An analysis of recharge mechanisms is recommended.

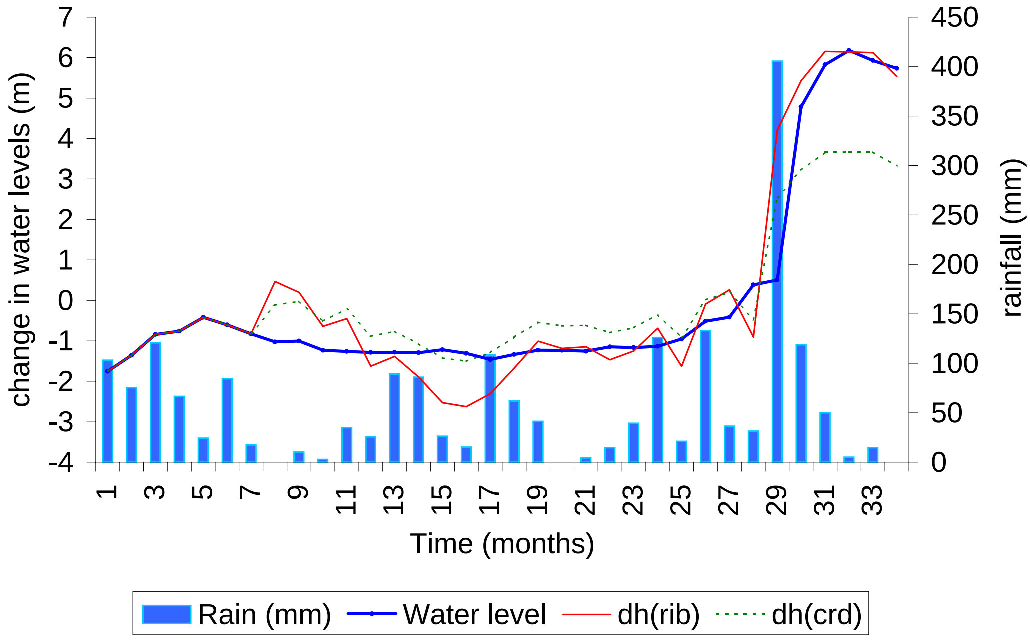

Figure 6.6 Simulation of groundwater fluctuation using the RIB method based on data from Dewetsdorp aquifer.

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PART IV

Role of Surface Water - Groundwater Interaction in Recharge Estimation

7. Surface Water – Groundwater Interactions in the Context of Groundwater Resources

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ABSTRACT Surface water-groundwater interactions need to be quantified in two ways in the context of groundwater resource estimates, namely recharge of groundwater by surface water, adding to the groundwater resources, and discharge of groundwater to surface water, as an alternative measure of groundwater recharge. Rivers are known to be the major route for recharge in arid and semi-arid zones and accurate methods of assessment of recharge are vital. River flow measurements are often the only high quality data available for water resource analysis, and have the advantage that they integrate the catchment's overall behaviour. Methods of analysing groundwater contributions to streamflow will be helpful in cost effective resource management. The intended theme of this paper is the need to develop a conceptual model of the catchment of study before beginning to apply techniques to resource estimation because surface water- groundwater interactions do not always conform to the classical concepts of recharge and baseflow. There are techniques available for estimating recharge from rivers, although it is much easier to estimate transmission loss than recharge. The most commonly and successfully used methods are water balances, particularly channel water balances, and catchment-scale models. Calculations of water table rise can be erroneous. Hydrograph analysis to obtain estimates of groundwater discharge is feasible. The best methods use rules to ensure consistency over time and between operators, but do not necessarily calculate total groundwater discharge.

7.1 Introduction

Are the differences between surface water and groundwater as clear as we would like to think? We recognise groundwater when digging a well and incepting the water table. We know surface water when we see a flowing river. But when they are in close contact they are not always separate and distinct, which is unfortunate when we are trying to quantify the interactions in order to estimate water resources.

Surface water-groundwater interactions need to be quantified in two ways in the context of groundwater resource estimates:

- Recharge of groundwater by surface water, adding to the groundwater resources. Rivers are known to be the major route for recharge in arid and semi-arid zones and accurate methods of assessment are vital.
- Discharge of groundwater to surface water, as an alternative measure of groundwater recharge. River flow measurements are often the best data available for resource analysis, and integrate the catchment's behaviour. Methods of analysing groundwater contributions to streamflow will be helpful in cost effective resource management.

This paper will concentrate on estimating recharge from and discharge to surface water, but these are not mechanistic exercises. They require a sensitive understanding of the hydrological processes that operate. Hence I start with discussions of river types, runoff and recharge mechanisms, and water resource concepts.

7.2 River Types

Two classification factors for rivers are useful in the present context - vertical positioning and streamflow characteristics. A similar but more comprehensive and locally focussed analysis for South Africa is provided by Xu et al. (2002). Firstly, rivers can be at different elevations in relation to underlying groundwater, as shown in Figure 7.1. Remote or perched rivers are associated with dryer climates, smaller or upstream parts of catchments, and changes in geology such as the emergence of a mountain river onto an alluvial fan. Groundwater does not generally contribute to surface flow. Recharge is the normal process and is controlled by stream flow and particularly by bed characteristics.

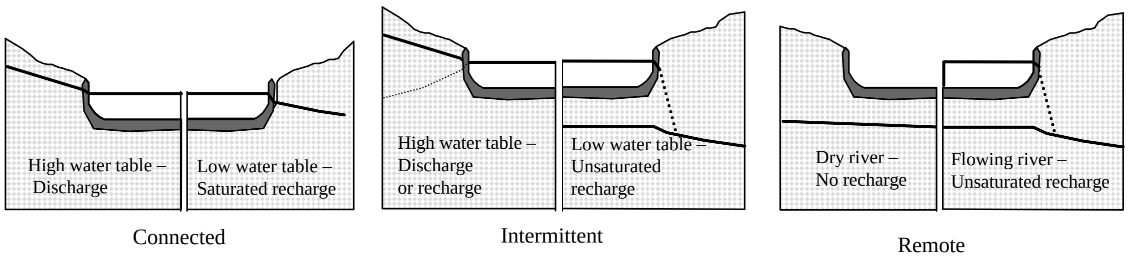


Figure 7.1 Classification of rivers by vertical positioning relative to the water table.

Connected rivers are the norm in temperate and humid climates, particularly in the lower parts of a catchment. They are the base elevation of the flow system, providing the discharge route for groundwater. Recharge is the exceptional process, occurring at changes in geology, near pumping wells, or transiently during periods of high flow. Interchanges between surface and groundwater are controlled by stream flow, water table configuration and geology. At a local scale, these interchanges can be variable in space and time. Figure 7.2 shows a long section of a river in New Zealand, with a plan view drawn alongside. At this local scale, flow occurs both ways between surface and groundwater, with the river losing upstream of each riffle, and gaining downstream.

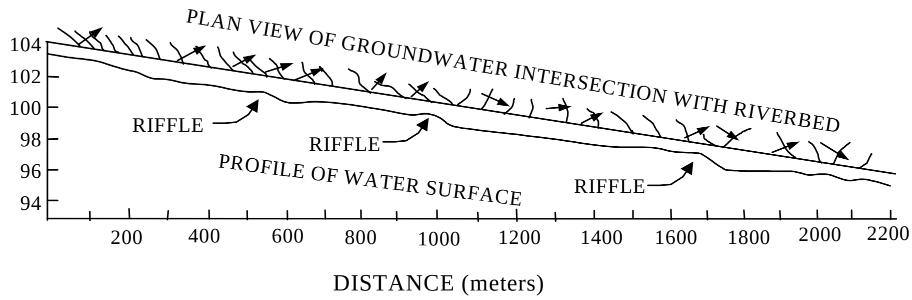


Figure 7.2 Water surface profile (in vertical section) and groundwater flow directions (in plan view) along the Waimakariri River, New Zealand. Based on Figure 8, van't Woudt et al. (1979).

Some rivers switch between perched and connected as the water table elevation changes seasonally or in response to recharge events. These intermittent rivers are the most difficult to analyse as illustrated by the example from North Africa in Figure 7.3. Analysis of monthly channel water balances initially shows an apparent lack of correlation between transmission loss and upstream flow (Figure 7.3a). Closer inspection (Figure 7.3b) reveals an envelope of 100%

losses which occurs in summer months. Some winter months show groundwater discharge to the river, and other months fall between these extremes. This pattern can be interpreted as a river connected, at least seasonally, to groundwater with the relative elevation of river stage and water table being a major control on the direction and quantity of flow in or out of the river.

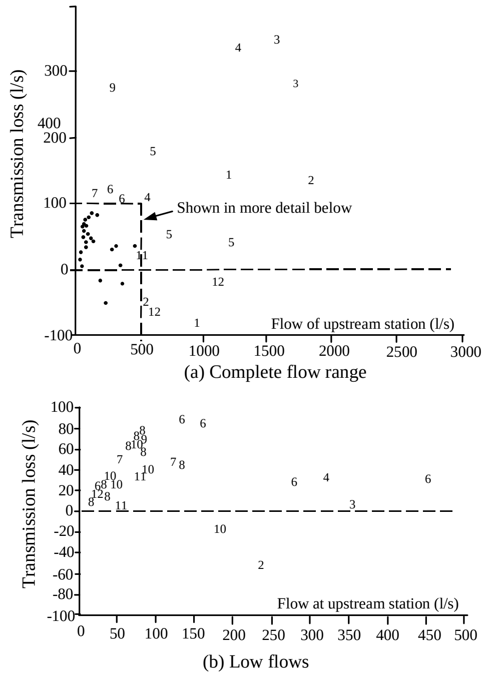


Figure 7.3 Relating transmission loss to river flow for a seasonally connected river in North Africa (Lerner et al., 1990).

The second classification is by streamflow characteristics, into ephemeral (event dominated) and perennial (continuous) rivers (Figure 7.4). Perennial rivers are normally connected, associated with groundwater discharge, wetter climates, and larger catchments. For example they are the norm in the UK. In the upper reaches, perennial rivers may become seasonal, because the upper springs only flow at times of high water tables. Conversely, ephemeral rivers are associated with dryer climates and are normally perched systems.

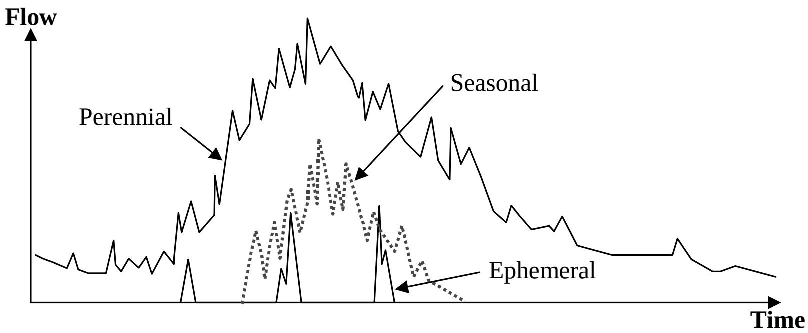


Figure 7.4 Classification of rivers by flow characteristics.

As man continues to develop and manage water resources, his influences can dictate the characteristics of a river. Examples include:

- damming of rivers, with continuous release of water to give perennial flows,
- removal of water from a catchment, reducing flow,
- abstraction of groundwater, reducing baseflow and possibly inducing recharge,
- discharge of effluents, sustaining river flow in the dry season,
- alterations to beds by lining, channelisation, gravel extraction, recharge lagoons, and dams.

Hence in lowland Britain, there are no rivers with natural hydrographs. In some cases, dry season flows are virtually constant because sewage discharge has replaced baseflow as the dominant low flow. In other cases, previously perennial rivers are dry for most of the year due to groundwater pumping, Some rives are only kept flowing by wells pumped directly into them, sometimes accompanied by channel lining, to preserve some aesthetic or environmental value.

7.3 Runoff Mechanisms

The conversion of rainfall to runoff is a combination of several processes. These are illustrated schematically in Figure 7.5 and described below:

1. Hortonian overland flow or infiltration-excess overland flow (IOF) is generated when rainfall intensities exceed infiltration capacity. It is associated with high rainfall rates and sealed surfaces such as bare rock, soils compacted by overgrazing, intensive agriculture, traffic or rainfall, and with man-made coverings such as roads.
2. Saturation-excess overland flow (SOF) occurs when soil becomes saturated from below. Rainfall is rejected and throughflow from upslope re-emerges on the surface. SOF occurs at lower intensities than IOF, and over a temporarily variable area as the area of saturated ground varies.

3&4. Unsaturated and locally saturated flows within slopes are mainly related to layered permeability or the presence of macropores that form preferential pathways. The importance of such flows is their rapid transmission of water to help saturate lower slopes, and create the conditions for SOF.

5. Groundwater flow (SSQ) occurs below the main water table and provides the continuity of flows between rainfall events. However groundwater can show a storm response to rainfall.

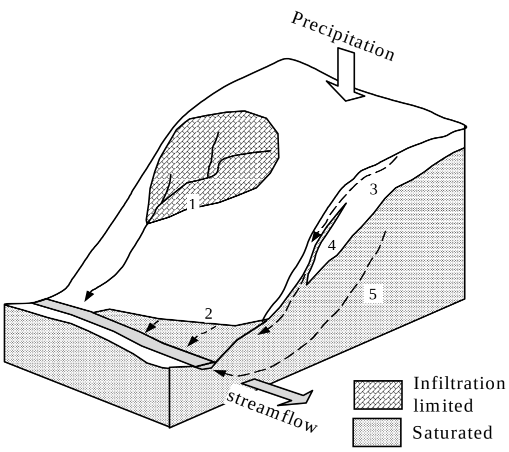


Figure 7.5 Modes of runoff generation. Based on Figure 9.11, Church and Woo (1990).

Figure 7.6 relates the occurrence of the three main flow mechanisms to catchment characteristics. Catchments in lowland Britain typically lie near the SOF-SSQ axis, while classical arid or semi-arid catchments lie closer to the IOF apex.

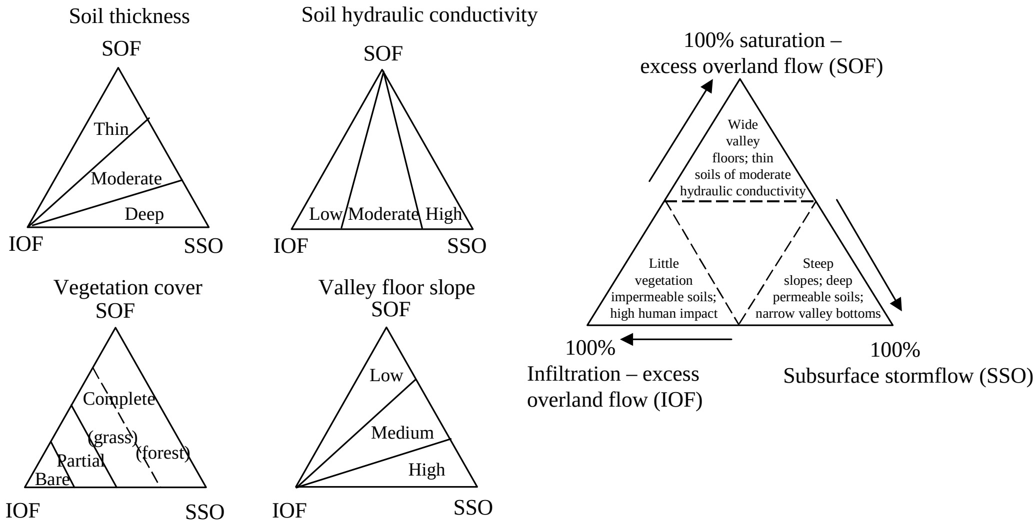


Figure 7.6 Relation of storm runoff origins to catchment characteristics of soil thickness, soil hydraulic conductivity, vegetation cover, and valley floor slope. The large triangle integrates the four components. Based on Anderson and Burt (1990).

The response of groundwater in valley bottoms and near to connected rivers is an important part of surface water-groundwater interactions, and is schematically illustrated in Figure 7.7. Without rainfall, the discharge is of regional groundwater flow; this is the baseflow of the groundwater resources, the water resources which is potentially exploitable and which has to be estimated. The figure shows the moisture profile in the valley bottom with the capillary fringe near to ground surface. Only a small volume of water is required to saturate the ground. Once rain falls, the ground quickly saturates with two consequences. SOF is generated. Secondly the large rise in the local water table displaces a pulse of groundwater towards the river. The storm response of the river may be mostly groundwater.

If the water table in Figure 7 were initially below the riverbed, the consequence of rainfall would differ. Runoff in the river will initially be able to infiltrate (recharge), but the limited storage available will cause the water table to rise rapidly. Once it reaches the surface, recharge will be rejected. Figure 7.8 illustrates this effect with an example from Namibia.

The purpose of this discussion is to illustrate that groundwater and surface water are not distinct, but are end points on a continuum of hydrological responses. Understanding is necessary before tools can be applied for resource estimates.

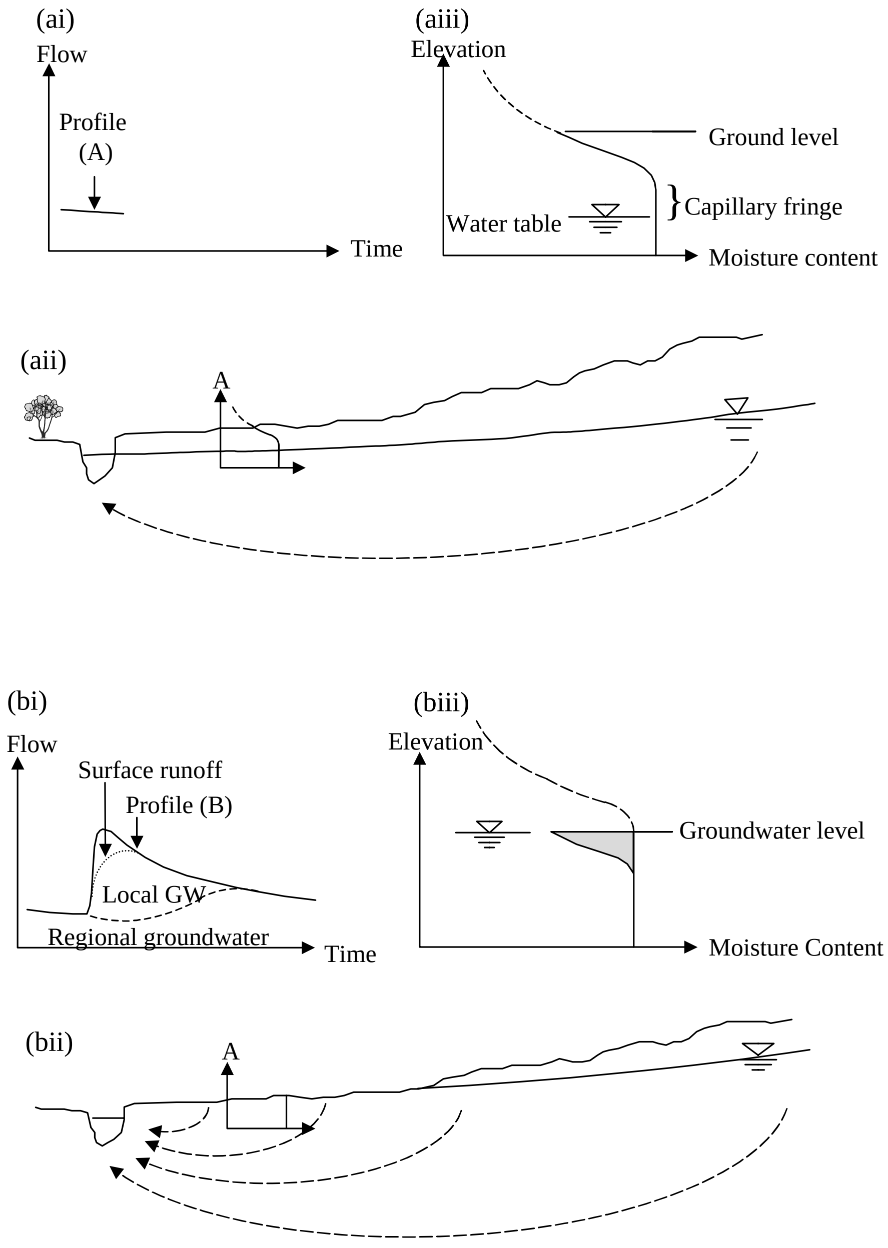


Figure 7.7 Runoff generation near connected rivers, emphasising the groundwater component. (a) prior to rainfall, showing (i) river flow hydrograph, (ii) cross-section, (iii) soil moisture profile at point A. (b) after rainfall showing (i) river flow hydrograph and flow components, (ii) cross-section with raised water table and increased groundwater discharge, (iii) fully saturated soil moisture profile at point A.

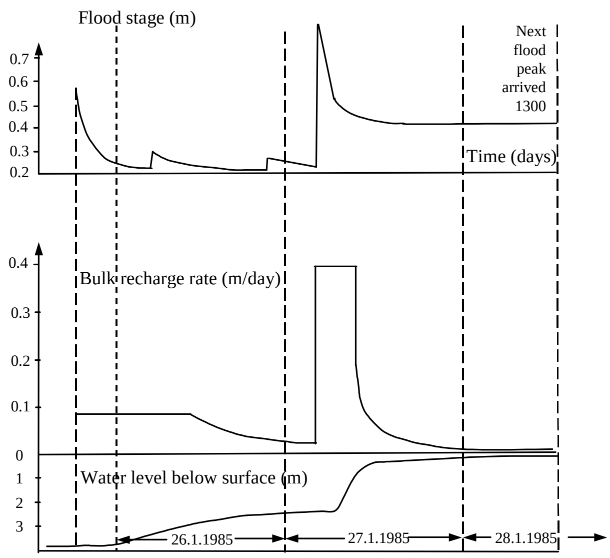


Figure 7.8 Flood and groundwater hydrographs at Gros Barmen, Namibia, with calculated recharge rates. Based on Figure 3, Crerar et al. (1988).

7.4 Groundwater Resources

There are various definitions of groundwater resources. I will define RESOURCES as the water available for beneficial uses. The beneficial uses will include agricultural and potable water supply, environmental uses such as riparian vegetation and amenity flows in rivers, and downstream demands.

Resources differ from YIELD, which is the water that can be taken for supply, within the applicable constraints. These may be legal, economic, technical or environmental.

In an unexploited, steady state groundwater system, the water balance can be written as

RECHARGE - OUTFLOW = 0 (1)

where RECHARGE includes all inputs to groundwater, that is from rainfall, surface water and inter-aquifer flows. OUTFLOW includes phreatophyte evapotranspiration and discharges to surface water, and can be equated to the RESOURCES. This simple definition is not always useful because of the changes when exploitation starts. When wells pump, the water balance should be rewritten as

ADJUSTED -	REDUCED -	PUMPING +	STORAGE = 0	(2)
RECHARGE	OUTFLOW		LOSS	
	to S_W and E_E			

Equation (2) acknowledges that recharge may have increased and outflows decreased in response to lowering of the water table. The latter usually dominates over recharge increases and the yield of a groundwater system is at the expense of surface water. For sustainable and environmental acceptable development of groundwater, these losses must be balanced against the beneficial use of pumped groundwater.

7.5 Methods of Estimating Recharge from Rivers

Recharge has been defined as the water added to the saturated groundwater body; in the context of river recharge it is the water that leaves a river and crosses the water table. Estimating recharge is often difficult, and many studies and methods find it easier to estimate transmission losses, that is the water that leaves the river downwards. Storage in the unsaturated zone, bank storage, evapotranspiration, perched water tables and shallow lateral flow can lead to large differences between recharge and transmission losses.

Hughes and Sami (1992) give an interesting example of some of the issues of transmission loss and recharge. They analysed two flood events over a 2 km reach in the Bedford research catchments near Grahamstown, SA. Transmission losses were 22 and 75% of runoff, and were estimated from neutron tube measurements of soil moisture changes. These losses replenish the valley alluvium. Groundwater is found in fracture zones in bedrock, generally >20 m below surface. What happens to the transmission losses? How much recharges deep groundwater, becomes underflow in the alluvium, or is evaporated by the phreatophyte vegetation?

Lerner et al. (1990) put recharge estimation methods into five groups:

- direct measurements,
- correlation methods,
- tracer techniques,
- Darcian approaches, and
- water balances.

7.5.1 Direct measurements and correlation methods

Direct measurement is limited to lysimeters for direct recharge and seepage meters for open water bodies such as lakes and canals. Although seepage meters (Kraatz, 1977) could, in concept, be used for rivers, variability of flows and bed conditions make their use impractical for everyday use.

Correlation methods are those which relate recharge to some other, more easily measured variable, such as groundwater level or streamflow. They are most applicable for transmission losses from perched rivers. One of many examples from the literature is shown in Figure 7.9, based on detailed studies of 53 storms on rivers in Kansas. This example shows a very good correlation, with the data points being within a factor of 2 (multiply or divide) of the central estimate; many other studies show wider error bands.

Besbes et al. (1978) show how a groundwater response function, similar in concept to a unit hydrograph, can be derived for river recharge from single events. This is another type of correlation method.

Several general points of caution should be noted about correlation methods. The error bars are always wide, due to natural variability and measurement difficulties. In order to develop correlations, good estimates of recharge are needed, calculated by another method; is it not possible to continue to use these good estimation methods? Finally, there is always a temptation to transfer them to rivers for which they were not originally derived. All hydrologists know that two rivers and their alluvial aquifers can have broadly similar characteristics but have numerically different flow duration curves and hydraulic properties; transposition is likely to increase errors!

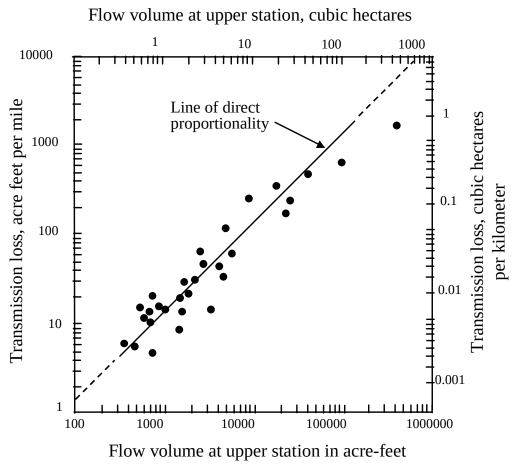


Figure 7.9 Relation between upstream flow and transmission loss for 14 ephemeral perched rivers in Western Kansas (Jordan, 1977).

7.5.2 Tracer techniques and Darcian approaches

Tracers, both applied and environmental, are widely used to estimate direct recharge from precipitation (for example Sukhija et al., 1996). They are also valuable in identifying recharge sources, and can be used to identify zones of groundwater derived from rivers. There have also been attempts to measure seepage rates through canal beds (Bouwer and Rice, 1988). However there do not appear to have been any successful uses of tracers to quantify river recharge. Zellweger (1994) attempted to measure influx to and efflux from a river with applied tracers. He succeeded in estimating influx by dilution calculations, but could not estimate efflux from tracer data because it made no difference to concentrations in the river.

In principle, the rate of spreading into groundwater of an applied tracer could be measured, and yield a recharge estimate. In practice, large tracer injections, extensive sampling networks, and long measurement durations would be needed to obtain more than a point estimate for a single event.

Darcy's law allows the development of infiltration equations and flow nets, the use of field data on hydraulic properties and heads in the saturated and unsaturated zones, and the application of numerical models of surface-groundwater interaction. These methods have been used in desk and hypothetical studies, and in research studies of cross-sections and short reaches of rivers. For resource studies, we require data over long reaches, whereupon variability of channel and bed become large. It is unrealistic to expect to gather sufficient field data to calculate recharge by application of Darcy's law in a vertical plane at a regional scale.

7.5.3 Water balances

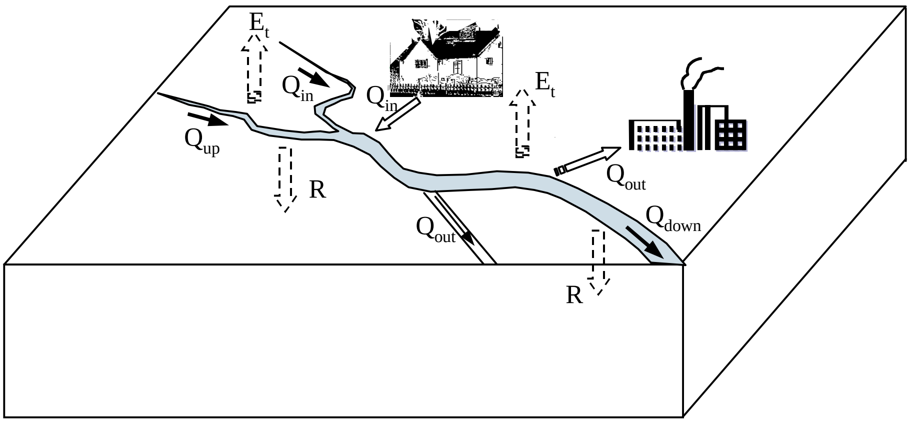
From the above discussion of other methods of estimating river recharge we can see that few will be useful, and that our last chance is with water balances approaches. Fortunately these can work! There are a variety of approaches, such as:

- channel water balance,
- channel flow routing,
- water table rise,
- catchment modelling, and
- aquifer modelling.

7.5.4 Channel water balance and flow routing

Figure 7.10 illustrates the channel water balance, in which recharge - more accurately transmission loss - is found as the difference of all other flows. Sound in principle, the approach has potential errors which must be controlled:

- Measuring up and downstream flows can be difficult and inaccurate, especially for ephemeral rivers.
- Differencing large numbers gives rise to large errors.
- Tributary inflows are often unknown, and are a particular problem in arid environments where rainfall is spatially variable. Often an empirical rainfall-runoff model must be used to make estimates.
- In urban or agriculturally developed areas, abstractions for use and waste water returns may be major parts of the balance.
- Evaporation from the bed or by phreatophyte vegetation may be a major part of the balance.
- Only transmission loss is estimated.



$$Q_{up} + \Sigma Q_{in} - \Sigma Q_{out} - R - E_t = Q_{down}$$

Figure 7.10 Schematic of the channel water balance method for recharge estimation for rivers.

This approach has several advantages, not least that it does attempt to calculate a water balance and not leave unaccounted for water. It is easy to understand. River flow data are often the only good data available for catchments and, if a good correlation can be established, recharge records can be extended from historical flow data.

The water balance is the hydrologist's approach. A river modeller might wish to account for the hydraulics of flow and times of travel, and use a flow routing model. Knighton and Nanson (1994) used a Muskingham routing procedure to model transmission losses along a complex river channel in Australia, with little success. A more complex flow routing model was used on a 47 km reach of the Wadi Najran in Saudi Arabia (FAO, 1981, Section 18.3.3.4). It was successfully calibrated against stage data for one well-documented flood, then used to simulate seven years of historical floods to estimate transmission loss.

7.5.5 Water table rise

Water table rises can be expected in response to recharge events on ephemeral rivers and recharge periods on seasonal rivers. It seems apparent that the volume of recharge can be calculated from the volume of the rise and the specific yield. This approach is widely used, especially on a regional scale. Sometimes the approach is used in reverse to estimate specific yield. Unfortunately there are some theoretical problems and misconceptions in its use that may give rise to serious errors; good results should be possible if these can be overcome.

The first problem is the choice of a specific yield value. Personal experience of groundwater modelling has shown that there is often no relation between the values provided by pumping tests and those needed to simulate long term, regional groundwater flow. Short-term tests usually give a lower value. Nwankor at al. (1992) have convincingly demonstrated that retention in an extended capillary fringe is responsible. What specific yield value do we use for recharge estimates?

The second problem is that the peak of the water table rise does not signify that recharge has ended. Figure 7.11 attempts to illustrate this point schematically with a groundwater level hydrograph, a hydrograph of recharge reaching the water table, and the response of groundwater outflow from the zone of interest. Recharge peaks and gradually declines as drainage through the unsaturated zone becomes slower. The peak of the level hydrograph occurs when the recharge rate equals the outflow rate, shown by the dashed line. There may be substantial recharge after this time. Conventional analysis gives recharge up to the peak ($S_y \Delta h = Y_1 - Z_1$ where Y_1 and Z_2 are the flow volumes shown), whereas the flow volume $Y_1 + Y_2$ (or $Z_1 + Z_2$) is required. This is seen on the outflow hydrograph if only it could be calculated. The recession curve displacement method of baseflow analysis has some similarities and is discussed below.

The importance of delayed drainage may seem hypothetical, but an old example from Wilson and de Cook (1968) demonstrates that it is a real and significant effect. Figure 7.12 shows their data from a field site in the USA. They measured water table responses and moisture content of the vadose zone. Two perched water table mounds are created by the recharge events and take several months to dissipate. This contrasts with the main water table responses which die away within days.

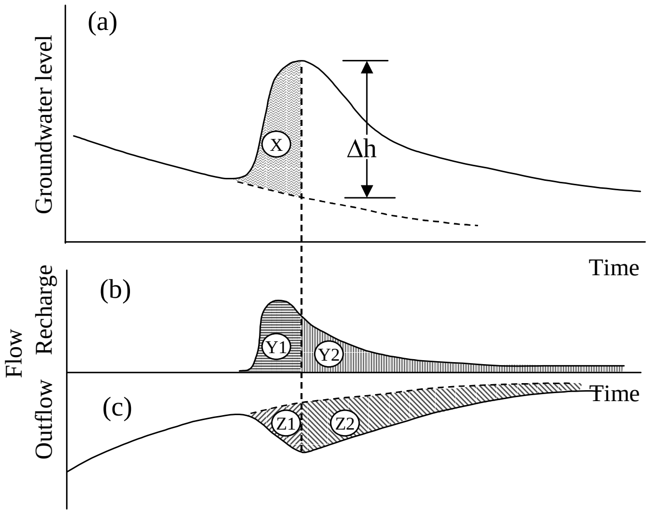


Figure 7.11 Responses to a recharge event in a remote ephemeral river. (a) groundwater level hydrograph, (b) recharge rate at the water table, (c) groundwater outflows from the recharged zone.

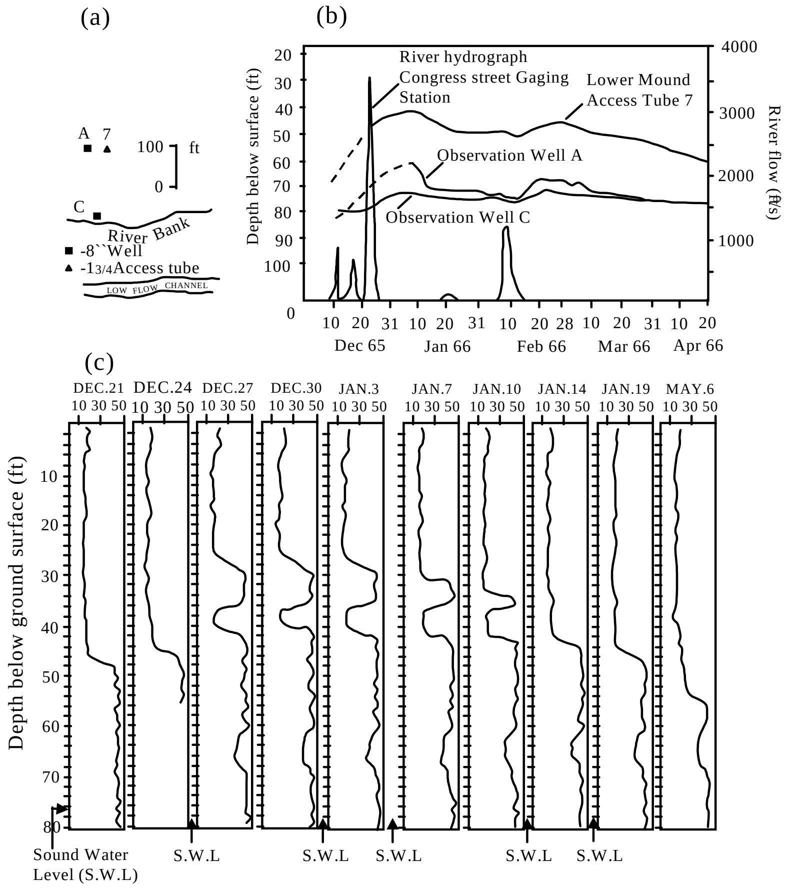


Figure 7.12 Delayed drainage of river recharge to the water table on the Santa Cruz River, USA. (a) location of observations wells relative to the river, (b) flood hydrograph and resulting responses in perched and deep water tables, (c) moisture content profiles in access tube 7. Based on Wilson and de Cook (1988).

7.5.6 *Catchment and aquifer modelling*

Catchment models vary from simple "bucket-and-pipe" models to distributed soil-groundwater-channel models. They include transfers to and from groundwater within them. Aquifer models are bounded by the water table, so transfers to and from groundwater are part of their boundaries. Both types of model estimate water balance. They have the advantages of integrating all available data (hydrology, groundwater, geology) and being of use for predictive studies of resources. Arnold et al. (2002) apply the SWAT model to 131 catchments in the upper Mississippi basin and argue that the results agree with a baseflow separation method. Sami and Hughes (1996) report good agreement between an uncalibrated surface-subsurface model and a chemical mass balance for a semi-arid catchment in South Africa. My experience is that such models are a valuable aid to quantifying recharge, even when few data are available.

7.6 Estimating Groundwater Discharge from River Hydrographs

For an exploited aquifer,

$$\text{AVERAGE DISCHARGE} = \text{AVERAGE RECHARGE} + \text{RATE OF STORAGE DEPLETION} \qquad (3)$$

Thus an alternative to estimating recharge is to estimate discharge. In an arid environment, the components of discharge that may need to be measured and aggregated include:

- Springs, that is discrete discharge points. Flows can be measured, although corrections may be needed for consumption by riparian vegetation.
- Lakes and seas may receive diffuse seepage or discrete discharges, depending on geology. Direct measurement is difficult, although seepage meters have been used on lakebeds, and Darcy's law estimates are the most promising way to proceed.
- Sabkas or saltpans are internal drainage points for which Darcian flow estimates of upward flow can be used.
- Evapotranspiration, particularly by phreatophytes, may occur over large areas and is not easy to estimate.
- Abstractions by wells. Much of this water may return as irrigation return or wastewater infiltration, and only the net consumption should be evaluated.
- Rivers, which receive both point and diffuse discharges. Channel water balances or tracer dilution can measure flow changes.

As an alternative to measuring each component of discharge, river flow hydrographs can be analysed to estimate the groundwater component, sometimes called baseflow. Of course this approach will not estimate sabka, wells and evapotranspiration discharges, but these can be approached by other means. More seriously, it may be difficult to distinguish surface runoff from groundwater. In fact there is an issue of definition of baseflow - the discussion of runoff generation argued that much of the storm response of a river is composed of groundwater by a displacement mechanism. This applies to connected rivers, which will include any river that has a groundwater component in its flow.

So what is baseflow? At one extreme, it is groundwater, which is the water that has been below the water table. At the other extreme it is the slowly responding part of the flow hydrograph, which we hope is groundwater from outside the zone of influence of storm runoff mechanisms. I am unable to offer a useful and conceptually sound definition of baseflow in the context of groundwater resources and we must use a pragmatic approach, based on the conceptual model implied by the method of hydrograph separation.

Three approaches to baseflow separation seem to be useful conceptually sound:

- graphical separation,
- baseflow rating curves, and
- recession-curve displacement.

Halford and Mayer (2000) argue, from an analysis of 13 sites in the USA, that these methods can be unreliable if used alone. In contrast, Arnold and Allen (1999) claim to have had good success with correlation between a separation technique and catchment mass balances for six USA streams. Wittenberg and Silvapalan (1999) successfully analysed streamflow to determine all the main components of groundwater balances, including seasonal variations in evapotranspiration, for a catchment in the humid part of Western Australia. Healy and Cook (2002) provide a recent review of techniques to use groundwater levels to estimate recharge. As our interest is in long-term water resources, we are not interested in event-based methods, such as those used to study flood hydrographs.

7.6.1 Graphical separation of baseflow

Hand-drawn separations have been common amongst hydrologists, typically using plots of daily flows. This approach identifies slowly responding flows, but consistency between years and between hydrologists is likely to be a problem. The Institute of Hydrology (1980) came up with a computer based graphical separation technique for its studies of low flows in the UK. A series of rules identify turning points in the hydrograph, which are connected and used to calculate a baseflow index (BFI). The BFI is not claimed to be an estimate of groundwater discharge, but it is a consistently derived measure of slowly responding flows.

7.6.2 Baseflow rating curves

The easy part of any separation is during the dry season, when all flow is from groundwater. Conversely, the hard parts are during the wet season or storms, when the regional groundwater component is masked by storm responses. The baseflow rating curve uses groundwater levels to pick out the groundwater component. Flows during the dry season are correlated with groundwater levels in a representative observation well to create a rating curve. During the wet season, the curve is used in reverse, with field observations of groundwater levels being used to estimate baseflow, hence achieving a systematic separation in these difficult periods. There are few published reports on these techniques. Avery et al. (1999) applied them successfully to a shallow, recharge sensitive spring. Ketchum et al. (2000) applied a related method called the cumulative storage accumulation curve to a spring in West Virginia, USA.

7.6.3 Recession-curve displacement

Rutledge and Daniel (1996) propose an automated procedure to estimate baseflow by studying the displacement of the hydrograph during each event. It is based on a related manual method, which is illustrated in Figure 7.13. The steps of the process are (1) develop a master recession curve and calculate the recession constant, (2) calculate a critical time from the hydrograph peak, after which the recession is log-linear, (3) extrapolate the pre and post event recessions to the critical time, and determine the hydrograph displacement at this time, and (4) calculate groundwater discharge volume from this displacement.

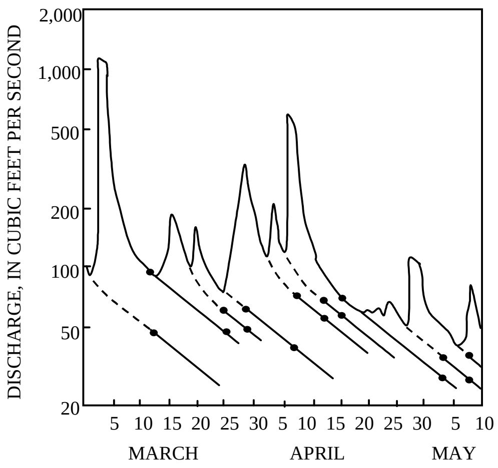


Figure 7.13 Example of recession-curve displacement method for estimating recharge, showing extrapolated recessions and vertical displacement of recession with each event (between dots). Based on Rutledge and Daniel (1996).

The procedure is a graphical analysis of the hydrograph, and there is no theoretical justification for claiming that it estimates groundwater discharge. However, just as the BFI, it is a consistent procedure, and we can hope that the result is at least related to discharge, and so a suitable starting point for resource estimation.

7.7 Discussion

The intended theme of this paper has been the need to develop a conceptual model of the catchment of study before beginning to apply techniques to resource estimation. Surface water-groundwater interactions are complex and do not always conform to the classical concepts of recharge and baseflow.

There are techniques available for estimating recharge from rivers, although it is much easier to estimate transmission loss than recharge. The most commonly and successfully used methods are water balances, particularly channel water balances and catchment-scale models. Calculations of water table rise can be erroneous.

Conversely, hydrograph analysis to obtain estimates of groundwater discharge is feasible. The best methods use rules to ensure consistency over time and between operators, but do not necessarily calculate groundwater discharge. Nor do they include routes of discharge such as phreatophyte evaporation.

Finally, we must remember that recharge estimation is only part of the story of resource management. Not all recharge is available for abstraction by wells, for both technical and environmental reasons.

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8. Recharge and Stream Flow

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ABSTRACT In order to prescribe interactions between rivers and groundwater, a classification system is proposed. Examples are also given based on author's observations. Groundwater issues related to South Africa's surface water resources are duely discussed.

8.1 Introducton

8.1.1 Czechoslovakian stream flow study

According to Knezek and Krasny (1990) groundwater runoff assessment is generally considered the best way of estimating regional groundwater resources. It is assumed that under natural conditions in Central Europe, groundwater runoff equals the long-term available resource. Knezek and Krasny mapped groundwater runoff in Czechoslovakia, which is almost entirely underlain by the Bohemian Massif. The Massif consists largely of late Paleozoic igneous, metamorphic and well-cemented sedimentary rocks. Groundwater runoff accordingly originates from a zone of weathered and fissured hard rocks not more than a few tens of metres thick.

They found that runoff is not so much dependent on favourable water-bearing characteristics of rocks, than on morphology and height above sea level, precipitation and evapotranspiration. The highest mountains along the Czech frontier reach elevations of between 1200 and 1600 m.a.m.s.l. and the mean annual precipitation fluctuates between 800 and 1600 mm. Groundwater runoff ranges from more than 10 l $s^{-1}km^{-2}$ that is 315 mm a^{-1} at the highest elevations to between 3 to 7 l $s^{-1}km^{-2}$ (95 to 220 mm a^{-1}) in lower mountainous areas with a precipitation of less than 800 mm a^{-1}. In the flatter areas groundwater runoff decreases to between 0.5 and 2 l $s^{-1}km^{-2}$ (16 to 63 mm a^{-1}).

8.1.2 Stream Flow Studies in Crystalline Basement areas of Malawi and Zimbabwe

During the British Geological Survey's research project on the hydrogeology of crystalline basement aquifers in Africa (Wright and Burgess, 1992), base flow from 26 Malawi and Zimbabwe catchments was analysed, as a means of providing a minimum estimate of recharge. Observed groundwater base flow contributions to stream flow expressed as a fraction of rainfall (Farquharson and Bullock, 1992), range in Malawi from 0.02 to 0.25 (i.e. from 16 to 370 mm a^{-1}, for mean annual rainfalls of 900 and 1500 mm, respectively). The corresponding figures for Zimbabwe are zero to 0.09 (zero to 80 mm a^{-1} and rainfalls of 600 and 900 mm a^{-1}). The ratio, base flow to rainfall, correlates significantly positive with mean annual rainfall and with mean relative relief. The latter is the mean value of relative relief calculated for each square kilometre of the catchment. The relationship with the mean dambo perimeter, on the other hand, is as one would expect, negative. Dambos are areas of seasonally water-logged bottomlands i.e. areas where groundwater is lost through evapotranspiration.

It would appear that the greater the relief, the steeper the piezometric gradients that can develop, the faster the groundwater flow and the lesser the volume of groundwater lost through evapotranspiration en route to the stream Other factors associated with greater relief undoubtedly have to be considered as well such as higher rainfall, greater stream density, thinner soil cover (larger fraction of precipitation infiltrating) and differences in vegetation cover between mountainous and hilly country compared to that of the plains.

8.1.3 South African streamflow studies

Both the Czechoslovakian and African groundwater base flow studies relate to hard -rock areas, where groundwater is principally contained in a near-surface zone of weathered and fractured rock. In this respect, conditions over most of South Africa are comparable. With the exception of some localised occurrences of permeable porous Cretaceous and Karoo sandstones, pre-Tertiary Formations do not feature as primary aquifers.

As early as 1949 Enslin, estimated groundwater recharge from base flow in the upper reaches of a number of rivers rising along the eastern escarpment (see Table 8.1). The upper reaches of these rivers drain a variety of geological formations ranging from Swazian granitic rocks through Randian, Transvaal and Karoo Sequences. The positions of the stream flow gauging stations which were used in this exercise are not given. Consequently it is not possible to give the geological composition of the different catchments except for stating that the Blyde River catchment is largely underlain by Malmani dolomite, and that of the White River lies on granite and of the upper Umgeni River on Karoo strata.

Table 8.1 Estimated recharge in a number of rivers.

Catchment	Area km^2	Mean Base Flow $10^6\ m^3\ a^{-1}$	Mean Base Flow per unit area mm.	Percentage of Mean Annual Rainfall
Luvuvhu River	953	51.4	54	4.2
Groot Letaba R.	570	45.1	79	5.7
Blyde River	518	118.8	229	19.8
White River	109	14.7	134	13.5
Nels River	622	64.3	103	10.7
Kaap River	1606	71.5	44	4.6
Komati River	8625	454.6	53	5.2
Ngwempisi River	881	45.3	51	5.1
Ishlelo River	699	28.6	41	4.0
Assegai River	1891	42.9	23	2.3
Pongola River	7097	206.3	29	2.9
Umhlatuzi River	1243	62.5	50	6.6
Umgeni River	4061	128.6	32	3.1

Since 1935 when the Jonkershoek Forestry Research Station was established, research into the effects of afforestation on stream flow from small mountain catchments has been conducted not only in the southwestern Cape but amongst others also at Cathedral Peak in the Natal Drakensberg and Makubulaan in the eastern Transvaal.

Apart from demonstrating that afforestation with exotic species undoubtedly reduces streamflow, an important finding of these studies has been that overland stormflow comprises but a small fraction of the total runoff. Midgley and Scott (1994) concluded from the use of stable isotopes D and ^{18}O that stormflows in Jonkershoek catchments are generated by the rapid displacement of subsurface water. Only 8.5% or less of the stormflow consists of rainwater that has not followed a subsurface route. The rest is presumably accounted for by so-called interflow which has to be viewed as distinct from groundwater. According to figures supplied by Forestek and given in the manual of Bredenkamp et al. (1995) base flow averages about 75% of the total annual runoff of three small catchments in the southwestern Cape and accounts on average for about 28% of the rainfall.

8.1.4 National base flow map

As adjunct to the Water Research Commission's project 298 "The surface water resources of South Africa" the groundwater component of river flow was determined in each of the approximately 2000 quaternary catchments into which the country with inclusion of Swaziland and Lesotho has been divided. The map was published as part of a set of national groundwater maps (Vegter 1995a).

8.2 Relationships between Groundwater and Stream flow

This paper was solicited by the organisers of the 1996 Warmbath workshop on surface and ground water issues in South Africa on account of the base flow map that was produced as part of a set of national groundwater maps and the use to which it was put in the construction of a national recharge map. The idea of the organisers was ostensibly that this paper should deal with base flow as a means of estimating recharge. However to put base flow in its proper context, it is deemed necessary to draw attention to the different relationships between groundwater and streams.

8.2.1 Types of streams

Streams are usually divided into three classes depending on their runoff characteristics namely Ephemeral, Intermittent and Perennial. Although such a division also distinguishes indirectly and to a large degree between the relationships of streams to groundwater, a more direct approach is essential. Three main types may be distinguished:

1) Piezometric surface at all times below streambed level - characteristic of ephemeral streams, though not necessarily limited to them only. One of two conditions may occur:
a) Material between streambed and piezometric surface is pervious - the stream is influent and the piezometric surface slopes downward away from the stream. The stream acts as a sink and recharges groundwater. Little or no work has been undertaken in South Africa to quantify stream recharge.
b) Intervening material more or less impervious -very little or no recharge takes place - i.e. a detached stream (author's terminology).

It would appear that conditions over much of the country militate against rivers being all but minor localised sources of recharge for the following reasons:

- The hard-rock environment and dearth of laterally extensive alluvial deposits below riverbed level;
- The fact that the watertable is a sub-dued replica of the topography over the greater part of South Africa inhibits the lateral expansion of the recharge mound that is being built up below the river by infiltrating water; and
- Rocky riverbeds and silty channels limit infiltration.

2) Piezometric surface slopes laterally down towards the stream. One of the following conditions may be encountered:
a) Groundwater reaches and emerges into the stream at all times. The piezometric surface at the stream is permanently above the stream stage and the material between it and the streambed is pervious - porous or fractured. The stream acts as a drain; is effluent and perennial.
b) Groundwater from the catchment area emerges into the stream at intervals i.e. for a while after recharge episodes - the stream is intermittent. During dry periods groundwater storage is depleted by the effluent seepage or in combination with evapotranspiration from the stream banks and within the catchment. Groundwater may be replenished to a certain extent in the immediate vicinity of the stream by storm runoff. In the absence of rechargeable alluvial deposits and / or porous decomposed rock, replenishment from storm runoff would appear to be of minor importance compared to the volume of water recharged over the catchment area. Recharge from storm runoff is restricted in its lateral extent as well as volumetrically by low storage capacity.
c) Groundwater does not reach the stream because it is permanently being dissipated along its flow path towards the stream by evapotranspiration - a famished stream (author's terminology).

3) The piezometric level fluctuates alternately above and below the stream stage. The stream is underlain and bordered by alluvial deposits and / or porous decomposed rock. The stream is alternately in- and effluent. Comparatively speaking, groundwater flow from the hard-rock catchment towards the stream generally is of minor importance and in certain cases of no consequence at all. The inter-action between alluvium and stream is virtually all that matters.

Note that a particular designation does not necessarily apply to the full length of a stream. Most rivers exhibit different characteristics along different reaches. A good but somewhat extraordinary example is that of the Wonderfontein Spruit on the Far West Rand which is influent and disappears along certain stretches and effluent along others where fed by spring flow.

8.2.2 Examples of types

Some examples of stream types are:

- Influent: Kuruman River downstream from at least Frylinckspan; Molopo River downstream from at least Tshidilamolomo, Phepane, Kgokgole, and other "laagtes" in the catchments of the Kuruman and Molopo River;
- Detached: Relatively steeply graded and dry, rocky stream beds particularly in the arid northwestern parts of the country;
- Effluent: Upper reaches of perennial rivers rising on the eastern escarpment such as the Vaal, Olifants (Tvl), Tugela, Blyde, Komati etc;
- Intermittent: Streams in the Karoo such as the upper reaches of the Salt River (Beaufort West), the Kamdeboo, the Sundays, the Brak (De Aar);
- Famished: Rocky sections of the Limpopo River such as alluvium-free stretches between Stockpoort 1 LQ and Sannandale 9 LQ; and the steeper graded section between the junctions with the Lephalala and Motlouse Rivers. The bordering country which is underlain by granulite-gneiss of the Limpopo Mobile Belt is very poorly endowed with groundwater;
- In-/effluent: Wide stretches of relatively unexploited alluvium along the Limpopo River between the confluence of the Marico and Crocodile Rivers and its junction with Mahalapswe River. Under conditions of heavy exploitation, as is presumably the case downstream along the Limpopo at Weipie and along the Crocodile River between

Koedoeskop and Thabazimbi, the stream may change its dual character to influent only. The latter has been declared a Subterranean Water Control Area and has been the subject of a number of studies.

8.3 Runoff Process and Hydrograph Analysis

8.3.1 Components of runoff

Runoff from rain (or snowmelt) reaches the stream channels by several routes. It is generally recognised that there are four different components of runoff, each of which reaches the stream by a different path:

(1) Surface runoff is the residual after interception, surface ponding and infiltration have been extracted from precipitation. It reaches the streams by traveling over the soil surface. In this context the term stream includes not only the larger permanent streams but also the small rivulets that carry water only during and immediately after rains. Surface runoff therefore occurs over relatively short distances to the nearest channels.

(2) Interflow is the water that infiltrates the soil surface and moves laterally through the upper soil horizons until it is intercepted by a channel or until it returns to the surface downslope of its point of infiltration. Wet weather seeps and springs are the result of interflow.

(3) Groundwater flow originates from water that percolates down to the groundwater table. Groundwater flow follows a more devious route to the stream than any other component. As a result the water represented by the groundwater accretion from a particular storm is discharged into the stream over a long period of time.

(4) Channel precipitation is that precipitation which falls directly on the water surfaces of lakes and streams. As it is generally a small component, it is usually included with surface runoff.

8.3.2 Hydrograph Analysis

Because of the differing characteristics of the four components of runoff, the shape of the runoff hydrograph is dependent, among other things, on the relative proportions of each component present. Local variations in rainfall, infiltration and antecedent conditions preclude any attempts to identify each component of runoff.

It is common, therefore, in the solution of practical problems, to separate only two portions. Surface runoff, channel precipitation and interflow are usually grouped into a single item designated as direct (or storm) runoff or quickflow. Groundwater, or base flow, is treated as a separate item.

Several procedures exist for separating the hydrograph into the quickflow and base flow components. All involve an element of subjectivity while most, if not all, rely on measured flows at relatively small time steps of the order of 1 hour to 1 day.

One of the products of the WR90 project (Midgley et al, 1990) is a time series of monthly flows for each of the approximately 2000 quaternary catchments into which the study area is divided. Each time series covers the 70 years period from 1920 to 1989 (hydro years).

In order to split the monthly flows into surface and groundwater components, a simple procedure first developed by Herold (1980) was adopted. This procedure is described below:

Let Q_i = total flow during month
QG_i = groundwater contribution
QS_i = surface runoff

Then: $$Q_i = QG_i + QS_i \qquad (1)$$

The assumption is made that all flow below GGMAX is groundwater flow, thus:

$$QS_i = Q_i - GGMAX \qquad (\text{for } Q_i > QGMAX)$$
or $$QS_i = 0 \qquad (\text{for } Qi \leq QGMAX) \qquad (2)$$

and hence $QG_i = Q_i - QS_i$

The value of GMAX is adjusted each month according to the surface runoff during the preceding month and is assumed to decay with time, hence:

$$GGMAX_i = DECAY \cdot GGMAX_{i-1} + PG \cdot QS_{i-1} / 100 \qquad (3)$$

where subscripts i and i -1 refer to the current and preceding months;
DECAY = Groundwater decay factor (0 < DECAY < 1)
PG = Groundwater growth factor (0 < PG < 100)

An added constraint is that GGMAX may not fall below a specified value, QGMAX.

Calibration of this model is achieved by selecting appropriate values of DECAY, PG and QGMAX so that a realistic split between surface runoff and groundwater is obtained. Calibration is facilitated by graphical output of the total and groundwater hydrographs to the computer screen (Figure 8.1). As the process is somewhat subjective it was tested against a more vigorous analysis on the daily flow record at DWA&F gauge X3H001 on the Sabie River. The results indicated that the method proposed was sufficiently accurate.

8.4 Base flow from SA Quaternary Catchments

8.4.1 WR90 overview

The primary objective of the Water Resources 1990 (WR90) project was to provide a basis for preliminary planning of water resources development.

It was decided to provide the water resources information at a quaternary sub-catchment level. South Africa is divided into 22 primary drainage regions, further subdivided into secondary and tertiary sub-catchment areas. The WR90 team defined the quaternary (i.e. fourth-level) sub-catchment boundaries to create close on 2000 sub-catchments. The average area of the quaternary catchment is about 650km^2, varying from around 100km^2 in some mountainous areas to several thousand km^2 in the arid north – western interior. All catchment boundaries were digitized to form the basic GIS coverage for the project. Information provided for each quaternary catchment includes rainfall, evaporation, runoff, land cover and water-use, soil types, natural vegetation and underlying geology.

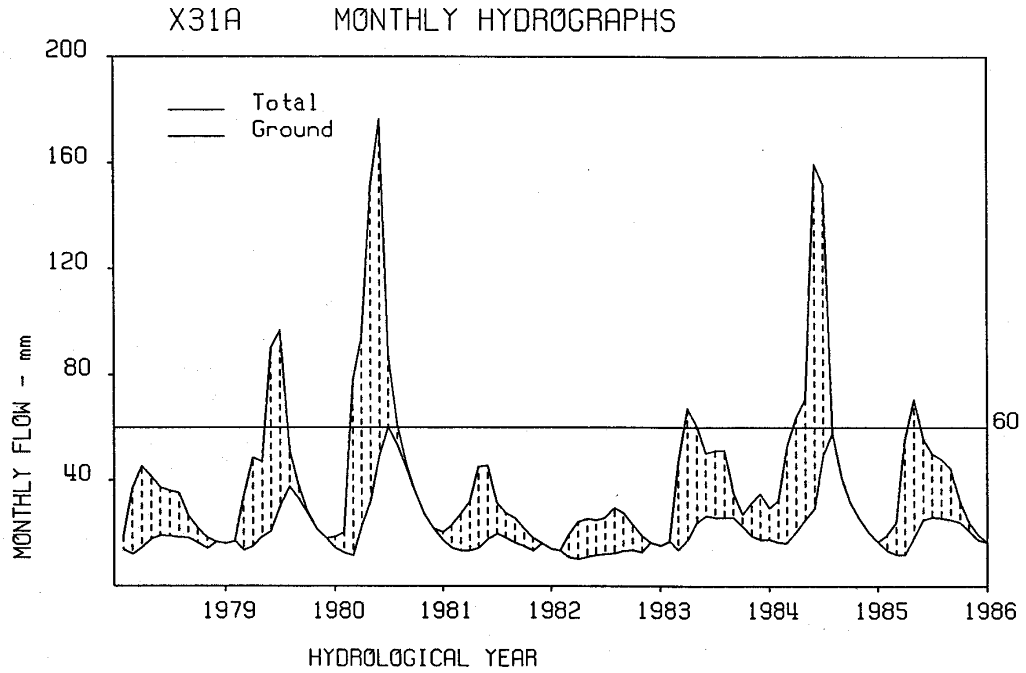

Figure 8.1 Graphic output of hydrograph separation based on Herold (1980).

The publication is in 6 volumes. The information and time series of the hydrometeorological data are provided in a set of appendices, whereas the spatial information is shown in a set of 10 maps. A user's manual is provided which gives the procedures used to create the information and maps, outlines problems and pitfalls and also provides a user's guide to the documents, including worked examples.

Perhaps the most important aspect of the WR90 project was the procedure used to estimate runoff for each quaternary catchment. Basic information on runoff was in the form of measured flows at nearly 500 suitable stream flow gauges. With few exceptions the flow regions of the gauged catchments have been impacted by upstream developments.

The rainfall-runoff model originally developed by Pitman (1973) was improved to accommodate for the effects of upstream developments. Calibration of the model, named WRSM90 (Pitman & Kakebeeke, 1991), provided sets of model parameters for each gauged catchment. Regionalisation of the parameters was accomplished with reference to supplementary information on topography, rainfall, geology, soil type and vegetation.

The regionalized parameters were then used to generate synthetic long-term (70 years) virgin flow sequences for each catchment. These (monthly) flow sequences formed the primary input to the base flow analysis described in Section 8.3.2.

8.4.2 Base flow characteristics for SA rivers

Table 8.2 lists hydrological information (incl. base flow) for each of the 22 primary drainage regions in South Africa. For the country as a whole base flow accounts for just over 20% of the total runoff of approximately $51x10^9$ m^3. The total base flow is less than 2% of the rainfall and this percentage is almost insignificant in the drier regions (e.g. F, J, L, N, P, & Q). The highest percentages are to be found in the (mostly) well-watered regions of the western and southern Cape (G, H, & K) and the eastern escarpment (T, U, V, W, & X). Examination at a quaternary catchment level shows that base flows are greatest in rainfall areas underlain by dolomite. This is the case for the upper Blyde and Sabie rivers, where mean base flows approaches 20% of rainfall and accounts for roughly half of the total runoff.

At the opposite end of the scale there are large areas where base flow is effectively zero. In the summer rainfall region no base flow is found where the MAP is less than 500 mm, whereas in the winter and year-round rainfall regions the threshold is somewhat lower. A MAP of only 300 mm is sufficient to yield some base flow in steep mountain catchments, but elsewhere the cut-off is around 400 mm.

From year to year the base flow is far less variable than the quick flow component of runoff. In extreme dry years the base flow is about one-quarter to one half of the mean and in very wet years the base flow is about two to three times the mean. Addition of the quick flow component renders a total runoff that is far more variable, with dry year flows in the range of zero to one-third and wet years in the range of three to twenty times the mean. In the case of both base flow and total flow the lower variability is associated with the higher rainfall areas, and vice versa.

Table 8.2 Base flow for primary drainage regions.

Drainage region	Area (km^2)	MAP (mm)	Mean annual runoff			Mean annual base flow			
			(10^6m^3)	(mm)	%(MAP)	(10^6m^3)	(mm)	%(MAP)	%(MAR)
A	109610	528	2176	19.9	3.8	690	6.3	1.2	31.7
B	73550	620	2651	36.0	5.8	758	10.3	1.7	28.6
C	196293	571	4298	21.9	3.8	606	3.1	0.5	14.1
D	409621	315	6987	17.1	5.4	947	2.3	0.7	13.6
E	49063	212	1008	20.5	9.7	102	2.1	1.0	10.1
F	28623	129	24	0.8	0.6	0	0.0	0.0	0.0
G	25312	476	1986	78.5	16.5	250	9.9	2.1	12.6
H	15530	545	2059	132.6	24.3	245	15.8	2.9	11.9
J	45134	260	662	14.7	5.6	50	1.1	0.4	7.6
K	7220	763	1307	181.0	23.7	298	41.3	5.4	22.8
L	34731	283	495	14.3	5.0	46	1.3	0.5	9.3
M	2630	555	151	57.4	10.3	10	6.6	1.2	6.6
N	21428	330	279	13.0	3.9	2	0.1	<0.1	0.7
P	5322	560	174	32.7	5.8	4	0.8	0.1	2.3
Q	30243	410	519	17.2	4.2	29	1.0	0.2	5.6
R	7936	675	580	73.1	10.8	87	11.0	1.6	15.0
S	20485	610	1043	50.9	8.3	209	10.2	1.7	20.0
T	46684	863	7397	158.4	18.4	1526	32.7	3.8	20.6
U	18321	935	3128	170.7	18.3	868	47.4	5.1	27.7
V	29046	829	3994	137.5	16.6	770	26.5	3.2	19.3
W	59200	825	6533	110.4	13.4	2000	33.8	4.1	30.6
X	31157	715	3361	107.9	15.1	1370	44.0	6.1	40.8
TOTAL	1267100	474	50812	40.1	8.5	10868	8.6	1.8	21.4

8.5 Is Base Flow a Measure of Groundwater Recharge?

8.5.1 Recharge defined

To obviate any misunderstanding, it is necessary to define the term recharge. The sense in which it is used here agrees with the definition in the AGI Glossary (1973), namely recharge comprises the processes involved in the absorption and addition of water to the zone of saturation. It does not include water reaching the belt of soil water or the intermediate zone of aeration. The term also means the volume of water added to the zone of saturation.

8.5.2 Flow paths

Based on the premise that in the long-term there is neither accretion nor depletion of groundwater storage, replenishment must be balanced by an equivalent loss of groundwater. The movement of groundwater to streams and lower-lying areas is evident from groundwater contour maps. The existence of such hydraulic gradients does however not necessarily imply, uninterrupted groundwater flow paths to the stream and emergence of groundwater in the stream. Flow paths in hard-rock areas are often discontinuous. Under natural conditions all of the recharge therefore need not find its way to the stream, being discharged locally within the catchment through small springs and seepages and/or less obvious areas of evapotranspiration. Abstraction from boreholes and wells will generally further detract from the volume reaching the stream.

8.5.3 Groundwater loss other than in stream

Examples of natural losses are the vlei's, small springs and seepages found in Kwazulu-Natal catchments. Natural groundwater loss is however not necessarily confined to shallow groundwater occurrences and lack of base flow should not be taken as an indication of zero recharge. Evapotranspiration may completely dispose of recharge in areas of shallow groundwater while transpiration by deeper-rooted phreatophytes may account for loss where the watertable is deep (see Vegter 1995b).

8.5.4 Base flow underestimates recharge

It should be clear that the groundwater component of river flow referred to as groundwater base flow or base flow for short, provides a minimum estimate of recharge. Recharge estimates obtained from a variety of groundwater studies within and just outside the base flow areas were compared with those derived from base flow. The mean difference between the two sets of values indicates underestimation of recharge by base flow by about 30 mm a^{-1} (Vegter 1995a).

Temporal variability of base flow

A deficiency of the national base flow map should be noted. Whereas annual mean base flow and its annual variability are portrayed, no indication is given about its variability during the course of a year.

In following section the question will be discussed as to whether the mean base flow from any particular catchment may be taken as a reliable - albeit somewhat under-rated - indicator of its exploitation potential.

8.6 National Recharge Map

The compilation of the National Recharge Map was based on:
1) The national base flow map. Because of underestimation, base flow figures at the bottom end of the scale, were adjusted upwards by 25 mm a^{-1}. For higher base flow rates, the addition was gradually reduced - being zero for mean base flow figures above 100 mm a^{-1}.
2) Recharge was estimated at eight localities in the Northwestern Cape using the relationship between recharge and rainfall in excess of 15 mm d^{-1} according to the De Aar model of Vegter (1992).
3) Carlsson et al. (1994) state that based on the results of two approaches, the soil-water and the chloride mass balance methods - recharge in the south-western half of Botswana, has been assessed at less than 1 mm a^{-1}. Considering that values as low as 2.6 and 2.9 mm a^{-1} have been calculated for a thin soil cover at Pella and Garies, according to the De Aar model, it appeared reasonable to map the West Coast sand belt, north of the Olifants River, the Koa valley of western Bushmanland and the western Kalahari as areas with a mean recharge of less than 1 mm a^{-1}.
4) In the no-base flow area, 28 scattered recharge estimates from groundwater studies are available.

Recharge contours were drawn according to adjusted base flow values, and were interpolated between the edge of the base flow areas and the 1 mm a^{-1} recharge contour, using the point recharge figures mentioned under 2) and 3), and the ACRU effective rainfall map.

It should be realised that the contouring could not strictly adhere to the local recharge values, the base flow and effective rainfall contours and that discretion had to be used in adjusting and smoothing recharge contours. The recharge map should be seen as depicting broad trends rather than portraying recharge numerically with a degree of assurance. On a local scale considerable variation and divergence from that presented on the map, may be expected. The map provides a very rough idea of the magnitude of the replenishable groundwater resources but cannot be used for estimating exploitation potential or safe yield even if it were further refined.

8.7 Recharge and Exploitation Potential

The raison d'être or driving force for concentrating attention on recharge is the growing demand for potable water supplies, especially for rural communities, which calls for assessment of the exploitation potential of groundwater resources. The question however is whether exploitation potential can be deduced from a quantitative determination of recharge, prior to development of the groundwater supply.

8.7.1 Requirements for a groundwater supply

In developing a groundwater supply the following requirements generally have to be met:

- a minimum quantity of water must be provided at all times. This quantity, which has variously been termed "safe yield"; "firm yield"; "harvest potential"; and "exploitation potential", obviously is limited by recharge. Note that these terms refer to distinct hydraulic units / groundwater bodies / compartments, not to boreholes. The terms "exploitation potential" etc. imply determination prior to full development of the resource.
- maintenance of an acceptable water quality, and

- no undesirable environmental effects such as deterioration of vegetal cover and resultant onset of gully erosion, subsidence and the formation of sinkholes.

In the following paragraphs the problems of predicting exploitation potential will be discussed.

8.7.2 The concept of exploitation potential

Exploitation of groundwater means salvaging groundwater which otherwise would be discharged naturally through springs, seepage into streams and evapotranspiration. These discharges, which in terms of groundwater exploitation are viewed as losses, are in the long-term balanced by recharge. In other words, by salvaging as much as possible of the losses, abstraction approaches recharge. In addition, abstraction lowers the watertable, thereby creating additional storage space and possibly inducing additional recharge. The effectiveness of salvaging natural discharges, and of approaching the limit set by recharge, depends on

- the hydrogeologic character and structure of the saturated zone: the configuration and relationship of aquifers, aquitards and aquifuges to points / zones of groundwater loss.
- the practicability of siting and pumping boreholes to suppress all losses.
- the economics of sinking and pumping boreholes.

8.7.3 Problems associated with the concept of exploitation potential

Several problems are immediately clear:
a) Suppression of evapotranspiration losses may require clearing of vegetation or may lead to deterioration of vegetal cover and eventually soil erosion.
b) At any locality, the inhomogeneous hydraulic nature of saturated weathered and fractured hard-rock formations, can only be established from detailed hydrogeologic examination, drilling and pumptesting.
c) As the rate of natural groundwater discharge hardly can be expected to be constant, its suppression entails a corresponding varying rate of pumping. A steady supply is obviously not realised in this manner.

The last problem can be overcome only if fluctuating recharge can be regulated by storage i.e.:

- the volume of groundwater in storage must always be adequate to maintain the maximum possible steady withdrawal rate; i.e. the transmissivity should not decrease as result of the lowering of the piezometric level to the point where flow rates to boreholes and pumping rates can not be maintained.
- the storage capacity at any time must be large enough to accommodate recharge, and
- the watertable must be prevented from rising above the level for the onset of natural discharge.

8.7.4 Can these requirements be met?

To gain some ideas whether these conditions can be satisfied, the following conditions need to be considered:

1) Large volume of groundwater in storage and large storage capacity.
2) a - Limited volume of groundwater in storage and b - Limited volume of groundwater in storage as well as limited storage capacity.

Certain dolomitic groundwater compartments and certain coastal sand deposits fall in the first category. The saturated thickness of water-bearing and transmissive dolomite and sand deposits is considerable, up to 100 metres, and the volume of groundwater in storage is several tens to a hundred times mean annual recharge Lowering of the watertable to prevent loss will therefore not effect borehole yields to any appreciable degree. Theoretically it should be possible to develop a supply equivalent to mean recharge.

The ability of suppressing natural discharge, particularly in the case of the dolomite however, cannot be gauged prior to local hydrogeological investigation and the drilling of boreholes. An effective borehole spread and pumping regime can only be developed experimentally. This means that even if recharge is a known quantity, exploitation potential cannot be predicted reliably beforehand. See in this connection also the conclusion reached by the Committee on Ground -Water of the American Society of Civil Engineers (1961) as quoted by Kazman (1988) pp 216 - 217.

In South Africa categories 2a and b are almost ubiquitous. Both consist of a thin zone of saturated weathered and fractured hard-rock. The volume of groundwater stored is of the same order of magnitude as recharge. The difference between 2a and 2b lies in the different thicknesses and / or storage capacities of the zone of aeration:

2a. The overlying zone of aeration poses no restriction on recharge - the watertable is deep or the zone consists of porous decomposed rock. These conditions occur where: the rock formations are relatively deeply weathered and fractured and where
i) there is little recharge - typical of semi-arid to arid conditions or where the greater part of infiltration is dissipated from the zone of aeration through transpiration.
ii) groundwater is transpired by deep-rooted vegetation e.g. Bushveld in the northern parts of country. Under these conditions loss of groundwater through transpiration can be prevented only through bush-clearing.
iii) there is topographic relief' and groundwater drains out of the aquifer system at lower levels
Under 2a conditions borehole yields drop rapidly as transmissivity is reduced with dewatering of the saturated zone and a limit is placed on the extent to which water in storage can be drawn on. Losses are only partially recoverable.

2b. The storage capacity of the zone of aeration is limited owing to a shallow watertable and / or it consisting of fractured hard-rock. This is the case over large areas of South Africa, which are underlain by Karoo rocks.

Potential recharge is not realised when the available storage has been filled to capacity. This condition is not restricted to a particular region. Within a single area this condition may exist alongside that with adequate storage capacity.

Another 2b characteristic is that for a short period after a recharge event a considerable part of the recharge usually is discharged by springs and seepage into streams. Where groundwater is being recharged at fairly regular intervals, as happens in the humid higher rainfall areas, perennial flow is possible. As noted in the Introduction, regular rainfall and recharge, rather than favourable storage characteristics, are responsible for sustaining stream flow. In the arid and semi-arid parts with irregularly and widely spaced recharge events, the remaining volume in storage cannot sustain a steady rate of supply, equal to the mean recharge rate. Under these circumstances the available rate of supply, after a recharge event, unavoidably exceeds the mean rate of recharge, and falls short thereafter during dry periods.

It should be evident that at any locality the maximum possible steady borehole supply, "exploitation potential" that may be salvaged from natural losses, can not be estimated with any degree of assurance without at least detailed local hydrogeological investigation, sinking and pumptesting of boreholes. In addition knowledge of the temporal variability of recharge is necessary.

Taking further into account that the natural groundwater regime is changed by groundwater exploitation:

- the creation of additional storage space may enhance recharge;
- groundwater exploitation is inextricably accompanied by changes in land-use which in turn may either promote or reduce recharge.

It is clear that recharge cannot be seen as an external parameter that may be determined independently.

8.7.5 Summary

The question of exploitation potential has to be looked at from the recharge as well as discharge points of view.

Considerations from the recharge point of view are:

- Storage and storage capacity must be adequate at all times to maintain the maximum possible withdrawal rate and the necessary minimum transmissivity accommodate for potential recharge at any time;
- Creation of additional storage capacity through abstraction may enhance recharge;
- Land-use changes accompanying groundwater exploitation may either enhance or reduce recharge.

From the discharge point of view the capability of salvaging unwanted discharges, so-called losses, has to be considered. This depends on:

- Identification, location and assessment of natural discharges;
- The hydrogeological character and structure of the saturated zone and its relationship to points of natural discharge;
- The practicality of siting, sinking and pumping boreholes to suppress losses;
- The economics of sinking and pumping boreholes;
- Avoidance of undesirable / detrimental environmental effects of groundwater abstraction.

8.8 Conclusions

Perennial exploitation of groundwater entails the availability and use of a steady supply of acceptable quality without detrimental / undesirable environmental effects.

While providing an upper limit, recharge determinations are inadequate by themselves for estimating the development potential of groundwater sources.

Recharge is not uniquely determinable in terms of rainfall. It depends on the status of an aquifer system, whether in a virgin condition or a state of exploitation as well as ground surface conditions - natural or disturbed by man.

Exploitation potential is not a purely hydrologic quantity that can be determined on the basis of natural factors only and prior to groundwater development. It is inescapably tied up with practical and economic exploitation considerations.

The extent to which a particular aquifer system / groundwater unit / compartment can be developed is only determinable by developing it and by approaching its full exploitation potential in a stepwise fashion. Development of a groundwater supply entails detailed local hydrogeologic investigation, drilling and pumptesting followed by monitoring its performance under exploitation and adjusting abstraction rates accordingly.

Rather than accentuating recharge as a research priority, future attention should be aimed at determining actual rates of groundwater exploitation and exploitability at a number of places, especially those where the maximum has apparently been reached or exceeded, and by guiding and monitoring exploitation in those areas where critical water supply situations are to be solved by groundwater. A byproduct of such studies will be estimates of recharge and less irrecoverable losses under reigning practical and economic circumstances.

The impact of groundwater development on the natural environment needs to be monitored simultaneously.

The following maps are available at a scale of 1: 2 500 000 (Vegter, 1995a):

- Groundwater component of river flow
- Mean Annual groundwater recharge
- Effective Rainfall (ACRU)
- Base flow as a percentage of mean annual precipitation
- Base flow as a percentage of mean annual runoff
- Quaternary catchments

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9. Recharge Estimation in Fractured Rock Aquifer from Rainfall - Spring Flow Comparisons: The Uitenhage Spring Case

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ABSTRACT This paper presents a synthesis of recharge estimation in the fractured Coega Ridge Aquifer northwest of Port Elizabeth in South Africa. The aquifer comprises quartz arenites of the Table Mountain Group and is partly drained through the Uitenhage Spring. Various recharge estimation methods based on rainfall – spring flow relationships, such as a simple water balance, groundwater dating (^{14}C), chloride mass balance, cumulative rainfall departure and a 'moving average' water balance approach were applied. The latter approach simulates the Uitenhage Spring flow precisely for a recharge area of 60 km^2 and suggests that outflow is represented by a much smaller time span of 23 years of rainfall infiltration than the circulation period of hundreds of years which was derived from groundwater dating. The 'moving average' approach was also used for assessing recharge during undisturbed natural conditions (prior to 1909) and recharge during 'disturbed' conditions (accommodating for borehole drilling and abstractions since 1909).

9.1 Introduction

9.1.1 Historical account of the Uitenhage Spring

The Uitenhage Spring presently consists of nine eyes with a total flow of 45 l/s and is the largest spring in the Uitenhage Artesian Basin (UAB). It is important both from a historical and strategic viewpoint. Stone-age artefacts and a pre-historic mammal tooth found at the spring indicate that the eyes have been a constant supply of water to early inhabitants for at least 200000 years (Hickson, 1989). Spring flow, however, has decreased gradually as shown in Figure 9.1. Hickson (1989) reports that in 1773 the total yield from 20 different eyes was estimated at 105 l/s, whereas in 1829 this flow had decreased to 80 l/s. The first official and reliable gauging was in 1867 when a flow of 89 l/s was recorded. Since 1899, spring flow has varied over time, especially in the period of early 1900s to 1960s, mostly due to abstractions from the aquifer (Maclear and Woodford, 1995).

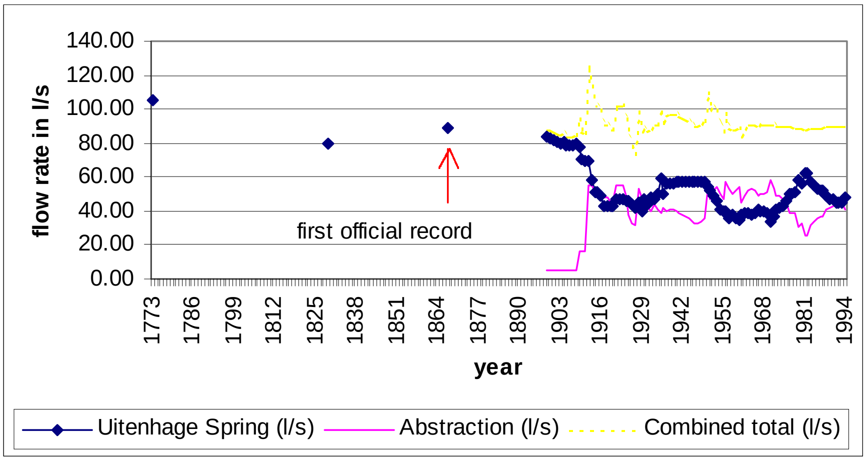

Figure 9.1 Flow variations at Uitenhage Spring (1773 to 1994).

In the mid 1960s the Spring supplied 25% of Uitenhage's water, and presently it supplies about 15% of Uitenhage's bulk water requirements (Maclear, 2002). Note that Uitenhage has one of Eastern Cape's largest industrial areas. The following two phases of spring flow are recognised (adapted from Maclear and Woodford, 1995):

I Before commencement of borehole drilling in the Coega Ridge Aquifer (CRA) in 1908, the Uitenhage Spring was in a natural flow status.
II From 1909 and onwards, the impact of (uncontrolled) borehole drilling in the CRA resulted in an increase in total abstraction and thus in a 'disturbed' flow status.

From 1993 to present, spring flow recovery occurred, which can be attributed to the Department of Water Affairs and Forestry programme of sealing old artesian boreholes.

It is of interest to note that the major increase in boreholes drilled in the Coega Ridge Aquifer (CRA) did not increase the total yield of the unit, which remained relatively constant at 80 l/s, but rather increased the depth to the piezometric level, thus changing pressure conditions from artesian to sub-artesian. The yield from the unit decreased slightly towards the end of the 1960s, as a result of increased leakage (through rusted borehole casings) of groundwater, under artesian pressure, into the confining overlying Cretaceous sediments.

With the introduction of the ecological reserve concept and licence application under the new South African Water Act of 1998, the status and sustainability of the Uitenhage Spring has become important and of interest to hydrogeologists and water managers.

9.1.2 Hydrogeologic setting of the Coega Ridge Aquifer

The Coega Ridge Aquifer (CRA), within which the Uitenhage Spring occurs, is located in the north-central portion of the Uitenhage Artesian Basin (UAB; Figure 9.2). The aquifer comprises quartz arenites of the Table Mountain Group (TMG). East of the Uitenhage Spring the aquifer is overlain by impermeable mudstones and siltstones of the Uitenhage Group. The aquifer stretches further eastward along the Coega Ridge to the coast. The steep (50°) northerly dipping TMG-Bokkeveld Group contact to the north forms the northern boundary of the aquifer.

The CRA is economically important as the source of artesian to semi-artesian groundwater for irrigation of citrus (for export) and lucerne, especially at sites of large-scale abstraction, viz. Uitenhage, Sandfontein, Amanzi Estates, Coega Kop and Wells Estate (Figure 9.2). Although boreholes along the Coega Ridge are hydraulically connected (Maclear and Woodford, 1995), the nature of connectivity is not well known.

The eyes comprising the Uitenhage Spring are fault-controlled and lie along a 120 m N-S seepage front, conform the strike of the contact between the Table Mountain Group (TMG) and the Uitenhage Group (Maclear, 1996). The spring is fed from the unconfined section of the TMG aquifer system. Its catchment boundaries are limited by the M10C catchment to the west and the north, the contact zone of the Table Mountain – Uitenhage Groups to the east, and the impermeable Coega Fault to the south.

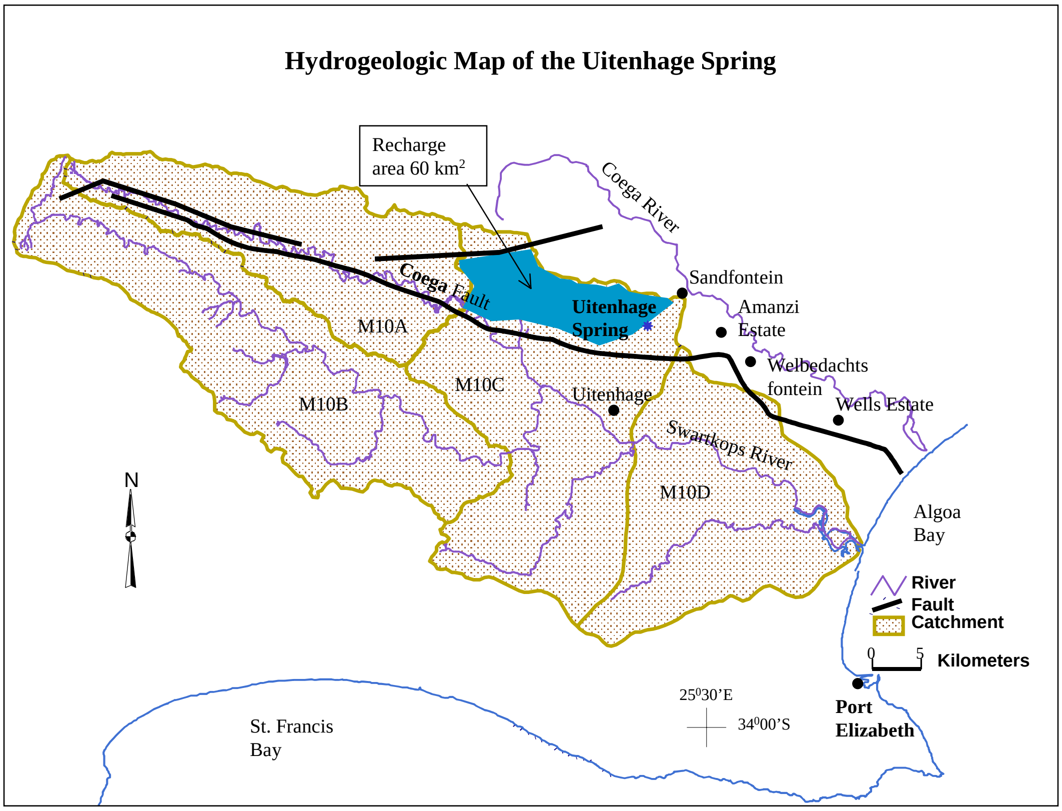

Figure 9.2 Hydrogeology of the Uitenhage Artesian Basin and the Uitenhage Spring location.

9.1.3 Aquifer characteristics

Groundwater temperature data from boreholes drilled into the TMG Aquifer in the UAB are related to circulation depth. Based on a geothermal gradient of 37 m/ºC in the TMG (from Jones, 1992) and an ambient temperature for recharge water of 18 ºC, the temperature of Eye 2 at Uitenhage Spring suggests a groundwater circulation depth of 185 m (Maclear, 1996). A much higher circulation depth of 1130 m is calculated for the 54 ºC artesian groundwater that flowed out of the old Zwartkops Spa borehole (Maclear, 1996). Precise determination of circulation depth, however, is complicated by varying thickness of the overlying Cretaceous sediments, which influences the geothermal gradient.

Groundwater from the Uitenhage Spring is of an excellent quality with a salinity of 15 mS/m and is suited for drinking purposes. Groundwater samples from the confined zones of the CRA did not contain measurable Tritium (Venables, 1985), suggesting that these groundwaters are older than 55 years and that no recent recharge has taken place at these locations.

Coega Ridge Aquifer characteristics are as follows: transmissivity varies between 50 and 400 m^2/d, storativity is about $2*10^{-4}$, yields vary between 3 and 23 l/s, average TDS is 170 mg/l, groundwater temperature varies between 23 and 33 ^oC and the overall size of the aquifer is 470 km^2.

9.2 Groundwater Recharge Estimation

9.2.1 Simple water balance

Kok (1992) estimated recharge (Re) at 83% of the average annual rainfall based on the following simple water balance:

$$\mathrm{Re}(\%) = \frac{Q_s}{Rf * A} 100 \qquad (1)$$

where Q_s is annual Uitenhage Spring flow (=2.4 Mm^3/yr) and R_f is average annual rainfall (=460 mm) over the recharge area A (=6.3 km^2). The relatively high recharge estimate probably originates from an underestimation of the size of the recharge area.

9.2.2 Groundwater dating

Talma et al. (1982) demonstrated aging of groundwater in an easterly, down-gradient, direction along the strike of the Coega Ridge, based on a detailed hydrochemical and isotope (^{14}C) investigation in the recharge area and a confined section of the TMG aquifer near Uitenhage. Calculated groundwater ages range from 1350 years at the Uitenhage Spring immediately east of the recharge area, to 28000 years at the Coega Kop discharge area near the sea (Heaton et al., 1986). From these dates, the flow rate along the flowpath in the CRA was calculated by Maclear (1996) at 0.76 m/yr. Heaton et al. (1986) furthermore estimated that most recharge water discharges in the Uitenhage Spring at the edge of the unconfined area and that less than 3% of the total recharge flows into the confined section of the TMG aquifer to the east.

9.2.3 Chloride mass balance method

Maclear (1996) estimated recharge between 24 to 55% of average annual rainfall using the chloride mass balance method:

$$\mathrm{Re}(mm) = \frac{Cl_{rf}}{Cl_{gw}} Rf \qquad (2)$$

where Cl_{rf} is the rainfall chloride concentration, Cl_{gw} is the chloride concentration of Spring water and Rf is the annual mean rainfall over the recharge area. Maclear (1996) justfied the use of Spring water chloride content for estimating recharge based on the following reasoning:

- The fractured aquifer has experienced “flushing” by low TDS groundwater for millennia (Hickson, 1989).
- Any leachable Cl has been flushed from the fractures.
- Cl contamination from the Uitenhage Group deposits is not possible, as these deposits are confined to the more distal (central) regions to the east, down-gradient of the Spring.

Maclear (1996), however, acknowledged that the results from his study were based on Cl concentrations from only two rain samples and the average of these may not be representative for a long-term average value.

9.2.4 'Moving average' water balance approach

Maclear and Woodford (1995) applied the cumulative rainfall departure (CRD) method in an attempt to determine the time lag between rainfall recharge and spring discharge and thus to validate the age determination of Talma et al (1982) and Heaton et al. (1986). It was not possible, however, to obtain a good correlation between cumulative rainfall departure and spring discharge. A new relationship between rainfall and Spring flow is proposed with the aim to determine periods of rainfall events that can be attributed to specific Spring flow:

$$Q_{si} = \mathrm{Re}\left(\frac{1}{m}\sum_{j=0}^{m-1} Rf_{i-j}\right)A - Q_{pi} - Q_{leaki} \quad \text{(i, m = 1, 2, 3…)} \tag{3}$$

where Q_{si} is spring flow in m^3/d during month i, Q_{pi} and Q_{leaki} are groundwater pumped out of boreholes and the amount of water leaked from the aquifer system during the i-th month, respectively, Re is the percentage of rainfall attributed to recharge, and A is the aquifer recharge area in m^2.

The above relationship assumes that the present spring flow is fed by the current and past rain water infiltration in the recharge area over m time intervals. Flow from the Uitenhage Spring may be simulated based on this 'moving average' approach, provided that Q_{pi} and Q_{leaki} can be determined. In the case of the Uitenhage Spring, parameter Q_{leaki} is ignored. Parameter Q_{pi} is obtained from historic records (Marais, 1965; Maclear and Woodford, 1995; Maclear, 1996). Eq. (3) was programmed in an Excel spreadsheet, which allows for visual comparison and error tracking of the spring flow simulations in a user-friendly manner. The simulations enabled the following:

- Determining the size of the recharge area of the Uitenhage Spring as 60 km^2, northwest of the Uitenhage Spring. This is an order of magnitude different from the 6.3 km^2 of the catchment area around the Uitenhage Spring, as delineated by Kok (1992).
- Comparison of the rainfall moving average with spring flows, without taking into account abstractions. The result is the same as an earlier similar result of Maclear and Woodford (1995), using the CRD method, where no good correlation was found. On many occasions simulated and observed spring flows are out of phase. This confirms that the relationship is masked by the effect of large-scale abstractions from the CRA.
- Taking into account information from Marais (1965) and estimates from Maclear (2002), the simulations were repeated using Eq. (3). A unique match between calculated and observed flows is found at a recharge rate of 10.8% of average annual rainfall. The time-span for groundwater circulation, from rainfall infiltration and recharge to discharge at the Uitenhage Spring, is determined at 23 years.
- In the case of the Uitenhage Spring, an average lag time of 12 years is derived from the spring flow simulation.

The result of the simulation is shown in Figure 9.3. Note that the recharge rate of 10.8% of average annual rainfall is a conservative estimate, since the parameter Q_{leaki} was ignored.

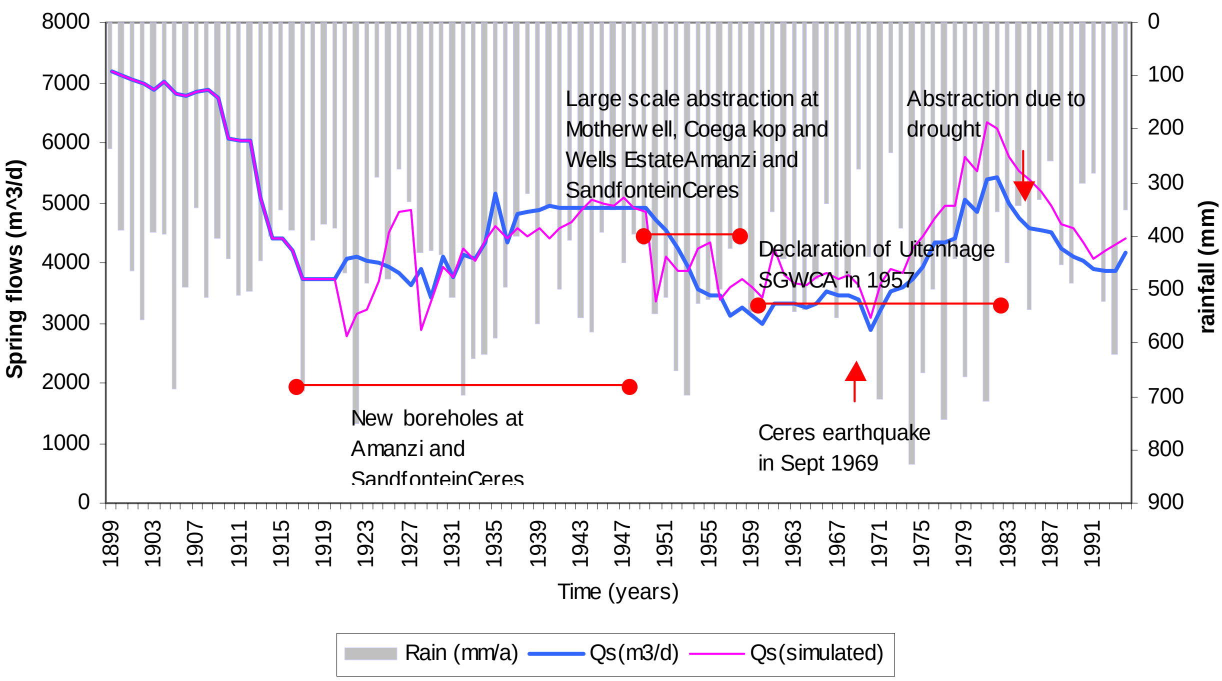

Figure 9.3 Flow variations at Uitenhage Springs: 1899 to 1995 (after Maclear, 1996).

9.2.5 Ratio of cumulative flow to rainfall

The ratio of cumulative flow and cumulative rainfall may be used to verify the above linear relationship. If rainfall events are directly responsible for the Spring flow, the ratio would be the recharge percentage of average annual rainfall. To establish a relationship between rainfall and the Spring flow, the following equation is applied to all 96 annual records:

$$\mathrm{Re}_i = \frac{\sum (Q_{si} + Q_{pi})}{A \sum Rf_i} \quad (i = 1, 2, 3\ldots) \qquad (4)$$

with the parameters as defined previously. Two scenarios are tested, one taking abstraction into consideration and one without abstraction. If abstraction is excluded, there is no linear relationship (Figure 9.4), whereas incorporation of abstractions gives a recharge of 10.6% of average annual rainfall and a correlation coefficient of 0.9993.

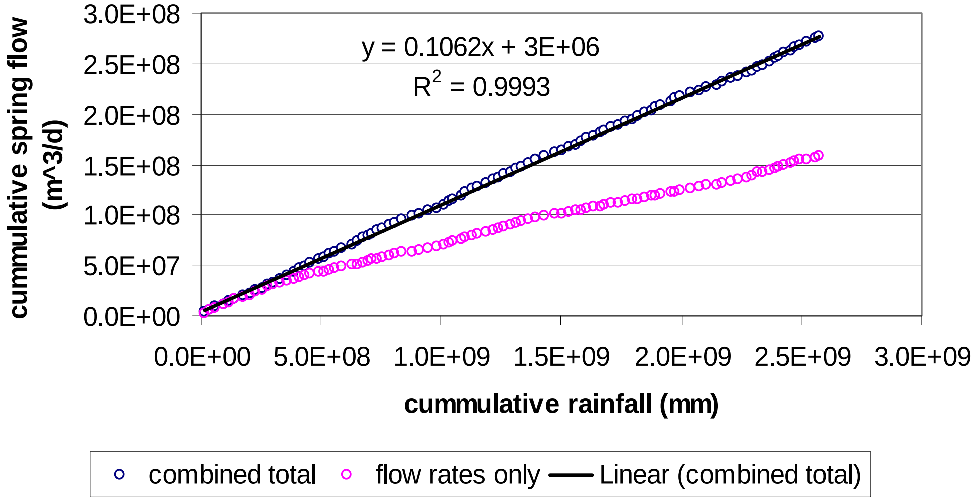

Figure 9.4 Comparison of Uitenhage Spring flow with rainfall: combined total (incl. abstractions) and excl. abstractions.

9.3 Reconstruction of Historic Spring Flow

The flow history of the Uitenhage Spring was reconstructed by using the same parameter values for the recharge area and recharge rate and using historical rainfall data. Rainfall records for Uitenhage for the time period of 1773 to 1898, however, were not available. Simulation of spring flow for that period of time requires the rainfall events as input and therefore, a rainfall generator was designed and programmed in Excel. Figure 9.5 shows the result of the simulation. The generated flow history reflects a similar trend of flow at the Uitenhage Spring, and suggests a higher recharge rate than in the 20th century. This higher recharge rate is expected, since groundwater abstraction by means of boreholes, beginning in the early 1900s, would trigger induced recharge at a rate that is higher than that of the aquifer system in its natural undisturbed state prior to drilling. The induced recharge is determined at 1% for the Coega Ridge Aquifer.

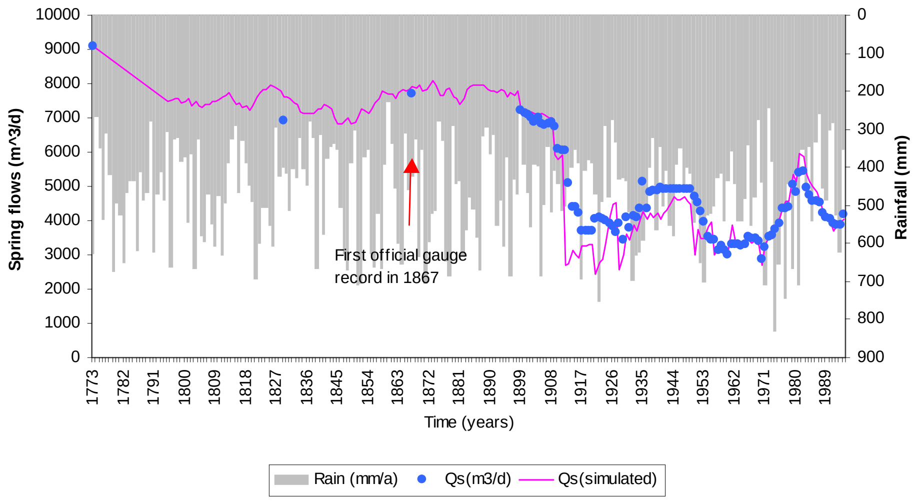

Figure 9.5 Reconstruction of the flow at Uitenhage Spring.

9.4 Discussion

The rainfall 'moving average' approach gives a more accurate recharge value than any of the other methods used. In comparison with the CRD method, no aquifer storativity, which is difficult to determine in fractured rock aquifers, is required in Eqs. (3) and (4). The simulated recharge rate of 10.8% of average annual rainfall was verified by a statistical regression analysis between cumulative rainfall and cumulative spring flow. When abstraction is excluded, the regression gives a recharge rate of 6.8% of average annual rainfall. If abstraction is included in the equation, the regression analysis gives 10.6% of average annual rainfall, as shown in Figure 9.4. The latter is essentially the same as that obtained from the spreadsheet simulation, using the moving average approach.

The simulation shows that the Spring flow-rate is dependent on the preceding 23 years rainfall events, with an average residence time of groundwater circulation before discharging at the Uitenhage Spring of 12 years. If the above is true, then the relatively young water interpreted by Talma et al. (1982), based on one measurement of the Spring sample with a Tritium value of 17.4 TU (March 1963), can be as old as 23 years old. The ^{14}C-derived-ages of more than 1350 years at the same spring (Talma et al., 1982) are doubtful, although these ages may represent groundwater from a deeper circulation depth that was intercepted by a borehole, which was drilled in Eye 2 at the Spring in an attempt to increase the total flow from the eye. Using the age data from Talma et al. (1982), Maclear (1996) calculated the groundwater flow rate along the flow path in the CRA at less than 1 m/yr. Assuming that this flow rate is correct, it would take at least 400 years for rainfall to percolate to groundwater, circulate to a depth of about 185 m and then flow upward under pressure to discharge at the Uitenhage Spring. This implies that the rainfall record from 1899 to the present is not applicable and irrelevant with respect to controlling the flow at the current Uitenhage Spring. This does not seem to be supported by the simulations presented in this study, however.

The Uitenhage Artesian basin is the only extensive aquifer with free-flowing artesian conditions in South Africa and is considered an ideal 'working model' for the fractured TMG aquifer. It contains all the variables found in similar Cape Fold Belt aquifers within a relatively small and well defined area, such as artesian and semi-artesian conditions, cold to hyperthermal water and macro- to micro-scale fracturing. Further investigations of such a 'working model' would result in a better understanding of TMG aquifers in South Africa, to the benefit of groundwater users and advancement of the hydrogeological science.

With the available information, the Auto Regression and Moving Average (ARMA) method was employed to establish the relationship between rainfall, abstraction and spring flow, for the purpose of water resource management within this artesian basin. Application of the ARMA method reveals a time lag of 13 years between rainfall events and spring responses, confirming that the estimated average residence time of 12 years in the study area is a realistic figure. Despite the promising results for recharge estimation it should be noted that the following issues have not been addressed in this paper and require further research:

- Is the Uitenhage Spring fault-controlled and how?
- More data should be collected to allow for 3D numerical simulations.

The relationship between rainfall, spring flow and water level fluctuations is conceptualised in Figure 9.6.

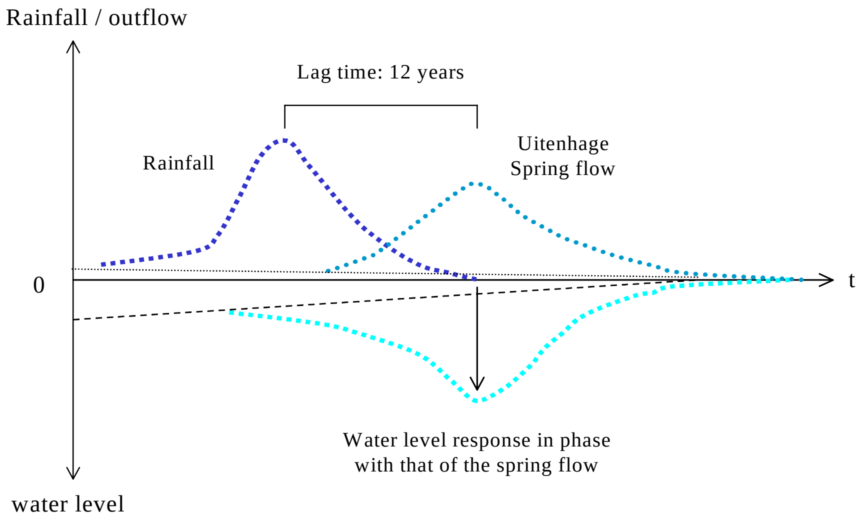

Figure 9.6 Conceptualisation of rainfall, spring flow and water level fluctuations of the Coega Ridge Aquifer.

9.5 Conclusions

In order to quantify the recharge of the Coega Ridge Aquifer, the Uitenhage Spring flow is simulated, using a simple statistical approach. From the simulation, the current spring flow is attributed to the current and preceding 23 years of rainfall events. The time frame for recharge may be divided into two periods. The first period is under natural undisturbed conditions, prior to the drilling of boreholes in 1909, during which time the total recharge based on the Uitenhage Spring flow is determined at 10%. The second period follows on from the introduction of borehole drilling, with the total recharge determined at 11%, based on a combination of the Uitenhage Spring flow and borehole abstraction. The difference between the two recharge values represents induced recharge.

The deep groundwater flow into the confined section of the Coega Ridge Aquifer can be quantified as follows. If 3% of the total recharge, i.e. 11% of the average annual rainfall based on the Spring flow simulation, is the deep flow component as suggested by the isotope results, then the annual deep flow component of groundwater in the CRA is estimated at 88 595 m^3 or 2.81 l/s.

9.6 References

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10. The Role of Interflow in Estimating Recharge in Mountainous Catchments

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ABSTRACT Interflow is generally ignored in recharge studies. Particularly in sloping areas this component of the hydrologic cycle, however, may significantly affect the estimation of recharge. Research is currently being undertaken that will address this issue for a variety of hydrogeological regimes. This paper is the first in a series of papers and concentrates on small mountainous catchments. An approach comprising three different methods based on river flow data is proposed for quantifying the influence of interflow on recharge estimation in this regime. The Vermaaks River Valley in the small mountainous Kammanassie area (21.5 km^2) of South Africa is used as an example to illustrate the approach.

10.1 Introduction

Interflow in mountainous catchments accounts for part of the baseflow in rivers. Figure 10.1 shows under which circumstances interflow may occur (see also Lehman and Ahuja, 1985; De Oliveira-Leite, 1985; Sunada and Tin, 1988; Wetzel et al., 1996; Flügel and Smith, 1998 and Seiler et al., 2000):

a- Occurrence of relatively impermeable soil horizons, hindering downward transport of moisture,
b- Favourable configuration of fractured networks,
c- Partially saturated flow formed via a perched water table, and
d- Preferential flowpaths in soils.

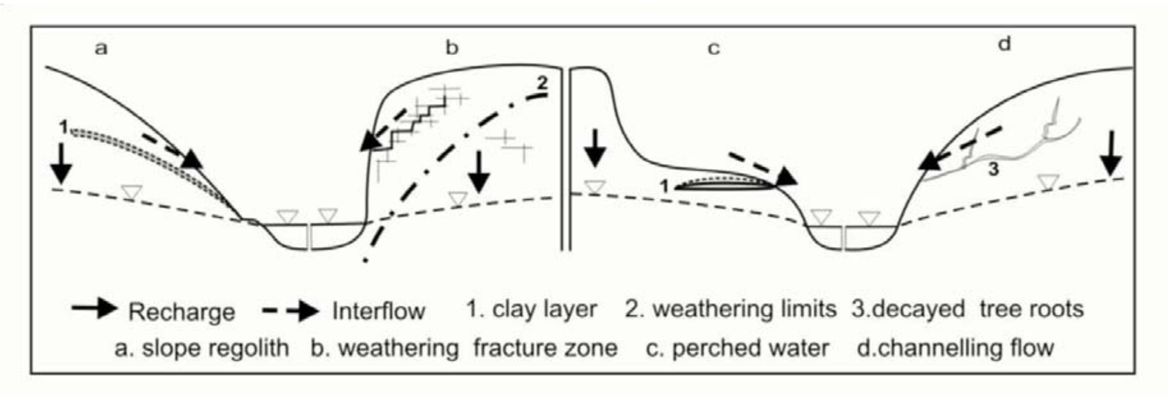


Figure 10.1 Various types of interflow in mountainous areas.

Research into interflow is important from an ecological point of view as this component of the hydrologic cycle often sustains local ecosystems. The zone in which interflow occurs may serve as a reservoir for storing water during the rainy season while at the same time it may allow for percolation to a deeper groundwater reservoir, often through a network of fractures. From a recharge point of view, ignoring interflow would result in overestimation of the recharge rate. There has not yet been, however, much research in Southern Africa on quantifying the role of interflow in recharge estimation.

This paper aims to narrow this knowledge gap for small mountainous catchments. To examine and quantify interflow in this type of catchments, the following approach is proposed:

1) Physiographic investigation at regional scale
2) Analysis of hydrologic data, e.g. flooding events and flow data at local scale
3) Conceptualise the hydrogeology of the study area
4) Water balance study of the unsaturated zone
5) Determination of interflow

Following the determination of interflow, recharge can be quantified using the 'Equal Volume' method. The above approach will be applied to a small mountainous catchment in South Africa and is described in subsequent sections.

10.2 Physiography

The Vermaaks River catchment forms a small part of the Kammanassie mountain ranges, which consists of quartzitic sandstones of the Table Mountain Group (TMG). One of the climatic features of the area is the large diurnal and seasonal fluctuation in temperature. Daily average minimum and maximum temperatures for summer and winter vary between 15 to 42 °C and -3 to 18.5 °C, respectively. Although rain falls throughout the year, the highest rainfall often occurs in autumn or spring. Figure 10.2 shows a west-east transect with various flow systems occurring in the area. The study area (Figure 10.3) is located in the Vermaaks River catchment and comprises a narrow valley with an elevation ranging from 560 to 1340 m above mean sea level (mamsl). The area is surrounded by the Kammanassie mountains with a maximum altitude of 1950 mamsl on the south side. The valley has an average slope from southeast to northwest of 1: 250. The mountain ranges rise steeply above the valley floor. The valley is drained primarily in an east to west direction and turns north towards Olifants River. The above physiographic conditions facilitate development of a weathered zone within the catchments.

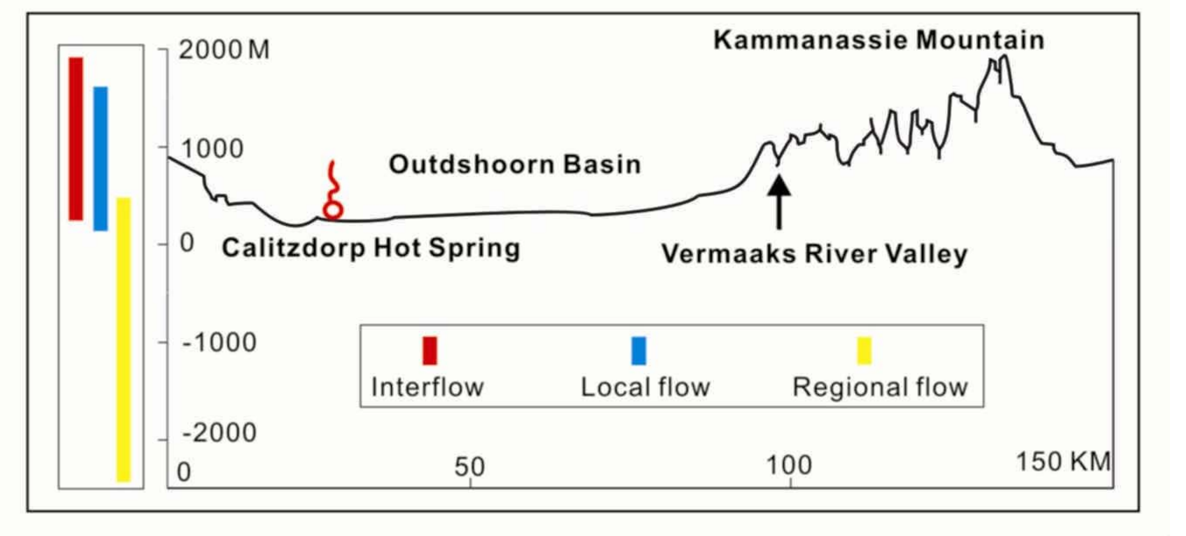


Figure 10.2 Relation between interflow, local and regional flow systems.

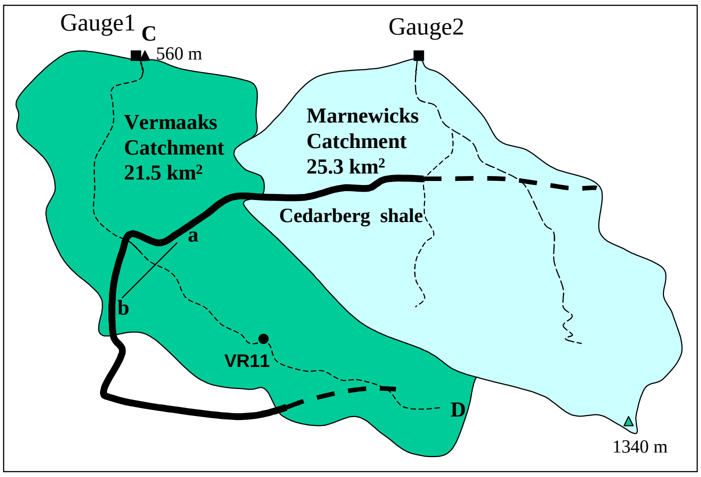


Figure 10.3 Vermaaks and Marnewicks Catchments.

10.3 Hydrologic Information

General hydrologic data for Quaternary Catchments up to Primary Catchments in South Africa is readily available from Midgley et al. (1994). Hydrographs of the Vermaaks River and Marnewicks catchments are shown in Figure 10.4. Geomorphologic conditions of the Vermaaks catchment favour the formation of interflow, which in turn maintains flow in the valley. A perennial small pond occurs at the geological contact zone between the Peninsula and Cedarburg formations. The pond is maintained by either shallow groundwater or locally saturated flow from the shallow weathered zone in the vicinity or both. Field geomorphologic and hydrogeological surveys confirmed that the pond is largely maintained by combined flows, collectively termed interflow, from the shallow weathered zone. Flows recorded at the Gauge station down-stream of the pond may therefore be used for quantification of the interflow component.

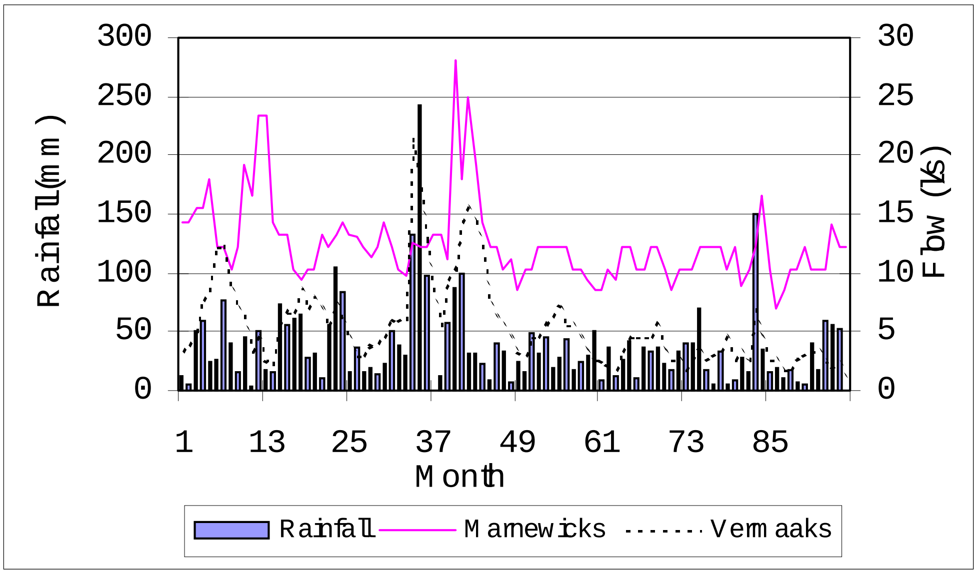


Figure 10.4 Hydrographs of the Vermaaks and Marnewicks Rivers.

10.4 Conceptual Hydrogeologic Framework

10.4.1 Flow system

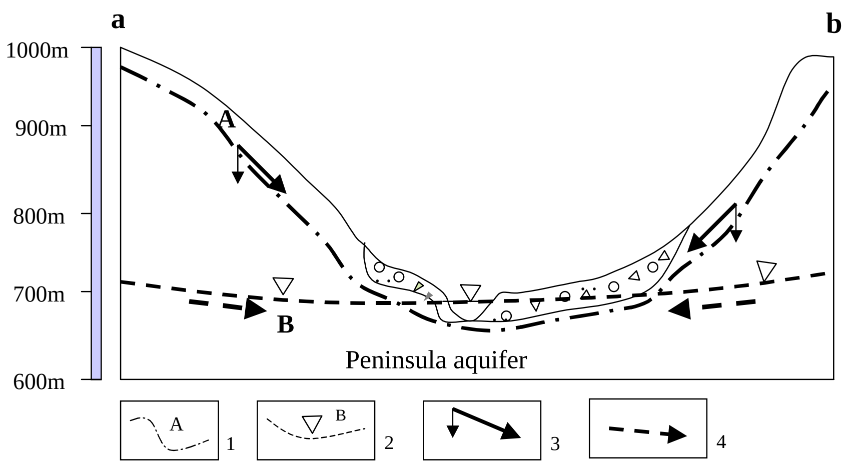
The TMG is subdivided into six formations, of which only the Peninsula Sandstone Formation, Cedarberg Shale Formation and the Nardouw Subgroup are present in the study area. The quartzitic sandstone formations of the TMG are aquifers with relatively high permeability due to a pervasive network of fracture sets, including bedding-parallel fractures, bedding-orthogonal and bedding-oblique jointing at various scales, and fault zones with variable length and displacement characteristics. The geological structures control groundwater flow.

The Peninsula aquifer system consists of both shallow weathered zones and deep fractured bedrock into which a number of boreholes, called the Vermaaks Wellfield, were drilled to supply water to a local community. The wellfield consists of four boreholes which are located along the valley within the upper catchment of the Peninsula Formation. Since 1993, the largest drawdown observed in the wellfield was 80m. The radius of influence gradually extended beyond several kilometers over a period of 10 years. Analysis of satelite imagery suggests that infiltration likely takes place within outcrops of the Peninsula. High concentrations of fracture intersections and frequencies appear to facilitate recharge to the deep fractured Peninsula aquifer system.

Shallow weathered zone
Alluvial and slope deposits, consisting of sand, gravel and other unconsolidated materials, are distributed at the foot of the mountain and along the valley floor. Thickness of these deposits is up to 15 m. Furthermore, a shallow weathered zone of variable thickness has been developed, especially in areas with a high density of fractures. Clusters of springs at different locations suggest that not all shallow weathered zones are hydraulically connected, neither may they be connected to the deep fractured rock aquifer within the Peninsula Formation. An implication would be that groundwater development in the Peninsula aquifer may not have a detrimental impact on ecosystems maintained by local flows in the unsaturated zone.

Fractured bedrock
The Peninsula fractured rock aquifer outcrops in the upper Vermaaks and Marnewicks catchments and subcrops the Cedarberg Formation and Nardouw Subgroup in the lower parts of the catchments. The Peninsula formation mainly comprises sandstone, breccia and cataclastic rocks. The aquifer is likely to receive infiltration from both open fracture networks and the shallow weathered zone in the upper part of the Vermaaks River catchment whereas in the lower parts of the catchment prevailing aquifer conditions are confined.

The relationship between the shallow weathered zone and fractured rock aquifer is illustrated in Figures 10.5 and 10.6. The lower part of the shallow zone is extensively weathered with fractures partly filled in by clayey materials and is less permeable than the upper zone. Groundwater outflow from this zone is in the form of gravitational springs and seepage zones, collectively called interflow. The perennial pond suggests beds of relatively low permeability at the bottom (Figure 10.5). An aquitard may thus occur between the shallow and deeper zones. Water levels in either of these zones may change if the deep aquifer is stressed through pumping the wellfield. Figure 10.6 illustrates the dominant role interflow may play in contributing to baseflow of the Vermaaks River catchment at the down-stream gauge.



Box 1: water level shallow weathered zone
Box 2: water level fractured rock aquifer
Box 3: when water level in deeper zone lower than water level
in shallow zone: percolation from shallow to deeper zone
Box 4: flow direction

Figure 10.5 Hydrogeology along transect a-b (see Figure 10.3 for location).

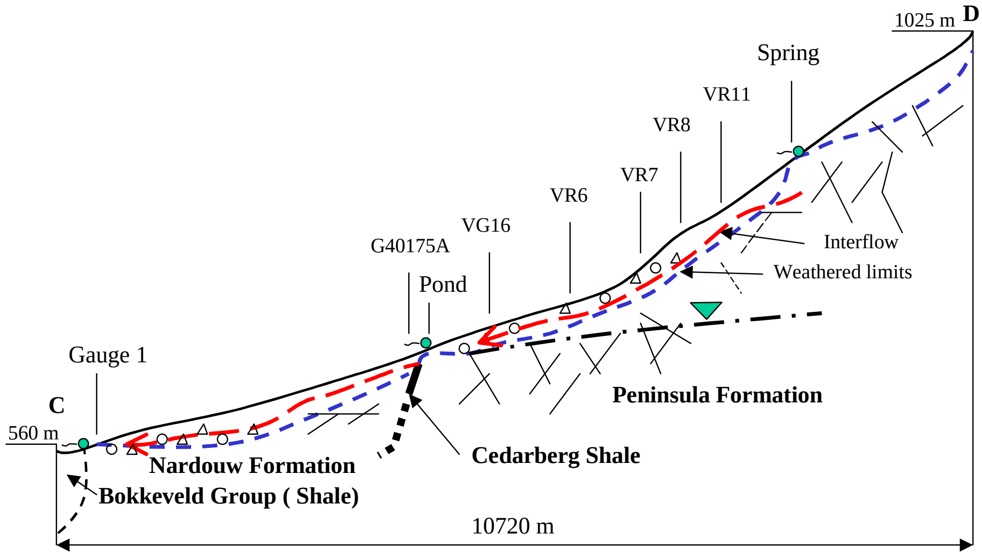


Figure 10.6 Longitudinal Section of Vermaaks River.

10.4.2 Hydrochemistry

Surface runoff in the mountain streams mainly originates from groundwater according to Jolly and Kotze (2002). No conclusion, however, could be drawn on the proportion of interflow versus groundwater contributing to stream flow based on available hydrochemical and isotope data. Main differences between the chemistry of groundwater from boreholes VR (Figure 10.6) and stream flow at Gauge 1 are the increased concentrations of cations and anions, particularly Na and Cl, in the stream flow. The increased concentrations and thus increased EC cannot be explained by evaporation only as isotopic contents are almost the same: stream flow at Gauge 1: ^{18}O = -7.49‰, ^{2}H = -45.05‰; Groundwater: ^{18}O = -7.36‰, ^{2}H = -43.36‰ (Kotze, 2001). Mixing of groundwater and interflow may explain the chemical composition of stream flow at Gauge 1 but a detailed hydrochemical study would be needed along flow paths that includes sampling in the zone of interflow to quantify the contribution of both components to stream flow.

10.5 Water Balance of the Unsaturated Zone

Assuming that the surface area of the (weathered) unsaturated zone is equal to the entire catchment area (Figure 10.3), a simple water balance over a given time interval i can be formulated as follows:

$$Q_{\inf i} = Q_{outi} + Q_{Eti} + Q_{\mathrm{Re}\, i} \qquad (i = 1,2,3 \ldots n) \tag{1}$$

where Q_{infi} is total infiltration to the unsaturated zone, Q_{outi} is total outflow such as spring flow, etc., Q_{Eti} is evapotranspiration of subsurface flow and Q_{Rei} is recharge to the deep aquifer. If the last two terms of the right hand side of Eq. (1) are ignored, Q_{infi} can be determined from flow measurements. If Q_{Rfi} represents rainfall per year and r is a minimum infiltration rate in terms of percentage of MAP, the minimal infiltration rate based on flow data from the Nardouw Gauge Station can be calculated as:

$$\sum_{i=1}^{n} Q_{\inf i} = r \sum_{i=1}^{n} Q_{Rfi} + c \qquad (i = 1,2,3 \ldots n) \tag{2}$$

10.6 Quantification of Interflow

Three methods are used to evaluate interflow mechanisms. These include 1) analysis of flooding events, 2) regression based on cumulative rainfall and flow data and 3) scenario examination of hydrographs.

Analysis of flooding events
Flow data from early 1997 to late 1999 were carefully examined to separate possible causes for increases in river flows. Four episodic surges of flows can be attributed to heavy rainfalls and their preceding rainfalls. The infiltration rate as defined by Eq. (2) is the ratio of river flows to their corresponding rainfall events. The decreasing influence of preceding rainfalls is weighted through a series of factors C^n, with C ranging from 0.75 to 0.95 and the power n (from n=1 to n=n) representing the maximum number of days still having an influence on the river flow. The analysis of the four flow events give consistant estimates of infiltration rates of about 1%.

Regression
Starting point for this analysis is the stream flow recorded at the outlet of the Vermaaks catchment (gauge 1) and rainfall over the catchment as shown in Figure 10.4. The flow for November and December 1996 has been interpolated. Note that there is no runoff in the

study area for most of the year. A statistical analysis was carried out to examine relationships between cumulative total flow (CTF), total rainfall (CTR), flow departure (CFD) and rainfall departure (CRD). Results of the analysis is given in Table 10.1.

Table 10.1 Statistical analysis between flow and rainfall.

Pair of Factors		CTF versus CTR	CFD versus CRD	Flow versus Rainfall	CRD versus CTF	CTR versus CRD	Rainfall versus CRD
r	Vermaaks	0.9963	0.9032	0.5756	0.4033	0.3770	0.1154
	Marnewicks	0.9959	0.7302	0.1454	0.2994	0.3770	0.1154

The analysis revealed that the best correlation is between CTF and CTR (Figure 10.7). There is also good correlation between CTF and rainfall in the preceding month. This suggests that both rainfall from the current and preceding month is important to the estimation of infiltration.

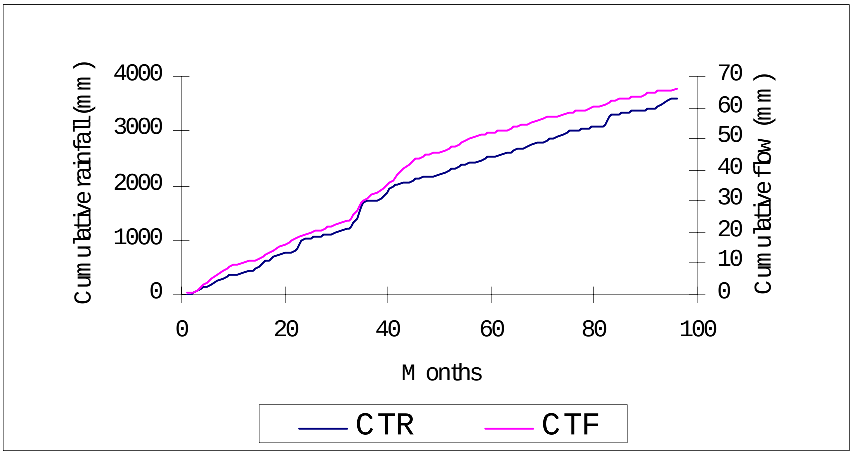


Figure 10.7 Relationships between CTR and CTF.

Additional regression analyses (Table 10.2) also reveals that the accuracy of regression is related to the time interval used. If the time interval of regression is very long, the ssresid/ssreg (the residual sum of squares and regression sum of squares, respectively) in Table 10.2 will change from 0.0128 (0.0302 after flood) to 0.00742. The constant in the regression equations after flood is larger than before the flood. If the time interval for regression is long enough, the influence of floods will diminish. Figures 10.8 and 10.9 show the results of regression analyses for different time intervals and for the two catchments as a whole.

Table 10.2 Results of regression analysis: 1994 to 2001.

Catchment	Eq. of regression	se_{RT}	se_b	r^2	se_y	ss_{reg}	ss_{resid}	Duration (Months)
Vermaaks	F_B=0.0182R_B+1.753	3.6E-04	0.28	0.9874	0.86	1844.55	23.56081	32
	F_A=0.0165R_A+5.7453	3.7E-04	0.39	0.9732	1.41	4059.90	111.70	56
	F_T=0.019R_T+1.6081	1.7E-04	0.38	0.9926	1.79	40748.09	302.27	94
Marnewicks	F_T=0.0361R_T+2.960	3.4E-04	0.76	0.992	3.60	147523.1	1220.13	94
Σ	F =0.0279R +2.0878	1.9E-04	0.55	0.996	1.97	87909.57	637.98	94

The slope of the regression equation or the infiltration rate is given in Table 10.2. The constant in the regression equation is the basis of base flow. The basis ranges from 1.60 mm to 5.75 mm whereas the infiltration rate ranges from 1.65% MAP to 3.61% MAP. The infiltration rate and the constant are 1.82% and 1.75 mm annually before the flooding event (Figure 10.8). After the flooding they become 1.65% and 5.75 mm annually. The results from 1994 to 2001 are 1.9% and 1.61 mm. From the above, the average infiltration rate is 1.79% and the average constant is 3.04 mm. It is obvious that the influence of the flooding is remarkable.

Due to the hydrogeological resemblence of the Marnewicks and Vermaaks Catchments, the infiltration rates are only slightly different as shown in Figure 10.9. The calculated infiltration rates are variable over short time intervals, such as 2 years. The infiltration rate converges to a constant value over a longer time interval.

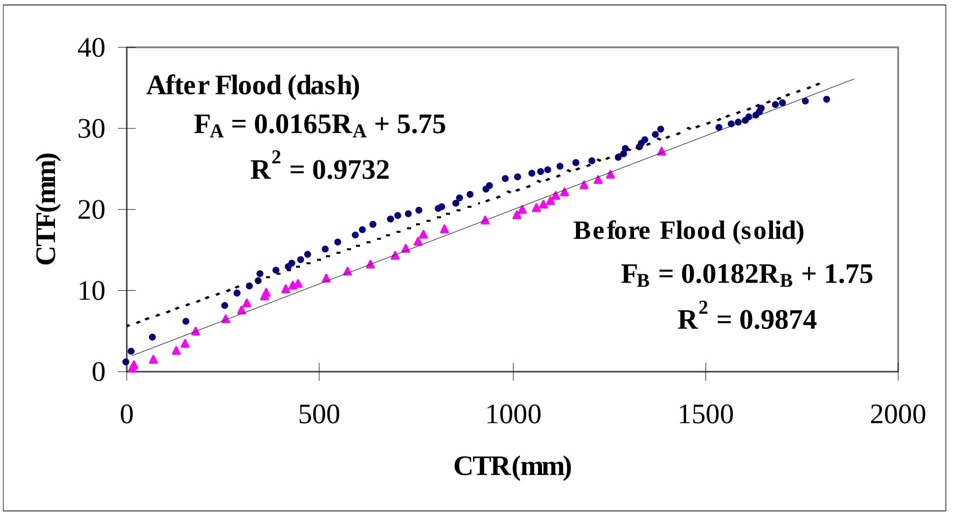


Figure 10.8 Regression analysis of Vermaaks stream flow.

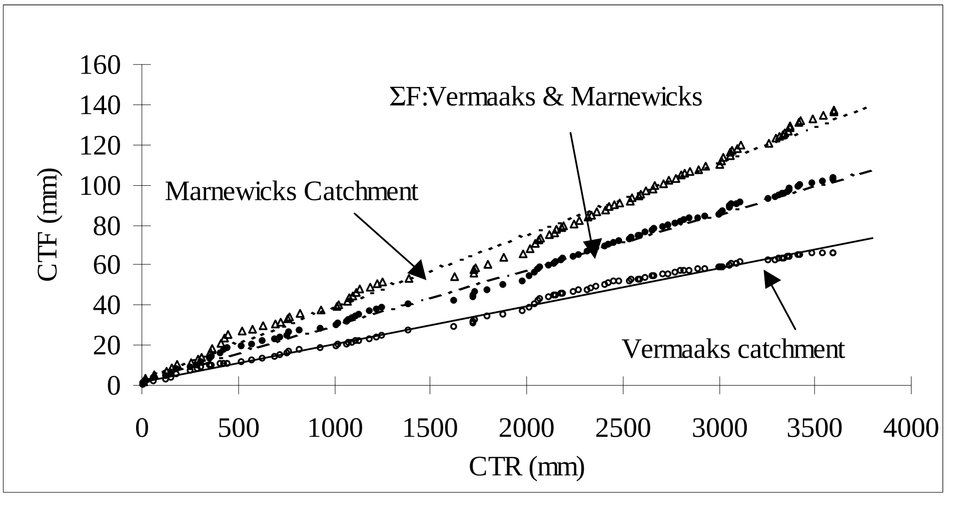


Figure 10.9 Regression analyses of Vermaaks and Marnewicks catchments.

Hydrograph simulation

There is no universally accepted way in which interflow can be accurately quantified. In this paper, however, we have adopted the Xu et al. (2002) approach of hydrograph separation to verify the estimated interflow component. For the quantification of interflow, extreme ranges were used for the decay factor D (0.0-0.5) of previous groundwater contribution and the factor of 'rainfall induced flow increment' I (0.0 to 1.0). The results are shown in Figure 10.10. The hatched zone represents the area of likely interflow and amounts to 33 to 66% of runoff and 0.59 to 1.17% of MAP.

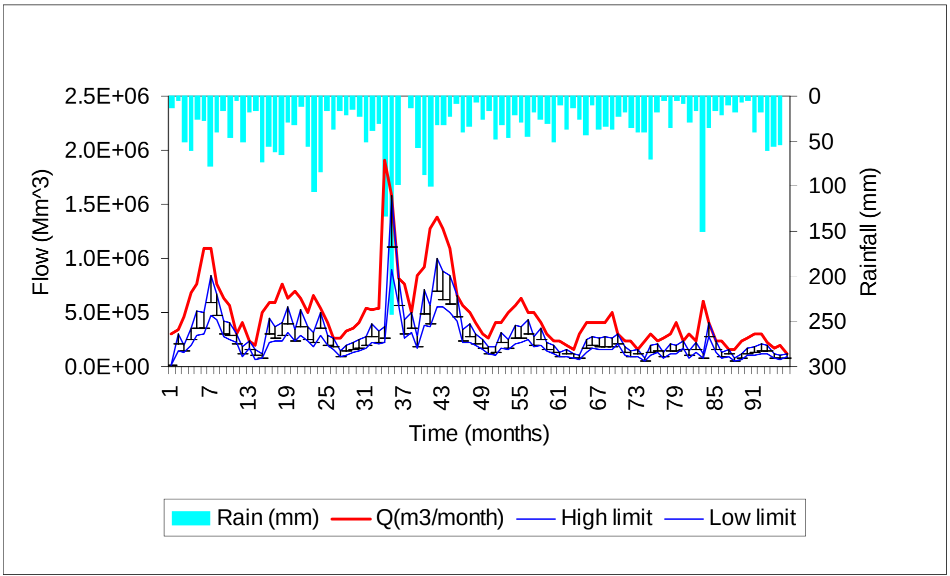


Figure 10.10 Interflow component of hydrograph of Vermaaks Gauge 1.

10.7 Role of interflow in estimating recharge

Many methods have been applied to estimate groundwater recharge in the area such as hydrogeochemical and environmental isotope methods and simulation of water level fluctuations (Bredenkamp et al., 1995; Kotze, 2001; Jolly and Kotze, 2002) but no attempt has so far been made to quantify recharge by taking into account the component of interflow. The influence of interflow on recharge estimation was evaluated in this paper for the Vermaaks River Catchment based on the Equal Volume method. The procedure followed was to first identify the lag time between interflow and rainfall events, and then compare Equal Volumes with rainfall over corresponding periods. As shown in Figure 10.11, the difference between including and excluding interflow in the calculations of recharge is 1%. In other words, the recharge is overestimated by about 10% if one does not take into account the component of interflow.

10.8 Conclusions and Recommendations

An approach is proposed to examine and evaluate the process of interflow in small mountainous catchments and applied to the Vermaaks River Catchment in the Kammanassie Mountains in South Africa. From the Vermaaks case study the following conclusions are drawn:

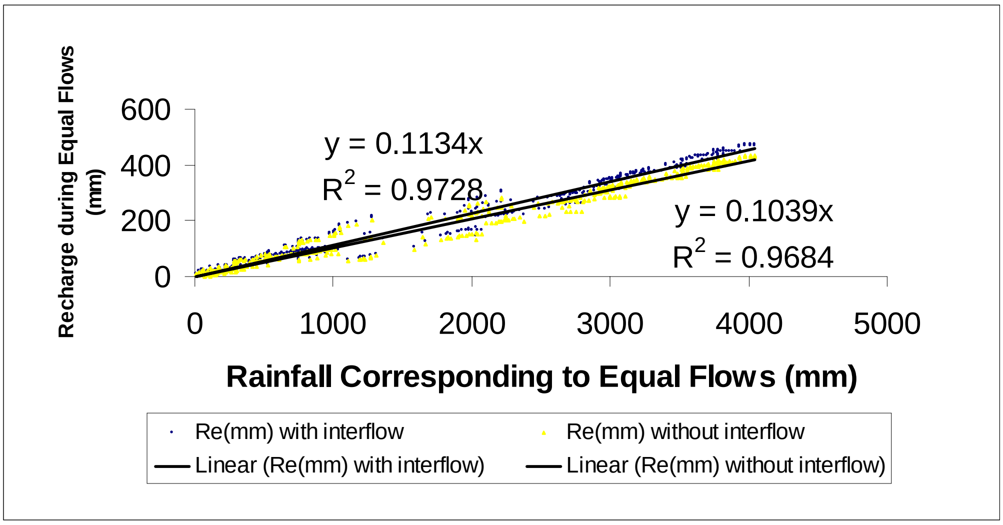


Figure 10.11 Comparison of Recharge Estimates with and without Interflow.

- A shallow weathered zone is responsible for generation of interflow towards Vermaaks River;
- Characteristics of interflow are associated with local hydrogeomorphologic conditions;
- Interflow from the unsaturated zone can be quantified using a simple regression technique;
- The error made when neglecting interflow in estimating groundwater recharge for the Vermaaks River Catchment is relatively small, i.e. ~10%.

Overall, we conclude that the role of interflow in the small Vermaaks River mountainous catchment in estimating groundwater recharge is not important. Whether the same holds true for other small mountainous catchments remains to be seen.

In the case of Vermaaks River Catchment we recommend that the contribution of interflow and groundwater to stream flow be verified by a detailed hydrochemical and isotope study. Induced recharge to the deeper aquifer due to groundwater abstraction should also be examined as well as the regional groundwater flow.

We furthermore recommend that the methods proposed in this paper be applied in similar small mountainous catchments to validate the results of the Vermaaks case study.

10.9 Acknowledgements

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PART V

Integrated Approaches to Recharge Estimation

11. Techniques for Estimating Groundwater Recharge at Different Scales in Southern Africa

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ABSTRACT Various applied techniques for estimating ground water recharge are described for local, hillslope, catchment, regional and national scales. It is clear that the scale of the problem being investigated must determine the method used. Recharge estimates are generally performed assuming a vertical profile. However, it is also evident that the hillslope and lateral processes inherent in generating event and low flow discharges must be defined and quantified so that the proper streamflow generating source, whether ground water aquifer or hillslope vadose zone, is identified. From this knowledge will emanate adequate quantification of ground water recharge mechanisms and rates.

11.1 Introduction

The definition of what constitutes the ground water recharge component in the hydrological cycle is often difficult and could easily depend on the perspective of the scale at which the process is observed. While a basin study of water resources may define the recharge component as the useable resource from a large aquifer, a user of a rural well may define the recharge as that component of interflow and percolation which replenishes the well's source during an irrigation season. Since the demand for localised sources of ground water, as well as regional catchment management are both pressing issues, methods for quantifying the recharge process at different scales are examined so that appropriate methods may be used for the quantification of recharge for specific purposes.

A number of techniques which quantify the recharge process at different scales are presented in this paper. We examine the local or (point scale), the hillslope scale, the catchment scale as well as regional and national scales.

11.2 Local Scale

Local scale recharge has been estimated using a daily hydrological water budget model, ACRU (Schulze, 1995). The processes and water volume accounting shown in Figure 11.1 have been used successfully to model the daily stormflow and baseflow throughout southern Africa at specific locations. The recharge to ground water has mostly been taken as that volume of water which is depleted from the subsoil and is added to the intermediate and ground water store below. While this water budgeting method may be appropriate for estimating the average diffuse discharge from the subsoil in typical profiles, this simplified method of estimating the volume of water recharging the ground water may not be appropriate for specific scenarios where hydraulic potentials and soil and ground water flux variations control the movement of water into an aquifer.

The simplification of the subsoil-ground water interaction used in ACRU, shown in Figure 11.2a, relies on user-defined parameters for apportioning the volumes of water allocated to the different horizons each day. Although the incremental volumes moving from one horizon to another are predominantly dependent on the storage states (and thus indirectly, hydraulic potentials) of the horizons, the transfer parameters either remain constant or take up predetermined values according to the storage state of the horizons. The physics of the interactions between surface-, soil- and ground water, which may play a predominant role in the recharge process is not always adequately simulated in this threshold-based approach. Rapid wetting front development resulting from extreme rainfall events requires a more detailed physically based approach (Gee et al, 1994; Stephens, 1994). Interactions between layers of different hydraulic properties and between capillary forces in the soil and saturated conditions in fractured or dual porosity media also require a revised model since these interactions could easily dominate the recharge process. Estimation of recharge in arid and semi-arid environments could be improved if the dynamics of extreme events were successfully simulated rather than relying on simplified water budgeting.

Attempts at modifying the ACRU soil water budgeting with a Green-Ampt wetting front model have proved successful in simulating the flux through the soil profile in response to rainfall events of different intensity (Lorentz et al., 1995; Howe and Lorentz, 1995). Innovations in applying Richard's equation make this an attractive method for simulating soil water dynamics (Short et al 1995). Considerable improvement in both infiltration and soil water redistribution dynamics has also been achieved in ACRU, with the application of a solution of the Richard's equation in the soil horizons. However, these approaches are always dependant on appropriate boundary conditions at the vadose zone- ground water or fractured media interface. Most early attempts have used a constant head condition at the zero matric pressure elevation, thus requiring a user input of the ground water depth. However the response to the recharge and consequent perturbations in the soil water potentials have been largely ignored.

An attempt to retain the physical basis of subsoil-ground water interaction simulation has been developed in the ACRU model as depicted in Figure 11.2b. The fluxes from one zone to another are modelled using the hydraulic characteristics of the porous media. The water dynamics is modelled in the soil horizons of the vadose zone using Richard's equation. At the interface of the soil horizons and a fractured bedrock, it is recognized that soil matric pressure heads must approach zero before flow is induced into the fractures. Allowance is provided for flux into the fractured bedrock prior to saturation of the soil interface if the bedrock properties allow for capillary flow. Once soil water matric pressures are zero at this interface, flow into the bedrock commences at a rate dependent on the hydraulic properties of the bedrock. Soil water contents may then increase in the soil as shown in Figure 11.2b and positive pressures may develop at the interface, forming a perched water table. Perched water tables may also develop in the soil horizons if clay layers are present.

Distinctly different recharge processes have been simulated in the Romwe experimental catchments near Triangle in Zimbabwe, some 150 km north of Beit Bridge. This modelling was undertaken in collaboration with the Centre for Ecology and Hydrology, Wallingford, who completed extensive monitoring of the recharge processes in these agricultural catchments (Butterworth et al., 1999). Ground water recharge was shown to be strongly controlled by the characteristics of the weathering profile, lending credence to attempts to model the recharge dynamics in semi-arid areas by considering profile hydraulic characteristics and boundary conditions. The Romwe catchments receive a mean annual precipitation of 581 mm with a standard deviation of this mean of 263 mm which is typical of the large interannual variation in semi-arid catchments. Subsistence agriculture is practised on two predominant soil types in the catchment. The first is a red clay with granular microstructure derived from the mafic pyroxene gneiss north of the stream (Figure 11.3). The second is a grey coloured sandy soil of coarse texture underlaid by a thick clay layer. This soil type is associated with the leucocratic gneiss south of the stream. Typical profiles and associated hydraulic conductivity characteristics are shown for the soils in Figures 11.5 and 11.6. The red clay soil has a saturated conductivity an order of magnitude less than the grey sandy soil but the clay which underlies this sandy soil has a saturated hydraulic conductivity an order of magnitude less than that of the red clay soil.

During an event on February 17th 1994, 141 mm of rain was recorded. The resulting recharge to the bedrock estimated from the data is dramatically different for the two horizons as indicated in Table 11.1. The model simulates the runoff and recharge adequately in the two profiles for the event. The recharge for the event was estimated from ground water elevation changes (Figures 11.5 and 11.6) and from hydraulic potentials and conductivities deduced from observed water content changes in the soils at a depth of 2.5 m. The results indicate that the clay layer below the grey sand effectively inhibits recharge while significant recharge occurs through the red clay soil during the week after the event. Agricultural practices may also contribute to the differences in recharge since far more tillage is practised in the heavier red clay surface soils and water retention bunds are often used. The maximum increase in ground water elevation on the north and south side of the stream is shown for the wet season, 1993/94 in Figure 11.4. It is clear that the predominant increases in ground water elevation are associated with the red clay soils overlying the leucocratic gneiss bedrock. It is anticipated that a more detailed simulation of the soil water fluxes will provide improved comparisons with other methods deemed in the past to be more accurate than simplified water budgeting (Allison et al., 1994; Barnes et al., 1994; Robson et al., 1992).

Table 11.1 Recharge estimates for the event of February 17th 1994.

Soil	Rain (mm)	Runoff (mm)		Recharge (mm)		
		Modelled	Measured	Modelled	Estimated from G/W elevation	Estimated from changes in water content at 2.5m
Red clay	141.0	6.2	7.0	20.4	16.0	18.6
Grey sand	141.0	52.8	46.5	0.3	*	0

* Observed ground water elevation increased due to flow from sand layer into unlined piezometer.

11.3 Hillslope Scale

Recharge to ground water occurs as one of the processes in hillslope sections and it is often important to define the travel times and pathways of water sources on a hillslope in order to distinguish between the components of event and low flow water. Lateral fluxes of water can be sufficiently slow and residence times sufficiently long on a hillslope that these sources are able to support low flows during dry periods. It is possible that accumulations from these sources could be mistakenly taken as deep ground water contributions to low flows. This would have serious consequences if the land use on the hillslopes were to change. Although lateral fluxes are assumed in the water budgeting models discussed, a more realistic account of the soil and ground water can be obtained by considering the processes on a hillslope scale (Sami and Hughes, 1993). These processes are being observed in an experimental catchment in the Mondi, North East Cape Forests area. The MAP of the area in which the catchment lies is some 850 mm, which is considered marginal for successful commercial forestry. Therefore, knowledge of the hillslope soil water processes is important to allow management to avoid planting trees in areas which may be continuously waterlogged or in areas which are prone to rapid depletion of infiltrating water. The dynamics of flow on the hillslopes in the experimental catchment have been observed under pristine conditions to determine the hillslope water budget prior to the introduction of trees. Monitoring has continued after the trees were established in 2002, so that the changes in the water distribution on the slopes can be quantified and modelled to predict long-term influences of afforestation on receiving water bodies. The catchment lies in a sequence typical of the Molteno formation with two outcrops of rock at different elevations on the slope.

The catchment topography is shown in Figure 11.7, along with the network of neutron probe access tubes, ground water monitoring piezometers, automatic tensiometers, weirs and profile pits. Hydraulic characteristics of the materials have been measured (Figure 11.8) in order to estimate the fluxes from the various sources. The sources and pathways of water on the hillslopes of the catchment have been identified and quantified from the responses of the observed soils water status (Figure 11.9). Typical response zones have been described as:

- Zone 1: Upper slopes of eastern half of the catchment, where delivery of water in a disconnected near surface macro-pore zone delivers water to bedrock outcrop at the toe of the slope. Soil matric pressure is not continuous between responses in near-surface layers and deeper layers near bedrock. Surface water runoff generation for no more than 20 to 30m upslope contributes to flow at the toe. Slow deep ground water from soil/bedrock interface recharges to toe and to lower slopes and bedrock. All water from this zone is delivered to Zone 2.
- Zone 2: Recharge from upslope zone and infiltrating water raise ground water levels at seepage lines and wetland areas. Some flow in near-surface macro-pores. However, it is normally associated with the resident ground water rising into the macro-pore layers, particularly adjacent to the stream and seepage lines leading to the stream. Some ground water ridging near the stream yields increased hydraulic gradients for short periods during moderate to intense events.
- Zone 3: Near stream surface and near-surface water runoff, dominated by ground water intersecting rapid delivery macro-pore layers.
- Zone 4: Some flow in near-surface macro-pore layers, but mostly due to intersection of soil/bedrock perched water. There is generally soil matric pressure continuity between upper and lower layers. Near the stream, water is delivered through ground water rising into macro-pore layers.
- Zone 5: No perched water tables are evident, even during intense events. Little macro-pore discharge in near-surface layers, even during intense events. Significant wetting to deep horizons with slow delivery of unsaturated water to lower slopes (Lorentz, 2001).

Simulation of the catchment runoff using response functions to describe the volumes and arrival times of the various sources reveals a dominant contribution from near surface macro-pore horizons (Figure 11.10). More complex model development may have to be made where observations of large macro-pore or pipe flow is evident (Nieber and Warner, 1991). Nevertheless, water accumulated on the interface of soil profile and bedrock as well as gravity driven, soil water accumulation contributes to discharge in the stream throughout the winter. These accumulations from many of the upland subcatchments appear sufficient to maintain the low flows observed in the receiving rivers, without having recharged a ground water aquifer

(Lorentz, 2003). The relative contribution to both low flow and event water from hillslope and ground water sources is the subject of a present study in which natural isotopes are used in conjunction with the hydrometric measurements, to identify the contributing sources, pathways and travel times.

11.4 Catchment Scale

The water budget approach depicted in Figure 11.1 has been successfully applied to hydrological studies of catchments up to 50 km^2 although larger basins have been simulated by subdividing them into smaller interlinked and cascading subcatchments (Kienzle et al., 1997). GIS is used to prepare area averaged parameters for input to the model. Since the recharge component is a small proportion of the water budget, estimates of recharge are liable to be compounded by errors in the overall water budget. This approach is, therefore, not always ideal for recharge estimates but since the water budget method is based on simulating physical processes, it is useful for comparative exercises in assessing recharge in areas differing in soil profiles, land uses and rainfall patterns.

The direct estimation of recharge on a catchment scale could well be improved by considering the integration of hillslope processes contributing incremental discharges to the conveying stream. Such approaches are becoming increasingly popular in catchment hydrology and can only contribute to a better understanding and simulation of recharge.

11.5 Regional Scale

Several methods of estimating ground water recharge and discharge for a catchment from streamflow records have been proposed in the literature. Many of these are graphical, subjective and do not necessarily reflect the true nature of either the recharge or discharge. However, the ACRU model is a deterministic model which attempts to physically model small catchment processes in a conceptual-physical manner.

Research into the development of a baseflow decision support system (BFDSS) to strengthen ACRU's standard water budget's ground water and streamflow generation routines have been undertaken (Hughes, 1997). By using data from 200 gauging weirs located throughout South Africa, all of which were deemed to be recording natural flow, this research established a link between baseflow recession characteristics and a basin's physical and geological properties.

Determining multiple segment master recession curves aided in the evaluation of the true shape of the master recession curves (MRC) thereby accounting for McMahon's (1995) suspicion that exponential recession theory would be inappropriate for baseflow recession modelling in South Africa. This was achieved using the procedure summarised in Figure 11.11 where all recessions are extracted (stage A) and eliminated (stage B) if they contain missing or suspect data or if significant rainfall occurs during the recession or if they are of less than 10 days in duration. The remaining data are converted from daily specific to discharge specific data (stage C) to facilitate the calculation of average recession constants for each discharge interval (stage D). These are used to construct the MRCs (stage E).

11.6 National Scale

An extensive database has been developed for hydrological simulations in South Africa by the School of Bioresources Engineering and Environmental Hydrology, University of Natal. Daily rainfall data have been assembled for each of the 1946 quaternary subcatchments in South Africa for the period 1951-1993. GIS techniques have been used to derive representative soils and vegetation parameters for each quaternary subcatchment and files of average monthly potential evaporation and maximum and minimum temperatures have been compiled. The database comprises a powerful resource for estimating components of the hydrological cycle throughout the area, comparison of irrigation supply and demand, crop production performance for different areas as well as for estimating the significance of certain extreme events or trends such as climate change.

Hydrological simulations have been undertaken to estimate the distribution of average annual recharge, as shown in Figure 11.12. The recharge simulated for a wetter than average year (La Nina, 1988/89) is compared with the average annual recharge by displaying the ratios of the recharge for these periods. While the 1988/1989 season produced wetter than average rainfall predominantly in a band from the south-east to the north-west of the country, only a few quaternary subcatchments in this semi-arid zone yield recharge estimates larger than average. Conversely, during a drier than average season, 1982/83, every quaternary subcatchment in the summer rainfall region of the country received drier than average rainfall and yet a significant number of quaternary subcatchments in this region were simulated to have above average recharge for that El Nino year. These simulations indicate that there may be merit in examining more closely the influence of extreme and often localised events and rainfall patterns in the resultant recharge.

11.7 Conclusions

Various applied techniques for estimating ground water recharge have been described. It is clear that the scale of the problem being investigated must determine the method used. It is also evident that the processes inherent in generating event and low flow discharges must be defined and quantified so that the proper streamflow generating source, whether ground water aquifer or hillslope vadose zone, is identified. From this knowledge will emanate adequate quantification of ground water recharge mechanisms and rates.

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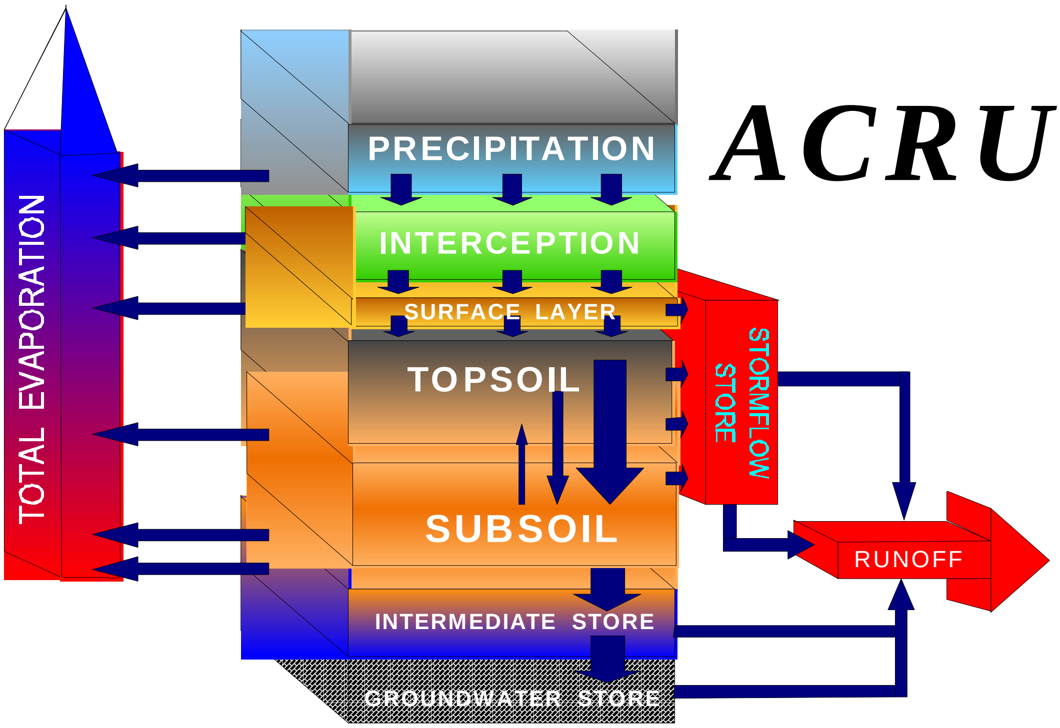

Figure 11.1 The ACRU water balance model structure.

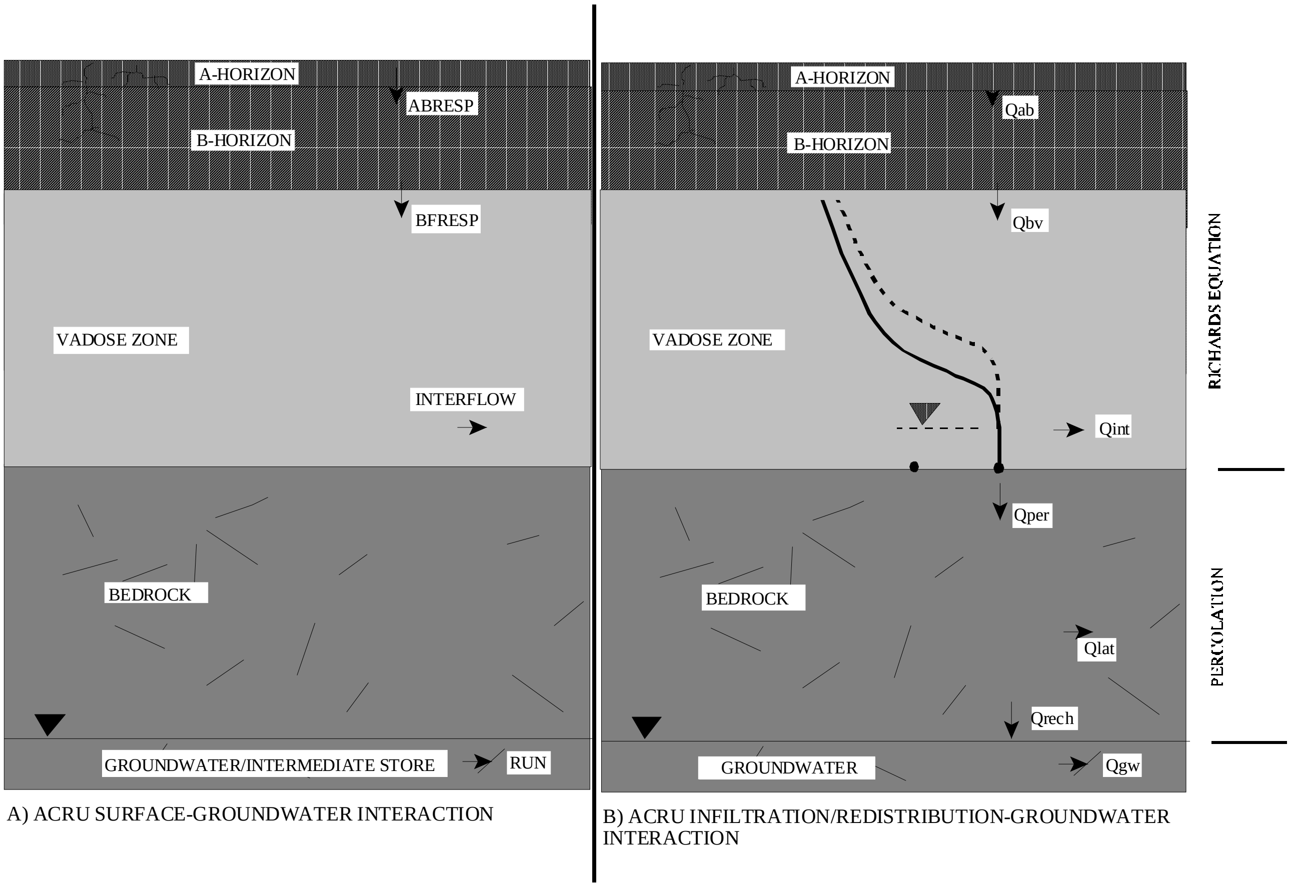

Figure 11.2 Schematic of the surface-groundwater interaction.

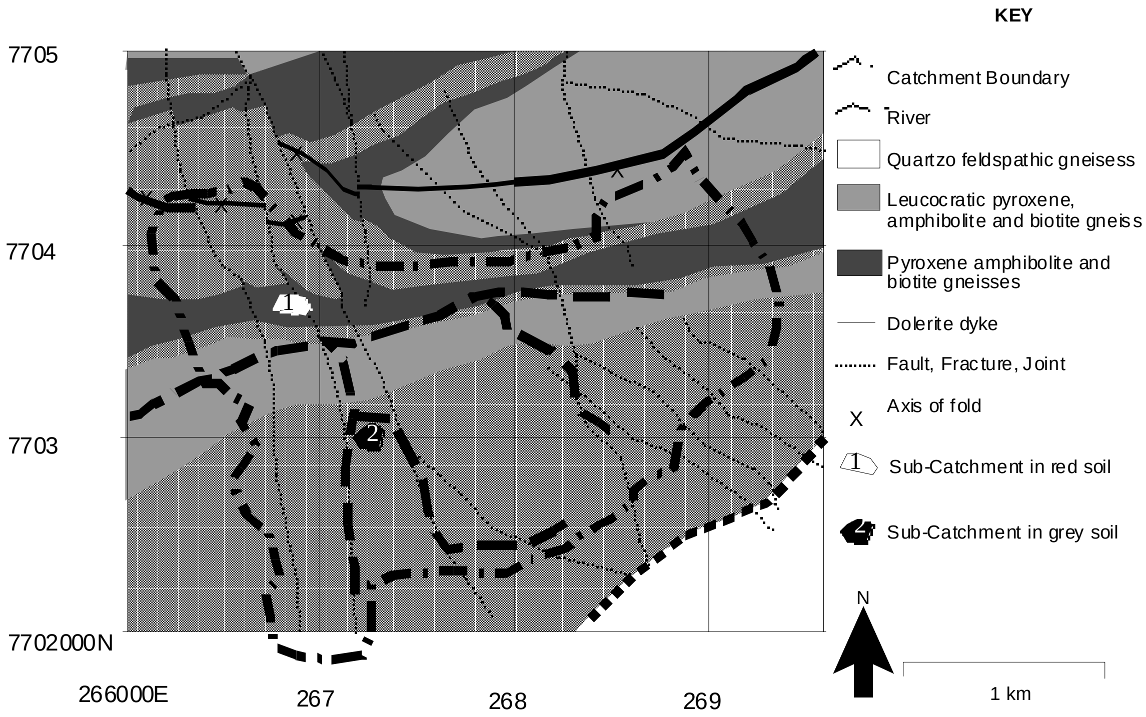

Figure 11.3 Geology of the Romwe catchment, Zimbabwe (after Butterworth et al., 1995).

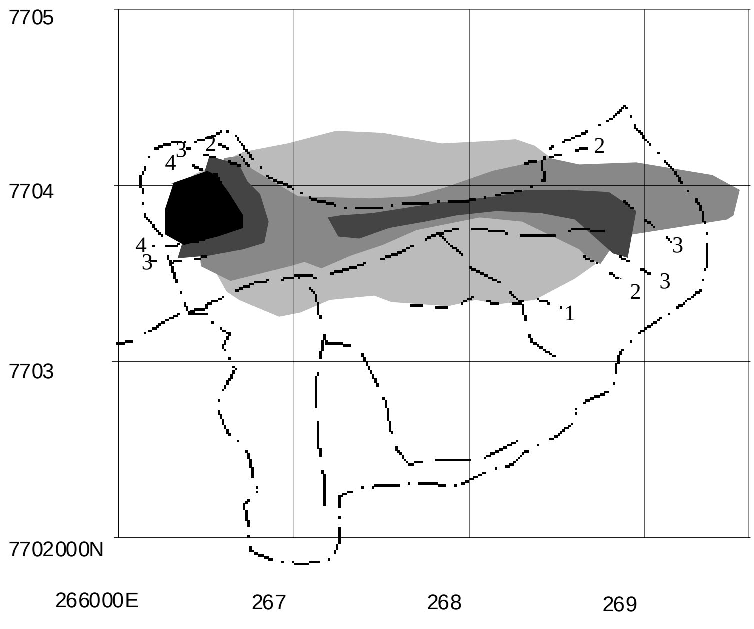

Figure 11.4 Groundwater level rise (in m), 14 November 1993 to 24 February 1994 in the Romwe catchment, Zimbabwe (after Butterworth et al., 1995).

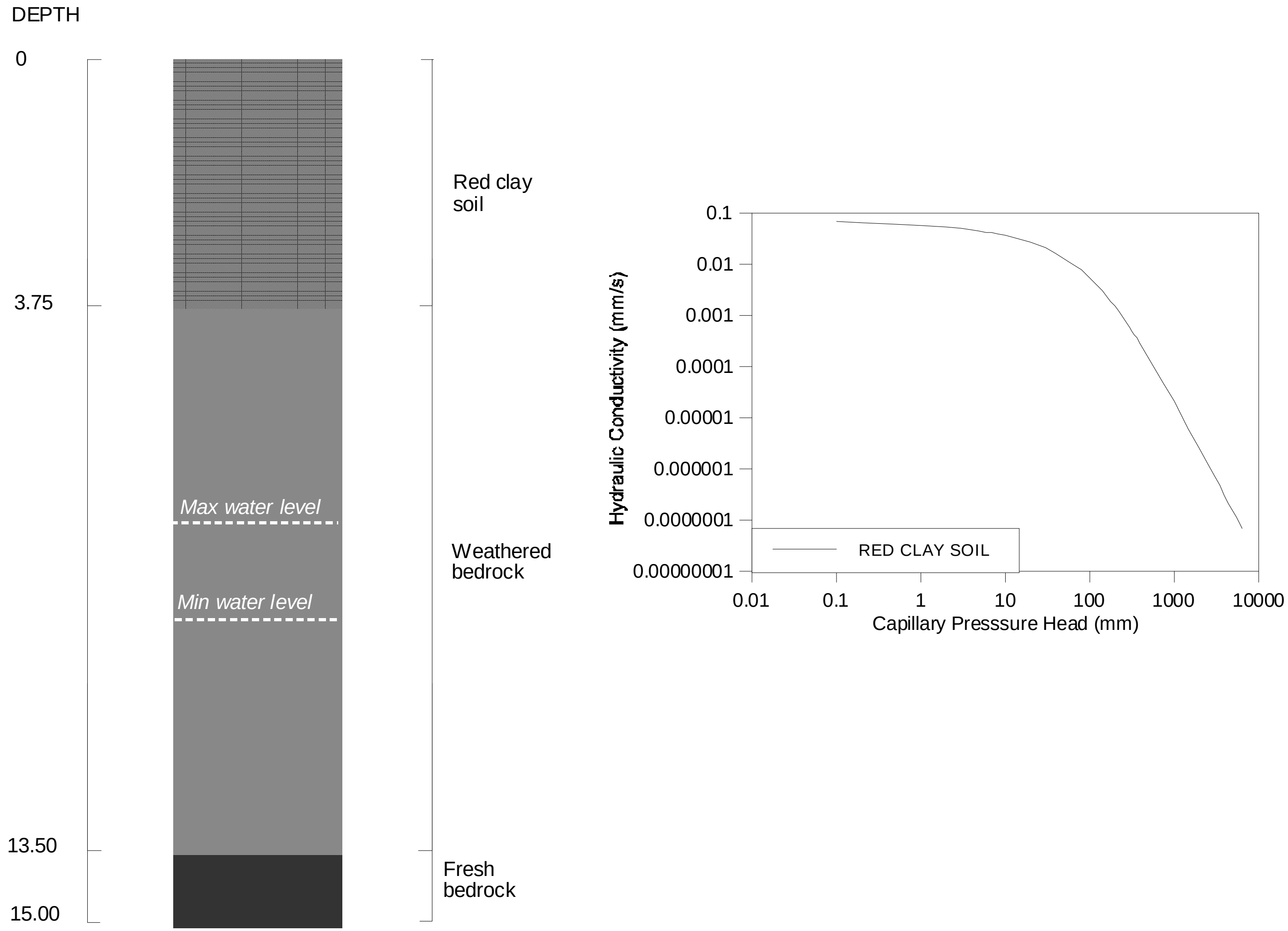

Figure 11.5 Profile of the red, clay soil horizon, Romwe catchment, Zimbabwe (after Butterworth et al., 1995).

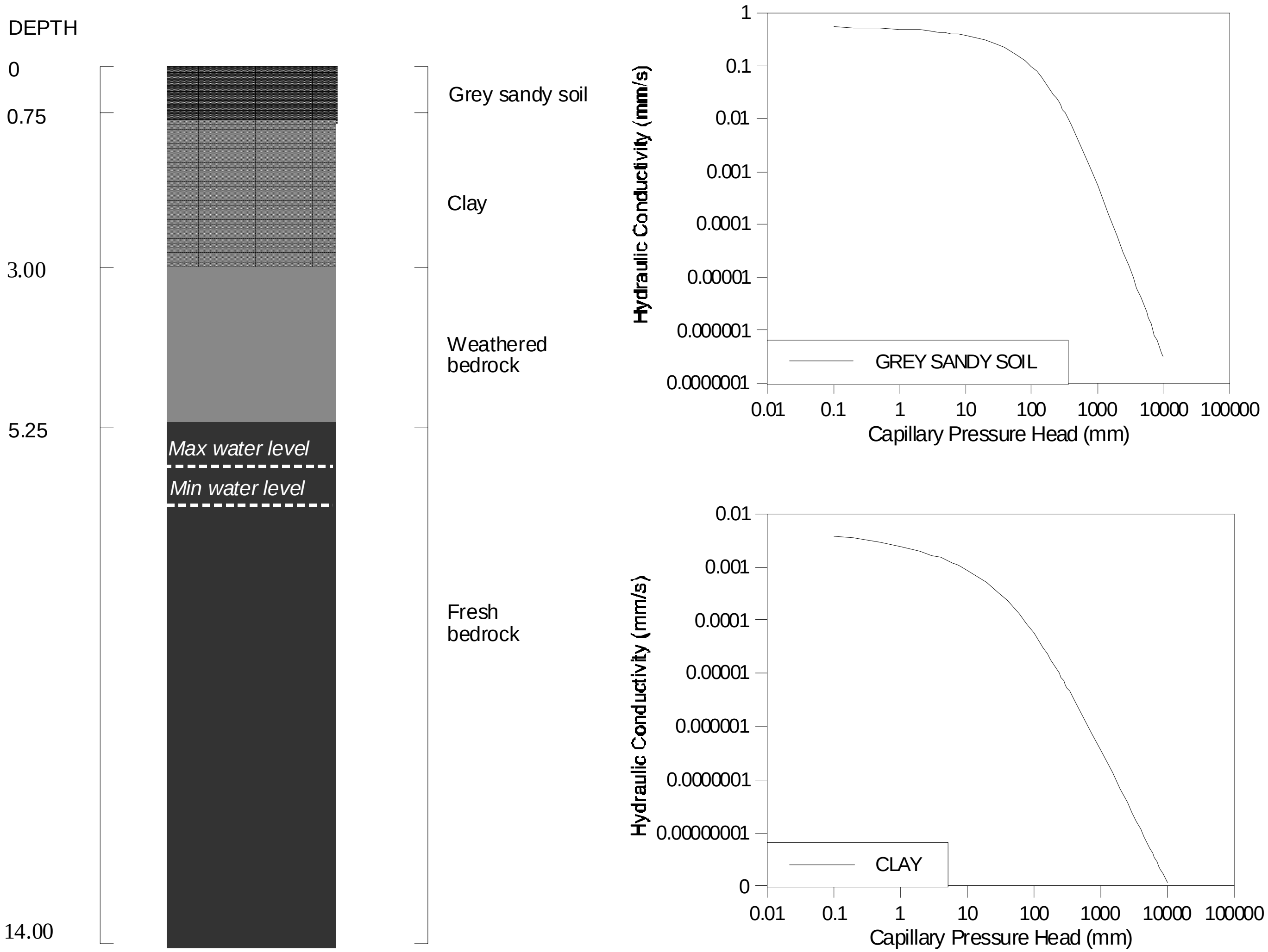

Figure 11.6 Profile of the grey, sandy soil horizon, Romwe catchment, Zimbabwe (after Butterworth et al., 1995).

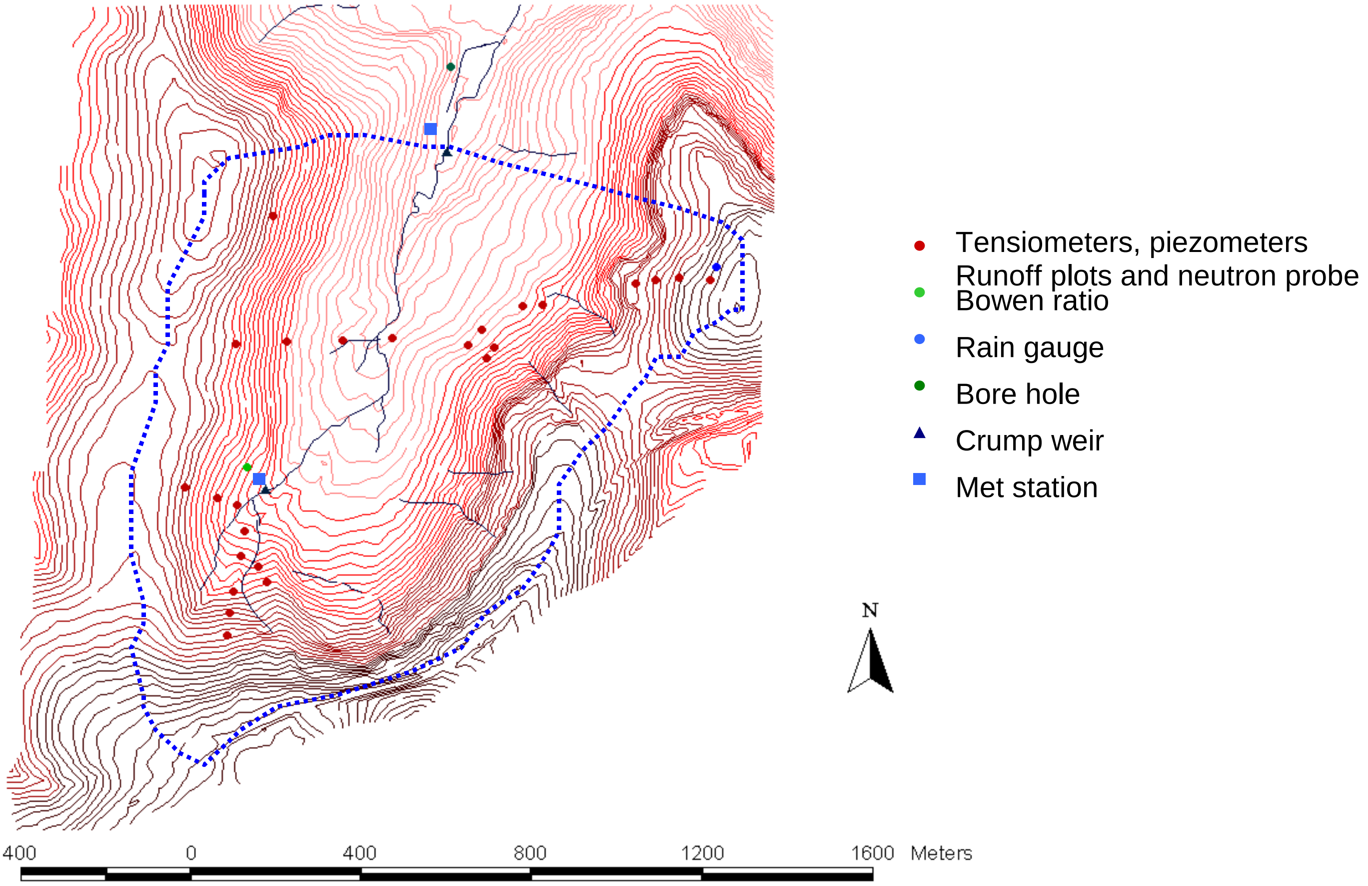

Figure 11.7 The Weatherly experimental catchment, North East Cape Forest.

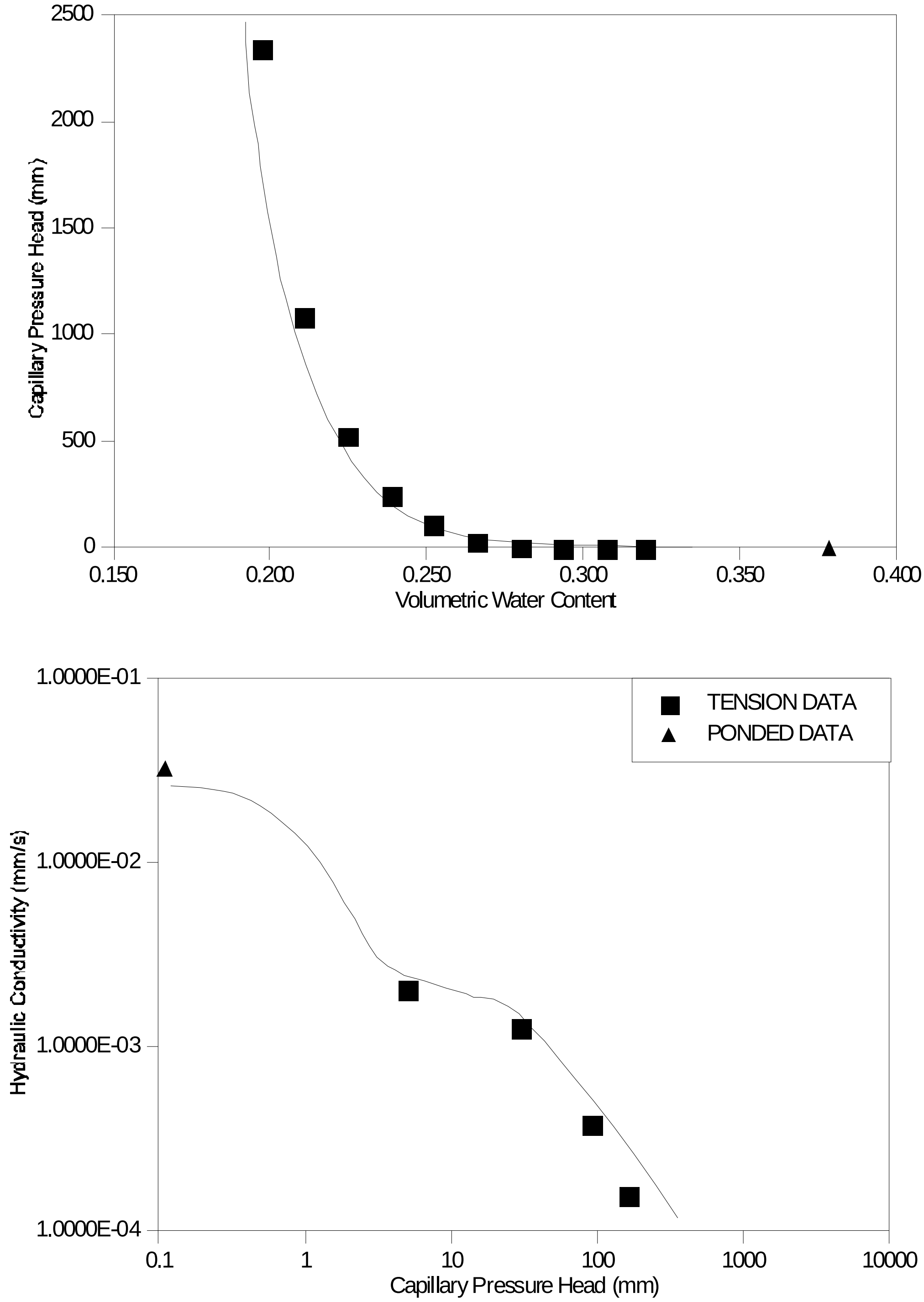

Figure 11.8 **Water retention characteristic (above) and hydraulic conductivity characteristic (below) of one of the profiles, Weatherley research catchment.**

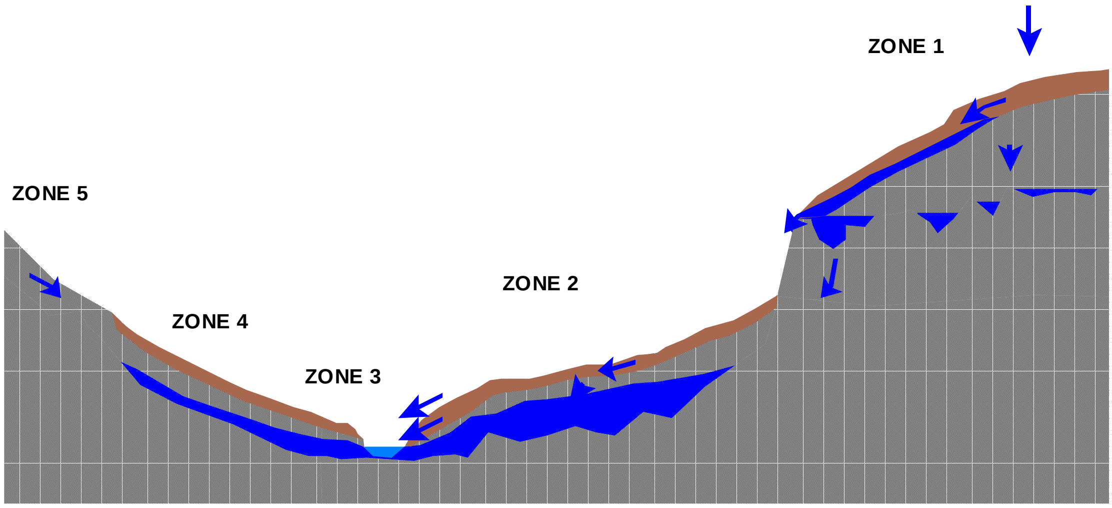

Figure 11.9 Section of the dominant flow processes and flow pathway zones (facing upstream from the lower weir), Weatherley research catchment, (Lorentz, 2001).

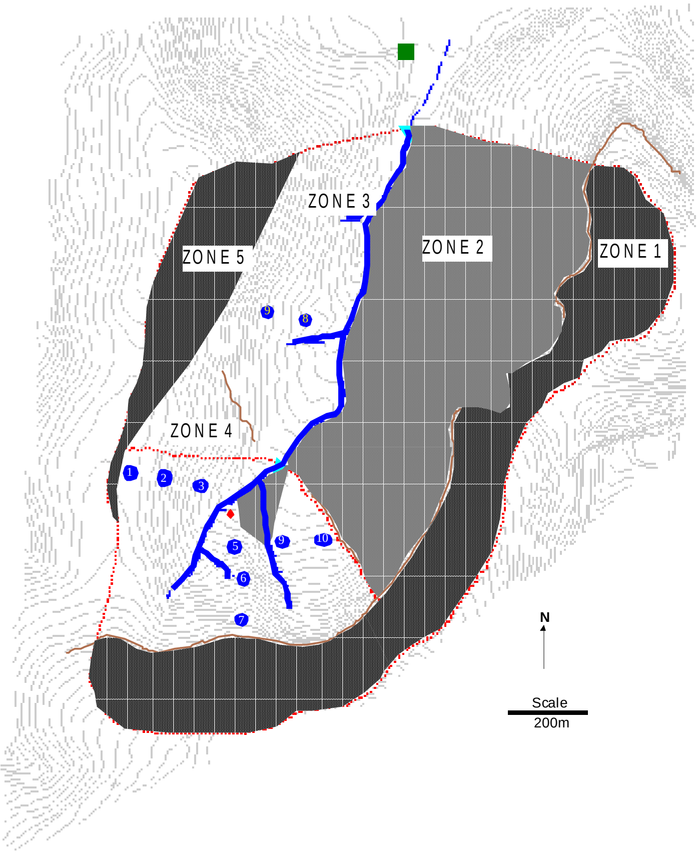

Figure 11.10 Contributing zones based on simulated catchment runoff, Weatherley research catchment (Lorentz, 2001).

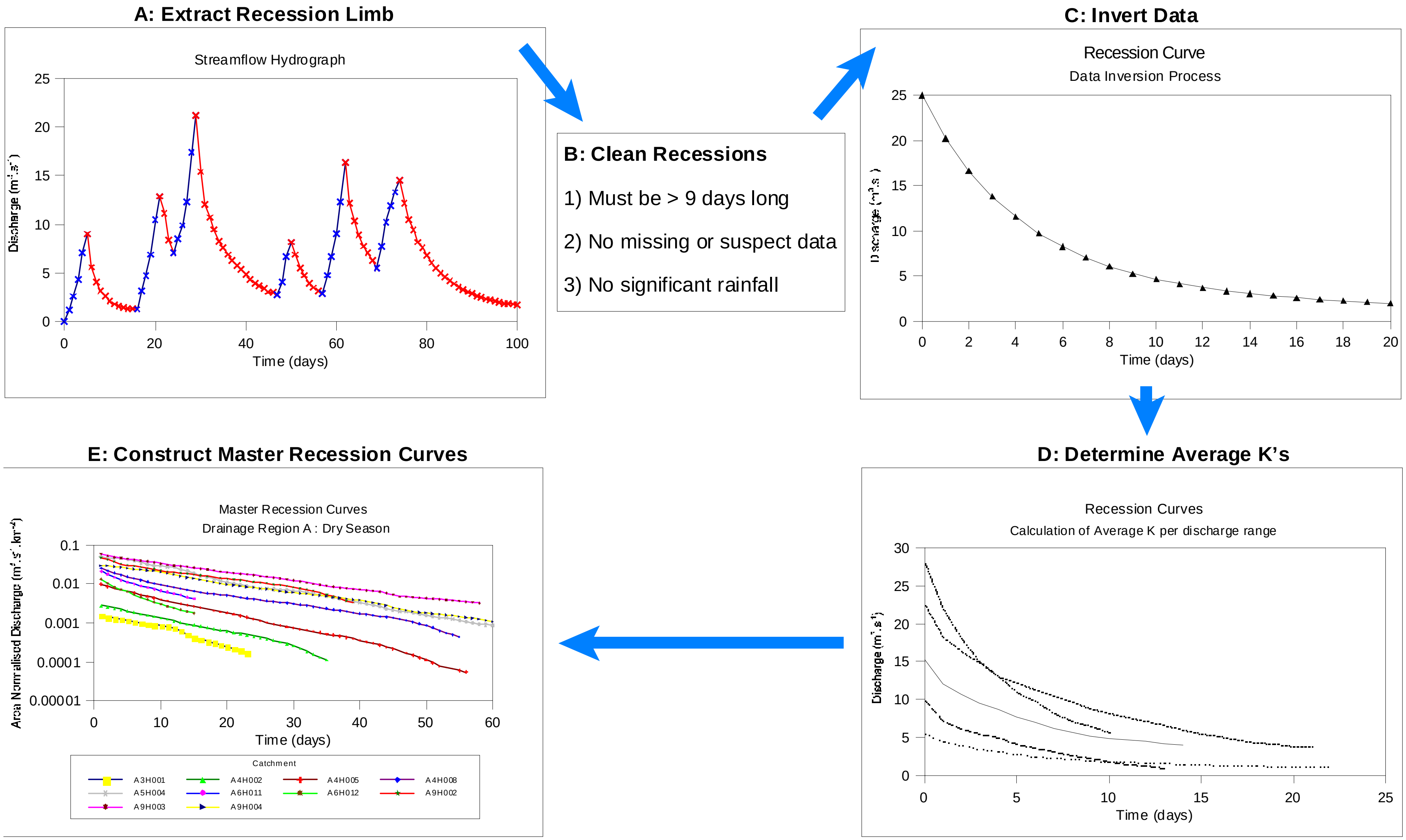

Figure 11.11 Hydrograph recession analysis (Hughes, 1997).

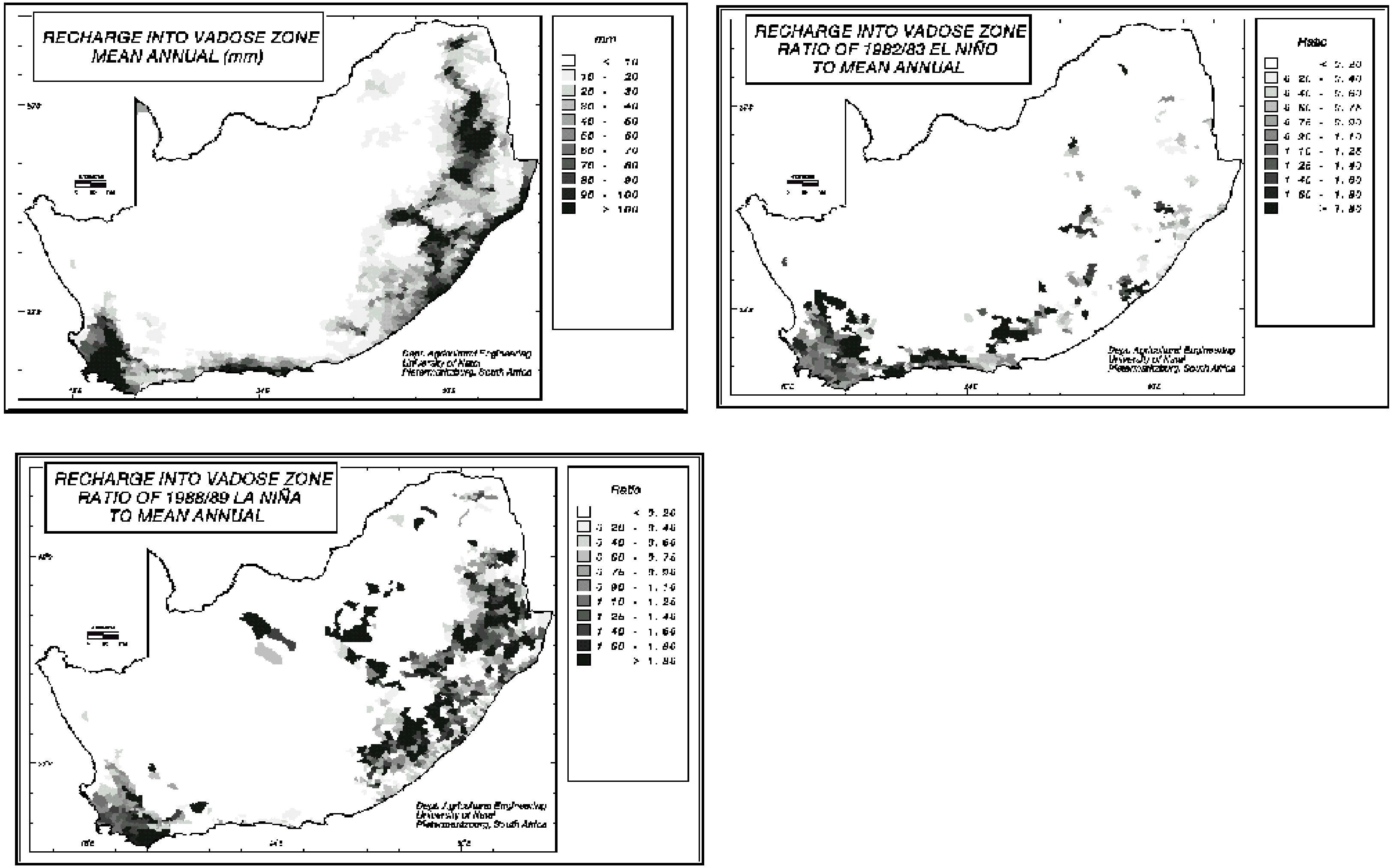

Figure 11.12 Average and extreme year recharge estimates for South Africa.

12. A Comparison of Recharge Estimates in a Karoo Aquifer from a Chloride Mass Balance in Groundwater and an Integrated Surface-Subsurface Model

Karim Sami
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ABSTRACT The paper describes the use of an integrated surface-subsurface modelling approach to estimate recharge to a semi-arid Karoo aquifer and a comparison of the results with a chloride mass balance on a sub-area scale. The chloride balance suggests a mean annual recharge of 4.5 $mmyr^{-1}$ based on a mean annual rainfall of 460 mm. The model predicts 5.8 $mmyr^{-1}$ from a mean annual rainfall of 483 mm during the simulation period. This is substantially lower than other estimates of recharge to Karoo aquifers. Recharge was found to be highly variable in space with a positively skewed distribution in time. Spatial variability was attributed to differences in the distribution of soil moisture content resulting from variations in slope, soil texture and depth. Both methods suggest that recharge is considerably less than the mean of 9.9 $mmyr^{-1}$ estimated from the currently prescribed rainfall-recharge relationship for Karoo aquifers.

12.1 Introduction

The quantification of regional recharge is of great importance in the semi-arid Karoo regions of South Africa. These regions are predominantly underlain by fractured rock aquifers of low storage capacity, which receive limited and highly variable recharge. Consequently, their groundwater exploitation potential during prolonged droughts is largely controlled by the temporal variability of recharge.

Most recharge studies in semi-arid areas based on the chloride method have been based on chloride profiles in the unsaturated zone, but these also only provide point estimates of recharge. Studies based on unsaturated zone chloride profiles equate recharge to the ratio of chloride in rainfall to that of soil water below the zero-flux plane. Hence they presume that recharge occurs through a matrix of deep unconsolidated sediments and that soil water concentrations reflect the degree of evapotranspiration. Some researchers assume steady state conditions so that the flux of chloride below the root zone equals the input to the surface (e.g. Johnston, 1987; Edmunds et al., 1992), while others use transient state profiles (e.g. Walker et al., 1991; Thorburn et al., 1991). Input rates are based on current atmospheric ascession, thereby assuming that all precipitation inputs percolate vertically through the soil. As lateral redistributions of the tracer are ignored, the method cannot be used where significant overland flow occurs.

These methods cannot be applied in South Africa because deep unconsolidated deposits are rarely encountered, except for those of limited lateral extent. In addition, very little water percolates through the soil matrix to any significant depth, even during high rainfall events (Sami, 1992; Kirchner and Van Tonder, 1991). South African aquifers are recharged primarily by indirect flowpaths and estimation methods which more accurately reflect our understanding of these physical processes on a regional scale are required.

Geochemical tracers in the saturated zone provide an attractive alternative since they reflect an areal integration of long term average tracer inputs, thereby circumventing complexities resulting from diverse recharge pathways. Yet the use of tracers in groundwater to quantitatively evaluate recharge has received less attention (Erriksson & Khunakasem 1969; Adar et al., 1988). As variations in geochemistry are more distinct in semi-arid areas than humid areas (Fontes et al., 1989), especially in fractured rock aquifers where lateral mixing is restricted, they may also be more easily related to variations in recharge.

A method to obtain historic mean annual recharge based on chloride concentrations in groundwater and rainfall is described in this paper. The method has been applied to a semi-arid catchment overlying a fractured Karoo aquifer. Ensuing recharge estimates are compared to the mean of a time series derived from a variable time interval rainfall-runoff model into which surface-subsurface interaction functions have been incorporated (Hughes and Sami, 1994).

12.2 Study Area

12.2.1 Physical description

The study area is a 111.7 km^2 network of gauged semi-arid catchments located in the Great Fish River Basin surrounding the town of Bedford in the Eastern Cape. The topography is one of rolling hills and ridges with gentle slopes (less than 8%) over more than 80% of the area. Local relief varies between 90 m and 150 m. Surface drainage consists of a dendritic system of ephemeral channels developed on bedrock with occasional alluvial or colluvial patches where significant transmission losses may occur (Hughes and Sami, 1993).

The mean annual rainfall is approximately 460 mm. Annual mean pan evaporation (based on regionalised Symons pan figures) is of the order of 1600 mm. Rainfall varies between short duration, high intensity convective storms mainly occurring during summer, to longer duration events of variable intensity and total depth which can occur at any time of the year. The regional long-term runoff coefficient is 3.19% (Tordiffe, 1978). Surface drainage is thus a small component of the long-term total water budget.

The soils are predominantly alkaline sandy clay loams, clay loams and sandy loams. The depth of weathered material varies from less than 500 mm on hillslopes to in excess of 6 m in some valley bottoms (Hughes and Sami, 1993). Fifty infiltration tests conducted using a sprinkling infiltrometer indicate that infiltration rates after 60 min range from less than 5 mm hr^{-1} on crusted and overgrazed slopes to between 10-20 mm hr^{-1} on more densely vegetated valley bottoms.

12.2.2 Geology

The study area is underlain by upper Permian age sandstones and mudstones of the Middleton Formation of the Karoo Supergroup, as well as by younger intrusions of Jurassic dolerites. Hydraulic gradients and fracture orientations generally follow the surface drainage. Static water levels vary between 5 m and 35 m below the surface. A geochemical and isotopic study by Sami (1992) found that chloride in groundwater is of meteoric origin and that residence times throughout the aquifer were relatively uniform. Geochemical variations therefore reflect spatial differences in recharge volumes.

12.3 Methodology

12.3.1 The Chloride Balance Method

In catchments where chloride can be shown to be of uniquely atmospheric origin its concentration in groundwater relative to that in rainfall can be used to calculate mean annual recharge. Estimates of the chloride flux to the catchment can be obtained from bulk precipitation samples so that:

$$J_p = C_p P \quad (1)$$

where J_p is the steady-state vertical flux density (mg m^{-2} yr^{-1}), C_p is the weighted mean chloride concentration in rainfall (mg l^{-1}) and P is mean annual precipitation (mm).

Under steady-state conditions where surface runoff is negligible, the chloride flux to the surface equals the flux to the water table. A removal of chloride during percolation is unlikely since the ion does not form salts unless concentrations are extremely high, it is not adsorbed on mineral surfaces and plays few biochemical roles (Dettinger, 1989). Fluxes to groundwater may be lost over the short term by the precipitation of chloride salts following rain events that are completely evapotranspirated. Being the most soluble, however, chloride salts are the last to be precipitated and the first to be leached in subsequent rain events. As a result, virtually all chloride salts are dissolved in whatever volume of water reaches the groundwater and its concentration in recharging water reflects the fraction of rain water reaching the water table.

Under steady-state conditions where hydrodynamic dispersion can be ignored (Allison et al. 1985) and flow is essentially vertical this relationship can be expressed as:

$$R = J_p/C_r \quad (2)$$

where R is the mean annual recharge (mm) and C_r the mean chloride concentration of percolating water (mg l^{-1}). This relationship is not strictly valid in groundwater as it does not take into account lateral differences in C_r resulting from variations in the rate of horizontal transport. Hence recharge cannot be calculated solely from local groundwater unless R and C_r are spatially uniform.

According to Eriksson and Khunakasem (1969), one dimensional horizontal chloride flux in an aquifer that is well mixed vertically, J_c (mg m^{-1} yr^{-1}), would be related to the accumulated vertical flux from a horizontal distance x (m) beginning at the catchment boundary and along the flow path so that:

$$J_c = J_o^x J_p \, dx \quad (3)$$

since the lateral flux of chloride can also be expressed as:

$$J_c = 1000 \, Q_g \, C_g \quad (4)$$

where C_g is the chloride concentration of groundwater at distance x, the steady-state groundwater flow Q_g (m^3 m^{-1} yr^{-1}) at point x can be calculated. Since under steady-state conditions groundwater flow is related to recharge by:

$$R = dQ_g / dx \quad (5)$$

an integration of R across horizontal unidirectional flow lines yields the recharge volume.

The major disadvantage of this method of computation is that Eq. (3) assumes uniform conditions in the aquifer so that chloride is only accumulated along a unidimensional flowpath. If flow convergences exist, or if some lateral redistribution of salts occurs at the surface prior to infiltration, the relationships discussed above would not be strictly valid. In semi-arid environments, however, lateral redistribution from the re-infiltration of runoff into localised recharge zones and from channel transmission losses are common recharge mechanisms (e.g. Lloyd, 1986). Furthermore, in fractured rock aquifers groundwater flow tends to be concentrated in the more permeable fracture zones, usually located in the valley bottoms. Chloride fluxes are therefore related to the chloride load of the area from which the fracture zone collects water. The transport of salts to valley bottoms by surface runoff suggests that contributing areas can be defined by the surface topography. Under steady state conditions lateral chloride exports would be related to the vertical flux and the sum of the contributing sub-catchment areas by:

$$J_s = J_p \sum A_j \tag{6}$$

where J_s is the rate of chloride export (mg yr^{-1}) and A is the sub-catchment area (m^2). If it can then be assumed that outflow across a sub-catchment's seepage face is of a uniform chloride content, annual sub-catchment groundwater discharge Q_d (m^3 yr^{-1}) can be calculated by:

$$Q_d = J_s / 1000\, C_g \tag{7}$$

and sub-catchment recharge is:

$$R = 1000\, (Q_{doj} - Q_{dij}) / A_j \tag{8}$$

where o and i are groundwater outflows and inflows to sub-catchment j.

12.3.2 The Variable Time Interval (VTI) Model

The semi-distributed VTI is an integrated surface-subsurface model which runs within a hydrological modelling system (HYMAS). Although it incorporates some relatively complex hydrological processes, the model has been developed for practical application purposes and is applicable over a wide range of climatic types. The theory and structure of the model have been completely described in Hughes and Sami (1994).

12.3.3 Data availability and preparation

Groundwater samples were collected from 12 boreholes on a monthly basis for a period of 18 months between 1989-1990. During the study period many small events and one large recharge event were recorded, none of which resulted in any changes in groundwater chemistry.

Using the ARC/INFO GIS package, 12 sub-areas were defined according to borehole locations and topography, with each borehole considered to be the outlet of a subsurface catchment (Figure 12.1). These represent the surface area (A_j) contributing chloride to each of the sub-catchments characterised by a borehole.

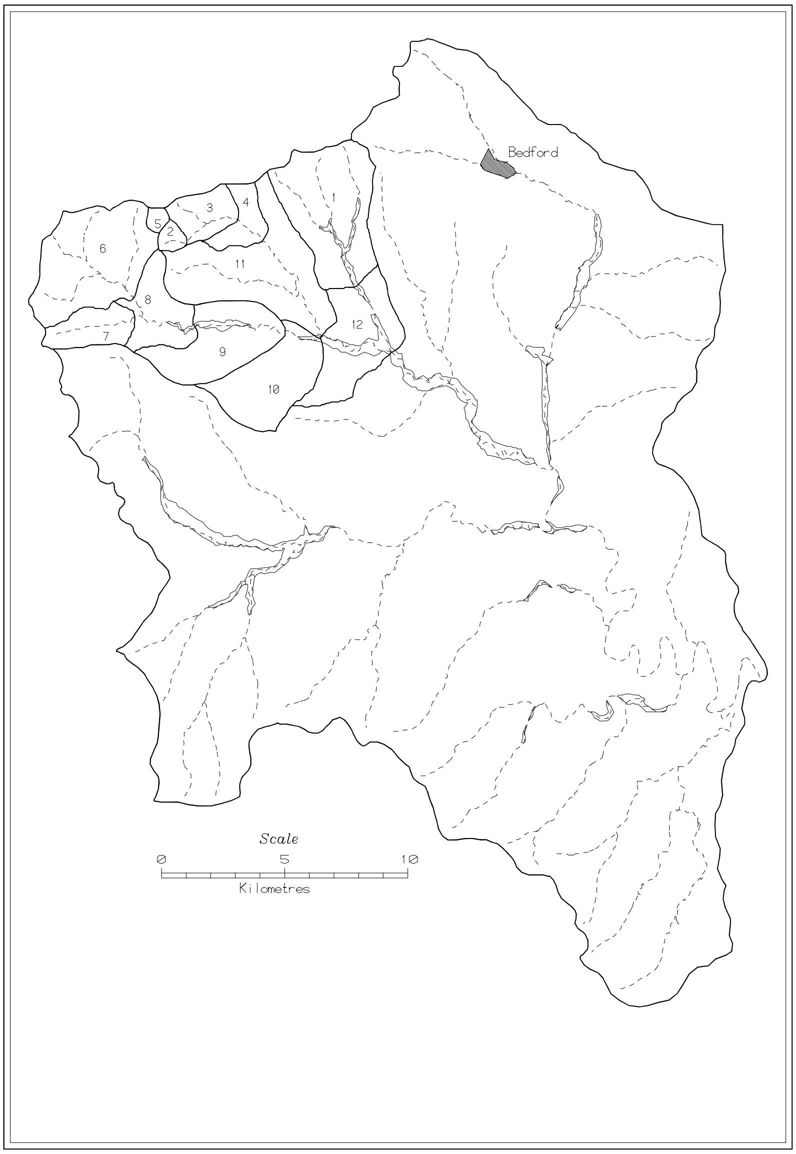

Figure 12.1 Location of study area and subdivison into sub-areas.

Sub-areas were progressively integrated from the catchment boundary along the regional hydraulic gradient in order to calculate the rate of chloride export J_s through each sub-area. Fluctuations only occur in response to large rain events and are highly variable spatially. Hughes and Sami (1993) recorded responses of 0-5 m following 130 mm of rain. In addition, the delays to borehole hydrograph peak varied from 50-520 days following the event, generally increasing eastward. This was attributed to the lateral migration of the recharge pulse through the aquifer.

Rainfall samples were collected monthly from 8 tipping bucket rain gauges. These were fitted with plastic collectors and therefore represent a composite of the monthly chloride load. One South African Weather Bureau rainfall gauge (076/884) is located in sub-area 9. The mean annual rainfall at this gauge since 1956 is 483 mm. Rainfall from this gauge was used to run the VTI model from 1956-1991.

Observed runoff data was available for the outlet of sub-area 9 for the period 1989-1991 and was used to calibrate the VTI model. A significant tract exists in subareas 8 and 9, where Hughes and Sami (1992) estimated that transmission losses into this alluvium diminished storm runoff volumes by up to 181 700 m^3 during a 130 mm event. This represented 22% of the total generated runoff volume and is equivalent to 6 mm of runoff from the upstream sub-areas. These losses recharge groundwater in the immediate vicinity of the alluvium, since large and rapid variations in groundwater levels have been observed in a borehole located in the vicnity (Hughes and Sami, 1993).

12.4 Results and Discussion

12.4.1 Chloride mass balance

During the 729 day study period between 622 and 996 mm of rain were recorded at the various tipping bucket gauges, with the average being 780 mm. The chloride concentrations in rain water varied from barely detectable amounts to 10.3 mg l^{-1}, with a weighted mean value of 4.11 mg l^{-1}. In general, much less variability was recorded in the chloride concentration of large rain events than in the smaller convective rain events. If the recorded weighted mean concentration is representative of the long term rate of chloride deposition, then J_p is 1.99 g m^{-2} yr^{-1}. From Eqs. (1) and (6)-(8) sub-area recharge rates were calculated (Table 12.1). The area weighted mean recharge rate is 4.45 mm yr^{-1}.

Table 12.1 Recharge estimates from a chloride mass balance in the 12 sub-areas.

Sub-area	Contrib. sub-areas	Area (km^2)	G.water Cl (C_g) (mg l^{-1})	Sub-area Cl load (T yr^{-1})	Total Cl load (J_s) (T yr^{-1})	G'water flow (Q_d) (Ml yr^{-1})	Change in flow (Q_{do}-Q_{di}) (Ml yr^{-1})	Sub-area recharge (mm yr^{-1})
1	1	17.74	395	35.2	35.2	89.1	89.1	5.03
2	2	1.07	546	2.1	2.1	3.9	3.9	3.64
3	2 & 3	4.45	336	8.8	10.7	31.9	28.4	6.40
4	2 - 4	3.52	416	7.0	17.7	42.6	10.6	3.03
5	5	0.73	509	1.4	1.4	2.8	2.8	3.90
6	5 & 6	17.69	272	35.1	36.5	134.4	131.5	7.44
7	7	4.91	261	9.7	17.0	65.3	38.4	7.84
8	5 - 7	6.94	412	13.8	57.6	139.9	0.0	0.0
9	5 - 9	11.65	346	23.18	80.8	233.4	93.59	8.03
10 & 11	5 - 11	30.89	425	61.3	159.8	376.0	100.0	3.24
12	1 - 12	12.18	537	24.2	152.0	283.1	0.0	0.0

12.4.2 Variable Time Interval Model

Table 12.2 gives a summary of the simulated mean annual water balance of each of the 12 sub-areas and the entire catchment.

Table 12.2 Mean annual water balance simulated by the VTI model for the period 1956-1991.

Sub-area	1	2	3	4	5	6
Actual Evap.	466.7	449.9	457.3	462.1	442.0	449.5
G'water arge	5.8	6.4	7.0	8.2	4.7	3.0
Transmission losses ^{3}*10^3)	0.53	0.00	0.04	0.11	0.00	0.35
Sub-area Routed Runoff	10.4	26.6	18.6	12.6	36.2	30.4
Sub-area	7	8	9	10	11	12
Actual Evap.	455.2	463.4	465.2	465.1	465.3	467.2
G'water Recharge	4.7	3.2	6.9	7.2	7.4	6.0
Transmission losses ^{3}*10^3)	0.05	29.08	19.34	1.50	0.77	0.97
Sub-area Routed Runoff	23.0	16.3	10.8	10.6	10.2	9.7

The statistics of the temporal variation of recharge are given in Table 12.3. The skewness values demonstrate that recharge is positively skewed. By plotting these distributions on a probability plot it was found that the sub-areas with the highest skewness coefficients (1, 6, 8, 9 and 12) had log-normally distributed recharge, while the remaining sub-areas are best approximated by a square root-normal distribution.

Table 12.3 Annual recharge statistics for the 12 sub-areas and the entire catchment expressed in mm.

Sub-area	1	2	3	4	5	6	7	8	9	10	11	12	Tot
Mean(mm)	5.80	6.44	7.00	8.17	4.67	2.98	4.7	3.18	6.87	7.15	7.37	6.0	5.8
Med.(mm)	5.22	6.03	6.84	7.88	4.23	2.42	4.24	2.75	6.19	6.78	6.82	5.52	5.29
S. Dev. (mm)	3.25	2.82	3.05	3.61	2.19	1.90	2.7	1.79	3.38	3.48	3.49	3.36	2.90
C. Skewness	1.0	0.15	0.15	0.55	0.41	0.79	0.43	0.73	0.77	0.62	0.66	0.9	
Mn(mm)	1.19	1.17	2.14	2.26	1.14	0.47	1.12	0.61	1.62	1.88	1.86	1.41	1.36
Max(mm)	13.7	11.3	12.9	17.8	9.0	7.8	10.1	7.2	14.7	15.4	15.6	13.6	12.0
CV(%)	56.1	43.8	43.5	44.2	46.8	107	57.5	56.3	49.2	48.8	47.3	55.9	50.1

12.4.3 Comparison of recharge estimates

Figure 12.2 shows annual rainfalls and the area weighted recharge simulated for the entire catchment. The model simulates a mean annual recharge of 5.8 mm yr^{-1}, while the chloride method yields 4.5 mm yr^{-1}. Figure 12.3 compares the estimates by sub-area. Comparable results have been obtained for sub-areas 1, 3, 5, and 9, while discrepancies occur in sub-areas 4, 6, 8, 10, 11 and 12. The discrepancies can be partially explained by processes affecting the horizontal redistribution of chloride by surface runoff generation, which would influence the accuracy of the chloride balance on the local sub-area scale. For example, in sub-areas 6 and 7 a significant proportion of the chloride deposited by rainfall is transported to sub-area 8 by overland flow, where surficial salts flushed in the initial wave of surface runoff infiltrate into the upper reaches of the alluvium in sub-area 8 as transmission losses. As a result, the chloride model over-estimates chloride inputs to groundwater while under-estimating those to sub-area 8, implying that recharge estimates in sub-areas 5-7 could be too high while those in sub-area 8 too low.

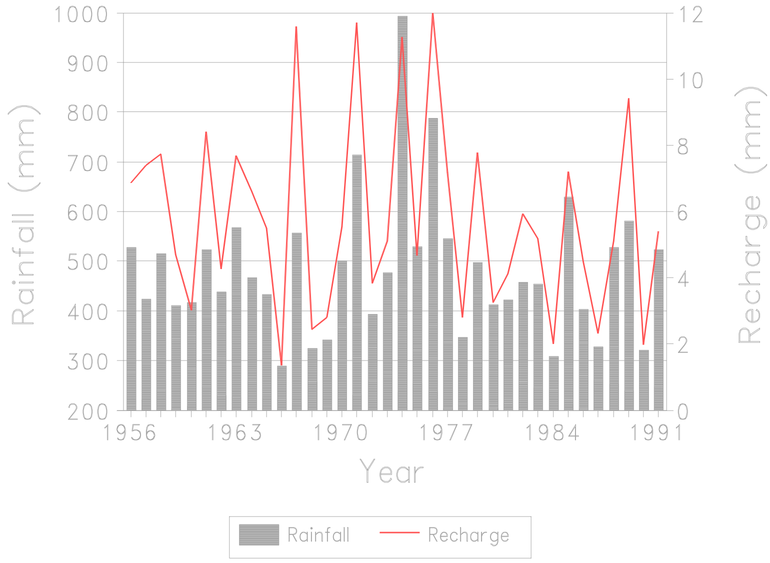

Figure 12.2 Rainfall and recharge between 1956-1991 as determined by the VTI model.

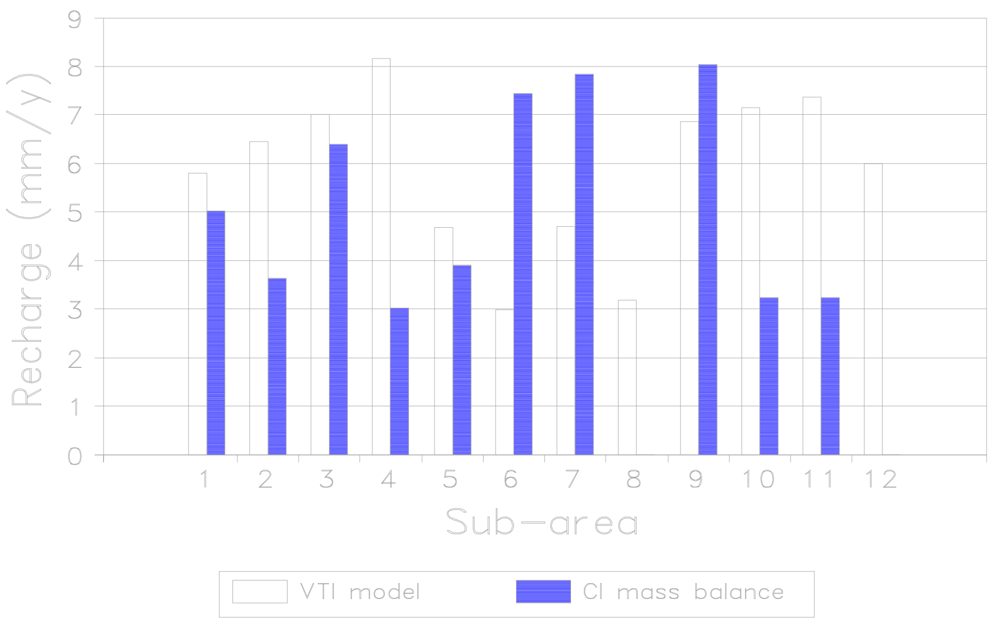

Figure 12.3 Comparison of recharge estimate by the VTI model and the Cl method.

12.4.4 Rainfall-recharge relationships

A regressional relationship between simulated annual recharge and rainfall is shown in Figure 12.4. A regression of rainfall and recharge predicts that:

Recharge (mm) = log (Rainfall (mm)) * 22.4 - 54.1 mm (9) with an r^2 of 0.79 and standard error of 1.97. The relationship exhibits a poorer fit during high rainfall years than low rainfall years because such a simplistic approach ignores the temporal distribution of the rainfall. Using this relationship and the long term mean annual rainfall of the catchment, a recharge rate of 5.5 mm yr^{-1} is obtained.

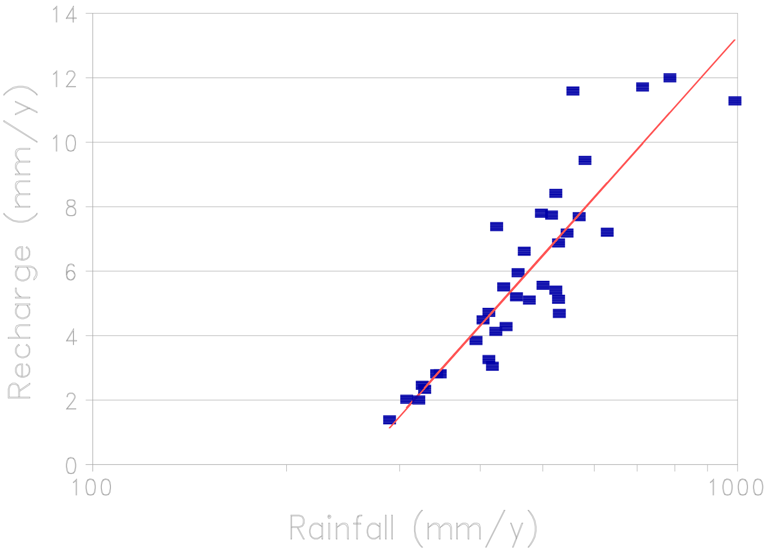

Figure 12.4 Relationship between annual rainfall and recharge.

Annual variations in recharge for selected sub-areas are shown in Figure 12.5, which illustrates that there is more spatial variation during wet years than dry years. Recharge ranges between 2-14 mm, which is equivalent to 0.7% of rainfall during dry years and 1.2% during wet years. This is substantially lower than the currently prescribed 2-4% by the Kirchner and Van Tonder regressional equation (1991) for Karoo aquifers with soils greater than 200 mm thickness. The relationship presented by Kirchner and Van Tonder, suggests that catchment recharge should vary between 5-22 mm yr^{-1} with a mean of 9.9 mm yr^{-1}. The mean annual recharge of the time series obtained from the VTI model is in much closer agreement with results from the chloride mass balance than the Kirchner and Van Tonder relationship, which indicates the inherent danger of using regionalised relationships.

12.5 Summary and Conclusions

This paper presents a time series of groundwater recharge simulated by an integrated surface-subsurface semi-distributed model and a comparison with mean annual recharge rates derived from a chloride mass balance. The model was applied to a semi-arid Karoo region of South Africa covered by rolling grassland and underlain by fractured sedimentary rocks. The chloride balance shows that mean annual recharge is of the order of 4.5 mm yr^{-1} from a mean annual rainfall of 460 mm. The model predicts a mean annual recharge of 5.8 mm yr^{-1} based on a mean annual rainfall of 483 mm between 1956-1991, but suggests that recharge can be highly variable in space with a positively skewed distribution in time. The spatial variability of recharge was found to be higher in wet years than in dry years.

Both methods suggest that recharge is considerably less than the mean of 9.9 mm yr^{-1} suggested by the currently prescribed rainfall-recharge relationship for Karoo aquifers in South Africa with soils deeper than 200 mm.

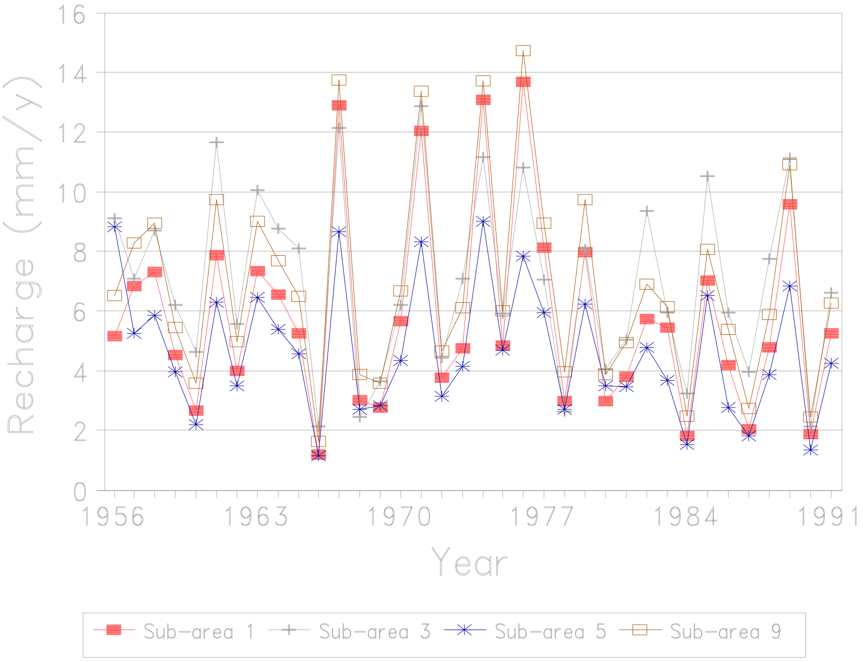

Figure 12.5 Recharge for to selected sub-areas as determined by the VTI model.

12.6 Acknowledgements

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PART VI

Towards Sustainable Development of Groundwater Resources

13. Changing Rainfall – Changing Recharge?

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ABSTRACT The assumed consequence of rising temperatures of the earth's surface is a shift of the climatic belts towards the equator. Rainfall has decreased in parts of Southern Africa over the past twenty years and high rainfall events that are mainly responsible for aquifer recharge have become scarce. Although these changes are statistically not relevant hydrogeologists have to react to indications that groundwater recharge is seriously affected. The situation also requires revision of some commonly accepted assumptions underlying recharge and assured yield determination techniques.

Considering the increasing scarcity of water resources inclusion of environmental topics in primary to tertiary education is propagated. Better understanding of the environment and the processes that underlie the replenishment of groundwater resources should in the long-term help to reduce, if not eliminate, unconsidered changes to the hydrological cycle.

Proposals are made, how the reduced resources may be stretched and used more efficiently.

13.1 Introduction

Given the chemical composition of the Earth and the external energy supplied by the sun geology controls our entire surrounding. Excluded from this control are the rare impacts of meteorites and also the new power our species has obtained with the development of natural sciences and their application in the engineering environment[1].

Geology and the sun's energy have shaped the surface and are mainly responsible for climate and rainfall distribution. Slope, soil, sunshine, temperature and water determine largely the plant cover.

All processes on earth strive towards an equilibrium situation. While internal forces create mountain ranges and oceans, erosion models the finer details of the surface and fills the depressions.

Groundwater too is constantly adapting to changes. These changes are induced by recharge (from precipitation and surface water) as well as losses (through outflow and consumption).

This paper intends to show the challenges that the groundwater community may face in a changing environment. Some assumptions we make in our day-to-day groundwater management decisions seem to be no longer true. Potential solutions to growing demand shortages in a holistic, conjunctive use approach may help to ease the consequences.

13.2 Groundwater under Arid and Semi-arid Climate Conditions

Neglecting consumption for the moment the water level in an aquifer reflects the varying recharge conditions. Under arid and semi-arid conditions the water level drops during the year

[1] Whether, from the viewpoint of the Earth, meteorite impacts and human activities must be regarded as catastrophic shall not be discussed.

due to outflow (or the balance between outflow and inflow). During a normal rainy season the water level raises somewhat and then starts receding again. It is only during the major rainfall events that occur every thirty-, fifty-, one hundred years or longer that the aquifer is fully recharged again. This principle is shown in Figure 13.1. The actual water-level behaviour is somewhat modified through the following processes:

- Evapotranspiration by plants exerts a higher strain in the summer months.
- Abstraction also tends to be high in the summer months but is reduced during cooler periods and after rainfall events.
- Where aquifer recharge occurs from surface water and river alluvium the available head influences the water-level rise.

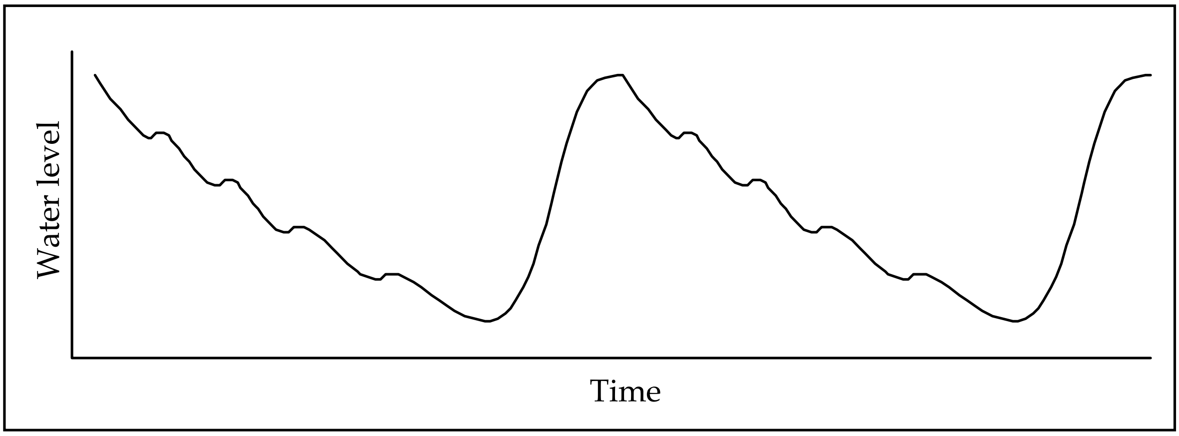

Figure 13.1 Schematic water-level changes with time.

In addition one has to realise that recharge, especially in semi-confined and confined aquifers, not only occurs in a vertical but predominantly in a ± horizontal direction. The water-level rise at a given point in an aquifer therefore also depends on the distance from the recharge area; the aquifer transmissivity; and the storativity. Figure 13.2 shows schematically the high and immediate response of the water level near the recharge area and the delayed lesser response farther away.

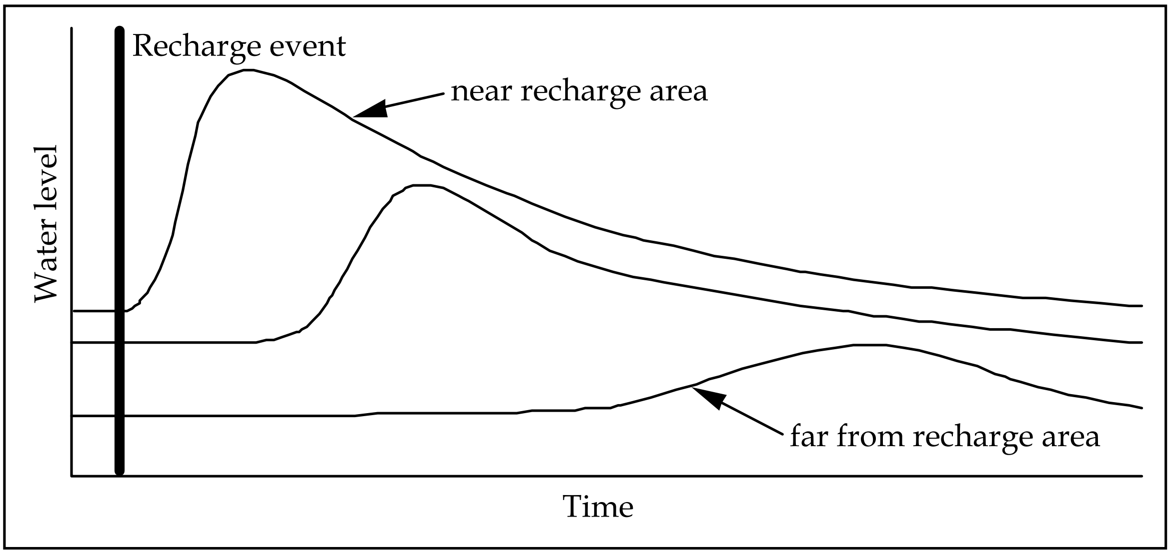

Figure 13.2 Water-level response to recharge event.

This means that the lower part of an aquifer is only recharged months after the event, partly depleting the upper part of that aquifer. Experience has shown that it takes two good rainy seasons following each other with not more than one lesser season in between to completely recharge an aquifer. This as also implied by the short and long memory of the CRD method (Bredenkamp et al., 1985).

Most aquifers in Southern Africa are fractured aquifers. This means that the water is mainly stored in joints, fractures and faults and to a much lesser degree in the matrix of the rock. Permeability and storativity in fractured aquifers vary with depth and it has been found that the water-bearing capacity of most of these aquifers decreases considerably when the water level is lowered by more than 20 to 30 m.

Abstraction causes an additional drawdown of the water level in addition to the natural recession. The degree of the additional drawdown depends on the amount withdrawn and the storativity of the particular aquifer.

13.3 Rainfall and Recharge

Aquifer management can be described as the art to abstract not more water than is replenished with due consideration of the unavoidable losses. Good housekeeping is therefore essential within the time span major recharge events occur. Many of the present efforts in hydrogeology aim at the determination of the recharge.

All the methods used to define the rate at which groundwater is replenished, imply that we deal with a stable climate. Although we know that the climate has changed in historical times it is only during the past ten or twenty years that the influence of increasing CO_2 levels and depletion of the ozone layer is entering discussions. It is, however, still unclear, what influence this may have on the weather pattern in Southern Africa. Clear is that the temperature of the earth's surface is rising.

Figure 13.3 shows a comparatively slow rise of the temperature during the first three decades of this century followed by a relative stable period until the end of the seventies.

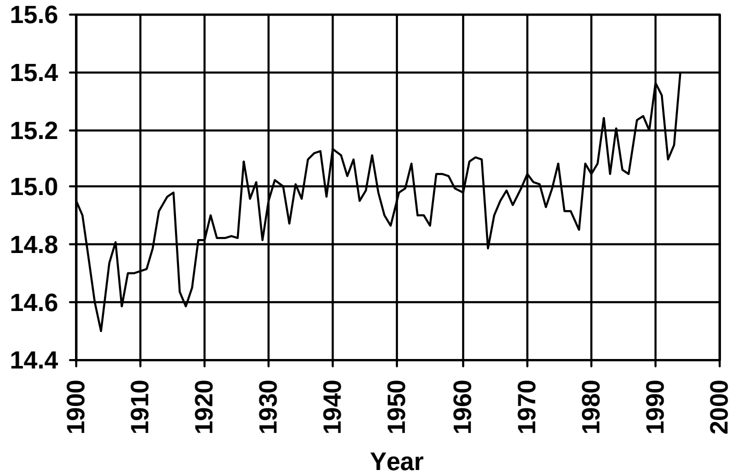

Figure 13.3 Mean global temperature.

Since then temperatures have risen at an increasing rate, only shortly interrupted by the activities of Mt. Pinatubo in the early nineties. Climatologists assume that rising global temperatures change the atmosphere's circulation pattern and cause a shift of the climatic belts towards the equator. Decreasing rainfall and runoff in the central and northern parts of Namibia and increasing precipitation in the South (e.g. Karasburg) do not contradict this theory. The reported southwards extending Sahel zone in North Africa also fits into this model.

Analysis of rainfall data is difficult and it is often maintained that the time series available are not sufficiently long. Dyer (1978) showed in his analysis of rainfall and runoff data that in some of the investigated catchments runoff follows the sunspot cycle; in others it is in anti-phase; and in still others it changes abruptly from in-phase to anti-phase. He came to the conclusion that a hypothesis of no relationship between river flow and the solar cycle could not be accepted.

The additional 18 years of rainfall and runoff data that have been recorded since Dyer's analysis have not helped to find statistical proof of cycles or trends. Van Langenhove and Rukira (1995) studied recent river flow regime changes in the Southern African Region and the possible reasons. During the past 20 years, runoff in Namibia's northern border rivers and the Zambezi has decreased to about half of what it was before the eighties. They found that longer periods of even lower flows have occurred in the past and concluded that the present situation should not be seen as an exceptional new development and may continue for an indefinite period.

It may equally well be that recent rainfall series reflect changing rainfall patterns. Analysis of the Windhoek rainfall data suggests, that the rainfall pattern is changing (Figure 13.4):

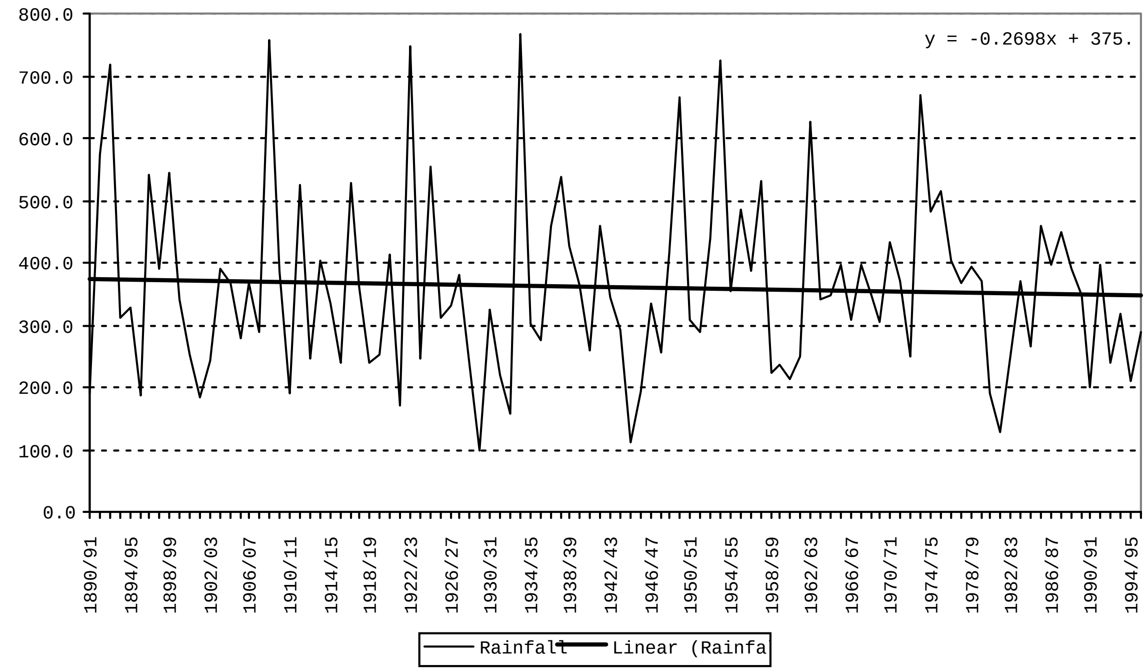

Figure 13.4 Windhoek rainfall.

1. The mean annual rainfall shows a slightly falling tendency for the past hundred years from about 375 to 350 mm.
2. Since about the 1950's the peak values measured in the extraordinary rainfall years have decreased.

3. For the first time during the record period the ten-year mean rainfall has fallen for prolonged periods below the median ten-year rainfall (see Figure 13.5).

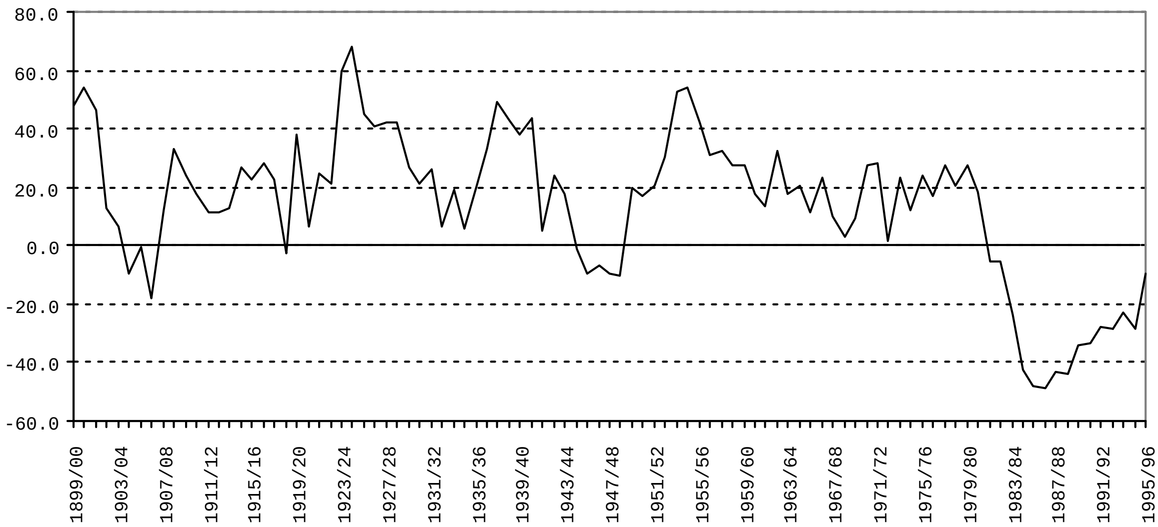

Figure 13.5 Windhoek: ten-year Mean minus Median rainfall.

This means that the drought years have become relatively more severe although the rainfall during the "median" rainfall year has slightly risen. The immediate consequence of this observation is that those rainfall events that are responsible for the major recharge events (see Figure 13.1) have become less frequent; less pronounced; or both: we must therefore expect considerably less recharge than the long term mean recharge.

Again using the Windhoek data as an example Figure 13.6 shows the decrease in the number of months with rainfall above 50 mm, that is, those events that contribute most to the groundwater recharge. In the short term the consequences for surface runoff may be less severe because with the reduced peak rainfall values dams may still occasionally be filled. Only the number of times they overflow may be reduced.

The question arises whether the observed changes are significant. Statisticians say, "No, we need much longer data series to assess the variability". Data generation based on existing (short) data series and assumptions (like: the mean rainfall is constant) does not provide an answer either[1]. Water supply is a relatively short-term business. We have to respond to today's situation irrespective of what the probabilities are that the next decade or the decade thereafter may reverse the present situation. In Germany the dikes along the North Sea are raised in spite of the fact that observed sea levels rise is not statistically significant because of the otherwise potential disastrous consequences. The exceptionally high number of hurricanes in the southern United States in 1995 and the intensity of hurricane Andrew (Anonymous, 1992 and Weber, 1996) have convinced American insurance companies that the climate is changing.

[1] Assumptions lying behind data generation influence the outcome. Above average rainfall during the Seventies has led to over-optimistic judgement of the MAR in the White Nossob River, where the Otjivero Dam was built.

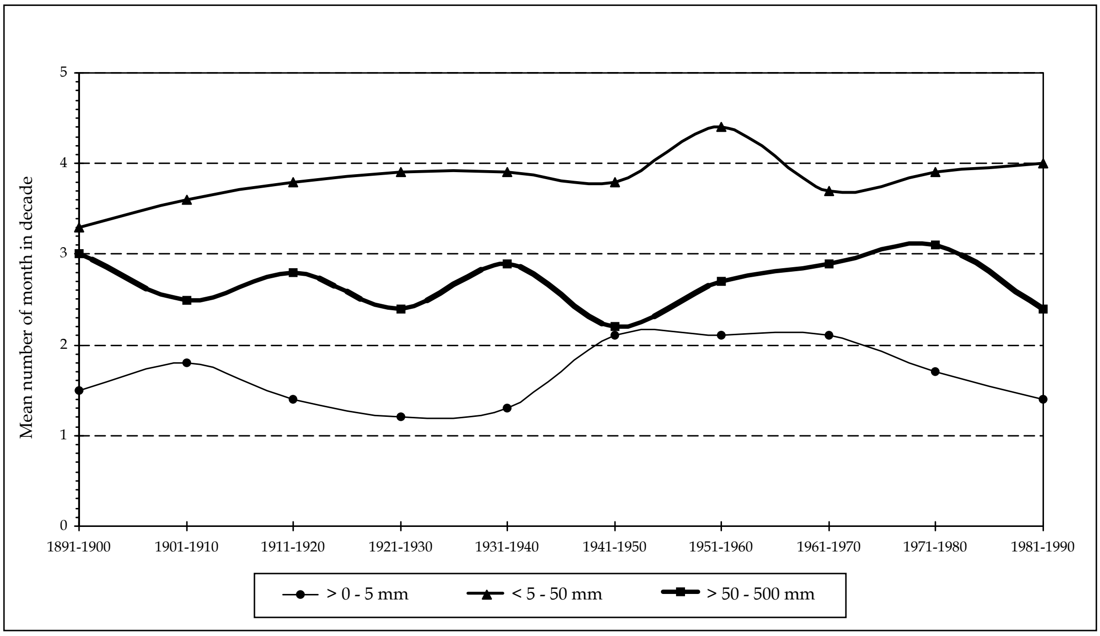

Figure 13.6 Mean number of months per decade with specified rainfall totals.

13.4 Climate Changes

As geologists we know that the climate does vary over long periods of time. If we look at the climatic history of the Pleistocene in Europe we find that there were at least six glacial periods leading to a mean annual temperature drop of about 16°C (see Figure 13.7, after J. Büdel in Brinkmann, 1954)[1].

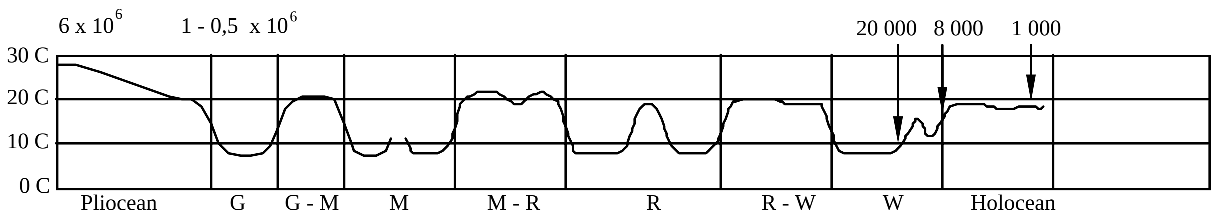

Figure 13.7 Climate changes in Middle Europe.

Since the end of the last glaciation about 20 000 years ago several medium and longer-term variations have also been recorded: In the 11th century there was a climate optimum. In the 17th century mining activities in the high mountain regions of the Alps came to a halt because adits to the mines were covered by progressing glaciation. Glaciers in the Alps had their greatest extension in the late 1850's and have since receded and have lost half of their volume during the past 140 years.

The decreasing runoff in parts of Southern Africa, e.g. in the tributaries of the Zambezi, the Okavango, Cunene and others, has been mentioned above. Prevailing winds and currents have changed in the Southern Atlantic. For water supply we cannot wait for sufficiently long time

[1] The European observations reflect a worldwide change as glaciation was also observed in North America, South America and Antarctica. With a decrease of about 4° C temperature changes in the tropics were less drastic

series data to be collected. For the short-term future we have to take the observed changes into consideration and adjust our estimates accordingly.

13.5 Water Resources Management

At least in some parts of the Southern African subcontinent the consequences of climatic changes on recharge will probably be:

- lesser replenishment of aquifers during extraordinary rainy seasons;
- less frequent complete recharge.

13.5.1 Adjustment of methods

Most of the methods that are used to estimate recharge are based on the assumption that the climate is stable, that is for instance, mean rainfall or mean chloride deposition do (and did) not change.

In the case of the Windhoek example it makes a considerable difference whether one uses a mean of 375,5 mm/a for the cumulative rainfall departure curve (CRD: Figure 13.8) which is valid for the start of the available time series in 1890 (see Figure 13.4); 361 mm being the mean for the 105 year record period; approximately 347 mm/a (the present value on the trend line); or the value for the particular year on a sliding trend line. No particular justification can be seen in using the long-term mean value because that is completely arbitrary: should recording have started in 1950 one would use another value than for a record series starting in 1970 or 1890.

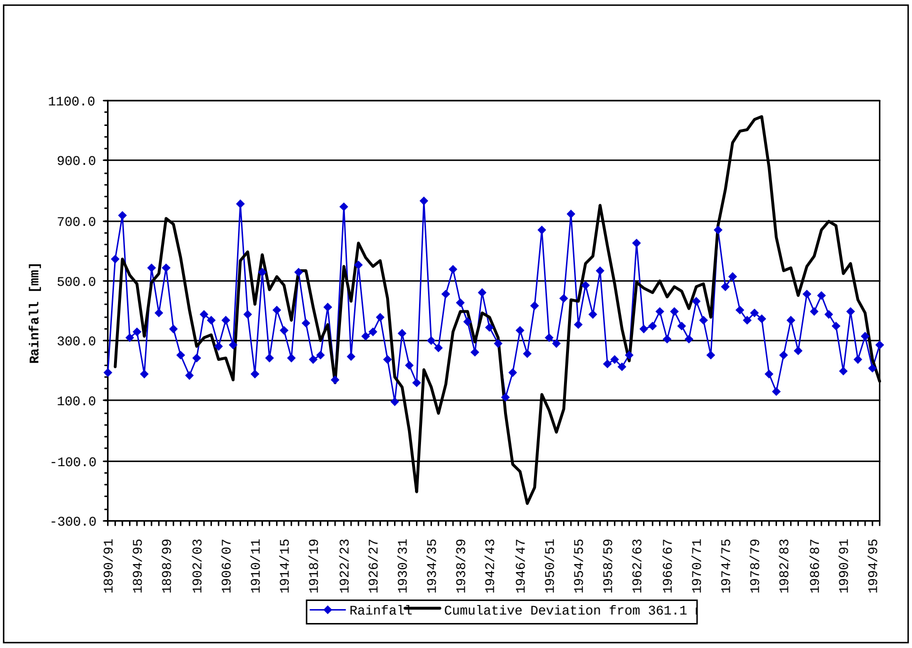

Figure 13.8 Cumulative rainfall departure diagram.

It is proposed that for management purposes rather mean values relevant for the immediate past are used because they reflect the present aquifer status much better.

13.5.2 Adjustment of approach

Another aspect that needs to be addressed is mankind's interference in the environment and, as regards our immediate field of interest, especially in the hydrological cycle. If we look for the underlying reasons for the present situation we find that:

- Development of science since the Renaissance had opened the door for the changes that have come with the industrialisation age. James Watt was the first who laid the foundation for the replacement of horse and human power through mainly fossil energy sources. The invention of cement carried this development further.
  With the "empowerment" of the human race we have obtained the means to change our environment on a global scale. Unfortunately this has happened without considering these changes in comprehensive feasibility studies that normally precede larger projects.
- Statistics offer the means to cost optimise actions but not their (unwanted) results.
- The world has become increasingly rectangular. With increasing urbanisation and the advent of the computer age our race is losing contact with nature. This trend has lead to a split between the environment and our modelled perception of it. Remote sensing and software more and more replace nature.
- If modern techniques are used to supplement our observations we will largely benefit. If, however, they replace our perception of the world around us we shall have to pay for the difference between reality and apprehension.

In the fifties and sixties young bright people were eager to become civil engineers, a profession they thought contributed more than other engineering professions to human welfare. This viewpoint also reflected the attitude of society. A sudden change in attitude occurred in the beginning of the seventies when people began questioning the benefits of large civil engineering works. A new consciousness for the impact of society's activities on the environment developed (James and Niemczynowicz, 1992).

James found the absence of well-organised and workable communication between civil engineers and nature particularly disturbing. Civil engineers are crucial to reduce the undesirable consequences of the civilisation machine. Consequently, their widening working sphere has to be mirrored in future education, so that engineers are prepared and able to cooperate in a constructive manner with ecologists and other necessary scientists.

13.5.3 Adjustment of priorities

In view of growing demands and diminishing water resources it seems appropriate to review the present situation. The tendency to build greater and more expensive schemes can still be noticed. It is the author's opinion that time has come to:

- More thoroughly consider the environmental effects and consequences of our actions.
- Do away with simplifying perceptions and try to minimise the difference between nature and our understanding of it.
- Make better use of the available resources and minimise losses.

In the field of hydrogeology and hydrology the losses comprise:

(a) evapotranspiration from open water surfaces (from approximately 40% upwards) and near surface groundwater bodies,
(b) unaccounted for water losses from reticulation systems that may typically amount to 20% or more of the water supplied,
(d) losses from the sewage network potentially to be used for water reclamation, or
(d) loss of suitable resources because of pollution,
(e) irrigation in water deficient areas; possibly on not really suitable soils; and with wasteful irrigation methods

Solutions that come to mind include:

- conjunctive use of water resources,
- artificial recharge to reduce evaporation losses and possibly improve groundwater quality; construction of sand storage dams,
- upgrading of reticulation and sewage networks, and in general
- avoidance or reduction of man-made activities that have a negative impact on the hydrological cycle such as increasing losses, reducing natural recharge or polluting water resources,
- water reclamation, and
- rainfall harvesting.

13.6 Conclusions and Recommendations

Indications are that recent changes in the weather pattern in parts of Southern Africa reduce the available water resources. Depending on the rainfall distribution and intensity surface water resources may receive slightly higher runoff in the "average" year. Major flood events become less likely. For drought periods there will be fewer reserves.

Aquifer management will have to accept both, smaller "assured" yields and lesser stored reserves for longer periods. This will also affect the conjunctive use of surface- and groundwater resources. Water seems to become increasingly scarce and we will have to learn to consume less.

It seems necessary to generate a wider awareness of the impact that man's activities have on the environment in general and on water resources in particular. To that purpose environmental topics should be included in the curricula of primary and secondary schools. On tertiary education level environmental hydrology and hydrogeology should be compulsory topics for engineers and scientists that will work in the water resources field.

To overcome the looming water shortages we must look at water reclamation and desalination as one way out of the dilemma as long as the rainmakers do not achieve greater success. Rainfall and dew harvesting may be other unconventional methods to look at. What we can do further is:

- Concentrate on artificial recharge and construction of sand storage dams to reduce evaporation losses thus making better use of the resources we have.
- Prevent "Losses unaccounted for" in our conveyance and reticulation systems.
- Care for the environment and reverse the damages done by us.

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14. Impact of Climate Change on Groundwater Recharge Estimation

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ABSTRACT Past observations and future projections of the global climate indicate that precipitation patterns are changing. Rainfall is a key factor in determining groundwater recharge and changes in the amount, frequency, duration and intensity of rainfall events will thus have a significant impact on groundwater resources. Groundwater responses to rainfall events have a longer lag time than the corresponding hydrological response in surface water systems. The impacts of rainfall variations on groundwater are generally smoothed over the short term. As a result, groundwater supplies are reasonably well buffered against the impacts of climate variability and provide a valuable resource to cope with short-term drought conditions, particularly in a conjunctive use situation. The presence of palaeo-groundwater in several arid zone aquifers, however, illustrates that groundwater systems are not resilient to long-term climate change.

Emerging trends from climate models for the next 50 to 80 years indicate that the subcontinent will experience both increases and decreases in precipitation. Using a simplistic empirical relationship between mean annual rainfall and recharge, we anticipate that a decrease in rainfall over the central parts of Southern Africa could have dire consequences for groundwater-dependent communities. A 20% decrease in mean annual rainfall volumes could translate to an 80% decline in recharge for areas that currently receive 500 mm rainfall per annum or less. In contrast, increased rainfall and increased rainfall variability, expected for the eastern Southern African coastal regions, could give rise to increased recharge and associated problems of rising water tables, increased baseflow to surface water systems and extreme flooding events.

14.1 Introduction

Climate change is a reality of life on Earth, and the geological record has many examples of both warmer and cooler periods. There is now well-founded concern that human activities over the past two centuries have caused unprecedented changes in the global climate over and above the natural climate variation. Observations over the past century, particularly over the northern hemisphere, indicate a trend that the world is warming, snow and ice cover decreasing and global average sea level rising (IPCC, 2001).

Along with growing awareness of global change, interest is emerging in determining the potential impacts of climatic variability and climate change on water resources (Loáiciga et al., 1996; Bonell, 2001). Intra-annual and inter-annual variability of water resources are directly related to rainfall variability. The impact of inter-annual rainfall variation on surface water resources is apparent in the variation in stream flow and increases with increasing size of the catchment. Short-term climate variations also affect groundwater recharge, but the impact on groundwater resources is generally smoothed in comparison to surface water.

The interest in determining impacts of climate change on groundwater resources is growing rapidly, particularly in Southern Africa, due to the increasing scarcity of water resources, as a result of increasing water demand for industrial, agricultural and drinking water use. The South African Water Research Commission, for example, recently commissioned a project relating to recharge processes in the Table Mountain Group aquifer systems, with one of the specified objectives to evaluate the sensitivity of recharge to spatial and temporal rainfall patterns with a view to understanding the impacts of climate change on water resources.

14.1.1 Groundwater recharge

In Southern Africa, and especially in the arid and semi-arid areas, groundwater is a strategic resource of great importance. Groundwater recharge is a key factor in determining the sustainable management of groundwater resources. The South African Water Act (1998), for example, states that only if recharge exceeds the sum of basic human and environmental needs in a catchment plus groundwater outflow necessary to sustain the same needs downstream, may groundwater be allocated for other uses.

A wide range of techniques are employed for estimating groundwater recharge, including direct methods of measuring infiltration (e.g. lysimeters) and indirect methods (e.g. catchment water budgets, soil moisture balance in hydrological models or chloride mass balance methods) in which the groundwater recharge is calculated, inferred or simulated (see e.g. Lerner et al., 1990; Gieske, 1992; Bredenkamp et al., 1995; Beekman and Xu, 2003). Other techniques such as groundwater dating, or the use of stable isotopes and chemical tracers, provide supplementary methods for recharge estimation by providing insight that may help to constrain the hydrological processes and transport times of the groundwater. With increasing aridity, recharge estimates can only be made with confidence when several independent techniques are applied and used for cross-calibration (Beekman et al., 1996; Beekman et al., 1999; Scanlon et al., 2002).

Some of the hydrological models that are currently in use for quantifying the impact of climate change on the state of groundwater resources in terms of changing groundwater levels or recharge are:

- EARTH (Gehrels, 1999): a lumped distributed parameter model for site-specific studies,
- CRD-Revised (Xu and van Tonder, 2001): relating water level changes to cumulative rainfall departures for regional studies, and
- Finch's model (2001): a spatially distributed soil water balance model for both site-specific and regional studies.

14.2 Methodology and Approaches

Within the framework of the International Hydrological Programme (IHP) of UNESCO, four approaches are being applied to evaluate the impact of climate change on water resources (Bonell, 1998):

a) using predicted outputs on precipitation and temperature from various General Circulation Model (GCM) scenarios (discussed in Section 14.3) as inputs into selected hydrological models,
b) improving predicted outputs by improving the land-surface parameterisation of GCMs by means of large-scale field experiments,
c) analysing longer term rainfall-runoff time series and separating anthropogenic effects (land use) from climate effects, and
d) adoption of climate change analogues from palaeoclimatic-historical climate research for application in water resources management. Groundwater resources generally contain rich archives of past states and fluxes. Analysis of these archives

can provide insight into changing conditions of climate and hence may be used in forecasting impacts of future climate changes.

Unfortunately, these approaches require considerable amounts of site-specific or regional data, which are not readily available for Southern Africa at this stage. A major obstacle in the application of the first approach is the vastly different scale of operation between GCMs and hydrological models. Currently the resolution of atmospheric models in typical GCMs is about 250 km in the horizontal and 1 km in the vertical (IPCC, 2001). Also GCM outputs usually need to be considered for three or more cells together, since this is typically the minimum skilful scale necessary to describe the smallest wavelengths in the model. Hydrological models, on the other hand, generally operate over catchment areas, which are much smaller than a single GCM cell. If, for example, the GCM predicts rainfall for a particular day, it does not have the resolution to specify where the rainfall occurs in the cell, which makes the information inadequate for recharge estimation. Future climate predictions are currently in the process of being downscaled from the GCM-level to regional level for Southern Africa (AIACC Scenarios Project, B. Hewitson, pers. comm.).

There are also numerous sources of uncertainty in the future climate projections (discussed in Section 14.3.1), which make it difficult to quantify potential climate change impacts on groundwater precisely. Hydrological models, such as the soil water balance model used by Finch (2001) are parameter-intensive and require considerable use of literature analogues and simplifying assumptions when applied to a specific catchment.

The approach of using palaeoclimate analogues or long-term records of past climate-recharge interactions is also difficult to apply in Southern Africa since there are very few (see Selaolo et al., 2003) locations where records of climate and groundwater have been kept in sufficient detail to allow such detailed analysis.

The lack of detailed climate projection data for the region as well as site-specific input data for hydrological models precludes the application of sophisticated modelling approaches at this stage. Instead, we have adopted a simplified approach in the form of a sensitivity analysis testing possible responses to extremes or averaged values on a wider scale, rather than detailed, modelled values for a particular site. The analysis in this paper is based on a simple empirical relationship between mean annual rainfall and recharge values for Southern Africa (See Section 14.5).

14.3 General Circulation Models

General circulation models (GCM) are the modelling tools traditionally used to simulate the global climate system and to generate quantitative predictions of future climatic changes due to human influences. Projections of future climate change are made in response to different scenarios of future radiative forcing agents, such as enhanced concentrations of greenhouse gases and anthropogenic aerosols (IPCC, 2001). Climate predictions benchmarked to “2×CO2”, i.e. when atmospheric carbon dioxide concentrations reach twice the pre-industrial level, have become standard. This is generally estimated to occur at some time from 2050 to 2080. Several GCMs have been able to reproduce the warming trend in the 20th century surface air temperature, when driven by radiative forcing due to increasing greenhouse gases and sulphate aerosols, and to simulate a limited number of palaeoclimates. This increases the confidence in the ability of the models to project future climates.

14.3.1 Uncertainty in GCM output

Uncertainty associated with GCM estimates of future changes in rainfall is much higher than with changes in temperature, since precipitation is dependent on a greater number of climatic variables. Note, however, that the effect of climate change on rainfall amounts and patterns of rainfall, particularly on intensity and increased evaporative losses due to higher temperatures, may be more important in its impacts on groundwater recharge than the rising temperature alone.

Near-future predictions of climate change are not necessarily more accurate than far-future predictions. Human-induced climate changes will occur against a background of natural climate variations that take place on various spatial and temporal scales. Forcings for the near-future are weak in comparison to far-future climate forcings and so the short-term anthropogenic signal is more difficult to separate from the background noise of natural climate variability.

Of particular interest is the outcome of a working group of scientists linked with IHP Phase IV who produced a monograph on the incorporation of climate change and variability into hydrological models (van Dam, 1999). They concluded that outputs from different GCM scenarios gave different (sometimes contradictory) predictions for precipitation, even for the same data sets and that the spatial resolution of GCMs is too coarse for drainage basin studies. Therefore, it seems difficult at present to provide confident estimates of the impacts of future climate change on water resources.

14.3.2 Regional climate information

Modern, coarse-resolution GCMs generally simulate large atmospheric circulation features well, but at the regional scale the models display highly variable area-average biases from region to region and between different models e.g. modelled precipitation averaged over a sub-continental area may vary between -40 and +80% of observed data. Downscaling techniques are used to enhance regional detail based on either (Giorgi and Hewitson, 2001):

- empirical relationships between local climate predictors and GCM output, or
- dynamical regional climate models (RCMs) which improve the spatial detail of simulated climate compared to GCMs.

Daily precipitation and rainfall values for the Southern African region (on a ¼ degree grid cell) obtained from empirical downscaling of output from 3 GCMs for 2 SRES emission scenarios are now available. Dynamical regional modelling estimates on a 30x30 km grid are planned for the end of 2003 (B. Hewitson, pers. comm.).

14.4 Impact of Climate Variability and Climate Change

14.4.1 Climate predictions

Projections of future climate indicate that globally averaged water vapour, evaporation and precipitation are predicted to increase. There appears to be a strong correlation between mean precipitation and the interannual variability, so future increases in precipitation could be accompanied by increases in variability, while areas with less precipitation, will likely have more reliable rainfall. At the regional scale, both increases and decreases in precipitation are expected (IPCC, 2001).

Preliminary patterns emerging from downscaling of GCM modelling over Southern Africa for the next 50 to 80 years indicate that the Western Cape (South Africa) is likely to experience an extended summer with a slight reduction in rainfall (possibly 10%). Rainfall in the Western Cape will likely become more coastally-focussed with less precipitation falling in the interior and more along the coast. Mean temperatures are likely to rise about 2ºC in the coastal areas. In the interior, warming is likely to be more severe, with temperatures in the Karoo rising up to 40% more than the global average which could mean up to 6ºC increase in average temperature. The Karoo could expect a 10 to 20% reduction in rainfall with summer convective rainfall events becoming less frequent, but more intense over the interior. Rainfall in KwaZulu-Natal is likely to increase by up to about 15% coastward of the escarpment along the east coast (B. Hewitson, pers. comm.).

14.4.2 Impact on water resources

The changes in precipitation predicted by GCMs will have a significant impact on the water resources of a semi-arid country such as South Africa. In regions where rainfall amounts and the intensity of rainfall events are expected to increase, more surface runoff may become available. Extreme events are expected to occur more frequently, thus droughts and floods may become more common and more severe, especially in the eastern parts of the country. Changes in the botanical diversity and its distribution (Midgley et al., 2002) resulting from changes in precipitation will also change the spatial and temporal variation in evaporative losses and thus the contribution of rainfall to surface runoff and recharge.

Decreased rainfall amounts over inland areas of the Western Cape and Karoo will decrease the amount of surface water available for consumption. This will place more pressure on the groundwater resources. Already some Western Cape irrigation schemes are only able to supply 80-90% of the planned water to individual farmers. Several Karoo towns are wholly dependant on groundwater, e.g. Beaufort West, Strydenburg and Victoria West, and the combination of increased water demand and decreased rates of recharge to the local aquifers will place these towns at risk of water shortages.

14.4.3 Impact on groundwater recharge

Changes in recharge will result from changes in effective rainfall as well as a change in the timing of the rainfall season (Gleik, 2000). In general, under a scenario of a warming global climate, increasing temperature results in decreasing precipitation over the central continental areas causing decreasing recharge and thus depletion of groundwater resources. Indirect impacts on groundwater resources may also arise from climate change impacts on vegetation and human activities e.g. groundwater abstraction patterns. Rainfall – recharge relationships may be used in a first attempt to assess impacts of climatic change on groundwater resources.

Figure 14.1 shows recharge rates as a function of annual rainfall for Southern Africa based on data from various studies in Botswana (Gieske, 1992; Beekman et al., 1999), South Africa (Bredenkamp et al., 1995) and Zimbabwe (Houston, 1988). These studies were conducted in many disparate locations over the entire subcontinent using different methods.

Comparison of the results revealed that in the area with annual rainfall less than 500 mm/yr, large differences exist between recharge values found. The figure shows that groundwater recharge becomes negligible for rainfall lower than about 400 mm/yr. Note that the annual rainfall threshold value, below which recharge becomes negligible, is related to the recharge estimation method applied: recharge estimated from baseflow separation tends to overestimate the threshold value, whereas the chloride mass balance method tends to underestimate this value.

With regard to assessing the impact of climate change on groundwater resources in Southern Africa, predicted longer-term changes in annual rainfall can be translated into longer-term changes in groundwater recharge using the rainfall-recharge relationship as depicted in Figure 14.1.

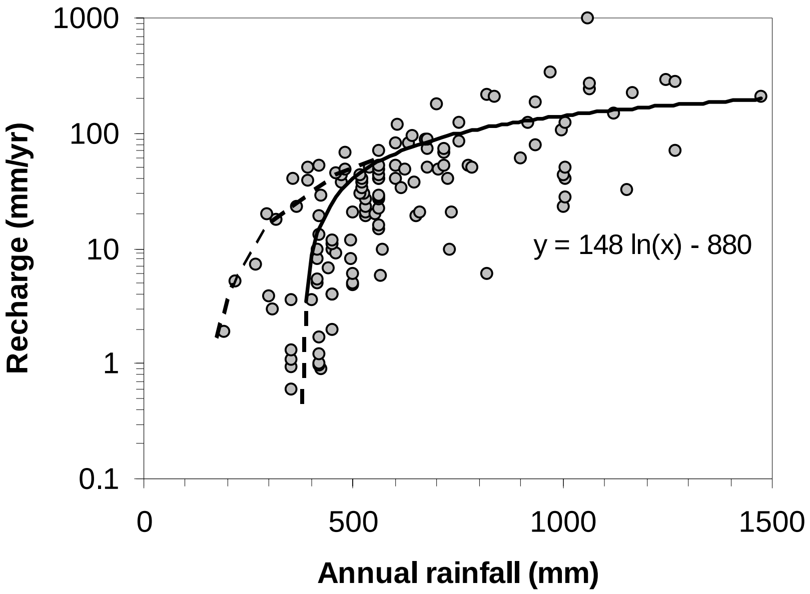

Figure 14.1 Recharge rates in Southern Africa (after Beekman et al., 1996).

The observed rainfall-recharge relationship is used to examine possible trends in groundwater impacts that might be expected if projected extreme changes in mean annual precipitation occur as a response to human-induced climate change. As an illustration, a groundwater-dependent town in the eastern Karoo may currently receive about 500 mm rainfall per year. According to Figure 14.1, about 40 mm/year of recharge should reach the local aquifer. In 50 years time, the mean annual rainfall is likely to have decreased, possibly by as much as 20% i.e. to 400 mm. This would cause a significant decline in the annual groundwater recharge, which may fall below 10 mm per year. Decreased recharge will lead to lowering of the groundwater table and increase both drilling and pumping costs for groundwater users.

Declining rainfall will also affect groundwater storage over the longer term. If 40 mm/year recharge occurs over a 100 km^2 area under the present climate, assuming steady state conditions of groundwater flow, and this value decreases to 10 mm/year once a new equilibrium is established under a scenario of significantly warmer, drier climate, then the volume of groundwater taken into storage on an annual basis will be lower by $3 \times 10^6 m^3$. Assuming a maximum of 50% of the annual recharge can be abstracted on a sustainable basis, the change in groundwater storage represents a loss of 1.5×10^6 m^3 of water resources for the area each year.

Climate change affects not only the volume of rainfall, but also the frequency, duration and intensity of rainfall events, which also have an impact on recharge. Summer convective rainfall events over the Karoo are expected to become less frequent, but more intense. This makes the net impact on groundwater recharge at a particular site difficult to predict. The frequency of rainfall events will affect antecedent soil moisture conditions. Infrequent events are less likely to cause saturated conditions and the drier top soils before the rainfall event may allow increased infiltration but decreased recharge, depending on the soil type and structure. Increased event intensity and duration may have the opposite effect, causing increased surface runoff and recharge.

In general, a decline in groundwater recharge and resources would be expected over the semi-arid and arid regions of Southern Africa under currently accepted climate change scenarios. These changes will require alternative groundwater management practices to control impacts, particularly in situations of groundwater dependency. Artificial groundwater recharge may become a useful technology under these conditions.

On the eastern coast of South Africa, the picture may be quite the opposite, with rainfall predicted to increase by up to about 15%. The Zululand coastal sand aquifer, for example may experience rising groundwater levels and increased groundwater storage, but also the associated problems of increased flooding events. Deterioration of water quality and related health impacts may be a secondary effect.

Coastal aquifers are also vulnerable to the impacts of changes in sea level. A predicted global rise of 0.1 to 0.9 m over the next 100 years combined with high intensity storm surges would increase the risk of seawater intrusion in the coastal areas of the Zululand and Cape Flats aquifers.

14.5 Challenges

Increased knowledge and expertise in climate change and spin-off effects are needed for proper long-term catchment management. Our current concepts and understanding of natural groundwater recharge processes under current conditions also need to be strengthened. Future research should therefore focus on:

- Upscaling physically-based, process-orientated hydrological models from micro- (<10 km^2) to mesoscale (10-10^4 km^2) and/or downscaling of the large spatial resolution (>10^4 km^2) of GCM results to match the two modelling approaches,
- Better understanding water transfer processes as part of land surface parameterisation of GCMs (e.g. soil-vegetation-atmosphere transfer models for recharge simulation coupled with land-use),
- Improved knowledge of stream-aquifer interactions,
- Palaeoclimatic and palaeohydrological reconstructions: e.g. hydrochemical reconstruction of aquifer flushing in the past and present by means of solute transport modelling and reconstruction of palaeoclimatic conditions by means of multiple tracer profiling of unsaturated zones as a key to the future,
- Studying the impact of episodic events on groundwater resources,
- Development and refining of predictive (stochastic) techniques to forecast groundwater recharge (e.g. using EARTH, CRD and GIS, remote-sensing (e.g. microwave), modelling, rating or regionalisation techniques), and

- Remedial actions to mitigate the impacts of climate change, e.g. investigating the buffering potential of artificial groundwater recharge in fractured rock aquifers against climatic variability (floods and droughts).

Large uncertainties still trouble quantitative assessments of climate change and groundwater recharge, neither of which fields could be considered an exact science at this stage. This means that the impacts of climate change on groundwater recharge cannot be predicted with a high level of confidence. What is known with certainty is that the climate is changing, that this will have an effect on water resources and increased efforts will be needed to plan for and manage the impacts of these changes. In the meantime, while our understanding of both climate and recharge processes and their interrelationship improves, good monitoring of current trends, for example trends in groundwater levels and a thorough synthesis of the collected data for Southern Africa is needed.

14.6 Acknowledgements

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15. Groundwater Perspective on Integrated Water Resource Management – Recharge, a Critical Indicator of Sustainability

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ABSTRACT The role of groundwater, with recharge as a critical parameter for determining its sustainable use, is becoming increasingly important in the emerging Integrated Water Resource Management (IWRM) paradigm. Practical examples are given in this paper from the major water law review in South Africa following democratisation in 1994. Emphasis is placed on the importance of recharge assessment within IWRM.

15.1 Introduction

During the past few decades there has been a major change in water management thinking. At the 1977 United Nations Conference at Mar del Plata the focus was squarely on water supply: “All people have the right to have access to drinking water in quantity and quality equal to their basic needs”. At the 1992 International Conference on Water and the Environment in Dublin a new view was endorsed - that of water resources of this planet being finite and fragile and in need of careful management to serve the need of its population. Water is now seen as an economic good and a participatory approach to management of this scarce resource is required, a view that goes beyond the old view that left the provision of services largely up to governments.

In between these two landmark water conferences lay the International Drinking Water Supply and Sanitation Decade with its ambitious objective of meeting everybody's needs for safe drinking water by 1990. Despite the efforts that have been made, this goal has not been reached; a third of the world population does not yet have safe drinking water and some 50 000 deaths occur everyday from waterborne diseases. One of the reasons for this failure is the increasing demand due to population growth, urbanization and industrialization.

Increasing water scarcity, i.e. the available water resources in a country in relation to its growing population, has led to strong international calls for sustainable development of our limited water resources. Many arid and semi-arid countries are already facing a water crisis and the problem is likely to become more serious and will continue well into the 21st century.

These changes are particularly significant for groundwater, because of its growing importance, not only as an essential source of water supply, but also because of its active role in the natural environment. At the same time groundwater has a history of local mismanagement in the form of overdraft and contamination due to ignorance and lack of regulation. Deciding how best to allocate groundwater for use either as a commodity for abstraction or for its in-situ services, especially in the face of externalities such as population growth, economic expansion and climate change, will require hydrologists to broaden their core business to integrate their traditional earth science expertise with disciplines such as economics, policy and regulatory analysis, social and ecological science (Rajone, 2002). This will provide them with a “seat at the table” where decisions are made about the future of the world. At the same time it will give licence for hydrological research to provide a more fundamental understanding of groundwater in a hydrological, ecohydrological and geological context.

15.2 Integrated Water Resource Management

Recognition of the finite and the vulnerable nature of water resources and of the increasing and varied impacts on their utility have led to the adoption of the vision of Integrated Water Resource Management (IWRM). This imperative has been best captured in Chapter 18 of Agenda 21. The often quoted paragraph 18.3 says that "The widespread scarcity, gradual destruction and aggravated pollution of freshwater resources in many world regions, along with the progressive encroachment of incompatible activities, demand integrated water resources planning and management. Such integration must cover all types of interrelated freshwater bodies, including both surface water and groundwater, and duly consider water quantity and quality aspects." According to Agenda 21 an integrated approach to water resources management requires an assessment of the resource, the development of various mechanisms, e.g. forecasting models, economic planning models and methods for water management and planning, and institutional reform in order to achieve sustainable development. The inclusion of environmental needs of water is also emphasised in Chapter 18.

The democratisation in South Africa in 1994 came at an opportune time to incorporate the latest international thinking into its fundamental water policy and law review of 1997 and 1998 respectively. Extensive reference will be made to the South African National Water Act, 1998 (NWA) to illustrate how IWRM is being put into practice.

Integration should be seen within and across three major sub-systems, i.e. the *natural resource system*, the *managed system* and the *institutional system*:

- **Natural resource system**: This should include not only the physical and chemical characteristics of the water itself, but also those components that ensure the functioning of healthy ecosystems, including the plant and animal communities and their habitats. These together give a natural water resource its water and in-situ service on which people depend.
  In the NWA these attributes of a natural water resource are defined as the "resource quality" which is pro-actively managed through the setting of "resource quality objectives."
- **Managed system**: This should include all uses, which impact on the utility of a resource.
  The NWA lays the basis for regulating water use in South Africa, and defines eleven different uses of water, amongst others taking water from a water resource, storing water, waste discharge and disposal and activities which reduce streamflow, i.e. afforestation. Reference is also made to activities outside the control of the NWA, such as land use practices, which can impact, on a water resource.
- **Institutional system**: This is required to ensure effective water governance and entails those social, political and economic organizations and their relationships which are in place to regulate the development and management of water resources at all levels and includes mechanisms, processes and institutions for this purpose.
  The Dublin principles of seeing water as an economic good and requiring a participatory approach for its management are major elements of an institutional system.

Water resources in South Africa are managed according to principles that underpin the National Water Act. The principles outlined below are fundamental tenets of water management in this country (Ministry of Water Affairs and Forestry and DANCED, 2002):

National asset

- All water, wherever it occurs in the water cycle, is a resource common to all, the use of which should be subject to national control. All water should have a consistent status in law, irrespective of where it occurs.
- There shall be no ownership of water but only a right to use it.
- The national government is the custodian of the nation's water resources, as an indivisible national asset, and has ultimate responsibility for and authority over water resource management, the equitable allocation and usage of water, and the transfer of water between catchments and international water matters.

Integrated management

- In a relatively arid country such as South Africa, it is necessary to recognise the unity of the water cycle and the interdependence of its elements, where evaporation, clouds and rainfall are linked to underground water, rivers, lakes, wetlands, estuaries and the sea.
- Water quality and quantity are interdependent and should be managed in an integrated manner, which is consistent with broader environmental management approaches.
- Water resources development and supply activities should also be managed in a manner, which is consistent with broader environmental management approaches.
- While the provision of water services is an activity distinct from the development and management of water resources, water services should be provided in a manner consistent with the goals of water resource management.
- The variable, uneven and unpredictable distribution of water in the water cycle should be acknowledged.

Sustainability

- The objective of managing the quantity, quality and the reliability of the nation's water resources is to achieve optimum long term, environmentally sustainable, social and economic benefit for society from their use.
- The development and management of water resources should be carried out in a manner, which limits to an acceptable level of danger to life and property due to natural or man-made disasters.

The Reserve

- The water required to meet people's basic domestic needs should be reserved.
- The quantity, quality and reliability of water required to maintain the ecological functions on which humans depend should be reserved so that the human use of water does not individually or cumulatively compromise the long-term sustainability of aquatic and associated ecosystems.
- The above-mentioned water required to meet peoples' basic domestic needs and the needs of the environment should be identified as "the Reserve" and should enjoy priority of use.

Equity

- In as far as it is physically possible, water resources should be developed, apportioned and managed in such a manner as to enable all user sectors to gain equitable access to the desired quantity, quality and reliability of water, using conservation and other measures to manage demand where this is required.
- The right of all citizens to have access to basic water services (the provision of potable water supply and the removal and disposal of human excreta and wastewater) necessary to afford

them a healthy environment on an equitable and economically sustainable basis should be supported.
- The location of the water resource in relation to land should not in itself confer preferential rights to usage.

Neighbourliness
- International water resources, specifically shared river systems, should be managed in a manner that will optimise the benefits for all parties in a spirit of mutual cooperation. Those allocations agreed to for downstream countries should be respected.

Institutional framework
- The institutional framework for water management should, as far as possible, be simple, pragmatic and understandable. It should be self-driven, minimise the necessity for state intervention, and should provide for a right of appeal to or review by an independent tribunal in respect of any disputed decision made under the water law.
- Responsibility for the development, apportionment and management of available water resources should, where possible, be delegated to a catchment or regional level in such a manner as to enable interested parties to participate and reach consensus.
- Since many land- uses have a significant impact upon the water cycle, the regulation of land-use should, where appropriate, be used as an instrument to manage water resources.
- Beneficiaries of the water management system should contribute to the cost of its establishment and maintenance.

In summary, the development, apportionment and management of water resources should be carried out using the criteria of public interest, sustainability, equity and efficiency of use in a manner, which reflects the value of water to society while ensuring that basic domestic needs, the requirements of the environment and international obligations are met.

15.3 Groundwater's Changing Role

Groundwater is a renewable and finite resource vital for social and economic development and moreover a valuable component of aquatic ecosystems. Its strategic importance worldwide lies in its role as reliable source of drinking water, both locally and regionally. Its special characteristics also make it ideally suited for various forms of conjunctive use, together with other water resources, thus aiming at an optimal use of the total resource.

Groundwater has been neglected in many countries, largely because of its hidden nature and because of water planners' past focus on bulk water supply. South Africa is a typical example, where the predominant hardrock groundwater systems only contributed about 12% to the total bulk water supply and groundwater was therefore defined in the law as “private water,” i.e. water which is not controlled by government.

With the recent policy shift towards development of the whole population, the recognition of impending national water scarcity and the acceptance of the imperative of sustainable development of available water resources, groundwater's role is changing dramatically in South Africa.

This change was brought about through the democratisation of the country and the 1994 government policy paper which set a minimum domestic water supply standard of 25 l/person/day within a walking distance of 200 m. Follow-up surveys indicated that 12 million, i.e. 30% of the population did not even have this most basic water supply.

Groundwater has played a major role in already reducing the backlog by half and it is recognized that it is contributing between 45-60% to domestic water supply, whereas its contribution to rural areas, where the largest part of the basic water services backlog remains, could potentially be 90%. New experience of hardrock systems show that yields of these systems could be locally increased by orders of magnitude under favourable fracture conditions and with appropriate exploration and development investment. While natural recharge may account for only 5% of total annual rainfall in a typical South African catchment, aquifers typically store 90% of all the water available in a catchment. With a greater understanding of the catchment in three dimensions there is considerable untapped potential as contribution to IWRM (MWAF with DANCED, 2002).

This raised profile of groundwater is reflected in a recent survey of major groundwater management issues in South Africa:

- Groundwater as a strategic resource for community water supply, even in surface water-rich catchments,
- Groundwater as a conjunctive use source throughout the country and particularly in water-stressed areas,
- Groundwater as a driver of sensitive aquatic ecosystems, e.g. coastal lakes, spring systems, river base flow and habitat and wetlands,
- Deep groundwater systems cutting across surface water divides, and
- Need for pro-active protection of groundwater resources, because of widespread pollution impacts; particularly in urban and mining environments, but also in the rural agriculture and settlement situation.

15.4 Groundwater Management as Part of IWRM

The move from a "vicious circle" of groundwater mismanagement to a "virtuous circle" of pro-active groundwater resource management is well illustrated in Figure 15.1 from a series of World Bank groundwater briefing notes (Tuinhof et al., 2002).

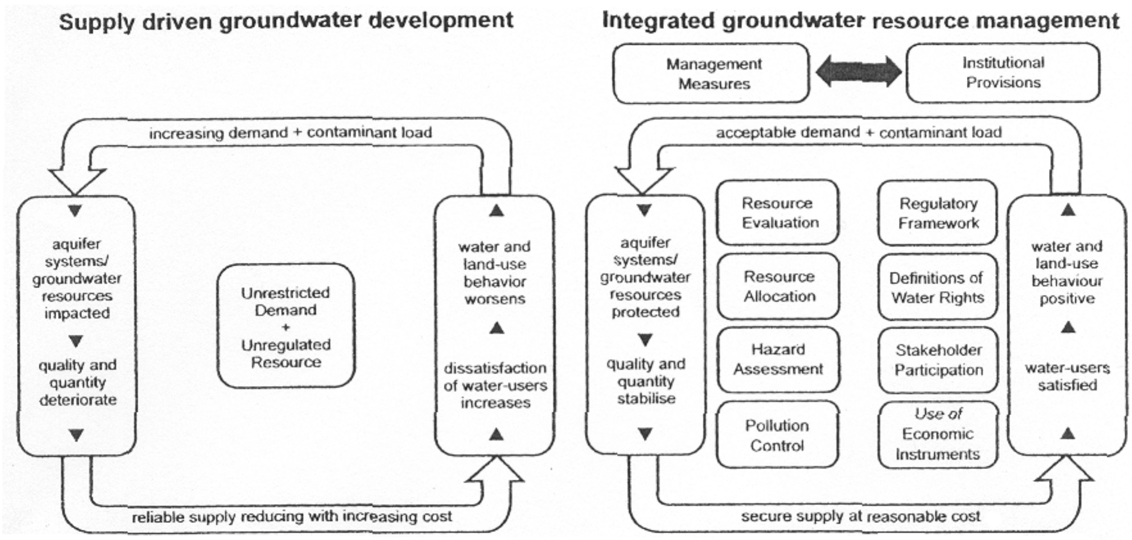

Figure 15.1 Trends in groundwater management (from Tuinhof et al., 2002).

In the IWRM framework it is important to emphasise the need for integration with all components of groundwater management, IWRM and catchment as illustrated in Figure 15.2 (Ministry of Water Affairs and Forestry and DANCED, 2002).

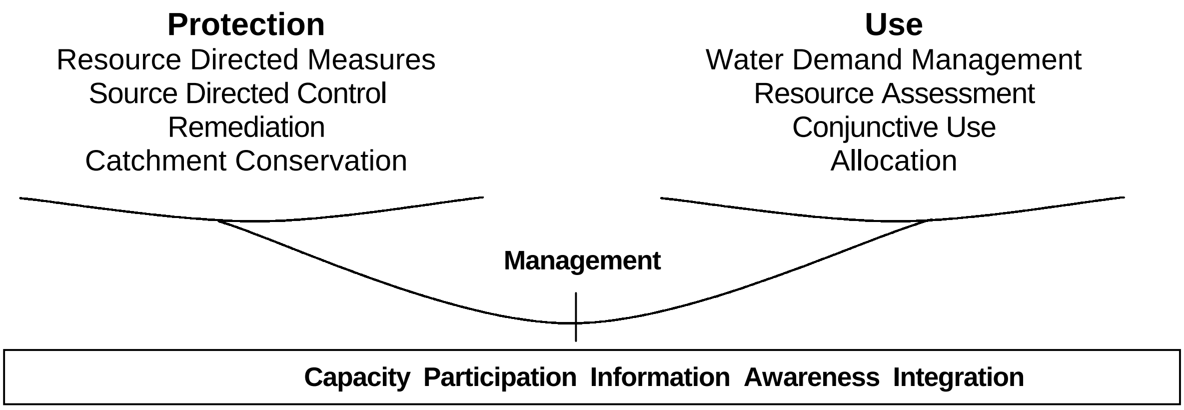

Figure 15.2 IWRM – The Balance of Protection and Use (from MWAF and DANCED, 2002).

The NWA places emphasis on the protection of water resources for their sustainable use. All water use must be authorized and no authorization can be obtained without prior securing of the resource base.

Pro-active protection of water resources is achieved through a novel set of instruments called Resources Directed Measures or RDM, which are specified in the NWA as:

- A national classification system for water resources, including groundwater,
- Determination of a management class for each resource,
- Determination of the "Reserve", which includes the basic human needs reserve and the ecological reserve, which must be determined for all parts of any significant water resource such as rivers, wetlands, lakes and estuaries as well as groundwater. Its purpose is to protect aquatic ecosystems in order to secure ecologically sustainable development and use of the relevant water resources. The Reserve represents the only right to water in terms of the NWA, whereas all other uses require an allocation, which is restricted in time, and
- Resource quality objectives, which operationalize the desired level of protection of a water resource.

Implementation of these measures requires an excellent understanding of the ecohydrological system in order to quantify the Reserve and manage for the resource quality objectives. It should be noted that the total quantity to be reserved in this way is estimated at 20% of the country's water resources and can be as high as 30% in certain areas. During the dry season and drought periods the portion that needs to be reserved can be considerably higher. It is for this reason that the setting of a management class for each resource as "natural", "good" or "fair" must be a highly participative process.

15.5 Institutional Changes

A major purpose of the Act is to achieve the establishment of suitable institutions for appropriate and participative management. This is particularly important for highly localized groundwater resources, which cannot be physically managed centrally. The Act provides three levels of management, namely national government, nineteen catchment management agencies at the new regional management level, and water user associations acting co-operatively at the local level.

All three levels hold tremendous opportunities for improved groundwater management. Water user associations can practice conjunctive use of water resources and artificial recharge of aquifers, put catchment conservation into practice on their land and reverse over-exploitation and degradation of water resources. They can also become an opportunity for community development through the sharing of skills and resources.

The catchment management level with devolved management responsibilities is crucial for addressing groundwater in a planned and integrated way in the public interest. It will also have to provide the essential support for local level management, which had been completely lacking in South Africa for groundwater management.

The national level has for the first time got clear duties defined in the Act, which can be challenged in public, e.g. the national water resource strategy and the need to monitor and provide national information systems. While the authorization of water use will in future be devolved to catchment management agencies, the protection of water resources through the resource-directed measures will remain the responsibility of national government.

The new institutions offer tremendous opportunities for groundwater management. However this role cannot materialize without a major capacity-building and support programme from national government, which should include:

- A regulatory framework within which an equitable allocation of water can occur,
- Specific support for previously disadvantaged groups to establish their water entitlements,
- Appropriate information on groundwater and its beneficial use and protection, and
- Appropriate institutional structures in which WUAs, operating under similar conditions, support each other and are in turn supported by their respective CMAs.

15.6 Information for IWRM

Information will be the cornerstone on which implementation of the new water policy and legislation will rest and on which new institutions can function effectively. New instruments in the Act, like catchment management plans, the reserve and controlling of streamflow reduction activities will be very information intensive. Integrated water resource management will create the need for much greater integration of the relevant information systems, e.g. hydrology, geohydrology, water quality and land type, use and cover.

With the greater focus on the hydrological system in water resource management, it becomes essential to critically address groundwater's place in this system. This was well illustrated in the World Conservation Strategy - Caring for the Earth of 1991:

“Continuously moving above and below the soil surface, water maintains and links the planet's ecosystems.”

The groundwater balance, in its simplest terms, in an unexploited, steady state, can be described by:

RECHARGE - OUTFLOW = 0

However, linked to the flow system are various ecosystems, in the areas of groundwater recharge and discharge as well as in the area of through flow. These systems are supported by groundwater in various ways and / or support the groundwater resource stability. Toth (1999) concluded that groundwater's active role in nature occurs simultaneously at all scales of space and time through its ability to interact with the ambient environment and through the systematized spatial distribution of its flow. To this must be added the emerging field of groundwater ecology (Gibert et al., 1994), which puts particular emphasis on the ecotones. These are the interface zones between different ecosystems, in which ecological processes are much more intense and resources more diversified. It has also become recognised that groundwater fauna, in particular microfauna, bacteria and fungi, play an active role in groundwater system functioning (Marmonier et al., 1997).

The many new participants in a devolved management system will need to share in these information systems which are presently still very centralized. Widespread public participation will require information, which is transparent and user-friendly, and systems, which can be used inter-actively to facilitate multi-user decision-making. Information will also have to stand up in a court of law, given the increasing competition for a scarce resource. The basis for these increased information needs will have to be a more intensive and highly systematic monitoring of water resources status and trends as well as of compliance with policy and regulations at all levels, from national to local.

Different levels of management will have different information requirements:

National government:	Ensuring resource sustainability, equity and international requirements.
Catchment Management Agencies:	Sustainable and equitable allocation of catchment water resources (including water quality management). Integration with land and environment managers within catchment boundaries.
Water User Associations:	Management of common resources within the regulatory constraints.
Water Services Authorities:	Provision of water to end users and managing local impacts of use.

To achieve the necessary direction and co-ordination in this regard, the National Water Act, for the first time, puts major emphasis on monitoring and information. The Minister must establish national systems to facilitate the continued monitoring, recording, assessing and disseminating of information on water resources, addressing the patterns of resource use and the response of the resource to use and management intervention. The crucial link between monitoring and IWRM will be achieved through monitoring programmes focusing on improved resource characterization, such as assessment of groundwater recharge, and on the achievement of resource quality objectives. This resource management-driven information offers clear opportunities for advances in hydrological and ecohydrological system understanding.

15.7 Conclusion

Moving groundwater from a water supply focus to one of water resource system management will require a major rethink from managers as well as groundwater practitioners. It will require working across hydrological disciplines in future monitoring and assessment programmes. It will also require a broadening of hydrologists' traditional earth science expertise with disciplines such as economics, policy, social and ecological sciences.

This is essential in order to assess to what extent we are achieving water resources system sustainability. There can be no doubt, however, that in the context of IWRM, groundwater recharge remains a critical indicator of sustainability, and as such deserves increased attention.

15.8 References

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