GROUNDWATER MONITORING: A CRITICAL EVALUATION OF GROUNDWATER MONITORING IN WATER RESOURCES EVALUATION AND MANAGEMENT

DB Bredenkamp

WRC Report No. 838/1/00

Water
Research
Commission

GROUNDWATER MONITORING : A CRITICAL EVALUATION OF GROUNDWATER MONITORING IN WATER RESOURCES EVALUATION AND MANAGEMENT

Report to the
WATER RESEARCH COMMISSION

by

D B Bredenkamp
Water Resources Evaluation and Management CC

WRC Project No 838/1/00
ISBN 1 86845 600 5

Groundwater Monitoring

Foreword

Monitoring has always been a hydrologist's priority, because he understands the need for reliable and continuous time series of key components of the hydrological cycle. He knows that quantifying, modeling, predicting and managing the hydrological cycle and its interaction with man and the environment is impossible without good data.

Monitoring should therefore be the essence of any national hydrological services. The dilemma is, however that hydrological services throughout the continent of Africa are seriously declining and in some regions totally collapsing. This is for a variety of reasons, in particular declining budgets and obviously the unrest in many areas. This was shown by a World Bank survey of hydrological services in Africa in the mid nineties, initiated because available hydological data were often so inadequate that it did not allow anymore for water project feasibility studies and financing decisions.

In the light of this disturbing trend it is thus highly significant that new legislation in South Africa, i.e. both the National Water Act, 1998 and the Water Act Services Act, 1997, for the first time make explicit provision for national water resources monitoring, assessment and information systems.

This Water Research Commission-funded project with its fresh focus and thinking on groundwater monitoring, and monitoring networks is clearly very timeous. Because of the major momentum generated by the 1water law implementation, it is crucial that the work is taken to a practical stage so that it can help direct planning and implementation of groundwater monitoring country-wide at this moment in time.

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EXECUTIVE SUMMARY

Introduction

Much time, effort and money are devoted to the establishment and maintenance of hydrological monitoring networks. Even more attention is given to elaborate storage and retrieval facilities in sophisticated data banks, and to depicting characteristics of aquifers by GIS image processing, which is often performed with limited or unreliable data. Full appreciation of the importance of good measurements is lacking, which leads to poor motivation of personnel, unreliable data collection, irregular observations, poor control of data quality, and indiscriminate closing of monitoring stations resulting in a loss of valuable information. In other cases, more monitoring stations than necessary are maintained at high cost in limited areas or to study a particular type of aquifer.

The main objectives of the project with the Water Research Commission, entailed

1. an assessment of the spatial and temporal requirements of groundwater monitoring and the use of "patchy" data;
2. an assessment of new methodologies for evaluating groundwater monitoring data - in particular the Cumulative Rainfall Departure method;
3. an assessment of the reliability of data gathered during monitoring programmes;
4. development of a new approach for the characterisation of aquifers in terms of recharge and storativity;
5. development of guidelines on the broad design of a national groundwater monitoring network, with emphasis on:
   - extending the scope/application of monitoring;
   - revising the present groundwater monitoring programmes;
   - compiling the key issues, findings and pespectives into a manual/guide on monitoring;
   - assisting the Department of Water Affairs and Forestry to improve the quality of data on the National Data Base.

Although the study has focussed mainly on monitoring in the RSA, the problems, shortcomings, streamlining and optimisation of monitoring are of world-wide concern. Hence the outcome of this study could have a wide impact and should be of benefit to many developing countries where there is insufficient and often unreliable hydrogeological data, and a need to streamline monitoring.

Methods of analysis

A major component of the study has been the validation of the integrity of groundwater level observations and the extension of limited data series by means of rainfall records, employing the cumulative rainfall departure technique (CRD). This technique is not new but its application as a powerful hydrological tool has been expanded and is illustrated by several examples. The CRD method represents a simple yet useful hydrological model of the groundwater balance, from which the response of water levels to recharge can be determined, also aquifer characteristics and the impact of abstraction. This is implicit in the mathematical relationship of the CRD, namely

$$ {}^{m}_{n}CRD_i = \frac{1}{m}\sum_{j=i-m+1}^{j=i} Rf_j - \frac{k}{n}\sum_{j=i-n+1}^{j=i} Rf_j + {}^{m}_{n}CRD_{i-1} \qquad \text{eq. 1} $$

where m = the short-term memory of the groundwater system
n = the long-term memory of the groundwater system
k = coefficient representing abstraction
= $1+Q/(A \times Rf_{av})$ .... eq. 2
Q = abstraction
A = the areal extent of the aquifer

Derivation of the unit response and aquifer storativity or recharge follows from the linear relationship between groundwater levels and the CRD values:

$$ h_i = a.CRD_i + E \qquad \text{eq. 3} $$

where h_i = the groundwater level for month i
a = proportionality constant, representing response of water levels per millimetre of recharge, which could be derived by regression
E = constant representing the average level around which the water levels fluctuate if the CRD series contains equal positive and negative values.

If the best regression fit indicates that k >1, the water levels without the effect of abstraction (natural levels) could be determined by setting k=1.

From the CRD_{av} with k=1 it follows that:

$$ h_i = a\{(Rf_i - Rf_{av}) + CRD_{i-1}\} + constant \qquad \text{eq. 4} $$

approximating to

$$\Delta h_i = a(Rf_i - Rf_{av})$$

where Δh_i = the incremental change in water level from month i -1 to month i
Rf_{av} = average rainfall

This implies that:

$$\Delta h_i = \frac{b}{S}(Rf_i - Rf_{av}) \qquad \text{eq. 5}$$

where b = a lumped coefficient representing the fraction of rainfall being recharged, and
S = the aquifer storativity

By means of the ${}^{m}_{n}CRD$ technique a reliable reconstruction of the full series of water level fluctuations could be extrapolated from rainfall data which are generally more readily available and longer than water level records. The CRD method can be used not only to asses the integrity of the data but to ascertain the impact of abstraction on the observed water levels. The CRD method clearly indicates that not the full impact of abstraction, converted to millimetres of rainfall, is effected on the groundwater levels. The CRD method appears to be universally applicable and could be applied to aquifers ranging from hard-rock fractured to dolomitic and alluvial systems.

The CRD series mimics groundwater level fluctuations in all climatic regions, indicating that the groundwater balance is determined by the average rainfall, which controls the establishment of vegetation adapted to the rainfall fluctuations. Hence the average rainfall represents the reference value in the CRD series controlling the recharge which is manifested by the water levels fluctuations.

Groundwater levels in relation to the moving average rainfall

As indicated in the manual on recharge (Bredenkamp et al. 1995), the groundwater level of a specific month is also linearly related to the average rainfall over a number of preceding months. This can be represented by the equation:

$$h_i = \frac{\alpha}{n}\sum_{j=i-n+1}^{j=i} Rf_j + C \qquad \text{eq. 6}$$

where α = b/S also the unit response of the aquifer, and C a constant
with b = coefficient of rainfall representing recharge,
n = number of months
and S = aquifer storativity

The impact of abstraction could be incorporated by reducing the rainfall by an equivalent depth of precipitation representing the abstraction, either on a monthly basis if abstraction data is available, or as a constant factor (d in the following equation).

$$h_i = \frac{\alpha}{n} \sum_{j=i-n+1}^{j=i} (Rf_j - d) + F \qquad \text{eq. 7}$$

where **d** represents the abstraction (Q), which is converted to an equivalent depth of precipitation
with d = Q/A,
A = the effective aquifer area over which the abstraction is taking place.
F = the maximum depth of the aquifer for zero average rainfall in the present hydrological equilibrium

Even though the CRD and moving average rainfall represent a simple hydrological model of the groundwater balance of a catchment, it proved to be more reliable and effective than sophisticated models in determining the response of groundwater levels to recharge and abstraction in the case of the Letlhakane aquifer in an area of low rainfall. At the same time the recharge could be quantified more reliably, providing confirmation and a more reliable estimate of recharge than other methods such as tritium profiles and chloride profiles, both producing point estimates of recharge. By means of the CRD and average rainfall model the impact of different rates of abstraction in relation to different rainfall sequences could be simulated for individual boreholes or from an integrated series derived from several boreholes.

Characterisation of aquifers and further applications

By means of the m/n periods of the CRD series the characteristics of different aquifers could be obtained, and parameters such as the unit response (parameter a) could be compared and displayed by GIS maps. If the recharge has been determined in a reliable way or if it could be derived by means of the chloride method, the aquifer storativity could be calculated and represented in a similar way. The average rainfall method could yield an estimate of the maximum depth of the aquifer (F in eq. 7).

Analysis of spring flows

Monitoring of groundwater levels and spring flow is important in estimating the recharge of an aquifer, for example by means of the saturated water budget, by balancing the input and output over a selected time interval Δt according to the water balance equation:

$$Re_{nett} = Q_{abstr} + S\Delta V + Q_{outflow} - Q_{inflow} \quad \text{eq. 8}$$

where Re_{tot} = total recharge: evapotranspiration losses included
Q_{abstr} = pumpage from system or spring flow
$Q_{outflow}$ = lateral flow to lower compartments
Q_{inflow} = lateral inflow from higher aquifers
S = aquifer storativity or specific yield
ΔV = change in saturated storage volume (positive or negative)

The effective recharge, that is, RE_{nett}, represents the nett recharge after evapotranspiration losses from the aquifer have been accounted for. For $\Delta V = 0$ and $Q_{outflow} = Q_{inflow}$ the aquifer storativity does not come into play and the effective recharge is equal to the abstraction over periods starting and ending with the same saturated volume (i.e. equal volume interval). Relating the recharge to rainfall creates a problem in that the equal volume status at any point in time is determined by rainfall from a period prior to the specific month (see eq. 7).

The discharge rates of springs are related to the general piezometric level of the aquifer. Equal flow rates would represent a status of equal-volume and could be used to determine recharge in relation to rainfall. However, both in the case of groundwater hydrographs and springflow, the water-level/flow of a specific month is determined by the average rainfall over the preceding n-months. Therefore recharge determined by means of the equal-volume method has to be related to the rainfall over n-months. Where abstraction has affected the groundwater levels or spring flows, recharge estimated by the equal-volume method would appear to be discrepant - but this inconsistency can serve to confirm that some external factor, such as abstraction, or leakage from municipal water lines has affected the spring flows, or that the rainfall measurements are unreliable.

Chemical monitoring and recharge characterisation

Ancillary measurements are often included in hydrological studies, for example of natural isotopes to trace the flow of water through the unsaturated zone of an aquifer system. Aspects regarding the use of tritium, ^{14}C, and the stable isotopes of hydrogen (deuterium ^{2}H) and oxygen-18 (^{18}O) are discussed, also reasons why some inconsistent and contradictory results have been *obtained from these supportive techniques.*

Attention has also been given to chemical fingerprinting which is best portrayed by springs yielding an integrated sample of groundwater recharged over a long period of time and over a large intake area. Aspects covered in this report include the use of data on the chemistry of natural groundwaters to

- characterize an aquifer in relation to aquifer type or climatic influences;
- establish the degree of mixing of groundwaters differing in chemical composition;
- determine recharge by means of the chloride method, having established that an aquifer is still in its natural state; if it is not, the natural background levels of chloride can sometimes be inferred from the initial measurements.

The reliability of the chloride method hinges on accurate determination of chloride concentrations of rainfall sampled in an area. There are insufficient measurements of chloride in rainfall at chosen localities in the RSA but the concentrations could be estimated from measurements obtained from Botswana, and elsewhere in the RSA. There appears to be a simple relationship between rainfall and the average concentrations of chloride, but more measurements of rainfall chloride are required. Such sampling of rainfall needs to be carried out for a few years and the collection of integrated monthly rainfall samples would be adequate. Logistically such sampling would fit in well with the proposed collection of monthly integrated rainfall samples at selected key monitoring stations.

Monitoring of natural isotopes

Of concern is the discrepancy between ^{14}C concentrations of spring water, mainly dolomitic springs, which rapidly respond to recharge, yet show little evidence of recharge through higher concentrations of ^{14}C which has been introduced into the atmosphere by nuclear tests, and precipitates with rainfall.

The rather constant ^{14}C-concentrations of the springs point to a different mechanism of recharge from that conventionally assumed. The lower concentrations of ^{14}C (closer to 50% modern *carbon) in springs are probably due to rapid infiltration bypassing the biological reservoir and* dissolving carbon from the dolomitic aquifer without feed-back exchange which in normal juvenile groundwater is usually responsible for ^{14}C concentrations of about 78-85%. Therefore the ^{14}C-dating of springs appears to be unreliable.

The chemical interaction and exchange of ^{14}C with the aquifer matrix, especially in dolomitic aquifers, appears to be not fully understood and seems to be related to the turn-over time of the water in the aquifer and the mechanism of recharge. This is revealed by the ^{14}C concentrations of different dolomitic springs, indicating that higher initial ^{14}C concentration are present in the smaller springs with low turn-over times. Only at the Kuruman eye and Grootfontein (Rietvlei, Pretoria) has a clear pulse of bomb ^{14}C appeared, which could be explained in terms of recharge contributed from a dolomitic and non-dolomitic aquifer. Clear evidence has been found that the lower ^{14}C concentrations of some of the larger springs is associated with lower ^{13}C concentrations, indicating that recharge from the last glacial period is present in the spring flows.

In the case of the Sishen dolomitic aquifer the ^{14}C measurements proved to be a valuable regional tracer identifying a major influx of recharge with bomb ^{14}C. The ^{14}C and bicarbonate concentrations indicate that the chemistry of carbonate dissolution is not yet fully understood.

Monitoring of groundwater abstraction

Monitoring of abstraction from groundwater aquifers is one of the key elements in water balance assessments to derive recharge and aquifer storativity (using the SVF or Hill method, for example) and other water balance components, or in modelling the hydrodynamic response of an aquifer. Irrigation from groundwater is one of the more uncertain components but could be inferred from rainfall. Relating measured abstractions to electricity consumption has proved to be a reliable method of estimating pumping. Similarly monthly abstractions to supply Mafikeng and for irrigation show a high correlation with the average rainfall over the 12 months preceding a given month.

Guide on effective monitoring and utilization of data

A more condensed version of the technical report has been prepared as a guide to facilitate more effective monitoring and utilization of available data. It indicates the conditions and criteria to be applied in order to improve existing monitoring networks, or to check the integrity of data, or when extending/infilling existing records. The analyses of about 220 monitoring stations in the RSA have been collated separately, specifically for the benefit of Directorate Geohydrology of the Department of Water Affairs and Forestry.

Conclusions and recommendations

The study has provided greater insight and perspective concerning groundwater monitoring activities and the value of reliable long-term measurements. On the other hand data covering shorter periods can also be utilized effectively in deriving aquifer characteristics. As to the frequency of measurements it appears that monthly observations of water levels and of rainfall are quite adequate for most groundwater interpretations and assessments. Regarding the distribution of monitoring points, a higher density would be required for studies of individual aquifers in order to derive the spatial characteristics of recharge and aquifer storativity, and to serve as reference points for reliable hydrodynamic modelling of the aquifers. In the Grootfontein aquifer in the Bo Molopo area, about twenty boreholes have been monitored, whereas about six to eight well distributed observation points would have been sufficient to model the aquifer. The number of observation points to be used is left to the scientific judgement of hydrogeologists but also depends on the importance of an aquifer and budget constraints. It is recommended that monthly totalling rainfall collectors be installed to reduce the uncertainty of the spatial variability of rainfall. At the same time the water collected could be analysed for chloride to improve the reliability of the chloride input in quantifying recharge.

Fewer monitoring stations would be needed for long-term monitoring on a regional scale. These stations would have to be representative of the different aquifer types and climatic regions but could be reduced to about five per aquifer in any region. In areas of low groundwater productivity, usually the lower rainfall areas, there is no need for a high density of stations.

The CRD and moving average rainfall proved to be of great value in determining the response of water levels in different aquifers, and allow the characteristics of aquifers to be determined in a consistent way. It also enables the effect of abstraction to be identified and natural conditions to be simulated, and provides an easy model with which to asses the impact of abstraction and recharge. In addition the unit response of water levels to recharge, aquifer storativity and maximum depth of an aquifer can be derived and presented on regional scale maps. By means of the moving average method the total depth of the aquifer below surface can be inferred.

The chemistry of groundwater is an important indicator of the hydrological balance and provides better understanding of the contributions from regional aquifers feeding the larger springs. Hence from the chemical data it could be proved that some springs derive their recharge from different aquifers, which is also reflected in the natural isotopic concentrations.

The value of natural isotopes in groundwater studies, especially C-14, is less than is often claimed. This is due to the complexity of the carbon chemistry, and exchange between carbon in the water and carbon in the aquifer, which affect the reliability of C-14 dating of groundwater.

The cumulative rainfall departures (CRD) and average rainfall are also related to surface runoff and could be a valuable indicator of the severity of drought in the different hydrological regimes once the relationships between the CRD /average rainfall and runoff have been established. The long-term records of spring flow reflect the cyclic oscillations in rainfall and provide a better index of rainfall fluctuations because of the larger areal integration of rainfall effecting recharge. Therefore any studies of drought and assessment of its severity should incorporate an examination of spring flows which have been corrected for abstraction ane should be combined with a CRD and cumulative rainfall analysis.

Similarly the hydrological impact of cloud-seeding could best be assessed from the response of springs emanating from fairly small catchments rather than from a dense network of rain gauges or from runoff. The study of springs in areas where limited data is available should incorporate both the chloride concentrations of the spring water and the topographical area delimited as the likely recharge area. This would apply to WRC contract WRC 5/859.

Apart from its contributions to a systematic and effective analysis of existing groundwater level measurements, and characterization of aquifers and spring flows, the study has provided greater insight and perspective on hydrological processes and interactions, and improvements in monitoring groundwater. It has also demonstrated the value of integrating groundwater level monitoring, groundwater chemistry and isotopic measurements, and the possibility of deriving meaningful information even from limited observations. Hence the study will benefit and aid the hydrological fraternity and especially developing countries in establishing effective observation networks for groundwater monitoring and aquifer characterization.

ACKNOWLEDGEMENTS

The contributions of various people by way of references to publications, or consultations are acknowledged. The following contributions merit specific acknowledgement and appreciation:

- Dorothy Venter for many of the routine analyses and the hard work of preparing the report, figures and tables.
- Joan Whitmore, former Director of the Hydrological Research Institute of the Department of Water Affairs, for editing the report, and an evaluation of the presentation.
- Eberhard Braune and the staff of the Directorate of Geohydrology for assistance in supplying data on the National Groundwater Database.
- Prof. Balt Verhagen and Siep Talma for perusal of the chapter on environmental isotopes.
- The Water Research Commission and the following members of the Steering Committee of the project:

K Pietersen	-	Water Research Commission - new Chairman and Project Manager
Dr. Green	-	Water Research Commission - acting Chairman on behalf of A G Reynders
A G Reynders	-	Water Research Commission - late Chairman
Ms A Howman	-	Representative for Director Water Quality Studies of the Department of Water Affairs and Forestry
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M R Meyer	-	Council for Scientific and Industrial Research
Mr L Esprey	-	Independent Council for Forestry Research
Mr. M Simonic	-	Hydromedia Solutions
Minute Secretary	-	Water Research Commission

Special thanks are due to the Water Research Commission for assigning the project to Water Resources Evaluation and Management and for funding the project.

This report will be supplemented by a condensed guide on groundwater monitoring.

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LIST OF SYMBOLS

	A	
A	=	the areal extent of the aquifer (m^2)
a	=	proportionality constant in CRD regression
α	=	proportionality constant in *MA* regression
	B	
b	=	a lumped coefficient of rainfall representing recharge
	C	
^{14}C	=	carbon-14 concentration (*% mod*) - where *mod* = modern carbon
C_t	=	the concentration of radioisotope after a time interval of *t* (*% mc*)
C_0	=	the initial concentration (*% mod*)
$Cl_{groundwater}$	=	chloride concentration of groundwater (*mg/ℓ*)
Cl_{rain}	=	chloride concentration of rainfall (*mg/ℓ*)
CRD_i	=	the cumulative rainfall departure for month i (*mm*)
${}_{av}^{1}CRD$	=	the cumulative rainfall departure relative to the average rainfall for the entire series (*mm*)
${}_{n}^{m}CRD$	=	CRD relationship with short and long-term memory (*mm*)
	D	
d	=	depth of water strike (*m*) below datum
	E	
E	=	regression constant in CRD relationship
E_{vt}	=	evapotranspiration losses (*mm*)
	F	
F	=	regression constant representing depth of aquifer (*m*)
	H	
H	=	the total thickness of the aquifer (*m*)
h_i	=	the groundwater level for month i (*m*)
Δh_i	=	the incremental change in water level from month i -1 to month i (*mm*)
^3H	=	tritium concentrations (*T.U*) - *T.U* = tritium units
^2H	=	deuterium concentrations (*per mil*)
	I	
I	=	the hydraulic gradient
	K	
k	=	coefficient representing abstraction
	=	$1+Q/(A \times Rf_{av})$
	L	
L	=	the width of the inflow or outflow area (*m*)

	M	
MA	=	moving average rainfall (*mm*)
%mod	=	carbon-14 concentrations (in %) relative to modern carbon concentration
m	=	the short-term memory of the groundwater system
	N	
n	=	the long-term memory of the groundwater system
n	=	number of months of moving average
	O	
^{18}O	=	oxygen-18 concentration (*per mil*)
	Q	
Q	=	abstraction (m^3)
Q_{abstr}	=	abstraction (*mm of rainfall*)
Q_{inflow}	=	lateral groundwater inflow (*mm of rainfall*)
$Q_{outflow}$	=	lateral groundwater outflow (*mm of rainfall*)
	R	
RE	=	groundwater recharge (*mm*)
RE (%)	=	percentage recharge
Re_{tot}	=	total recharge (*mm*)
Rf_i	=	rainfall in month i (*mm*)
Rf_{av}	=	average monthly rainfall for entire rainfall record (*mm/month*)
Ro_i	=	surface runoff (*mm*)
	S	
S	=	the aquifer storativity
S_0	=	aquifer porosity at the water level
SMOW	=	standard mean ocean water
	T	
T	=	the respective transmissivities (m^2/d)
t_d	=	age of water at depth d (years)
t	=	the time lapse, also referred to as dating (*years*)
$T_{1/2}$	=	the half-life of the radio-nuclide (tritium or ^{14}C) (*years*)
	V	
ΔV	=	change in saturated volume of aquifer (m^3)
	W	
w	=	pumping represented as an equivalent depth of precipitation (*mm*)
w	=	average width of the aquifer flow cross section

1. **Preamble/Résumé:**

Although monitoring is a vital and integral component of all hydrogeological studies, full appreciation of the importance of reliable measurements is lacking. Much time, effort and money are devoted to the establishment and maintenance of hydrological monitoring networks, and to the collection of data from them. Even more attention is given to elaborate storage and retrieval facilities in sophisticated data banks, and to depicting characteristics of aquifers by GIS image processing, often performed with limited or unreliable data. Hydrological data are essential in providing greater insight into hydrological processes and their responses to the rainfall and climate which are characteristic of the systems, yet often the scope for interpreting the measurements and transforming monitoring data into valuable hydrological information, is not fully realised.

This lack of appreciation of the value and importance of hydrological data often leads to poor motivation of personnel, unreliable data collection, irregular observations, poor control of data quality, and indiscriminate closing of monitoring stations resulting in a loss of valuable information. In other cases the more frequent measurements and more monitoring stations are maintained than can be justified by the cost and limited value of the additional information acquired.

Hydrological measurements are usually instigated as part of an investigative project, and hence in an ad hoc and uncoordinated manner. This is due to poor insight into the significance of the data within the overall framework of hydrological data collection in the long term, hydrological objectives, and the cost of such measurements.

> *This guide presents a critical evaluation of groundwater monitoring and interpretation in a coordinated and structured manner. It incorporates new perspectives so as to broaden insights gained from a previous study of groundwater recharge and aquifer storativity (refer to Bredenkamp et al., 1995), and is a condensed version of a technical report submitted to the Water Research Commission (Bredenkamp, 1999)*

Specific objectives of the guide are to:

1) broaden the perspective on monitoring in relation to groundwater evaluation and management;
2) assess the spatial and temporal requirements of measurements, and the use of scarce data;
3) evaluate new methodologies and extend application of the cumulative rainfall departure

analysis and the moving average rainfall method;
4) examine the integrity of hydrological data;
5) characterize aquifers in terms of their recharge and storativity, using a simplistic approach;
6) extend the scope/applications of groundwater monitoring, and incorporate water quality data;
7) illustrate the application of the techniques to monitoring data pertaining in the RSA, which are stored in the National Groundwater Data Base;
8) identify aspects which require additional research and study.

Although the guide focuses mainly on groundwater monitoring in the RSA, the problems, shortcomings, streamlining and optimisation of monitoring are of world-wide concern. Hence the guide will have a wider impact and should benefit many developing countries where there are rainfall data but insufficient and often inadequate hydrogeological data.

In reviewing the present status of groundwater monitoring in the RSA, the following data elements have been examined and are illustrated by typical examples: groundwater levels, rainfall measurement, springflow, groundwater abstraction, and additional measurements such as natural isotopes and groundwater chemistry. These ancillary measurements are often included in hydrological studies, and are used to trace the progression of water from the time of recharge until it reaches the saturated zone of the aquifer system whereafter it is transported by dynamic flow. Aspects regarding the use of chloride, tritium, ^{14}C, and the stable isotopes of hydrogen (deuterium ^{2}H) and oxygen-18 (^{18}O), and reasons why these supportive techniques have given some inconsistent and contradictory results, are discussed.

In groundwater studies hydrogeological characterisation of aquifers has received high priority in recent years, especially in areas where appropriate groundwater data are lacking. Combined with GIS displays, the hydrogeological features of different aquifers are being portrayed, and different images of the characteristic parameters being overlayed by computer presentations. The present treatise demonstrates how characterisation could be greatly improved if reliable groundwater recharge estimates and aquifer storativity were to be derived using proper monitoring and innovative interpretation of groundwater hydrographs. Specific proposals and guidance on follow-up studies which could be carried out, are provided.

2. Introduction

The scope of hydrological monitoring can be represented by means of the schematic diagram shown in Table 2.1.

A major component of the study has been the validation of the integrity of groundwater level observations and the extension of limited data series by means of rainfall records, employing the cumulative rainfall departure technique (CRD) and moving rainfall average method (MA). These techniques which proved to be similar, are not new and have been included in the manual on recharge (Bredenkamp et al., 1995), but their application as powerful hydrological tools has been expanded and is illustrated by several more examples.

Table 2.1 Schematic diagram indicating the scope and applications of groundwater monitoring.

Purpose / Estimations	**Methods / analysis**	**Monitoring parameters required**
Hydrological modelling	Deterministic models (estimation of recharge)	- rainfall - runoff including base flow - spring flow - groundwater levels - evaporation/evapotranspiration
Groundwater recharge	Water balance methods - unsaturated zone - saturated zone - chemical balance - regression (CRD)	- rainfall . soil moisture . water levels, abstraction, storativity . chloride - rainfall, water levels, storativity
	- isotopic measurements	- rainfall - tritium - carbon-14 - ^{18}O and deuterium
Groundwater management	Hydrodynamic models	- rainfall - recharge - abstraction and springflow - groundwater levels
	CRD simulation	- rainfall - groundwater levels - abstraction

3. Methods of hydrograph analysis

3.1 Cumulative Rainfall Departure (CRD) Method

3.1.1 Normal CRD relationship

Groundwater level monitoring constitutes one of the major components of aquifer studies.
The cumulative departures of rainfall from the average rainfall have been shown to match groundwater level fluctuations fairly well and thus to mimic the hydrological balance of an aquifer based on the natural rainfall occurring in an area (Bredenkamp et al., 1995).

The classical CRD relationship, which has produced good correspondence with groundwater level fluctuations, is expressed by the following equation:

$$ {}_{av}^{1}CRD_i = Rf_i - k.Rf_{av} + {}_{av}^{1}CRD_{i-1} \qquad \text{eq. 3.1.1.1} $$

where	CRD_i	=	the cumulative rainfall departure for month i
	Rf_i	=	rainfall in month i
	Rf_{av}	=	average monthly rainfall for entire rainfall record
	${}_{av}^{1}CRD$	=	the cumulative rainfall departure relative to the average rainfall for the entire series
where	k	=	coefficient representing abstraction
		=	$1+Q/(A \times Rf_{av})$ when pumping occurs .... eq. 3.1.1.2
where	A	=	the areal extent of the aquifer.
and	Q	=	abstraction

If a linear relationship between groundwater levels and the CRD values holds, as it generally does, it can be expressed by the equation:

$$ h_i = a.CRD_i + E \qquad \text{eq. 3.1.1.3} $$

where	h_i	=	the groundwater level for month i
	a	=	proportionality constant which is the water level response to one millimetre of recharge
	E	=	constant, representing the average depth of the groundwater level below surface.

Hence if the best fit indicates that $k > 1$, the natural water levels without the effect of abstraction could be determined by setting $k = 1$.

From eq. 3.1.1.1 for $k = 1$ it follows that:

$$ h_i = a\{(Rf_i - Rf_{av}) + CRD_{i-1}\} + E \qquad \text{eq. 3.1.1.4} $$

as $h_{i-1} = a.CRD_{i-1} + E$

or $\Delta h_i = a(Rf_i - Rf_{av})$

where Δh_i = the incremental change in water level from month i -1 to month i.

This implies that: $\Delta h_i = \frac{b}{S}(Rf_i - Rf_{av})$ .... eq. 3.1.1.5

where b = a lumped coefficient of rainfall representing recharge, and
S = the aquifer storativity

According to eq. 3.1.1.5 the monthly recharge (represented by Δh_i) is effectively a fixed portion (= b) of the rainfall in a specific month, and the losses from the groundwater system are proportional to this coefficient (b) and the long-term average rainfall. The remaining portion of the total rainfall input is dissipated as runoff, evaporative losses from the soil, and transpiration by plants, and is not part of the groundwater balance. Any contribution to groundwater recharge by surface runoff is incorporated in eq. 3.1.1.5 and is not distinguished as a separate recharge component. Positive increases in groundwater levels occur if $Rf_i > Rf_{av}$, whilst water levels would decline according to eq. 3.1.1.5 if $Rf_i < Rf_{av}$. Hence the CRD relationship represented by eq. 3.1.1.4 would mimic the natural response of groundwater levels to rainfall.

As will be indicated later, the CRD series seems to mimic groundwater level fluctuations in all climatic regions, indicating that the groundwater balance is determined by the average rainfall, which controls the establishment of vegetation which adapts to the rainfall fluctuations.

3.1.2 CRD with different short-term and long-term components

In the ${}^{1}_{av}CRD$ relationship (eq. 3.1.1.1) the rainfall over one month is related to the long-term average monthly rainfall. This can be represented as ${}^{short\text{-}term}_{long\text{-}term}CRD$ or ${}^{m}_{n}CRD$. Comparisons with measured water levels have shown that better correspondence is obtained if the short-term rainfall is represented by the average rainfall over a period of m months prior to a specific month, and the long-term average rainfall by the mean rainfall over a longer period (n months). Hence the ${}^{m}_{n}CRD$ relationship becomes:

$${}^{m}_{n}CRD_i = \frac{1}{m}\sum_{j=i-m+1}^{i+1} Rf_j - \frac{k}{n}\sum_{j=i-n+1}^{i+1} Rf_j + {}^{m}_{n}CRD_{i-1} \quad \text{.................. eq. 3.1.2.1}$$

where m = the short-term memory of the groundwater system
n = the long-term memory of the groundwater system
k = coefficient representing abstraction

As will be shown in Section 6 the short-term and long-term memories vary for different aquifers. The short-term memory represents the period over which effective rainfall preceding a specific month determines recharge. The short-term average rainfall could be viewed as averaging the antecedent rainfall which determines the recharge of a specific month, and the carry-over for an extended period. Hence mapping the m and n memories provides a means of comparing the characteristics of the spatial variability of recharge of different groundwater systems. Due to poor areal distribution of groundwater hydrograph and rainfall stations, there are usually not enough monitoring points and rainfall series for reliable comparison of the various types of short-term and long-term memory.

3.2 Hydrographs in relation to the average rainfall

3.2.1 Moving average method

As indicated in the manual on recharge (Bredenkamp et al., 1995) the groundwater level of a specific month is also linearly related to the average rainfall over a number of preceding months. This can be represented by the equation:

$$h_i = \frac{\alpha}{n} \sum_{j=i-n+1}^{j=i} Rf_j + F \qquad \text{eq. 3.2.1.1}$$

where	α	=	b/S is the unit response of the water level to one millimetre recharge
with	b	=	coefficient of rainfall representing recharge
	n	=	number of months
	S	=	aquifer storativity
and	F	=	inferred depth of aquifer below surface

The constant F representing the water level for zero average rainfall is the base level below surface above which the groundwater levels oscillate. F therefore represents the total depth of the aquifer below surface and would be a characteristic level for the aquifer at that specific monitoring point. It should be independent of climatic conditions.

The aquifer storativity can be derived using both the CRD and MA series if the recharge coefficient has been determined. The latter can be derived in different ways, of which the ratio of chloride in rainfall to that of the groundwater provides a simple yet effective and independent estimate of recharge (see Section 11.4).

The high degree of correspondence which can be attained using eq. 3.2.1.1 is indicated for borehole C3N021 in the Bo Molopo dolomitic aquifer (Fig. 3.2.1.1). A correlation coefficient of 0,90 has been achieved between the observed levels and the average rainfall over the previous 24 months. A period of 36 months proved to be the dominant long-term memory of boreholes in the Bo Molopo area when applying the CRD relationship (see Table 3.2.1.1). Application of equation 3.2.1.1 to the monitoring boreholes listed in Fig. 3.2.1.2 yields the values of coefficient α and of the aquifer storativity S assuming the recharge coefficient b = 0,10 (see Table 3.2.1.2).

In the case of the Centurion dolomitic aquifer the same method has been applied but because of fewer measurements and a shorter period of observation the correspondence is not as good as in the case of the Bo Molopo boreholes. However, even with few observations good correlations can be obtained (see Fig. 3.2.1.3 and Fig. 3.2.1.4) for boreholes Dklf24 and Voor245. Their respective values of the unit response (a) and of S derived from the rainfall relationship, using a recharge coefficient derived independently (see Section 5).

The hydrographs of the Rietvlei dolomite and of the Rietondale and Beaufort West aquifers could be analysed in the same way. It appears that the CRD method produces a higher correlation than the MA rainfall relationship if the effect of abstraction is incorporated in both series. The abstraction could be incorporated in the MA by reducing the monthly rainfall by an amount representing the average abstraction (see eq. 3.2.1.2). In this way it is evident that the CRD and average rainfall methods become compatible. In order to convert the abstraction to cub m/d the area of the aquifer would be required, but abstraction could also be expressed as an average value by a regression fit of w:

$$h_i = \frac{\alpha}{n} \sum_{j=i-n+1}^{j=i} (Rf_j - w) + F \qquad \text{eq. 3.2.1.2}$$

where $\alpha = \frac{b}{S}$ and w represents the abstraction (Q), which is converted to an equivalent depth of precipitation abstracted according to the equation:

w	=	Q/A (millimetres),
A	=	the effective aquifer area over which the abstraction occurs.

It is evident that secular variations in rainfall would also be reflected in the parameter w, and would have an effect on the water level response of the aquifer. Records of groundwater levels and springflow extending over many years therefore provide a reliable and quite unique record of climatic fluctuations. This will be discussed in greater detail in Section 13.

Table 3.2.1.1 CRD parameters, correlation coefficients and aquifer characteristics derived by regression between groundwater levels and the CRD series.

Borehole	Farm	*n* Value mths	*m* Value mths	*Δk* Pump	Cor Coef	Unit Response mm	Regres Coef	Rech Coef	Storativity CRD
GN51	Grootfontein	72	9	0	0.710	21.40	0.0214	0.11	0.0051
GN52	Grootfontein	36	1	0	0.870	14.94	0.0149	0.11	0.0074
GN53	Grootfontein	36	1	0	0.800	5.64	0.0056	0.11	0.0195
GN54	Grootfontein	72	9	0	0.720	12.60	0.0126	0.11	0.0087
GN57	Grootfontein	36	1	0	0.830	11.21	0.0112	0.11	0.0098
GN58	Grootfontein	72	9	0	0.920	21.30	0.0213	0.11	0.0052
GN59	Grootfontein	72	9	0	0.805	8.30	0.0083	0.11	0.0133
D4N829	Grootfontein	36	1	0	0.736	8.26	0.0083	0.11	0.0133
MM73	Mooimeisiesftn	36	1	0	0.840	27.74	0.0277	0.11	0.0040
MM75	Mooimeisiesftn	36	1	0	0.750	9.15	0.0091	0.11	0.0120
KL13	Kliplaagte	36	1	0	0.840	13.60	0.0136	0.11	0.0081
LS16	La Rye Stryd	36	9	0	0.590	9.08	0.0091	0.11	0.0121
DT9	Doornplaat	36	1	0	0.860	1.40	0.0014	0.11	0.0786
DN16	Vaalkopje	36	1	0	0.730	-6.60	-0.0066	0.11	-0.0167
DN20	Doornplaat	36	1	0	0.885	15.00	0.0150	0.11	0.0073
VK16	Vaalkopje	36	1	0	0.805	5.24	0.0052	0.11	0.0210
VK18	Vaalkopje	36	1	0	0.809	7.44	0.0074	0.11	0.0148
WF32	Wonderfontein	36	1	0	0.790	4.18	0.0042	0.11	0.0263
WF35	Wonderfontein	36	1	0	0.650	4.62	0.0046	0.11	0.0238
D4N817	Kliplaagte	36	1	0	0.850	7.04	0.0070	0.11	0.0156
MALMANI	Malmani	36	1	0	0.627	-0.88	-0.0009	0.11	-0.1250
VL21	Verlies	36	1	0	0.619	2.52	0.0025	0.11	0.0437
MALMANI	Malmani	72	9	0	0.730	-9.50	-0.0095	0.11	-0.0116
MO9	Malmani	36	1	0	0.770	2.22	0.0022	0.11	0.0495
BULTFTN	Bultfontein	36	1	0	0.770	6.32	0.0063	0.11	0.0174
KAALPLTS	Kaalplaats	36	1	0	0.810	4.68	0.0047	0.11	0.0235
PUTJES	Putjesvlakte	36	1	0	0.710	12.81	0.0128	0.11	0.0086
STRYDFT	Strydfontein	36	1	0	0.850	5.77	0.0058	0.11	0.0191
D4N009	Verlies	36	1	0	0.840	6.50	0.0065	0.11	0.0169
SPRINGV	Springvalley	36	1	0	0.630	10.84	0.0108	0.11	0.0101
EZ18	Elizabeth	72	9	0	0.856	2.86	0.0029	0.11	0.0384
VF110	Valleifontein	36	1	0	0.720	10.07	0.0101	0.11	0.0109
VF109	Valleifontein	72	9	0	0.770	25.60	0.0256	0.11	0.0043
VF113	Valleifontein	72	9	0	0.975	17.95	0.0180	0.11	0.0061
D4N833	Trekdrift	36	1	0	0.800	2.10	0.0021	0.11	0.0524
D4N836	Elizabeth	36	1	0	0.780	2.06	0.0021	0.11	0.0534
D4N075	Grootfontein	36	1	0	0.770	29.97	0.0300	0.11	0.0037
D4N826	Valleifontein	72	9	0	0.790	44.00	0.0440	0.11	0.0025
D4N830	Grootfontein	36	1	0	0.788	12.35	0.0124	0.11	0.0089
D4N831	Grootfontein	36	1	0	0.700	7.30	0.0073	0.11	0.0151
BB32	Blaauwbank	36	1	sim nat	0.835	19.11	0.0191	0.11	0.0058
BB35	Blaauwbank	36	1	sim nat	0.770	12.50	0.0125	0.11	0.0088
BB36	Blaauwbank	36	1	sim nat	0.870	14.50	0.0145	0.11	0.0076
BB39	Blaauwbank	36	1	sim nat	0.830	17.20	0.0172	0.11	0.0064
BB40	Blaauwbank	36	1	sim nat	0.880	14.90	0.0149	0.11	0.0074
BB41	Blaauwbank	72	9	sim nat	0.800	8.00	0.0080	0.11	0.0138
BB42	Blaauwbank	72	9	sim nat	0.830	19.80	0.0198	0.11	0.0056
GN38	Blaauwbank	36	1	sim nat	0.810	12.10	0.0121	0.11	0.0091
GN40	Blaauwbank	36	1	sim nat	0.93	7.50	0.0075	0.11	0.0147
GN41	Blaauwbank	36	1	sim nat	0.73	3.63	0.0036	0.11	0.0303
GN42	Blaauwbank	72	9	sim nat	0.92	25.70	0.0258	0.11	0.0043
								Average	0.0128

Table 3.2.1.2 Aquifer characteristics derived from CRD interpretations for Bo Molopo dolomitic aquifers.

Area	Site	Cor Coef Avg Rf	Period of Avg Rf (months)	C Depth (m)	a Unit Response (m)	w=pump (mm)	S derived from Avg Rf
Bo Molopo	GN51	0.652	24	-21.100	0.178	0	0.01487
Bo Molopo	GN52	0.940	60	-65.300	0.932	0	0.00708
Bo Molopo	GN53	0.922	60	-58.800	0.822	6	0.00803
Bo Molopo	GN54	0.913	60	-70.500	0.913	0	0.00723
Bo Molopo	GN57	0.940	60	-61.900	0.810	2	0.00815
Bo Molopo	GN58	0.850	60	-41.290	0.555	0	0.01189
Bo Molopo	GN59	0.910	60	-46.400	0.623	0	0.01060
Bo Molopo	MM73	0.944	60	-74.000	1.054	0	0.00626
Bo Molopo	MM75	0.733	36	-12.700	0.169	0	0.02345
Bo Molopo	KL13	0.752	60	-31.660	0.190	8	0.03477
Bo Molopo	LS16	0.910	60	-43.100	0.406	8	0.01626
Bo Molopo	DT9	0.740	36	-17.040	0.058	0	0.06851
Bo Molopo	DN16	0.858	60	-24.870	0.378	8	0.01746
Bo Molopo	DN20	0.860	60	-33.250	0.485	8	0.01361
Bo Molopo	VK16	0.912	60	-39.000	0.653	2	0.01011
Bo Molopo	VK18	0.860	60	-23.270	0.238	0	0.02775
Bo Molopo	WF32	0.880	60	-54.800	0.332	2	0.01986
Bo Molopo	WF35	0.795	60	-49.460	0.186	8	0.03556
Bo Molopo	D4N817	0.713	24	-35.140	0.126	0	0.02099
Bo Molopo	VL21	0.885	60	-56.040	0.607	2	0.01088
Bo Molopo	MO9	0.790	60	-22.000	0.048	8	0.13895
Bo Molopo	BULTFTN	0.910	60	-17.260	0.258	0	0.02559
Bo Molopo	KAALPLTS	0.756	36	-18.540	0.083	0	0.04794
Bo Molopo	D4N009	0.677	60	-29.180	0.474	2	0.01394
Bo Molopo	EZ18	0.880	60	-55.200	0.597	2	0.01106
Bo Molopo	VF110	0.800	24	-19.180	0.212	0	0.01248
Bo Molopo	VF109	0.770	36	-53.100	0.780	0	0.00508
Bo Molopo	VF113	0.930	60	-46.000	0.683	0	0.00966
Bo Molopo	D4N833	0.600	24	-10.800	0.029	0	0.09199
Bo Molopo	D4N836	0.610	36	-14.260	0.033	0	0.12000
Bo Molopo	D4N826	0.620	24	-39.600	0.610	0	0.00433
Bo Molopo	D4N830	0.632	48	-27.060	0.117	0	0.04513
Bo Molopo	D4N831	0.690	48	-24.760	0.154	0	0.03429
Slurry	BB32	0.599	24	-36.427	0.234	0	0.01127
Slurry	BB35	0.815	60	-54.632	0.634	0	0.01041
Slurry	BB36	0.720	60	-32.394	0.217	0	0.03040
Slurry	BB39	0.734	48	-43.763	0.315	8	0.01678
Slurry	BB40	0.845	60	-57.796	0.649	8	0.01017
Slurry	BB41	0.854	60	-18.117	0.236	0	0.02792
Slurry	BB42	0.913	60	-72.124	0.868	0	0.00760
Slurry	GN38	0.939	48	-60.389	0.750	0	0.00704

3.3 Impact of abstraction/artificial recharge on hydrographs

3.3.1 CRD interpretations

Greater insight and perspective on the impact of abstraction and artificial recharge on the groundwater balance can be obtained from the CRD relationship.

Considering eq. 3.1.1.3:

$$h_i = a.CRD_i + E$$

It could be rewritten as

$$h_i = a\ [\ \frac{1}{m}\sum_{j=i-m+1}^{j=i} Rf_j\ -\ \frac{k}{n}\sum_{j=i-n+1}^{j=i} Rf_j\ +\ {}_n^mCRD_{i-1}\ ] + Constant \qquad \text{eq. 3.3.1.1}$$

which could be split up into:

$$h_i = a\ [\ \frac{1}{m}\sum_{j=i-m+1}^{j=i} Rf_j\ -\ \frac{1+\Delta k}{n}\sum_{j=i-n+1}^{j=i} Rf_j\ +\ {}_n^mCRD_{i-1}\ ] + Constant$$

where $$h_i = a\ [\ \frac{1}{m}\sum_{j=i-m+1}^{j=i} Rf_j\ -\ \Delta Rf\ -\ \frac{1}{n}\sum_{j=i-n+1}^{j=i} Rf_j\ +\ {}_n^mCRD_{i-1}\ ] + Constant$$

and $$\Delta Rf = \frac{\Delta k}{n}\sum_{j=i-n+1}^{j=i} Rf_j$$

It turns out that the effect of abstraction is to decrease the input rainfall in the CRD simulation model, so that only a fraction a is manifested in the water level response. This implies that the system compensates for abstraction with a reduced effect on groundwater levels. This is regarded as a very significant result, and in the case of artificial recharge (negative abstraction) its effect is similar to an increase in rainfall of which only part would be manifested as recharge. It means that in the case of abstraction, effective water level decline is decreased due either to higher lateral inflow to the system, or to lower evapotranspiration. On the other hand, considering that the losses are higher for higher rainfall i.e. for shallower groundwater levels, artificial recharge would obviously be more effective if applied after depletion of the aquifer by pumping, but the rate of losses from the system would be increased as the water levels rise.

It is also evident that recharge from a given amount of rainfall would be more effective after a prolonged drought because of the lower losses related to a lower long-term memory (n-months preceding). This explains part of the dramatic recovery of water levels and spring flows after severe droughts effected by prolonged higher than average rainfall, resulting in lower evapotranspiration and abstraction when good rains occur.

According to the moving average relationship (eq. 3.2.1.2) the response of abstraction or artificial recharge on the groundwater level is also manifested by the fraction (= a) of the equivalent depth of precipitation (w) abstraction represents.

Therefore both the CRD and MA simulations will show a diminished water level response to abstraction, which in hydrodynamic models is effected by boundary inflow. The CRD and MA relationships, despite being "regression models", represent simple hydrological models and allow for simplistic and reliable simulation of groundwater levels. Further investigation of the simplistic approach is required for simulation of the piezometric response of individual boreholes, incorporating also interference between boreholes.

4. Analysis of spring flow data

4.1 Introduction

A standard and frequent monitoring activity is the measurement of spring flow. The springs indicated in Fig. 4.1.1 have been gauged in the RSA. In the case of the Grootfontein spring (near Pretoria) a complete record of the flow is available from before 1930. For Maloney's eye, and the Pretoria Fountains and Sterkfontein eye, flow records are also available since about 1931.

Several of the spring flow records are intermittent and in some cases the reliability of the data is uncertain, either because of poor measurements or because flows have been affected by abstraction. Only a few records of non-dolomitic springs are available in the RSA. A valuable series is that of the Uitenhage springs, but abstraction of groundwater and the drilling of artesian boreholes in its catchment have affected the flows. By means of the CRD method the natural flow record could be reconstructed (see Section 7).

4.2 Characteristics of spring flow

As springs provide a visual, measurable indication of the response of groundwater recharge to rainfall fluctuations, an examination of the flow records reveals the characteristics of the recharge and of the aquifers sustaining their flows. In the following section the procedures of spring flow analysis are discussed.

4.2.1 Flow of springs in relation to rainfall

4.2.1.1 Introduction

In South Africa the flow of dolomitic springs in relation to rainfall has been investigated by Temperley (1978), Fleisher (1981), Bredenkamp and Zwarts (1987) and Bredenkamp et al. (1992). Temperley indicated that the discharge of dolomitic springs fluctuates according to the cumulative departure of rainfall from the average ($^{1}_{av}CRD$). Bredenkamp et al. (1992) have shown that the flow of springs in the Bo Molopo area corresponds even better to the $^{m}_{n}CRD$ relationship and to the water level of the Wondergat because of similarities in rainfall distribution in the summer rainfall area (see Figures 4.2.1.1.1 to 4.2.1.1.5 for Olievendraai, Schoonspruit, Malmani, Stinkhoutboom and Dinokana eyes).

4.2.1.2 Spring flows in relation to the Cumulative Rainfall Departures (CRD) from the long-term average rainfall

Mathematically the relationship between spring flows ($Q_{i\ spring}$) of a specific month and the cumulative departures of rainfall from the average can be explained according to the simple Darcy equation:

$$Q_{i\ spring} = a.T.w\{h_{i\ av} - h_o\} \quad \text{eq. 4.2.1}$$
$$= b.h_{i\ av} - g$$

where

where	$h_{i\ av}$	=	integrated water level (mamsl) in the aquifer, for a specific month i. This level is linearly related to $h_{i\ (x,y)}$ i.e. water levels at any point x,y in the aquifer.
	b	=	$a.T.w$
	a	=	proportionality constant
	w	=	average width of the flow cross section
	T	=	average transmissivity of the aquifer
	h_o	=	the overflow height at the spring (mamsl)
	g	=	$a.T.w.h_o$

The groundwater levels of an aquifer at specific monitoring points are interdependent due to the similarities in the average rainfall, and spatial equalization of groundwater levels effected by interconnected fissures of higher transmissivity.

As $^{1}_{av}CRD$ represents a good simulator of groundwater levels

	$h_{av}(i)$	=	$d.CRD_i + C$
where	C	=	depth of the saturated aquifer from surface
	d	=	regression coefficient
	i	=	month i
and	$Q_{i\ spring}$	=	$z\{(Rf_i - Rf_{av}) + CRD_{i-1}\} + C_{flow}$
where	z	=	regression coefficient
	C_{flow}	=	long-term average flow of spring around which the flow fluctuates

	Rf	=	rainfall
	Rf_{av}	=	long-term average rainfall

hence $Q_{i\,spring} = \frac{J}{S}.\rho.CRD_i + C_{flow}$ .... eq. 4.2.2

where	ρ	=	coefficient of rainfall representing recharge
	S	=	aquifer storativity
	J	=	a constant incorporating the hydraulic coefficient and the width of the flow cross section delivering water to the spring

The correspondence between groundwater fluctuations at a single point and spring flow, can be demonstrated in the case of the Lower Fountain (Fig. 4.2.1.2.1), and Fig. 4.2.1.2.2 (Upper Fountain) and Fig. 4.2.1.2.3 (Sterkfontein eye). This confirms that there is spatial interdependence between piezometric levels measured at individual points in an aquifer. A remarkable degree of correspondence in both the finer oscillations and the general trend can be seen in the case of the Upper an Lower Fountains. In the case of the Sterkfontein eye the correspondence is not as good because the eye falls in a different compartment (Bredenkamp, 1999) which has been affected by abstraction. A graph of the Sterkfontein eye in relation to the hydrograph of Dklf28 reveals a poor match, showing unreliable water level measurements (see Fig. 4.2.1.2.4). A similar graph for the total flow of the Fountains in relation to the water levels of Dklf24 reveals a reasonable match, based on relatively few water level observations (Fig. 4.2.1.2.5). Typically for the Pretoria area small fluctuations of the water level show a large response in the spring flows indicating that highly transmissive fractures control the discharge. This is corroborated by the synchronous behaviour of the Fountains springs and Sterkfontein spring which exhibit pressure responses resembling those of a confined aquifer system. The pressure response in the highly transmissive fractures, according to a pumping test rather interestingly, reveals that the water level in the pumping borehole increases when pumping starts, indicating that the reduction of the pressure at the borehole causes an elastic compaction of the aquifer which results in higher water levels just after pumping starts. However, the effect gradually diminishes as the pressure in the fractures equalises with the average pressures (see Fig. 4.2.1.2.6).

4.2.1.3 Spring flow in relation to moving average rainfall

As has been demonstrated in the manual on recharge (Bredenkamp et al., 1995) the flow of a spring also corresponds surprisingly well to the average rainfall over several years prior to a specific month. This period conforms to the long-term memory in the CRD equation (n), yielding the relationship:

$$Q(i)_{spring} = p \cdot \sum_{j=i-n+1}^{j=i} Rf(j) + Q_{constant} \qquad \text{eq. 4.2.1.3.1}$$

where

where	p	=	$b.A$
	b	=	fraction of rainfall constituting recharge
	A	=	area of the aquifer
	$Q_{constant}$	=	constant flow if the average rainfall term is zero

Eq. 4.2.1.3.1 indicates a linear response between the spring flow and the average rainfall over the characteristic period which implies a characteristic cut-off average rainfall value for zero flow, depending on the average rainfall determining the hydrological balance, which controls the losses from the system. Zero average rainfall would therefore yield a negative flow equal to the average losses from the system for the prevailing climate. The losses would however adapt as the average rainfall fluctuates, yielding the new equilibrium ineffective (cut-off) rainfall at which the spring would apparently stop flowing.

4.2.1.4 Rainfall/recharge relationship

Based on eq. 4.2.1.3.1 the relationship between rainfall and recharge could be derived from spring flow records as in the case of the groundwater levels; unreliable spring flow records could be corrected for abstraction and the flows of several springs in relation to the average monthly rainfall could be plotted (see Fig. 4.2.1.4.1 for the Sterkfontein eye, and Fig. 4.2.1.4.2 for the Uitenhage springs which have been corrected for abstraction). The x-y plots indicate a linear relationship although points are scattered. However the flow of Buffelshoek eye shows an exponential response to the long-term average rainfall (see Fig. 4.2.1.4.3).

The cut-off average rainfall for zero flow represents an equivalent CRD or MA value, which explains why the relationship with the Wondergat shows zero spring flow if the water level in the sinkhole were to decline to a certain level. For springs in the Bo Molopo area this level is about 1411 mamsl (see Fig. 4.2.1.4.4), but it could be different for other areas depending on their hydrological equilibrium in relation to that of the Bo Molopo area (see Fig. 4.2.1.4.5). However, a linear extrapolation is only valid if the natural long-term hydrological equilibrium prevails. As this controls average natural losses from the system, these losses would diminish as the average rainfall decreases because the vegetation adapts to lower rainfall, which results in the occurrence of recharge even in arid areas.

Fig. 4.2.1.4.6 shows the total flow of the Pretoria Fountains plotted against the moving average rainfall over 120 months. The higher flows during the 1983 - 1993 period appear to be due either to recharge from leaky municipal pipelines or unreliable measurements, as will be indicated by the equal-volume analysis (see Section 8).

5. Application of the CRD and MA methods

5.1 Integrity, extrapolation and patching of hydrographic data

Groundwater assessment is constantly hampered by poor measurements or loss of data due to recorder failure or infrequent observations. By means of the ${}_{n}^{m}CRD$ and MA techniques a fairly reliable reconstruction of water level fluctuations can be generated from rainfall data which are generally more readily available. Likewise the reliability of groundwater level observations can be verified by means of these methods which appear to be of universal applicability and could be applied to aquifers ranging from fractured hard-rock to dolomitic and alluvial systems.

Groundwater level data can be checked by the CRD and MA series of a nearby rainfall station. This is especially useful in the case of data outliers and where water level responses differ from those indicated by the CRD series, e.g.

- i) water levels dropping instead of rising;
- ii) occurrence of spurious spikes;
- iii) water levels affected by abstraction, which require the value of k in eq. 3.1.2.1 and w in eq. 3.2.1.2 to be determined.

5.2 Utilizing historical data

In many instances regular groundwater level monitoring started long after boreholes had been drilled, and often only a few intermittent measurements were made. By means of the CRD method water levels measured at the time of drilling or at irregular intervals thereafter, can be compared to a series measured much later (see Fig. 5.2.1 for D5N519 at Williston). If good consistency is obtained, the likelihood is high that the reconstructed data series for periods of missing data are reliable; if they are not, corrections for pumping might have to be introduced.

The CRD and MA methods can be applied to all aquifers but the reliability of the water levels simulated in this way will depend primarily on the following:-

- i) the reliability of the rainfall record and the degree to which it is representative of a specific locality, as non-representative rainfall could result in a poor match between the CRD series and water levels;
- ii) the degree to which averaging rainfall of different stations reduces the correspondence between rainfall and groundwater levels at a specific locality;
- iii) the lag time between the recharge and the response of the groundwater levels - see Section 7.2.2 Orapa-Letlhakane.

iv) whether abstraction is known, or a constant average rate is substituted.

A high degree of spatial interdependence between groundwater levels in an area results from 1) the homogeneity of monthly rainfall and 2) hydraulic connection via large fractures or dissolution channels in dolomitic aquifers. In the latter case especially, differences in groundwater level responses due to spatial inhomogeneities of recharge and aquifer storativity are evened out. This causes both storativity and recharge derived from a single borehole, to be biased towards the average values of the system. However, the short-term response of water levels at a specific point would reflect local aquifer characteristics. An analysis of hydrographs from the Bo Molopo (Grootfontein area) illustrates application of the method (see Section 7.2).

The fact that CRD and MA analyses have shown good correspondence with different monitoring stations in the Grootfontein aquifer, even though rainfall data from only a single station, namely Slurry, was used, implies that the variability of monthly rainfall in the area is largely homogenized in spite of large variability of daily rainfall. Logically, disregarding influences such as topography and aspect, the spatial variability in rainfall stems largely from its temporal variability, which is determined by stochastic variability. Even in the case of a dolomitic aquifer recharging rapidly, the short-term memory (*m*) of water level responses of one month yields good simulations. Hence use of daily rainfall is essentially redundant, as monthly rainfall totals corresponding with groundwater level measured at the same time, would be adequate for the analysis of most groundwater reactions. However to obtain more representative rainfall data and limit the spatial variability, it is recommended that monthly totalling rainfall collectors be installed at each borehole in a monitoring network. Not only is it practical but it would reduce uncertainties regarding the representativeness of rainfall, and would also help to overcome the general scarcity of rainfall data for other hydrological purposes e.g. runoff assessments.

5.3 Application to Bo Molopo area

5.3.1 Rainfall

The rainfall at stations in the Bo Molopo region including some distant rainfall stations e.g. Pretoria, Kuruman, Ventersdorp have been correlated. The correlation coefficients listed in Table 5.3.1.1, have been plotted as a function of distance between Slurry and the other stations. It is granted that other factors such as topography and geographical/climatic factors also affect the rainfall correspondence between stations. As expected the closer the rainfall stations are to each other, the higher are the correlation coefficients between them. The correlation coefficients decrease exponentially at close range, but then linearly as the separation distance increases (see Fig. 5.3.1.1). This indicates the uncertainty and unreliability associated with using a rainfall record remote from a monitoring point.

Table 5.3.1.1 Correlation coefficients of rainfall between stations at different localities

Station		Rietvlei Pretoria	Waterklf Country	Irene	Pretoria Lyttelton	Rietondale	Zuurbekom	Rustenburg	Ventersdorp Kafferskraal	Coligne	Lichtenburg Manana	Lichtenburg TNK	Rooirandjies-fontein	Bo Molopo Slurry	Kuruman	Wonderwerk
		5135310	5134373	5133822	5133507	513404	4755288	5115234	473025	472/560	472/455	472/279	472/175	508/649	393778	358/049
		1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Correlation coefficient (Rsq)																
5135310	1															
5134373	2	0.856														
5133822	3	0.897	0.830													
5133507	4	0.836	0.839	0.887												
513404	5	0.794	0.792	0.792	0.791											
4755288	6	0.672	0.615	0.692	0.649	0.600										
5115234	7	0.697	0.655	0.676	0.675	0.640	0.627									
473025	8	0.553	0.525	0.522	0.483	0.484	0.546	0.495								
472/560	9	0.520	0.488	0.493	0.455	0.423	0.538	0.539	0.659							
472/455	10	0.562	0.518	0.545	0.569	0.527	0.601	0.562	0.705	0.733						
472/279	11	0.582	0.551	0.559	0.527	0.518	0.620	0.576	0.649	0.715	0.856					
472/175	12	0.441	0.397	0.419	0.380	0.348	0.474	0.467	0.589	0.736	0.640	0.693				
508/649	13	0.564	0.537	0.558	0.577	0.493	0.526	0.542	0.586	0.689	0.688	0.714	0.632			
393778	14	0.201	0.153	0.163	0.200	0.133	0.165	0.231	0.232	0.306	0.234	0.240	0.263	0.235		
358/049	15	0.165	0.183	0.173	0.207	0.180	0.166	0.235	0.185	0.234	0.203	0.231	0.211	0.195	0.109	
Correlation coefficient (Rsq)		3 months lag														
5135310	1															
5134373	2	0.904														
5133822	3	0.917	0.890													
5133507	4	0.863	0.885	0.928												
513404	5	0.894	0.872	0.886	0.880											
4755288	6	0.791	0.763	0.796	0.732	0.758										
5115234	7	0.806	0.777	0.767	0.745	0.776	0.766									
473025	8	0.710	0.713	0.682	0.642	0.693	0.729	0.695								
472/560	9	0.689	0.657	0.638	0.609	0.635	0.745	0.748	0.791							
472/455	10	0.705	0.667	0.679	0.702	0.697	0.770	0.765	0.827	0.841						
472/279	11	0.711	0.694	0.670	0.620	0.664	0.800	0.756	0.772	0.844	0.916					
472/175	12	0.594	0.571	0.547	0.485	0.514	0.636	0.657	0.682	0.830	0.742	0.813				
508/649	13	0.708	0.692	0.707	0.703	0.693	0.708	0.748	0.733	0.830	0.804	0.838	0.764			
393778	14	0.304	0.272	0.240	0.250	0.228	0.283	0.347	0.346	0.443	0.382	0.408	0.401	0.358		
358/049	15	0.405	0.420	0.385	0.410	0.417	0.351	0.450	0.499	0.547	0.460	0.454	0.481	0.472	0.311	

5.3.2 Groundwater levels

Groundwater levels in the Bo Molopo dolomitic area have been monitored for many years and are fundamental to the present appraisal of different techniques of assessing the reliability of data and deriving characteristics of an aquifer. A good spatial distribution of monitoring points allows the spatial variation of storativity or recharge of this important aquifer to be determined by the CRD and MA methods.

The different monitoring points in the Bo Molopo area (see Fig. 3.2.1.2) have been analysed. Correspondence between the CRD series and groundwater levels monitored in the Bo Molopo aquifers has been examined, and the short-term and long-term memories (*m* and *n*-values) and the *k* factors indicating natural conditions or the effect of abstraction have been determined. Similarly the long-term series of water level fluctuations in the Wondergat close to Slurry has been examined. The Wondergat data series is one of the most unique in the RSA, representing water level changes since 1922. It has been established that the Wondergat water level series not only represents groundwater responses and spring flow in its immediate vicinity (see Section 4.2), but appears to be linearly related to the water level fluctuations of the larger Bo Molopo area (Bredenkamp et al., 1995) and even corresponds to water levels and spring flow in dolomitic aquifers in the Pretoria region (Bredenkamp, 1996) situated more than 300 km from the Wondergat (see Fig. 4.2.1.4.5). This does not imply that there is a connection between the Wondergat and these aquifers, but merely reflects the similarity in the temporal variability of rainfall as the key variable determining recharge (Bredenkamp et al., 1995).

Initially it was assumed that the Wondergat water level series represented natural/unaffected fluctuations of groundwater in the Bo Molopo area, because the Wondergat falls within a dolomitic compartment with little abstraction and assumed impermeable boundaries in the form of diabase dykes. *This assumption was supported by the fact that the Wondergat water levels* show good correspondence with the flow of springs in the vicinity where little or no abstraction occurs, e.g. Buffelshoek eye (see Fig. 5.3.2.1). In fact, the good relationship between the Wondergat levels and the flow of Buffelshoek eye allowed the water levels of the Wondergat at the time it overflowed to be inferred.

However on analysing the flow of the Molopo eye in relation to the CRD series (Fig. 5.3.2.2) it appeared that the flow had been affected by abstraction. Different *k*-values pertaining to various parts of the CRD series had to be introduced to obtain a good simulation (see Fig. 5.3.2.3). The natural flow of the spring could be simulated by resetting $k=1$ in the CRD relationship (Fig. 5.3.2.2 and 5.3.2.3), indicating that the flow of Molopo eye had been reduced by more than 50% of its simulated natural flow in 1989. The good correspondence between the Wondergat levels and the flow of the Molopo eye (see Fig. 5.3.2.4) indicates that the levels of the Wondergat have also been affected by abstraction. As in the case of the Molopo eye, different *k*-values pertaining to different periods had to be applied in simulating the observed water levels of the Wondergat (see Fig. 5.3.2.4). The k-values <1 may indicate periods of high recharge or the combined effect of both recharge and reduced abstraction, or could represent a pressure response from afar. From the similarity between the water level of the Wondergat and spring flow responses over a large area it is concluded that 1) the dolomitic aquifer behaves as a confined system and 2) the diabase dykes are not impermeable boundaries as originally assumed.

It appears that a reliable and representative rainfall series is the independent variable needed to reconstruct the natural response of an aquifer. It was shown to be justifiable to use the monthly rainfall data of Slurry in analysing the groundwater hydrographs of monitoring boreholes in the Bo Molopo area. A CRD analysis of the boreholes shown in Fig. 3.2.1.2 has been performed (see Table 3.2.1.1). The degree of correspondence between measured and simulated values attained for some monitoring boreholes is revealed by the following figures:-

BB35 (Fig. 5.3.2.5): A striking similarity between measured water levels and values simulated by the ${}_{36}^{1}CRD$ series exists up to 1985. Since then water levels have declined due to increased abstraction.

GN41 (Fig. 5.3.2.6): This borehole is situated closer to the Grootfontein eye where large abstraction for Mafikeng is taking place. The effect of increased abstraction after the 1983 drought can clearly be seen and only partial recovery of water levels occurred after the high rainfall of 1988, before the decline in water levels resumed. For this hydrograph the ${}_{72}^{9}CRD$ series also produced the best simulation over the initial period when the effect of pumping was minimal. From the CRD analogy an S-value of 0,03 is derived which conforms to higher leaching of the dolomite in the vicinity of the eye.

5.3.3 Deriving the aquifer storativity (S) from the CRD relationships

The coefficients (parameter a = unit response) for the highest correlation of a regression between the CRD series and water levels representing natural conditions, are listed in Table 3.2.1.1. If the recharge of the aquifer cannot been determined e.g. by means of the chloride method (see Sect. 11.4) the recharge coefficient could be estimated. Assuming a uniform recharge of 11% for the entire Bo Molopo region, a value corresponding to that yielded by the chloride method (i.e. b = 0,11 in equation 3.1.1.5), the respective values of S for each hydrograph could be derived. If a more reliable estimate of recharge is available it should be used. The results of the individual boreholes are grouped more or less according to their localities, and the reliability of the CRD simulations is indicated by the correlation coefficients plotted in Fig.5.3.3.1. The aquifer storativity values (S) in relation to the different dolomitic formations and system of dykes are shown in Fig. 5.3.3.2. These S values compare reasonably well, but are slightly smaller than storativity values derived for the hydrodynamic model of the Grootfontein aquifer, which produced the best fit between measured and simulated water levels (Bredenkamp, 1996, and Bredenkamp and Nel, 1997). The storativity values also agree closely with the estimates based on the moving average rainfall method (MA method).

5.4 Hydrographs of the Pretoria dolomitic area

5.4.1 Centurion dolomitic aquifer

Groundwater hydrographs of the Pretoria-Centurion dolomitic area (see Fig. 5.4.1.1) have been analysed in the same way. The rainfall record of Irene has been used for the CRD analysis and an average recharge of 12% of the rainfall has been assumed (b = 0,12) to derive the aquifer storativity from eq. 3.1.1.5 (see Table 5.4.1.1).

Table 5.4.1.1 Aquifer characteristics derived from CRD and average rainfall interpretations- Centurion area.

Borehole	CRD method Unit response $a=b/s$ mm/mm	CRD method Derived S value RE = 12% of rainfall	Average rainfall method Unit response $a=b/s$ mm/mm	MA rainfall method Derived S value RE = 12% of rainfall
Dklf 28	2.86	0.042	1.37	0.087
Zwa 29	1.50	0.080	1.00	0.120
Dklf 24	5.00	0.024	0.76	0.156
Voor 245	4.36	0.028	1.13	0.106
Zwa 50	1.09	0.110	3.75	0.032
Zwa 46	2.78	0.043	1.31	0.091
Zwa 37	2.56	0.047	1.79	0.067
Va 3	3.73	0.032	1.12	0.107
Va 42	1.26	0.095	4.44	0.027
Average		0.056		0.088

Not only are these groundwater records of shorter duration than those in the Bo Molopo area, but measurements have been taken only every three months. This decreases the resolution of water level patterns and increases the chances of errors, especially for small fluctuations in water levels which are typical of the Pretoria dolomitic aquifer. The following examples are cited:-

Va3: (see Fig. 5.4.1.2): The best correlation occurs with ${}_{36}^{1}CRD$ series but a slight shift in water levels can be observed, due to the fact that the original monitoring borehole collapsed and had to be redrilled, causing a slight change of the collar elevation of the borehole. Some correlations between the CRD simulated values and groundwater levels in the Centurion dolomite were not as high as for boreholes in the Bo Molopo area (see Fig. 5.4.1.3 - Dklf 24). This is because water levels were measured at intervals of 3 months and the series of observations was shorter. The rise in groundwater levels in response to rainfall has been small, due to the higher storativity of the aquifer, increasing the effect of measurement errors. Figures 5.4.1.3 and 5.4.1.4 show the correspondence between the values simulated by the CRD series and the water levels for boreholes Dklf24 and Voor 245.

5.4.2 Rietvlei dolomitic aquifer

An analysis of some hydrographs of boreholes in the Rietvlei aquifer reveals a high correlation between the groundwater levels and simulations based on the CRD series (see Table 5.4.2.1, and Fig. 5.4.1.1, showing the locality). Assuming an average recharge of 12% of the rainfall, the aquifer storativity for each of the boreholes has been derived. It is clear that there is a large variation in derived storativity values ranging from 0,006 (A2N142) indicating confined conditions, to S = 0,15 (A2N143) indicating highly leached dolomite.

The large effect of abstraction on the combined flow of the Rietvlei eyes from about 1985-1992 is evident from Fig. 5.4.2.1. The pumping was introduced to effect greater utilization of the dolomitic groundwater by the local municipality.

Table 5.4.2.1 Results of CRD and MA interpretations for Rietondale, De Aar and Rietvlei dolomite.

Rainfall Station	Site		Best CRD			Pump			Unit response	*S* derived
			n	m	Cor Coef	dk	Constant E	Coef. b	a	
Beaufort-West	J2N051	De Hoop	24	2	0.8495	0.04	-166.8110	-0.2545	3.9291	0.0127
	J2N052	De Hoop	24	2	0.7518	0.02	-146.8527	-0.4312	2.3190	0.0216
	J2N066	Kuilspoort	24	5	0.8602	0.02	-84.2600	-0.2375	4.2114	0.0119
	J2N090	Springfontein	36	3	0.7563	0.02	-176.6590	-0.4288	2.3321	0.0214
	J2N093	Uitzig	36	6	0.7948	0.02	-784.1432	-1.7685	0.5655	0.0884
	J2N109	Plaatdoorns	120	2	0.4781	0.10	-421.9184	-0.3776	2.6487	0.0189
	J2N111	Brandwag	36	6	0.8704	0.10	-2481.3256	-1.3637	0.7333	0.0682
	J2N112	Lemoenfontein	36	3	0.7354	0.02	-421.6700	-0.9958	1.0043	0.0498
Pretoria	A2N034	Rietondale	120	4	0.9210	0.04	-986.0326	-0.4171	2.3974	0.0375
	A2N035	Rietondale	120	4	0.9444	0.04	-865.3975	-0.3621	2.7617	0.0326
Pretoria	A2N131	Rietvlei	60	3	0.8828	0.10	-3770.7509	-0.6932	1.4427	0.0832
	A2N136	Rietvlei	60	6	0.9400	0.10	-18107.2834	-3.2992	0.3031	0.3959
	A2N138	Rietvlei	60	4	0.8730	0.10	-4994.2536	-0.9141	1.0940	0.1097
	A2N142	Rietvlei	60	1	0.8880	0.00	-20.4430	-0.0530	18.8822	0.0064
	A2N143	Rietvlei	60	6	0.9222	0.10	-6975.2374	-1.2695	0.7877	0.1523

5.4.3 Rietondale shale-quartsitic aquifer

The results of two hydrographs of the Rietondale aquifer (Fig. 5.4.3.1) are listed in Table 5.4.2.1, those for A2N034 being presented in Fig. 5.4.3.2. The recharge of the Rietondale area has been derived by means of tritium profiles measured in the unsaturated zone, yielding an average recharge of 9% of the rainfall (Bredenkamp et al., 1995). The CRD series correlates very well with the water level fluctuations (r >0,9) if pumping is incorporated (Δk factor = 0,1). Using b=0,09 the storativity values have been calculated to be 0,038 and 0,033 for monitoring points A2N034 and A2N035 respectively.

5.4.4 Karoo aquifer in De Aar region (Table 5.4.2.1)

Except for borehole J2N109 (r = 0,48) the correlation between measured water levels and those simulated by the CRD series exceeds 0,75, and for all of these boreholes the k-factor indicates that water levels have been affected by abstraction. The aquifer storativity values have been inferred, assuming a rainfall/recharge coefficient of b = 0,03 which appears to be reasonable according to other studies. These storativity values range between 0,007 (J2N066) and 0,053 (J2N093). The latter value is rather high for the Karoo but could represent an alluvial aquifer.

5.4.5 Monitoring points stored in the National Groundwater Database (NGDB)

A similar analysis has been carried out on all groundwater level series of reasonable length stored in the National Groundwater Database (see Fig. 5.4.5.1), once the integrity of the data had been checked by means of the CRD regression method. Evaluation of the different monitoring data is discussed in Section 6. For most of the stations the rainfall fraction b representing recharge has been inferred by means of the chloride-ratio method (see Fig. 11.4.1.5 and also Fig. 11.4.2.1).

As is indicated in the next section, the CRD relationship allows a more meaningful characterisation of an aquifer by way of its storativity or unit response to recharge to be achieved, if reliable data on groundwater level fluctuations and rainfall are available. Although water levels are affected by pumping its effect can be determined, and the storativity and unit response can still be quantified.

6. Aquifer characterisation by means of the CRD and MA methods

6.1 Introduction

Characterisation of aquifers has been a major activity all over the world and hydrogeological maps have been produced at great cost. The classification and representation of aquifer characteristics could be much improved using the CRD and MA methods. The shortcomings of characterisation are often due to the fact that critical aquifer features such as recharge and storativity, cannot be quantified reliably. The compilation of hydrogeological maps often entails analysis of the occurrence of groundwater, its exploitation potential and aquifer characteristics, but often this is based on personal experience rather than on real information or an acceptable rationale and analysis of monitoring data.

The manual demonstrate how the CRD and MA relationship can provide information reflecting the hydrogeological characteristics of an aquifer. A basic requirement is that in deriving a record representing natural conditions, water level data should be analysed in a systematic way. This has been done for monitoring points shown in Fig. 5.4.5.1 to derive aquifer storativity, the unit response to recharge, m,n memories and aquifer depth. The spatial characteristics of the determinants have been plotted for four areas which have reasonable coverage of monitoring points. These were Beaufort West - Area 1, Bo Molopo dolomite - Area 2, Pretoria region - Area 3 and Pietersburg/Potgietersrust - Area 4.

6.2 Deriving aquifer storativity

The storativity of an aquifer (*S*) is one of the most important parameters of an aquifer, but also one of the most difficult to derive reliably, for the following reasons:-

1. The majority of aquifers are fractured and hence exhibit a combination of hydraulic response extending over large distances, and matrix response.
2. Classical interpretations of pumping tests do not always yield reliable estimates of aquifer storativity, as *S* derived from normal pumping test interpretations shows apparent distance dependency (Bredenkamp et al., 1995). The fractured nature of the aquifer matrix causes an initial rapid response of water levels along permeable fractures, followed by a delayed response due to matrix flow. The real average *S*-value can be inferred from an empirical calibration curve (Bredenkamp et al., 1995), or by modelling e.g. dividing the aquifer into two sub-systems, an upper matrix and lower fractured part, and then, for each subsystem, solving for transmissivity and storage coefficient by means of the RPTSOLV program (Van Tonder et al., 1996). The model requires that the ratio of the fracture thickness in relation to that of the normal aquifer, and boundary distance, be specified.
3. Deriving *S* and recharge from water balance interpretations is difficult due to the dependence of water levels on both aquifer storativity and recharge. This rules out the possibility of deriving these two parameters by optimisation. Determination of *S* is possible if certain simplifying assumptions are made, or if either of the two parameters has been determined by an independent method (Bredenkamp et al., 1995).

By utilizing eq. 3.1.1.4, or the MA or modified ${}_{n}^{m}CRD$ relationship the variability of *S* in different aquifers can be inferred if the recharge coefficient has been reasonably well established e.g. from a regional recharge relationship with rainfall such as the chloride rainfall-recharge response (Fig. 11.4.1.5). The most likely values of *S* are assumed to be those derived from the best fit between the CRD and MA series and the groundwater level series. The *S* values for the selected areas (1-4) are shown in Fig. 6.2.1 to 6.2.4.

If more reliable values of the aquifer storativity (S) are available (obtained, for example, from reliable pumping test interpretations), the spatial variability of recharge can be derived from the water level fluctuations in relation to the CRD and MA series. Alternatively the spatial variability of S can be narrowed down if monthly rainfall is measured at the monitoring borehole itself. Insertion of the average S-value would yield the areal variation of recharge if monitoring points are well distributed. The CRD and MA method would therefore also provide a check on storativity values determined by pumping tests and, if proven reliable, substitution of this S-value would yield the variability of recharge of the aquifer. It should however be borne in mind that even in pumping tests the water level response is averaged over the area affected by the cone of depression. This zone could extend quite far because of the highly permeable fractures or dissolution channels, for which the S-values turn out to be representative of confined conditions even though the aquifers are phreatic.

6.3 Unit response of groundwater levels to recharge

The coefficient ($a = b/S$) in eq. 3.1.1.3 represents the water level rise caused by one millimetre of recharge. To calculate this response it is not necessary to know the coefficient of rainfall representing recharge or the aquifer storativity S. However for each monitoring point the coefficient $a:b/S$ represents the characteristic response of the aquifer to recharge at that point, and collectively several boreholes from a specific aquifer would yield the average response of that aquifer. The unit response therefore provides a useful characteristic of an aquifer which could be displayed by mapping the regional variations or intercomparison with different aquifers in different climatic regions, all over the world.

The unit responses of four areas are indicated in Fig. 6.3.1 to 6.3.4. The sparsity of water level data limits proper featuring of the regional variations but clearly shows the higher unit responses associated with boreholes and aquifers of low storativity.

6.4 Variability of the short and long-term memory in water level series (m,n-values)

The m parameter in the ${}^{m}_{n}CRD$ relationship (short-memory) provides an indication of the relative rapidity of recharge and the response of water levels at individual monitoring points. The most common value even in the dolomitic areas is $m = 1$ and $n = 36$ indicating that water levels respond to rainfall within one month. A second group with $m = 3$ or 9 and $n = 60$ to 72 in the dolomite appears to relate to boreholes which have intersected major solution channels of higher storativity (S) and transmissivity (T). Because of the high T values the water level response is integrated over a large area and hence the water level responds to rainfall integrated over a longer period, as in the case of the Wondergat. The larger m and n values are more representative of the regional catchment response of the aquifer whilst the fast-reacting water levels relate to the aquifer characteristics of a specific locality. The *m/n* memories for the Bo Molopo region are plotted in Fig. 6.4.1.

6.5 Aquifer depth

The depth of an aquifer from surface, which can be derived from a regression between the MA rainfall series and a series of groundwater level measurements (see Section 3.2), is one of the most significant features of an aquifer, often difficult to derive even from drilling results. The depths from surface of the following aquifers have been derived:

De Aar - Karoo - Fig. 6.5.1
Bo Molopo Dolomitic - Fig. 6.5.2
Pretoria Dolomitic - Fig. 6.5.3
Pietersburg/Potgietersrus - Fig. 6.5.4

The greater depth of the dolomitic aquifers and a much shallower depth for the Karoo aquifer is clearly shown.

The difference between the depth and the water level represents the thickness of the aquifer at a specific time. Plotting of the thicknesses reveals its spatial variability for a specific aquifer (see Fig. 6.5.5), whilst comparison between different aquifers could be made in a similar way. Interestingly contouring of the aquifer depth indicates a deeper zone of leached dolomite in the Grootfontein dolomitic compartment (see Fig. 6.5.6). This accords with hydrological interpretations and identifies the zone of higher transmissivity along which the main flow to the eye occurs.

7. CRD and MA analysis in relation to groundwater management

7.1 Introduction

In heterogeneous aquifers such as dolomite and highly fractured ones, hydrodynamic modelling of the system has been only partially successful. Attempts to simulate the behaviour of dolomitic aquifers by means of 2-dimensional finite element models proved to be difficult and the calibration of 3-dimensional models is as yet not very successful. 2-D modelling was only reasonably successful as in most cases higher values of aquifer storativity and recharge had to be introduced to obtain good simulations of water levels and spring flows (Bredenkamp, 1997; Bredenkamp and Nel, 1997). These models also failed to reproduce rapid high response of water levels and springs to high recharge events. The springs were either introduced as variable rates of abstraction or were substituted by a constant-head outflow. In the latter case the overflow height of the spring had to be set lower than the actual outflow level to effect the right response. The unsatisfactory simulations are due to the fact that the water levels and spring outflows are governed by the rapid transmission of hydraulic pressure along major fractures or solution channels (confined flow), which can only partially be accommodated in a 2-D model.

7.2 Impact of abstraction simulated by CRD and MA relationships

7.2.1 Water level responses

For any series of groundwater level measurements which have been matched by the CRD and MA rainfall series in the appropriate way, the effect of variable abstraction and rainfall recharge could be simulated. This is illustrated by Fig. 7.2.1.1 indicating the impact of different rates of abstraction on the average groundwater levels in the Lichtenburg dolomitic aquifer.

In the case where spring flows are linearly related to the CRD and MA rainfall series they allow both the natural flows and those affected by abstraction to be simulated. In this way groundwater levels or minimum flow rates set as critical reserve constraints of sustained utilization of water resources, could be tested against various management options, in a simple way.

Fig. 7.2.1.2 indicates the relationship between the flow of the Grootfontein spring in the Rietvlei dolomite, simulated by the CRD series for different rates of abstraction. If the area of the aquifer is not known it could be estimated from the topographical drainage area above the point of outflow. The area of the aquifer could also be calculated according to the average yield of the springs and the percentage rainfall representing recharge. Because of the high transmissivity of the main channels feeding the springs, the spreading of abstraction over the intake catchment or its concentration at one point or in a small area, would not affect the simulated flows very differently. However if the abstraction occurs too close to the point of outflow the local cone of depression would have a greater impact on the flow of the eye.

In the case of the Upper Dinokana eye, the effect of abstraction at different distances from the eye could be simulated by means of the CRD method. Firstly the observed flow had to be matched to the CRD simulation. Only the initial series which has been least affected by abstraction was used for the calibration, whereupon different rates of abstraction were incorporated in the *k*-value (Fig. 7.2.1.3). Although the historical rainfall series has been used, any generated sequence of monthly rainfall from stochastic rainfall models, could have been introduced as input of variable recharge.

Fig. 7.2.1.4 shows the similarity between the abstraction (plotted inversely) and the combined flow of the Upper and Lower springs of the Pretoria Fountains Valley. The higher the rates of abstraction, the greater the effect on the spring discharge. Fig. 7.2.1.5 shows the impact of the abstraction on an x-y plot, indicating a linear reduction in the flow relative to the rates of abstraction. However as is indicated in Section 10, higher abstraction is related to lower rainfall and vice versa, so that the oscillations shown in Fig. 7.2.1.5 are actually the combined effect of abstraction and rainfall. The negative effect of abstraction on the flow of a spring has to be simulated by the *k*-factor in the CRD relationship. The correspondence between spring flows and the Wondergat levels over a large area suggests that the impact of abstraction could extend far, probably due to highly permeable fractures which propagate the hydraulic responses. One of the advantages of the CRD and MA methods is that they automatically incorporate the natural inflow/losses i.e. the boundary effects operating on the system.

7.2.2 CRD and MA methods applied in different rainfall regimes

It can be demonstrated that the CRD method mimics groundwater levels in all types of aquifer, implying that recharge of all aquifers is very similar. For example Fig. 7.2.2.1 indicates the high degree of correspondence between simulated and measured water levels in a mountain catchment in the Jonkershoek area. The best correlation was achieved for the ${}^{1}_{av}CRD$ series. It could be proved from the chloride concentration of the river flow in relation to that of the precipitation, that there is several months delay between the time of occurrence of rainfall and its appearance as runoff (see Fig. 7.2.2.2). However, the flow responds much faster to rainfall as groundwater is fed hydraulically to the river but nevertheless even in the Jonkershoek mountains the highest stream flows occur after the main rainy season. The CRD relationship therefore provides a strong indication that a significant portion of river flow in all areas is fed via groundwater.

The Orapa-Letlhakane well-field provides a good example of how the CRD method could be used to predict the regional response of groundwater levels to abstraction in a semi-arid environment. Studies over many years have failed to provide conclusive evidence that recharge by local rainfall occurs, or to quantify it reliably. Initially the simulation of the groundwater levels, by the ${}^{m}_{n}CRD$ relationship, showed a best fit for m = 12 and n = 36 (see Fig. 7.2.2.3) and a lag in the CRD of 4 months - the maximum tested at the time. This yielded a correlation coefficient of 0,87. However an even better simulation (Cor. coef. =0.97) has been achieved using the long-term average rainfall as input, with the actual rainfall over 36-months controlling the loss from the system (see Fig. 7.2.2.4; Verhagen et al., 1998). This CRD model represents the converse of the normal relationship and implies that recharge is related to the average rainfall whilst the losses are proportional to the actual rainfall over 36 months, which hydrologically would be difficult to explain. It is more logical that the recharge from a specific month is delayed, causing its impact on the water level to be manifested much later. A 14 months' delay in the ${}^{m}_{n}CRD$ series yielded the best correspondence (r = 0,99), where m = 12 months, n equals total length of the rainfall record (see Fig. 7.2.2.5), and abstraction is incorporated in the k-factor. In the first CRD simulation the delay had been only partly accommodated by the short-term memory as a delayed reaction of only 4 months. The best match (Fig. 7.2.2.5) yielded a recharge of 1,8% of rainfall, the aquifer storativity having been set at 0,003 according to the apparent *S*-distance relationship (see Fig. 7.2.2.6). A good fit of the water level response has also been obtained by introducing the abstraction as an average depth of precipitation, which is subtracted from the input rainfall.

In spite of the simplicity of the CRD 'model' it would predict the response of the groundwater levels in the Letlhakane-Orapa aquifers much more accurately than any of the more sophisticated models attempted in the past. For those models the variable recharge is an essential parameter and is difficult to determine reliably (Gieske, 1992). By means of the newly calibrated CRD model it would be easy to assess the impact of different rates of abstraction in relation to a range of rainfall sequences simulated by stochastic methods.

The fact that the long-term memory of a system (n-parameter) is shorter than the total length of the data series, and could be as short as 36 months, means that a long series of rainfall data is not essential for the application of the CRD method. The most important factor is that the monthly *rainfall data should be representative of the groundwater response at the monitoring point.*

It is clear that collection of monthly rainfall at or close to groundwater monitoring boreholes (but unnecessary for springs) would be adequate for application of the CRD and MA methods, and would reduce uncertainties resulting from the use of more distant rainfall stations. The spatial variability in monthly rainfall in a region, expressed as a statistical distribution, would reflect the uncertainty imposed by a non-representative rainfall record.

The rainfall/recharge response of springs (see Section 8) can also be used to derive recharge from an equal-volume analysis of spring flows. This analysis reveals that recharge of a spring, and hence the water level in the aquifer, are determined by the average rainfall over several preceding years. Linear dependence between groundwater levels and the flow of springs has been demonstrated (Bredenkamp et al., 1995).

8. Estimation of groundwater recharge by means of the Equal-Volume Method

8.1 Introduction

Monitoring of groundwater levels is important in estimating the recharge of an aquifer. This can be derived from the saturated water budget by balancing the input and output over a selected time interval Δt according to the water balance equation:

$$Re_{tot} = Q_{abstr} + S\Delta V + E_{vt} + Q_{outflow} - Q_{inflow} \qquad \text{eq. 8.1}$$

where

Re_{tot} = total recharge: evapotranspiration losses included

Q_{abstr} = pumpage from system, or spring flow

$Q_{outflow}$ = lateral flow to lower compartments

Q_{inflow} = lateral inflow from higher aquifers

E_{vt} = evapotranspiration losses directly from the aquifer

S = aquifer storativity or specific yield

ΔV = integrated change in saturated storage volume (positive or negative)

The effective recharge represents the nett recharge Re_{nett} after evapotranspiration losses from the aquifer have been accounted for. It can be obtained more easily than Re_{tot} because of difficulties of quantifying evapotranspiration. Hence effective recharge is usually obtained by applying the following balance to a selected time interval:

$$Re_{nett} = Q_{abstr} + S\Delta V + Q_{outflow} - Q_{inflow} \quad \text{eq. 5.2}$$

For $\Delta V = 0$ and $Q_{outflow} = Q_{inflow}$ the aquifer storativity does not come into play and the effective or nett recharge is equal to the abstraction over the equal-volume interval; if these components do not cancel out they have to be estimated according to the Darcy equation:

$$Q_{inflow/outflow} = T.I.L$$

where L = the width of the inflow or outflow area
I = the hydraulic gradient
T = the respective transmissivities

In most cases the shortage of water level data does not allow the hydraulic gradient and T to be inferred reliably. The difficulties of quantifying inflow and outflow reliably, and of proper delineation of the aquifer, are reasons why the method is usually not regarded as suitable for general application.

In the case of dolomitic springs in the Bo Molopo area, the fact that the flows correspond linearly to

- the water level in the aquifer, as is illustrated by the Wondergat, and to
- the cumulative departures of rainfall from the long-term mean,

implies that periods for which the flows at the beginning and the end are equal, represent an equal water level status of the aquifer and hence can be analysed according to the equal-volume method. Even more so would it apply to spring flows reacting to rainfall averaged in space and time. Spring flow therefore provides a more reliable manometer of the hydrological response of the system than individual boreholes can do. For this reason monitoring of spring flows is one of the most important hydrological activities.

8.2 Equal-volume method (EVM) applied to groundwater hydrographs

Whereas the flow of a spring is related to the integrated water level of the aquifer, individual hydrographs reflect the groundwater balance at specific points of the aquifer, but they also reflect a certain spatial integration of recharge.

In quantifying the recharge by the EVM, the "abstraction"over an identified period beginning and ending with equal water levels has to be known. This "abstraction" includes

- natural evaporative and transpiration losses which are usually difficult to quantify,
- outflow in the form of springs, which can be measured,
- groundwater abstraction either measured, or inferred from electricity consumption,
- differences between lateral inflow and outflow, which can be assumed zero for aquifers which are in a state of natural equilibrium.

In the present study the application of the equal-volume method (EVM) has been extended to derive recharge in relation to rainfall for periods of equal-volume represented by intervals starting and ending at the same water level. Delimitation of such equal-volume periods is the same, irrespective of whether or not an aquifer has been affected by abstraction, provided the abstraction is also incorporated. If the abstraction is not known, the natural groundwater levels have to be inferred by means of the CRD method (substituting $k = 1$). Just as groundwater levels in unstressed aquifers are related to the rainfall experienced during equal-volume periods, the losses would be proportional to the average rainfall over the characteristic long-term memory of the CRD relationship (see eq. 3.2.1.1).

The characteristics of recharge at a monitoring point could be determined by means of an equal-volume analysis of the groundwater hydrograph. Any external influence on the groundwater levels would show up as anomalous recharge, and provide an additional indication of the reliability of groundwater levels. This method would, however, be more elaborate than the CRD method.

8.3 Interpretations of spring flow

8.3.1 Equal-volume interpretations of spring flows

The rainfall/recharge response of springs can be used to derive recharge from an equal-volume analysis of spring flows because of linear dependence between groundwater levels and the flow of the springs. (Bredenkamp et al., 1995).

An equal-volume analysis of the Fountains of Pretoria has been carried out, using an adapted Fortran program originally compiled to analyse the saturated volume fluctuations of an aquifer based on the integrated water levels (Van Tonder, 1989). All periods starting and ending with equal flows have been identified (see Fig. 8.3.1.1). The flow during each equal-volume period has been plotted against the rainfall during the period, but the flows first had to be converted to millimetres of recharge by incorporating the size of the areas, and then expressed as percentages of the rain received.

Due to the large number of equal-volume periods, the spring flow record was subdivided into shorter periods which are listed in Table 8.3.1.1 along with the average rainfalls:

Table 8.3.1.1 Periods to which the equal-volume method has been applied to the Pretoria Fountains.

Period	**Average rainfall** (*mm/month*)
1948 - 1953	53,8
1953 - 1961	61,0
1961 - 1969	54,9
1969 - 1979	62,0
1979 - 1987	48,0
1987 - 1995	58,3

Fig. 8.3.1.2 shows that the recharge derived for equal-volume periods longer than 240 days, and plotted in relation to the rainfall during an identified equal-volume period, increased linearly irrespective of intensity of rainfall. However, converting this recharge to a percentage of the rainfall caused scattering of the points. For periods of lower rainfall i.e. the shorter periods, the percentage recharge increases, varying between 6% and 18% of the rainfall. However, for longer equal-volume periods the recharge percentages tend towards a constant value. This is because the average rainfall approaches the long-term mean. The higher percentage recharge for lower amounts of rainfall is not in accordance with hydrological principles. A clear discrepancy in the recharge results (see Fig. 8.3.1.3) can only be ascribed to incorrect flow measurements or to the addition of recharge, probably leaky municipal water/sewage lines. However, expressing the recharge in millimetres per month reveals a fairly constant value irrespective of the length of the equal-volume period (see Fig. 8.3.2.2.3 for Uitenhage springs). This results from the fact that the flow of the spring is caused by rainfall extending over a longer period than the identified equal-volume period. Therefore in the equal-volume analysis of a spring, the moving average rainfall pertaining to *n*-months (long-term memory) preceding a specific month should be used in determining the overall rainfall/recharge relationship. In the case of Buffelshoek eye the relationship conforms to an exponential variation of recharge in relation to this long-term average rainfall (see Fig. 4.2.1.4.3).

The impact of abstraction will show up as a discrepancy in the equal-volume recharge. It would however not be possible to differentiate between discrepancies caused by errors in rainfall, incorrect flow measurements or declining flows due to abstraction.

8.3.2 Flow of springs in relation to the average rainfall

8.3.2.1 Pretoria springs

All the Pretoria springs show a striking relationship with the moving average rainfall over a specific period except when the flows have been affected by abstraction (see Fig. 5.4.2.1), or other factors, or unreliable measurements. Fig. 4.2.1.4.1 suggests that the Sterkfontein springs would cease to flow if the average rainfall over 60 months declined to about 49 mm/mth. However, as indicated earlier on, this is not a valid extrapolation since the evapotranspiration losses during prolonged low rainfall periods would decrease and a new hydrological equilibrium would set in and control the flow of the spring.

Table 8.3.2.1.1 Flow of the different springs in relation to rainfall.

Spring	Average rainfall best corresponding with flow	Correlation coefficient	Apparent linear cut-off for zero flow
Pretoria dolomite			
Upper and Lower	120 months	0,76	46 mm
Sterkfontein	120 months	0,57	44 mm
Rietvlei	72 months	0,79	45 mm
Zeerust dolomite			
Buffelshoek	72 months	0,78	30 mm
Tweefontein Upper	36 months	0,55	25 mm
Tweefontein Lower	36 months	0,50	25 mm
Stinkhoutboom	72 months	0,69	35 mm
Vergenoegd	84 months	0,80	37 mm
Dinokana Upper	84 months	0,60	20 mm
Table mountain Quartzite			
Uitenhage springs	132 months	0,72	37mm

The gauging of springs in the Grootfontein (Molopo region) and Zeerust areas began around 1960. Although measurements of Molopo eye started in 1964 the first measurements proved to be unreliable when compared to the groundwater levels of the Wondergat (see Section 4.2.3). Reliable flow measurements of Molopo eye have been obtained since the new outlet works were constructed to divert the flow to Mafikeng, but the flows have been affected by abstraction (see Fig. 5.3.2.4.

The flows of different dolomitic springs in relation to the average rainfall corresponding best to the flows, are listed in Table 8.3.2.1.1.

8.3.2.2 Uitenhage springs

The combined flow of several Uitenhage springs emerging over a small area in which groundwater is recharged in the Table Mountain Group and Koega formations (see Fig. 8.3.2.2.1) have been monitored since about 1901. However the flows have diminished due to abstraction and leakage via artesian boreholes which have punctured the confined aquifer feeding them.

Correlations between the flows corrected for abstraction, and the ${}^{m}_{n}CRD$ series, indicate that the ${}^{1}_{132}CRD$ series (m = 1 and n = 132 months) yields a high correlation over the period 1965 to

1994. (Cor. Coef. = 0,874 - see Fig. 8.3.2.2.2). It therefore closely represents natural flow conditions because leakage from most artesian boreholes has been minimized. By means of the CRD relationship and incorporating the effect of abstraction in the k-value, the natural flow of the springs could be simulated. The real abstractions could be converted to k-values, after the area of recharge had been determined according to the rainfall/recharge coefficient, and produced almost the same response as the Δk-factor.

Using the equal volume method with the corrected flows of the Uitenhage springs and the average rainfalls for the preceding 132 months, the recharge during different periods has been determined (see Table 8.3.2.2.2 and Fig. 8.3.2.2.3).

Table 8.3.2.2.2 Results of equal volume analysis of the Uitenhage Springs and of chloride concentrations.

Period	Recharge (*mm/mth*)	Average rainfall (*mm/mth*)	Recharge %
1901-1911	2,5	39,8	6,20
1911-1924	2,4	39,5	6,00
1924-1936	2,0	39,7	5,80
1936-1948	2,3	37,2	5,90
1948-1953	2,2	43,0	5,50
1953-1961	2,5	41,8	6,25
1961-1969	2,4	39,8	6,00
1969-1979	2,5	42,0	5,90
Recharge derived from the Cl method = 6-8%			

The monthly flows of the Uitenhage springs in relation to the average monthly rainfall over the preceding 132 months could be interpreted as being part of an exponential relationship with rainfall (see Fig. 4.2.1.4.2). However according to the EV method the recharge is almost constant and shows only a slight linear increase (see Fig. 8.3.2.2.3). The average recharge according to the chloride method varies between 6-8% of the rainfall.

8.4 Conclusions

It is clear that recharge could be derived applying the EV method to the spring flows, provided that the long-term average rainfall of an appropriate preceding period is used. The EV method cannot be used to relate the flow to rainfall which occurred during EV periods. However if it is applied in a consistent way, it will show up anomalies in flow measurements, or flows affected by an external factor, as displacements of the relationship.

Both the CRD and MA methods reveal the characteristics of the recharge system and enable flow records to be checked, or missing data to be simulated. In the case of discrepancies, it is not possible to ascertain whether the flows or the rainfall have been measured incorrectly, or if it has been affected by abstraction.

As groundwater levels are related to the flow of the springs, it confirms, as is generally accepted in hydrological modelling, that even where no visible outflow occurs the losses would be higher for aquifers with shallow water levels.

The dependence of spring flow on rainfall over an extensive period preceding a specific month, has a significant effect on previous interpretation of isotopic concentrations of water emerging from springs, and requires re-examination.

9. Flow of dolomitic springs in relation to surface run-off

9.1 Introduction

Monitoring of surface runoff is one of the most important hydrological activities. Although measurements of both surface and groundwater are incorporated in hydrological data bases, they are treated as separate entities and their monitoring, data patching, evaluation and assessment are carried out independently. In this section similarities between surface and groundwater will be illustrated, which could facilitate intercomparison and infilling of either surface or groundwater records using records of the other.

9.2 Flow of springs in the Pretoria area in relation to runoff

9.2.1 *General*

The excellent records of springs in the Pretoria area and of runoff recorded for the Rietvlei dam allows of a comparison of the hydrological responses of surface flow, and of groundwater flow emanating from springs. The position of the Rietvlei Dam and the various springs is indicated in Fig. 5.4.1.1. Inflow into Rietvlei dam has increased as a result of sewerage disposal in the Kempton Park area, but has also diminished due to irrigation from groundwater. Therefore only the records for the period 1931 to 1978 of both Rietvlei Dam and the combined flow of the Rietvlei springs have been compared. The flow of the Sterkfontein eye has also been examined in relation to the inflow into Rietvlei dam.

9.2.2 Flow of Rietvlei springs in relation to Rietvlei Dam inflow

According to Fig. 9.2.2.1 the runoff is much more variable than the spring flow, due to faster and more direct response of runoff to rainfall. Yet if the variable inflow is integrated over a longer period the flow response improves quite dramatically, and a correlation coefficient of 0,8 is obtained if runoff is integrated over 60 months (see Fig. 9.2.2.2). This confirms that over this period both runoff and spring flow relate to the average rainfall of the preceding 60 months. Missing spring flow data could therefore be infilled from a regression between the average inflow into the dam and the flow of the spring, and vice versa.

It would be worthwhile investigating the relationship between groundwater levels and runoff over longer periods as a means of patching missing runoff data and correcting flows which have been affected by abstraction, afforestation or the construction of farm dams. These effects could be ascertained by a CRD relationship incorporating the k-factor, or MA method, and by comparing the average runoff over n months to groundwater hydrographs.

Likewise it would be worthwhile to investigate the rainfall-runoff relationship in relation to the cumulative departures of rainfall from the average (CRD method). The difference, however, would be that runoff would be controlled by the rainfall over a shorter period (a few days) in relation to the average rainfall over a longer period.

Assuming a linear relationship similar to the CRD groundwater response, the runoff coefficient could be derived, which would allow characterisation of runoff in different catchments. For a consistent comparison, surface runoff should be analysed during periods of equal volume status, according to the CRD series. Hence

$$RO_i = a(Rf_i - Rf_{av}) \quad \text{eq. 9.2.2.1}$$

where RO_i = runoff of day in mm per unit area
Rf_i = rainfall over preceding i-days
Rf_{av} = average rainfall of n-days
a = coefficient

Hence the total rainfall minus the runoff component and groundwater recharge would yield the fraction of rainfall lost as evapotranspiration. The latter would be an interesting characteristic to typify different catchments.

More study is needed of the influence of agricultural development, building of farm dams, urban development, afforestation and land management on natural runoff. The effects could be incorporated in a *k*-factor similar to that used for groundwater systems which have been affected by abstraction (see eq. 3.1.1.2 and 3.2.1.2). For that matter both positive and negative influences on the natural runoff could be related to a runoff *k*-factor. A factor of $k < 1$ would indicate an increased runoff response which could have been caused by catchment denudation or an extra runoff contribution due to urbanisation. A factor $k>1$ could indicate the effects of afforestation or agricultural development. Setting a *k*-factor, determined by regression equal to unity ($k=1$), would yield natural flows. As in the case of recharge, runoff over longer periods conforms to a linear fraction of rainfall, and by incorporating the size of the catchment the coefficient of rainfall representing runoff could be obtained. This would provide an easy way of characterising both runoff and recharge in catchments, on a comparative basis.

Such an extension of the CRD and MA methods would close the gap between groundwater and surface water assessments, and would allow hydrological entities to be characterized in a consistent way. At the same time the validation of hydrological data and infilling of missing data could be improved. As has been shown, the monthly response of groundwater levels is adequate for most assessments and management of water resources, and this would also apply to surface water impounded in a dam.

9.3 Runoff in relation to groundwater levels

According to Fig. 9.3.1 the simulated groundwater levels would theoretically have risen up to 3 metres above the ground surface theoretically e.g. during the period 1974 to 1980. During these periods the borehole would have flowed and if this were to happen over large areas it would cause a significant increase in runoff, because there would be no infiltration of rainwater and virtually 100% runoff in these areas. This is probably what happens in other dolomitic aquifers e.g. at Pering mine where surface water starts to flow over a large area during periods of prolonged high rainfall.

9.4 Overall conclusions regarding the CRD and MA methods

The main conclusions are that:

1. both the CRD method and mean rainfall over a characteristic period mimic groundwater levels and the flow of springs, as well as runoff integrated over longer periods;
2. the effect of abstraction or other impacts can be derived by incorporating a Δk value in the ${}_{n}^{m}CRD$ series, or an ineffective rainfall amount to be subtracted from the moving average rainfall;
3. in the case of unknown impacts the *k*-factor or ineffective rainfall could be obtained from a regression fit;
4. spring flow provides an integrated response to rainfall over the contributory catchment;
5. for a specific spring there is an apparently linear relationship between rainfall and spring flow, but this only applies for a specific average climatic equilibrium; as the average rainfall decreases, the losses from the system by evapotranspiration decline, enabling flow to continue even at lower rainfall;

6. application of the equal-volume method to spring flow allows the relationship between rainfall and recharge to be determined; the same technique could be applied in the case of runoff over periods of equal groundwater status, which would show up discrepancies in flow as abnormal responses in relation to rainfall.

10. Abstraction from groundwater

10.1 Introduction

Monitoring abstraction from aquifers is one of the key elements of water balance assessments to derive recharge and aquifer storativity, for example with the SVF and Hill method, or in modelling the hydrodynamic response of an aquifer. Irrigation from groundwater is a particularly uncertain component. Although attempts have been made to meter pumping (which all groundwater users in controlled areas are required to do), not all users have complied, claiming that the meters restrict the yields of the pumps and increase pumping costs. It is therefore necessary to have an alternative means of assessing abstraction reliably. Some results and experiences which have been gained in this regard will be discussed.

10.2 Irrigation abstraction

In the case of the Grootfontein aquifer the irrigation abstraction was derived from the electricity consumption of installations which had been calibrated (Van Rensburg, 1987). This is regarded as more reliable than estimating the abstraction from the areas under irrigation, and water application in relation to the rainfall, as it also allows past abstractions to be derived. The water pumped comprised 1) abstractions at the Grootfontein eye for urban water supply, and 2) pumping from several installations delivering water to irrigation pivots.

The electricity consumption of the different installations was available for two periods, namely from 1966 to 1973 and from 1978 to 1985. Pumping during the latter period had increased markedly as electricity had become more widely available, and irrigation from boreholes had expanded rapidly. The periodic droughts have also encouraged more farmers to resort to irrigation. Apart from the irrigation abstraction, groundwater has also been pumped at the Grootfontein eye to supply Mafikeng with urban water.

The abstractions from irrigation boreholes, derived from the electricity consumption, were summated for the different pumps to yield monthly totals. These were related to the rainfall, the best correspondence being between the abstractions and the average rainfall over the preceding 12 months. Although for both periods a linear relationship was applied, the regressions differed for the 1967 - 1973 and the 1978 - 1985 periods (see Fig. 10.2.1) as larger areas had been put under irrigation during the latter period. Simulation of abstraction in the first period of missing data (1974 - 1977) was based on the regression yielded by the period 1967 - 1973. Abstraction after 1984 was derived from the regression using the 1978-1985 data in relation to rainfall over the previous 12 months (see Fig. 10.2.1). These abstractions have been converted to areas under irrigation, according to the water requirements of different crops. The volumes of abstraction and the areas under irrigation conformed to a resurvey of the irrigation areas in 1997.

10.3 Urban consumption

For the Grootfontein eye the good correspondence between monthly abstractions and the average rainfall over the preceding 12 months allowed the missing data to be simulated (see Fig. 10.3.1). The higher rates of abstraction which have been measured latterly compared to pumping simulated from the rainfall relationship, are due to expropriation of irrigation rights in order to increase abstraction at the eye.

The total abstraction from all boreholes in the Grootfontein compartment was apportioned to the individual pumps according to the areas irrigated. This data was then used in a finite element model of the aquifer, and yielded good correspondence between the observed and simulated water levels (Bredenkamp, 1997).

11. Chemical monitoring and groundwater characterisation

11.1 Introduction

This guide also examines the rationale behind chemical monitoring and characterisation of groundwater with a view to broadening hydrological perspectives and demonstrating the value and use of chemical data in groundwater studies. Whereas the focus is usually groundwater pollution, the emphasis in this study has shifted to hydrological interpretations supported by the natural groundwater chemistry. These include the delineation of recharge areas, quantifying recharge, and chemistry as an aid in the interpretation of ^{14}C and other isotopes in groundwater.

Although the reliability of the chemical analyses is of vital importance, examining the integrity of the data is not covered in this study. Spurious outliers attributable to erroneous measurements or bad sampling, could be omitted, or corrected according to the predominant pattern or trend exhibited by the rest of the measurements. The main focus is chemical fingerprinting, best portrayed by springs yielding an integrated sample of groundwater which has been recharged over a long period and over a large intake area. The chemistry of single boreholes would be more representative of the characteristics of local groundwater.

11.2 Groundwater chemistry and monitoring

11.2.1 Overview

Aspects covered by this guide include evaluation of the usefulness of chemical data pertaining to natural groundwaters in

- characterising an aquifer in relation to aquifer type and climatic influences;
- indicating the degree of mixing between groundwaters of different chemical composition;
- determining recharge by means of the chloride method, having established that an aquifer is still in a natural state; if it is not, the natural background levels of chloride may sometimes be inferred from the initial measurements.

11.2.2 Monitoring

Chemical monitoring of several springs was initiated around 1960 in different regions of the country. The following table provides a list of the springs and the different aquifer types they represent (see also Fig. 4.1.1).

Table 11.2.2.1 List of springs sampled fairly regularly on a monthly basis.

Karoo and Quartzite	**Bo Molopo dolomite**	**Pretoria dolomite**	**West Rand dolomite**	**Ghaap Plato Dolomite**
Bloemhof	Olievendraai	Pretoria	Upper Turffontein	Vlakfontein
Trompsburg	Buffelshoek	Fountains	Gerhard Minnebron	Groot Koning
Mosterdhoek	Rhenosterfontein		Maloney's eye	Klein Koning
Toorberg 1 and 2	Dinokana Upper	Rietvlei		Manyeding
Kraalkop	Dinokana Lower			Tsineng
Mackies	Welbedacht			Thabasikwa
	Molopo			Boetsap
Uitenhage	Grootfontein			Bothetheletse
	Kareebosch			
	Malmani river			
	Malmani Upper			
	Doornfontein			
	Rietpoort			
	Kafferskraal			
	Vergenoegd			
	Tweefontein Upper			
	Tweefontein Lower			
	Wonderfontein			
	Skilpadfontein			
	Polfontein			

11.3 Interpretation of chemical data

11.3.1 Introduction

The chemistry of groundwater is determined primarily by rainfall and secondly by geochemical interactions which occur 1) in the unsaturated zone as the water moves down to groundwater level, 2) within the saturated aquifer where the water is in contact with the aquifer matrix, and 3) when pollutants enter the aquifer. Hence the chemistry of the groundwater reflects the outcome of the hydrological interaction from the time of recharge, along the flow paths in the aquifer, until the water is sampled.

The concentration of total dissolved substances in natural waters provides a measure of the recharge and its spatial variability because evaporative processes increase the concentration of dissolved salts. Thus water of a better quality is invariably associated with active recharge as is evident even in areas of low rainfall where the best quality waters are found at higher elevations receiving more rain than lower sites. Ridges and mountain ranges are higher because they comprise formations more resistant to erosion, such as quartzite, sandstone and dolerite which have little in the way of soluble constituents, and therefore do not decompose readily. Anomalies occur in or along river beds where groundwater of good quality is recharged at times of flooding, the Kuruman and Kuiseb Rivers being typical examples. Hence the chemical characteristics of groundwater provide an index of the relative ease of recharge in different regions, and of the variability of recharge within one and the same catchment (Sami, 1991). A study by Levin and Bredenkamp (1990) produced a general map of the total dissolved salts for the country based on chemical data contained in the National Groundwater Data Bank (NGDB). The rate of groundwater renewal is low if the concentrations of dissolved substances are high, whereas low concentrations signify regular recharge. For this reason a conservative element such as chloride can be used to quantify recharge (see Section 11.4).

In areas of similar rainfall but diverse geology the chemical fingerprinting of groundwater could help to establish whether recharge of springs occurs from the same geological formation or not. There is little intermixing of groundwater at the point of recharge. However, mixing does occur when groundwater is pumped, or emerges in the form of a spring or as base flow in rivers. In the case of the Pretoria Fountains, Grootfontein eye (Rietvlei-Pretoria), Maloney's eye, and Kuruman eye the chemistry of the spring water has provided clear proof of contributions of recharge from non-dolomitic aquifers falling within the topographical drainage areas of the springs (Bredenkamp, 1995). This provides evidence that springs are fed from groundwater draining from the topographical area lying at higher elevation than the spring outlet.

The mixing ratios of the different components can be obtained by means of a simple model, or a sophisticated model according to the method of Adar et al. (1992). The concept underlying the latter is the mixing of groundwaters of different characteristics, in different proportions and

Table 11.3.2.1 Comparison of the chemistry of different springs in relation to the characteristics of the recharge areas.

Dolomitic Spring	TDS (mg/ℓ)	HCO_3' (mg/ℓ)	Cl (mg/ℓ)	Sulphate SO_4 (mg/ℓ)	Comment
Elandfontein - Pretoria	210	120	4	5,5	Lower HCO_3' indicates contribution of non-dolomitic recharge
Bo Molopo springs					
Buffelshoek (100% dolomitic)	372	223	5,1	8,1	Pristine spring - very little contamination, low sulphate
Dinokana Upper (some recharge from quartzite)	351	207	4,9	6,6	Pristine spring - lower TDS and Cl indicating higher recharge
Molopo (100% dolomitic)	326	199	4,8	5,3	No contamination
Vergenoegd (100% dolomitic)	399	244	4,9	6,6	Natural spring - no contamination
Skilpadfontein (100% dol.)	442	272	5	6,5	Natural spring - no contamination
Rhenosterfontein (Transvaal shales and quartzites - only partially dolomitic)	395	209	25*	12	The chemistry is clearly different from that of normal dolomite - spring also indicates pollution
West Rand dolomite					
Upper Turffontein	700	178	49*	243*	High sulphate indicates contamination
Gerhard Minnebron	446	191	13,8*	83*	Lower alkalinity - possible non-dolomitic recharge or the effect of sulphide. Higher sulphate indicates pollution
Maloney's eye	202	122	4	5,6	Sulphate indicates slight pollution but little evidence at eye in spite of model indicating spread of sewage disposal (Barnard 1997)
Kuruman (partially dolomitic)	291	199	5	4,9	Clear evidence that recharge comes from two sources i.e. from dolomite and from Asbestos Mountain
Vlakfontein (100% dolomitic)	476	273	11	5,5	Pristine dolomitic spring - no contamination - recharge normal
Gt Koning (100% dolomitic)	419	240	10,2	4	Pristine dolomitic spring - Cl indicates normal recharge
Tsineng	731	395	53	15	Elevated HCO_3 indicating evaporative losses - Cl show lower recharge
Thabasikwa	711	396	41	28	Elevated HCO_3 due to evaporative losses - higher Cl indicate lower recharge or evaporation sink

* *Pollution present*

according to the substances they contain. The mixing ratios are varied until the concentrations of the chemical ingredients along the various flow paths linking different groundwater reservoirs match and also correspond to the chemistry of the outlet water at the spring.

11.3.2 Chemical interpretations of spring waters

Examination of the chemistry of the springs listed in Table 11.3.2.1 in terms of contamination and natural, pristine flow, reveals the following:

1) low sulphate concentrations (< 10 mg/l) in dolomitic aquifers is an indication of natural, unpolluted groundwater; and
2) low chloride concentrations indicate little contamination and could be used to derive the recharge;
3) low bicarbonate concentrations apparently indicate high recharge in dolomitic aquifers or that recharge is also contributed from a non-dolomitic aquifer(see Section 12);
4) plotting of chloride and sulphate on the same graph provides a useful comparison of the origin and extent of pollution (see Section 12.4.1)

11.3.3 Plotting of chemical data

11.3.3.1 Introduction

Plotting of ion concentrations against time often does not provide an adequate picture of the temporal change because of irregular sampling or missing data. In a normal plot of cumulative concentrations of chloride and sulphate, months of missing data show up as horizontal steps. Although this in itself is useful information, it does not give a good indication of changes in the concentration gradient. A new cumulative plot has been prepared in which months of missing data have been omitted. On the same graph the cumulative trend of the first 10 samples is compared to that of later samples. The deviation between these two cumulative trends would indicate whether the concentration of a specific ion has remained constant, or has increased or diminished. Presenting the measured concentrations on the second axis provides maximum information on one graph.

Plotting on a single graph the concentrations of the same ions in different springs oozing from the same geological formation, provides a comparison of possible anomalous characteristics of one system relative to others, as in Fig. 11.3.3.1.1. The high chloride concentrations of Bloemhof eye indicate that pollution has occurred, but the actual time series indicates a gradual improvement in quality i.e. decreasing concentrations of both sulphate and chloride (see Fig. 11.3.3.1.2). The latter graph shows a large scattering of points but yields a smoother graph on the new cumulative plot (see Fig. 11.3.3.1.3). The concentrations of chloride, reflecting pollution, cannot be used to determine recharge according to the chloride ratio method. The latter is a conservative chemical tracer and does not interact with the soil or aquifer matrix, but propagation of sulphate is usually delayed by adsorption in the soil. This delay is not present in the Upper Turffontein spring and probably reveals that the recharge mechanism is different (Fig. 11.3.3.1.4). It indicates that more direct recharge bypassing the soil zone is feeding the Turffontein eye, whereas greater interaction has taken place within the unsaturated zone in the case of Bloemhof recharge.

11.3.3.2 Dolomitic springs

The Upper Turffontein, Gerhard Minnebron and Maloney's eyes are three springs oozing from dolomitic aquifers in the same climatic environment, yet they show distinctly different chemical signatures according to the cumulative plots (Fig.11.3.3.2.1). The SO_4 concentrations of Upper Turffontein spring indicate a large increase from around 1979 when pollution had already started, and reached a high around 1990, clearly indicating contamination by sulphate which in normally unpolluted groundwater should be less than chloride. Interestingly the relative increase in chloride and sulphate concentrations on a normalized plot is very similar (see Fig. 11.3.3.2.1) indicating that the source of pollution contains a fixed ratio of these two constituents. This suggests that a single source is responsible for the pollution. The high sulphate concentration points to mining effluent from evaporation dams as the most likely source. Plotting the ratio of chloride to sulphate for different springs provides an indicator of the relative magnitude of sulphate pollution which can also be discerned from the cumulative plots of both ions on the same graph.

Disregarding some obvious outliers in Fig. 11.3.3.2.2, the Gerhard Minnebron spring has shown a steady increase in chloride concentrations from about 8 mg/l to 15 mg/l by 1992 whilst sulphate concentrations have increased from about 50 mg/l to 100 mg/l. The high level of sulphate at the start compared to chloride, as well as the steady increase in concentration, is clear evidence that the aquifer has been polluted from the start of the measurements. The cumulative plot would indicate the degree of additional contamination according to the deviation from initial conditions for Upper Turffontein eye.

Fig. 11.3.3.2.3 shows the low concentrations of both sulphate and chloride in the Maloney's eye, although according to an investigation by Barnard (1997) contamination of the aquifer has occurred due to sewage disposal but has apparently not yet reached the eye. Therefore judged from the chloride and sulphate concentrations the spring still represents almost uncontaminated water. Not only is this important for the estimation of recharge by means of the chloride method, but it explains why the ^{14}C concentration of the spring water is different from that of normal dolomitic aquifers, as will be discussed in Section 12.

However, the lower bicarbonate (HCO_3) concentration of Maloney's eye in comparison to that of other natural springs, such as Gerhard Minnebron and Buffelshoek eyes (see Fig. 11.3.3.2.4) suggests that part of its recharge is derived from a non-dolomitic aquifer. Such recharge could have occurred from the Magaliesberg or Witwatersrand quartzites, and, according to Barnard (1997), constitutes about 31% of the recharge; according to the chemical signature of Maloneys eye 80% of the recharge should originate from a non-dolomitic source which characteristically is low in bicarbonates. However there is also strong evidence (Bredenkamp, 1997) that low bicarbonate concentrations in dolomitic aquifers could result from fast recharge along zones of preferential recharge (see Section 12.4.3). The fact that virtually no pollutants have yet appeared in the spring water of Maloney's eye indicates either that contamination is still on its way and that the spring must have a large turnover time, or else that there is a significantly larger contribution of recharge from another source.

11.3.3.3 Other springs in the Bo Molopo area

The chemistry of the Renosterfontein spring shows that pollution (mostly by chloride) has occurred (see Fig. 11.3.3.3.1). The contamination entered the aquifer around 1981 whereafter concentrations increased to a peak in 1982. Thereafter both chloride and sulphate concentrations diminished exponentially. Chloride concentrations in 1995 were almost back to the starting levels of 1977, but sulphate concentrations have remained slightly elevated. It is uncertain what caused the pollution but it resembles a typical major influx of polluted water. It could not have been derived from fertilizer, but a more likely source of the pollution could have been the mining activity in the area. Although initially it was assumed that this aquifer is recharged from the dolomite, the lower bicarbonate concentrations indicate that most of the recharge is derived from a non-dolomitic source. This agrees with the topographical drainage of the area which indicates that the recharge area comprises deposits of alluvium/sand on the shales of the Transvaal Formations. The degree of pollution contrasts with the water quality of the Buffelshoek eye which still appears to be in a natural state although a slight relative increase in chloride concentration occurred from 1989 to 1991 (see Fig. 11.3.3.3.2).

The graph of Schilpadfontein (Fig. 11.3.3.3.3) indicates variable concentrations of sulphate and chloride. The chloride concentrations of Schilpadfontein improved slightly from 1981 and were more consistent from 1993 - 1995.

In the case of Polfontein only sporadic chemical measurements are available, covering the period 1981 to 1984, and 1990 to 1992 (see Fig. 11.3.3.3.4). Both chloride and sulphate concentrations are low, indicating a low level of pollution which is contrary to the general perception that significant pollution from the rural settlement around and township close to the eye has occurred. The lower chloride concentrations around 1983/4 are probably due to surface runoff entering the eye, while the relatively higher concentrations in 1990 could have been caused by a slight contribution of pollution and higher concentrations due to the severe drought at the time.

For the Upper Dinokana spring a longer series of sulphate measurements is available, but data for both chloride and sulphate only exist from 1989 to 1996. Because the chloride and sulphate concentrations during the latter part of that period responded in a similar way, and are low, the aquifer is regarded as being in a predominantly natural state (see Fig. 11.3.3.3.5). Hence for this spring, recharge could be derived using the chloride method (see Section 11.4).

In the case of the Molopo eye there are too few measurements of water quality to define a clear trend (Fig. 11.3.3.3.6). The variation in concentration is more likely caused by measurement variations or by fluctuations in recharge, than by pollution. Also the low concentrations of both chloride and sulphate indicate that this aquifer is in a natural state, and that a reliable estimation of the recharge, using the chloride method, could be obtained. The fact that the eye oozes entirely from a dolomitic aquifer supports the view that the chemical characteristics are those of a natural dolomitic aquifer.

As is discussed in Section 12 recharge from a non-dolomitic source or fast recharge to the saturated groundwater level causes the groundwater to be under-saturated in bicarbonates, which explains why no precipitation of calcite occurs at the outlets of these springs. In other parts of the world, carbonate in aquifers cause travertine depositioning (Issar, 1997). Unsaturated bicarbonate conditions also feature in the Pretoria Fountains, the Grootfontein springs and the Kuruman eye. The only spring indicating saturated bicarbonate dissolution and depositioning is the Thabasikwa eye near Reivilo (see Fig. 12.4.3.1.1). The alkalinity appears to have significant impact on the interpretation of ^{14}C and so-called dating of dolomitic groundwaters in the RSA.

11.3.3.4 Other types of aquifers

Uitenhage springs

For the Uitenhage spring fairly regular chemical data are available for the period 1980 to 1988. The cumulative plot indicates that the quality of the water remained intact (see Fig. 11.3.3.4.1) and is ideal for determination of recharge by means of the chloride method (see Table 8.3.2.2.2).

Toorberg springs

Two springs, referred to as Toorberg 1 and 2 were chemically analysed from about 1977 to 1991. The chloride concentrations of both springs were consistently low, except for a few obvious outliers, and less erratic than the sulphate concentrations. The chloride concentrations indicate that the springs are in a natural state, hence allowing the recharge percentage to be inferred by the chloride ratio method. The chloride concentrations in relation to sulphate are shown in the cumulative plot (see Fig. 11.3.3.4.2). The chloride concentration of the rainfall in the recharge area has to be inferred from the rainfall chloride graph shown in Fig. 11.4.1.1.

Mackies spring
Both chloride and sulphate concentrations in this spring are high in comparison to those of other springs but vary consistently except for a few outliers apparently resulting from poor measurements (Fig. 11.3.3.4.3). Application of the chloride ratio method in determining recharge is not reliable and the hydrological controls or pollution causing such high concentrations of chloride would first have to be investigated.

11.4 Determining recharge by means of the chloride ratio method

11.4.1 Introduction

As chloride is a conservative tracer and because it is introduced into an aquifer by the rainfall recharging the aquifer, the chloride concentration of groundwater in relation to that of rainfall can be used for quantitative estimation of recharge in Southern Africa, as has been indicated by several authors. Bredenkamp (1993), Sami (1991) and Gieske (1992) illustrated the application of the method to aquifers in South Africa and Botswana. The percentage rainfall representing recharge can be derived from the ratio of the chloride concentration in rainfall relative to that of the deeper unsaturated zone or first groundwater struck. In mathematical form the relationship can be represented as follows:

$$RE\ (\%) = Cl_{rain}/Cl_{groundwater} * 100 \quad \text{eq. 11.4.1}$$

The method can be applied by analysing the chloride concentration of soil moisture extracted from the unsaturated zone, or of shallow groundwater samples from boreholes. The method can be used to estimate recharge by using the chloride concentration of spring water, which integrates the recharge both spatially and in time. A specific requirement is that no chloride has been added by dissolution of aquifer material, or from salts contained within the aquifer matrix, or has entered the aquifer via pollution. The chloride injected by the recharging rainfall should only have been concentrated by the natural, hydrological, evaporative processes.

The accuracy of the chloride method obviously depends on the reliability of measurements of the chloride concentration of the input rainfall. This has been poorly monitored in the RSA and chloride measurements have been obtained at only a few stations and for limited periods. Fairly reliable data are available for Botswana, but in this report the concentrations have been represented differently from the method used in the manual on recharge. Instead of plotting the total chloride precipitation expressed as mg/sq m or Kg/ha, the concentrations (mg/l) have been used. The available data conforms to a definite pattern as is indicated in Fig. 11.4.1.1. This figure was derived from the average chloride values of Botswana, some measurements obtained for Pretoria and Bloemfontein, and chloride values for rainfall in the Zululand area. The main contribution of data from Botswana is represented in Fig. 11.4.1.2, indicating that the average chloride loads increase with increasing average annual rainfall. According to the chloride values obtained for the coastal regions of KwaZulu the chloride loads decrease exponentially as the distance from the sea increases (see Fig. 11.4.1.3). The coastal and inland rainfall samples show similar trends of chloride concentrations in relation to rainfall, running parallel to the Botswana data, but average concentrations are higher in coastal areas (see Fig. 11.4.1.1).

The relationship shown in Fig. 11.4.1.4 clearly indicating lower chloride concentrations (mg/l) for higher annual rainfall is still in agreement with higher chloride loads for higher rainfall, shown in Fig. 11.4.1.2. However, further measurements of chloride in rainfall are necessary to confirm the overall recharge relationship produced in Fig. 11.4.1.5. In the interim, application of the chloride method should not be based on chloride concentrations of individual rainfall measurements but of the average rainfall, because the groundwater levels responds to the average rainfall over several years according to the MA method. The unsaturated zone would even out temporal differences in chloride concentration, and groundwater oozing from springs represents a mixed sample integrated both spatially and in time, therefore yielding a more reliable average recharge of the spring drainage area.

The scarcity of specific measurements of chloride concentrations of rainfall in the RSA emphasizes the need to establish representative sampling. Monthly totalling rainfall samples obtained at groundwater monitoring stations could help to improve the collection of good data on the spatial and temporal variations of chloride, even if such measurements could only be carried out for a few years. Therefore cumulative rain gauges to measure both monthly rainfall amounts and for chloride determination should be installed at monitoring points which provide a good areal distribution.

It is also recommended that data of chloride measurements obtained in afforested areas should be analysed in greater detail as this would shed light on the rainfall-chloride and rainfall/runoff relationships. It would at the same time provide a direct means of quantifying the impact of afforestation on groundwater recharge (base flow)if water samples were to be collected below afforested areas.

11.4.2 Assessment of the recharge of springs in relation to rainfall, using the chloride method

Application of the chloride method to determine spring recharge in relation to rainfall can be illustrated very well in the case of various springs, provided they have not been contaminated. This emphasizes the importance of regular chemical analysis of spring water to confirm that natural conditions still prevail - a pre-condition to application of the chloride method.

Springs originating from the same geological formations are grouped together, and their recharge is derived by way of the chloride method. This is then plotted in relation to the average rainfall to derive a regional rainfall/recharge relationship which could be applied as a first approximation in groundwater studies (see Fig. 11.4.2.1 representing dolomitic aquifers). At this stage the composite graph (Fig. 11.4.1.5) is still regarded as a useful first approximation of recharge in relation to the average annual rainfall for the country as a whole.

11.5 Chemical sampling in boreholes

11.5.1 Introduction

Chemical monitoring of groundwater is an ongoing activity and many boreholes are being sampled but the data lack proper interpretation. The main purpose of these measurements is to establish reference levels of ambient water quality for the identification of pollution. Unreliable sampling could introduce spurious quality variations, complicating interpretation and the confirmation of pollution. This aspect, however, does not fall within the scope of the present project, but it is logical that sampling at points of abstraction would provide a more consistent picture of the chemical characteristics and of temporal changes of the groundwater than would be obtained by spot sampling. In the following section some interesting chemical results will be discussed.

11.5.2 Water quality of the Zuurbekom well-field

Fig. 11.5.2.1 represents the temporal change in groundwater quality at the Zuurbekom pumping station which was the first water supply scheme of the Rand Water Board. Chemical analyses have been performed of water samples collected from more than one production borehole, hence providing an integrated chemical picture similar to that of a spring. The graphs clearly show a large increase in ion concentrations, with sulphate as the main contaminant. Sulphate is a product of the gold-mining industry and has been released over many years in effluent, or has infiltrated from surface impoundments or slimes dams. In the Far West Rand some of the sulphate ends up in the streams and is transported along the Klip River, polluting the dolomitic aquifers along its course.

The chemical data of Zuurbekom boreholes, starting in 1956, is of great value when examining the temporal changes in the quality of dolomitic groundwaters and springs e.g. the Upper Turffontein eye, the Gerhard Minnebron and Maloney's eye. The Zuurbekom chemical record has been used in modelling pollution transport (Simonic, 1993). According to this model the pollution is derived mainly from the Klip River, and a good simulation of the concentration levels has been achieved (see Fig. 11.5.2.1). This figure also indicates the high concentration levels of sulphate in the Klip River.

Zuurbekom water quality in relation to the Cumulative Rainfall Departures (CRD) series

The CRD series of rainfall plotted against the sulphate concentrations (see Fig.11.5.2.2) indicates good correspondence with a two year lag between the two series, the latter being plotted inversely. It shows that the pollution levels of the groundwater increase when the natural recharge is low (actual CRD becoming more negative). The high concentrations occur when groundwater flow is directed towards the well-field whereas the flow is normally towards the river. Hence the sulphate concentration is diluted during times of high natural recharge, which coincide with a rise in the groundwater levels and lower influx of river water. The concentrations of sulphate in the rivers would also decrease during periods of high flow, although the total load of pollutants would increase.

Incorporation of the two year lag between the CRD values and the sulphate response increases the correlation coefficient from 0,35 to 0,6 (see Fig. 11.5.2.2). This could partly be due to retardation of sulphate ions caused by ion exchange, but is probably mainly due to later arrival resulting from lateral flow. The adsorption process effects the release of chloride ions and hence the chloride concentrations should also increase with increasing sulphate levels. That this clearly is the case in dolomitic springs can be seen by comparing the cumulative plots of chloride and sulphate on the same graphs. In cases where the chloride and sulphate concentrations respond differently, the pollution must be from an independent source or be due to a different mechanism of recharge. Hence in the case of the Upper Turffontein eye the predominance of sulphate indicates that the bulk of the recharge is occurring from the Wonderfontein spruit and not as diffused recharge via the unsaturated zone over the entire catchment. The constant chloride/sulphate ratio shown in Fig. 11.3.3.1.4 even at higher concentrations indicates a uniform source of recharge, probably infiltration or effluent from slimes dams.

11.5.3 Water quality of the Centurion Production Boreholes and the Pretoria Fountains

Chemical monitoring of three production boreholes in the Centurion area has been carried out since 1988, providing one of the best continuous records of chemical changes within the major dolomitic aquifer which feeds the Pretoria Fountains. The positions of the boreholes are shown in Fig. 5.4.1.1 and their chloride concentrations are plotted in Fig. 11.5.3.1. This shows slightly declining chloride concentrations for boreholes ZP13 and ZP16 but variable concentrations of chloride in the case of the Kentron borehole. The high concentrations compared to those of the Pretoria Fountains (see Fig. 11.5.3.2) proves that contamination of the aquifer in the Centurion area is under way, but fortunately the concentrations are diluted by predominantly uncontaminated water which is still feeding the Fountains. There has been an improvement in the quality of the water over the period 1988 to 1995. The fact that the chloride graphs of ZP13 and ZP16 run parallel indicates a similar temporal change due to consistent recharge and mixing. However, the chloride concentrations of the Kentron borehole - which captures polluted water during pumping - depend on the rate of abstraction, and probably on the variable rate of recharge from the spruit nearby. The cumulative plots of abstraction against chloride concentration of the Kentron borehole show a great similarity (Fig. 11.5.3.2) indicating that during periods of high abstraction more polluted water is drawn in by the pump. This could be due to either an influx of poorer quality water from greater depth or capturing a mixed concentration between that of ZP13 an ZP16 (see Fig. 11.5.3.1).

Fig. 11.5.3.3 representing the water quality of the Pretoria Fountains, clearly shows a small but significant increase in chloride and total dissolved solids (TDS) around August 1995, which is not evident in the case of sulphate. The water samples which have been analysed, however, represent a mixed water sample of three springs, viz. Sterkfontein, and the Upper and Lower eyes, sampled at the collecting chamber in the Fountains Valley. Hence the chemical changes of the individual springs cannot be discerned. Evidence of long-term change in the water quality of the Pretoria Fountains was apparent from chloride measurements presented by Vegter (1993). The first chloride measurements on record indicate very low concentrations of chloride (Cl = 1,7 mg/l) and that pollution had lately occurred, but measuring techniques for such low concentrations could have been unreliable at the time. As is clear from Fig. 11.5.3.1 (the Centurion boreholes), and Fig. 11.5.3.2 the Pretoria Fountains have been slightly contaminated and therefore the entire series of chloride concentrations has to be inspected to quantify the recharge by means of the chloride ratio method. Using the lowest level of chloride in Fig. 11.5.3.3 the recharge of the Pretoria dolomitic aquifer is estimated at 8% of the average annual rainfall, but based on the chloride values reported by Vegter (1993) the recharge is about 50% of the rainfall, which is too high. A recharge of 12% was used in the hydrodynamical model compiled by Bredenkamp (1996). In the case of the Rietvlei springs pollution is still small (see Fig. 11.5.3.4) and disregarding some outliers the average recharge based on the chloride method is about 25% of the rainfall.

11.6 Conclusions and recommendations

1. The natural chemistry of groundwater, especially that of springs which integrate recharge both spatially and in time, provides a means of identifying the source of the recharge. It provides proof that several of the dolomitic springs are partially being recharged from non-dolomitic aquifers. Pollution can be discerned from the temporal variation of the chemical compositions, and probable sources of contamination can be identified.
2. As has been illustrated in the case of the Zuurbekom samples, the negative of the CRD series, although representing a simple hydrological recharge model, corresponds to the change in water quality. This provides a means of assessing in a simple way the effect that higher or lower recharge from rainfall would have on the water quality in comparison to effects of pollution. Interaction between sulphate and the aquifer matrix is revealed by comparison of the lag between changes in sulphate concentrations and those of chloride.
3. As will be indicated in Section 12 the bicarbonate concentrations of dolomitic springs could indicate that preferential recharge is occurring. Both admixing and preferential recharge would affect the classical model of carbon interaction in

dolomitic aquifers which will affect the interpretation of ^{14}C in groundwater, especially radiocarbon dating of groundwaters.

4. With regard to the estimation of recharge, the chloride method can be applied with a high degree of reliability if the spring water has not been contaminated. In cases where contamination does occur the earliest measurements could be used. It is however important that the chloride concentrations of the rain input should be reliable. Although chloride measurements obtained from Botswana and elsewhere indicate that there appears to be a simple relationship between rainfall amounts and the average concentrations of chloride, more measurements of rainfall chloride have to be obtained. Such sampling of rainfall needs to be carried out for a few years, but obtaining integrated monthly samples would be adequate. Logistically such sampling would fit in with the proposed collection of monthly integrated samples of groundwater at selected monitoring points throughout the country.

12. Environmental Isotope Monitoring

12.1 Overview

The isotopic ratios of certain elements in the hydrological system are modified by processes such as fractionation and radioactive decay. These are used as tracers of natural water whilst large releases of artificial tracers in rainfall due to nuclear testing have provided a means for elucidating hydrological processes and flow paths. The monitoring, reliability and significance of these tracers are of importance in groundwater studies, but in many instances interpretation has proved difficult and inconclusive.

In natural waters the most important environmental isotopic tracers are stable isotopes {deuterium (^2H) and oxygen-18(^{18}O)} and radioisotopes {carbon-14(^{14}C) and tritium(^3H)}. The introduction of artificial tritium and ^{14}C by nuclear bombs has increased the natural concentration levels in rainfall and atmospheric CO_2, as indicated in Fig. 12.1.1. Whereas ^{14}C and tritium concentrations prior to the advent of bomb tritium and ^{14}C could be used to determine the age of groundwater from the decay of the radioisotope, elevated concentrations of tritium and ^{14}C in the groundwater indicating the presence of bomb-tracer proved that part of the groundwater recharge occurred after the introduction of the artificial tracer into the atmosphere. However the interpretation of ^{14}C concentrations in groundwater is in some instances uncertain and inconclusive and certain discrepancies which have been observed are discussed in the present report. Although tentative explanations have been put forward in this report an in-depth examination would be required to resolve the matter satisfactorily.

12.2 Radioisotopes

Dating of groundwater involves determining the decay-time of the radio-nuclide by the following equation:

$$C_t = C_o e^{-0,693t/T_{½}} \quad \text{eq. 12.2.1}$$

where
C_t = the concentration after a time interval of t
t = the time lapse, also referred to as the "age" of the water
Co = the initial concentration
$T_{½}$ = the half-life of the radio-nuclide (tritium or ^{14}C)

Depending on the mixing model used the time interval t allows the average turnover time of water in an aquifer to be determined. However, in practice results proved to be suspect because they yielded larger turn-over times compared to estimates based on water balance interpretations. The following complicating factors associated with the input, mixing during flow and obtaining representative samples are encountered:

- The initial natural concentration of these isotopes has been disturbed by inputs from nuclear bomb tests (see Fig. 12.1.1).
- Carbon-14 is being introduced into aquifers via the uptake of atmospheric carbon dioxide by plants and CO_2 releases in the soil zone, and the dissolution of carbonates of zero ^{14}C *concentration, in the soil or in the aquifers. The dissolution equation* indicates that starting with an initial concentration of 100% modern carbon (mc), representing the ^{14}C concentration in the atmosphere, the ^{14}C concentration of the bicarbonate in the groundwater should be 50% mc. However, the initial ^{14}C concentration of recent waters reaches 85% mc and higher as a result of the exchange-feedback equilibrium between CO_2 generated by the plant root zone, and bicarbonate in solution. This initial ^{14}C concentration is therefore somewhat uncertain and could depend on the degree of "openness" between the groundwater system and the soil CO_2-atmosphere.
- Variable mixing of waters of different ages could occur especially in heterogeneous aquifers which are being exploited, making it difficult if not impossible to obtain representative samples from pumping boreholes.
- Apart from the radioactive decay the ^{14}C concentration could be reduced by ^{14}C uptake during dissolution of calcite and by exchange with dead carbon (^{14}C = 0% mc) from the aquifer matrix. This could occur in an aquifer located in dolomite, limestone or calcrete, or non-dolomitic groundwater admixing or entering a dolomitic /limestone aquifer. Thus the resulting ^{14}C concentration could have been changed other than by radioactive decay and would be interpreted as having "ages" which are too high.

Tritium being a conservative tracer because chemically it behaves identically to stable hydrogen atoms in water molecules, should reflect the true flow picture of the groundwater if a representative sample has been obtained. However as a radioactive tracer the tritium unfortunately covers a much shorter time span than ^{14}C, but the higher bomb tritium and ^{14}C could be used as a tracer of recent water.

In view of various uncertainties associated with the dating of groundwater the focus has shifted to using these environmental isotopes as tracers incorporating the presence of the "bomb" component to reveal mixing and natural hydrological processes in groundwater (Verhagen et al., 1979 and 1991).

12.3 Stable isotopes

The significance of stable isotopes present in water i.e. deuterium (^{2}H) and oxygen-18 (^{18}O), lies in their different kinetic behaviour during evaporation/condensation of water, while ^{13}C could provide evidence whether or not carbonate exchange occurs in an aquifer or it could indicate different climatic conditions at the time of recharge (Vogel, 1963). The stable isotope concentrations are expressed as ratios which are expressed as deviations relative to a standard reference, according to the relationship:

$$\delta_{sample} = \frac{(R_{ref} - R_{sample}) * 1000}{R_{ref}}$$

where R is the isotopic ratios of $^{18}O/^{16}O$ or $^{2}H/^{1}H$ or $^{13}C/^{12}C$.

In the case of deuterium and ^{18}O the reference used is Standard Mean Ocean Water (SMOW). The stable isotope character of water which has condensed or evaporated in the hydrological cycle under natural equilibrium conditions, is illustrated by Fig. 12.3.1. This indicates a standard relationship between the deuterium and ^{18}O concentrations compared to SMOW with a slope of 8. Disregarding other effects, the average stable isotopic concentration of the water conforms to the average temperature of a specific locality, concentrations being more depleted in heavy isotopes the lower the temperature, and vice versa. Groundwater recharged by rainfall which, after precipitation, has been subjected to evaporation, is enriched in heavy isotopes due to preferential evaporation of the lighter isotope, causing a shift to an evaporation line (Fig. 12.3.1) with a slope of 4-5. Groundwater with this characteristic provides evidence that recharge in the semi-arid Kalahari must have originated from local infiltration (Verhagen et al., 1979; Vogel and Bredenkamp, 1969; and Gieske, 1992). The oxygen–18 ($\delta\,^{18}O$) and deuterium ($\delta\,^{2}H$) measurements therefore provide evidence of the origin of groundwater, as has been illustrated for Sishen (see Section 12.4.3) where some confined groundwater yielding artesian flow, appears to have been recharged during the last glacial period.

Assimilation of CO_2 by plants occurs by means of the Calvin (C3-cycle) or the C4-cycle (Hatch-Slack cycle) {Clark and Fritz (1979)} which appears to be a characteristic of the plant species Vogel 1963). Plants which have adapted to a colder Mediterranean type climate photosynthesize by means of the C3 cycle and yield ^{13}C ratios which are more depleted (lighter) than those of C4 plants (Vogel, 1993). These more depleted ^{13}C compositions are also reflected in the groundwater which has been recharged during prevailing climatic conditions. Thus older groundwater derived from the previous glacial period must bear the ^{13}C signature of that climate.

^{13}C concentrations on the other hand could provide an indication of whether the groundwater has been subjected to exchange with carbon from the aquifer. Groundwater would be lighter in ^{13}C composition than water having undergone exchange. Higher ^{13}C ratios (less negative) encountered, for example, in deeper/older water, could provide an indication that exchange with carbon in the aquifer has occurred. The apparent ^{14}C ages of such waters would therefore be too high and would yield too low estimates of recharge which would be too low and would have to be corrected for matrix exchange.

12.4 Estimation of recharge

12.4.1 Overview

Some simple interpretations based on the ^{14}C and tritium (^{3}H) concentrations could be made. It could provide

- a qualitative indication of recharge, as the presence of bomb ^{14}C or tritium would confirm that recent recharge has occurred locally;
- a measure of the turn-over time of the aquifer (ratio of storage to recharge) for well-mixed aquifers;
- an estimate of recharge by applying a relationship between recharge and the increase in age of the water with depth, derived by Bredenkamp and Vogel (1970). For such application it is essential, but difficult, to obtain reliable water samples from specific depths and to determine the storativity of the aquifer. Likewise a representative mixture of groundwater is not necessarily obtained when pumping from an aquifer, especially in view of differential mixing of contributions from different horizons and fractures in a heterogeneous aquifer. In a lumped-parameter approach, e.g. an exponential well-mixed model, reliable estimation of recharge, based on the increase in age with depth in an aquifer, has been rather qualitative (Bredenkamp and Vogel, 1970). In the present project the derivation of reliable quantitative estimates of recharge, based on the ^{14}C measurements of groundwater in the Sishen area, has been re-examined (see Section 12.4.3).

12.4.2 Difficulties in the interpretation of ^{14}C in dolomitic springs

Measurements over several years of different dolomitic springs in the RSA, have revealed that the ^{14}C concentrations of some springs have not responded as expected, e.g.:

1) The ^{14}C concentration in the range of 70-90%mc of several springs has remained surprisingly constant at each spring in spite of the introduction of bomb ^{14}C and an almost direct response of spring flows to high rainfall. An even better example is the Grootfontein spring in the Pretoria area (see Fig. 12.4.2.1) of which the ^{14}C remained close to 58%mc and experienced only a small peak in the ^{14}C concentration during the period 1993 - 1994. The best correspondence between the ^{14}C concentrations and the flow of the spring is obtained when the ^{14}C measurements are shifted backwards by 48 months and the natural flows of the spring are used. Similarly the ^{14}C concentration of Maloney's eye has remained surprisingly constant even though its flow responds in accordance with the average rainfall over 60 months, indicating active recharge (Fig. 12.4.2.2a). The best correspondence between ^{14}C and the flow is attained when the measured ^{14}C concentrations are lagged by 67 months (see Fig. 12.4.2.2b).

2) In several of the low and high-flowing springs the ^{14}C concentrations are higher than 85% mc, indicating the addition of recent recharge, but the presence of bomb ^{14}C is barely related to the average flow of the springs which is determined by the size of the aquifers. (see Fig.12.4.2.3).

3) A clearly detectable pulse of higher ^{14}C has only been observed in the case of Kuruman eye (see Fig. 12.4.2.4 after Bredenkamp et al., 1992). This is ascribed to a proportion of recharge containing bomb ^{14}C which was contributed from the western limb of the recharge catchment during the high rainfall period of 1974 - 1977 (Bredenkamp et al., 1995). Under normal conditions of flow the contributions from the two sub-catchments remain in balance according to the piezometric heads prevailing in each of the recharge areas. However, during the high recharge period the western sub-catchment received more recharge, causing the mixing ratio of spring water to be temporarily shifted towards a higher contribution of bomb ^{14}C and tritium. As the imbalance of the piezometric heads (recharge) equalizes over a period of time, the ^{14}C concentrations ultimately return to a more "normal equilibrium" value.

12.4.3 ^{14}C concentrations in dolomitic aquifers in relation to alkalinity

12.4.3.1 ^{14}C versus alkalinity in dolomitic springs

Talma et al. (1996) noticed that the alkalinities of dolomitic springs with low ^{14}C concentrations are lower than those with high ^{14}C (see Fig. 12.4.3.1.1), and Bredenkamp (1997) postulated that the alkalinity provides an indication of the rate/mechanism of recharge. This is supported by Fig. 12.4.3.1.2 in which recharge determined by means of the chloride method has been plotted against alkalinity. The graph indicates that higher recharge is related to lower alkalinity, and vice versa, indicating that the recharge of springs with low bicarbonate concentrations occurs more rapidly. This supports a hypothesis of rapid recharge whereby the water bypasses the carbonate reservoir of carbon dioxide in the unsaturated zone. This leads to under-saturation of dissolved HCO_3, resulting in an incomplete equilibrium level of ^{14}C which is normally attained between the recharge water and the soil CO_2. This represents an essentially closed system (Clark and Fritz, 1979) whereby the groundwater would then dissolve carbonate of zero ^{14}C concentration from the dolomite rock or soil. This would result in ^{14}C concentrations of lower than 85% mc (higher apparent age) in spite of the water having infiltrated virtually directly. However, the ^{14}C concentration of water emanating from a spring naturally depends on both the alkalinity of the water and on the turnover time of water in the aquifer system. For springs with high alkalinity and small turnover times the ^{14}C concentrations would conform more closely to the bomb signature, because the ^{14}C concentrations of the aquifer interface are gradually increased by reverse-exchange taking place between ^{14}C concentrations of the water and the aquifer matrix.

The relationship observed between alkalinity and ^{14}C concentrations (see Fig. 12.4.3.1.1) not only accords with this hypothesis but is also in agreement with observations and comments made by Issar (1997) regarding anomalies in the chemistry of the majority of dolomitic springs in the Republic of South Africa. He pointed out that these springs react differently from similar springs in other parts of the world, in that no depositioning of limestone occurs at the points of discharge. This implies that dolomitic waters in the RSA remain essentially unsaturated in bicarbonate because the partial pressure of the CO_2 - bicarbonate dissolution of the waters discharged by the springs is not in equilibrium. Incomplete saturation has been confirmed by Talma (1999 in press). In terms of the new hypothesis the recharging water remains partially saturated in bicarbonate and the degree of saturation will depend on the extent of exchange attained during infiltration through the unsaturated zone. Further dissolution of carbonate would occur depending on the presence of carbonates and on the partial pressure of CO_2 in the water. Thereafter during the course of flow through the aquifer, exchange of ^{14}C with carbon in the aquifer could take place. If the turnover time in an aquifer is small (low flowing springs), higher ^{14}C concentrations would be attained in the new equilibrium established between ^{14}C and the aquifer. This is likely to be a complex process but results in ^{14}C concentrations of the water discharged by smaller springs being higher than that indicated by the base line in Fig. 12.4.3.1.1. However, if during the recharge process a high alkalinity is already established when the water enters the saturated aquifer zone, modification of the ^{14}C concentrations would occur mainly via exchange. In more arid areas the depth to groundwater is greater and the root zone in which CO_2 is generated extends deeper. This causes longer exposure of infiltrating water to biogenic CO_2 in the unsaturated zone, causing springs in this climate to be less depleted in ^{14}C through exchange, than in high rainfall areas.

Under-saturation of bicarbonate could result from rapid recharge, but also from recharge contributed from a non-dolomitic aquifer, shallow soils and absence of readily soluble carbonate. This water could then dissolve carbonates with which it subsequently comes in contact, as would occur in a dolomitic aquifer, then yielding ^{14}C concentrations closer to half that of the atmospheric concentrations of ^{14}C. This appears to be the case in Maloney's eye, Pretoria Rietvlei (Grootfontein eye) and Kuruman eye. Hence the final ^{14}C concentration of the spring water would be lower than the values that would result according to the classical/theoretical uptake of ^{14}C in dolomitic aquifers. The present explanation, coupled with the size of the recharge area and the contributions of non-dolomitic recharge, appears to fit all of the ^{14}C measurements of water obtained from the various dolomitic springs, as is illustrated by the following examples:

- Only in the case of the Thabasikwa eye (No. 25 in Fig. 12.4.3.1.1) which has the highest alkalinity, has complete saturation of bicarbonate been attained, causing the depositioning of limestone at the point of outflow, and high ^{14}C concentrations which are consistent with a small turnover time.
- The low bicarbonate concentrations of both Grootfontein (Pretoria dolomite) and Maloney's eyes, are caused partly by recharge contributed from non-dolomitic aquifers and partly by rapid recharge. This also applies to the main spring of Kuruman but not to the second eye at Kuruman, which is fed entirely from a dolomitic aquifer.
- In the case of the Gerhard Minnebron and Turffontein eyes it can now be explained why the ^{14}C concentrations conform to the relationship indicated by Fig. 12.4.3.1.1 despite pollution which has been introduced from mining effluent without significantly affecting the ^{14}C concentrations. The acidified polluted water re-entering the dolomitic aquifer dissolves carbonate of zero ^{14}C concentration diluting the bomb ^{14}C which has been added by contact with the atmosphere in the evaporation/slimes dams.
- The fact that the initial ^{14}C concentration of dolomitic groundwaters appears to be related to the alkalinity of the water, and is affected by admixing of waters of different ^{14}C concentrations, has a profound effect on the reliability of "age" determination (refer to eq. 12.2.1) which assumes that radioactive decay alone has been responsible for the decline in ^{14}C concentration.
- The new approach indicates that virtually all the dolomitic springs in the RSA, notwithstanding large differences in ^{14}C concentrations, discharge water of fairly recent origin, despite low tritium concentrations. The low but detectable tritium levels being due to the admixture of a large volume of water older than 40 years

but still being young according to ^{14}C. The Tsineng spring which does not fit the general relationship indicated in Fig. 12.4.3.1.1, appears to be the only case of truly older water. However its apparently higher age could have resulted partially from carbon exchange in the aquifer and further measurements are required.

12.4.3.2 Verification of ^{14}C behaviour in other dolomitic aquifers

12.4.3.2.1 Boreholes in the Sishen-area

The characteristics of groundwaters in the Sishen area have also been examined in the course of verifying the hypothesis that low ^{14}C concentrations occur in groundwaters which are rapidly recharged. Several isotopic and chemical analyses are available from an investigation of groundwater in the Sishen iron-ore mining area which consists partly of a dolomitic aquifer and partly of conglomerates, banded ironstones and Kalahari sediments – see Fig. 12.4.3.2.1 (after Verhagen et al., 1979). The chemistry of the groundwater was used to establish if waters of different characteristics exist. As chloride is a conservative tracer, the chloride concentrations were plotted against alkalinity for the entire set of samples involving a large range of geohydrological conditions and aquifers. This shows a spatial distribution with a some outliers of high chloride in relation to bicarbonate (see Fig. 12.4.3.2.2). The ^{14}C concentrations plotted against depth of water strike of all boreholes indicate a trend of lower ^{14}C concentrations at greater depth, as would be expected (see Fig. 12.4.3.2.3), but the scattering of points is high. It is clear that subgroups derived from similar hydrogeological environments should be examined on their own. Therefore boreholes from Sishen and those from the Telele profile have been analysed separately.

A plot of ^{14}C concentrations of Sishen samples against alkalinity (Fig. 12.4.3.2.4) shows the results from groundwater samples spread over a large area. These samples, however, do not show a specific pattern but the samples selected from the Sishen open-mine quarry indicate that the higher ^{14}C concentrations are associated with lower bicarbonate (Fig. 12.4.3.2.5). Although these samples do not appear to fit the hypothesis put forward in Section 12.4.3.1, this groundwater appears to represent direct recharge which has had little interaction with carbonate rocks. That is feasible considering that a large proportion of recharge occurs when the Gamagara River is in flood. The groundwater samples of the Sishen mine and the cluster representing the artesian waters from the farm Woon indicate a slight decline of ^{14}C concentrations and bicarbonate concentrations at greater depths. The lower alkalinity (see Fig. 12.4.3.2.4) therefore does not manifest uptake of carbon from the matrix at greater depth but rather proves that the deeper artesian waters are derived from rainfall which probably recharged during the last glacial period, as is evidenced by the depleted ^{13}C ratios of the groundwater. To further the investigation the characteristics of groundwaters from the Telele profile, also representing a homogeneous hydrogeological regime, have been examined (see Fig. 12.4.3.2.1).

Telele boreholes

A series of exploration boreholes has been drilled across an ancient Kalahari trough at Telele to investigate the nature of the groundwater (Verhagen et al., 1979). Some of the boreholes (Telele-shallow) intersected water in the upper (shallow) Kalahari sediments lying above almost impermeable red clay (see Fig. 12.4.3.2.6). Underneath the clay, confined water was intersected in the so-called basal aquifer (Telele-deep). The isotopic composition indicated that this artesian groundwater is much lighter in deuterium and ^{18}O composition than contemporary water, signifying a colder climate, and therefore have probably been recharged during the last glacial period.

According to a plot of chloride against alkalinity (Fig. 12.4.3.2.7), exploration boreholes drilled along the Telele profile fall in two groups, namely one with chloride concentrations higher than 600 mg/l (Telele high Cl) and the second having less than 400 mg/l (Telele lower Cl). Fig.12.4.3.2.8 shows that if the higher chloride concentrations are plotted against the depth of water strike they still form a separate cluster but borehole 104 appears as an outlier. The lower chloride samples show a slight decrease in chloride with depth but chemically could be regarded as the same water and form a good set for examining the isotopic composition of the groundwater. The number of samples in the higher and lower chloride category has changed, as not all boreholes have been analysed for the same chemical ingredients or isotopic compositions.

The following graphs have been prepared:

- ^{14}C vs. depth of water strike (Fig. 12.4.3.2.9) indicating a decline of ^{14}C with depth, corresponding to an increase in age with depth. In this plot a distinction is made between the ^{14}C concentrations of the Telele samples of high and low chloride concentrations but they all seem to fall roughly on the same line. Dependence of ^{14}C on distances from the Gamogara appears to be small.

- ^{14}C vs. alkalinity (Fig. 12.4.3.2.10) : This graph indicates a linear decrease in bicarbonate associated with declining ^{14}C concentrations, which is in accordance with the hypothesis of lower ^{14}C concentrations at lower alkalinity - attributable to exchange, dilution or decay of ^{14}C.

12.4.3.3 Occurrence of carbon exchange

Confirmation as to whether or not carbon exchange has taken place in the aquifer could be obtained from the relationship between ^{13}C and ^{14}C and from ^{18}O in relation to ^{14}C or ^{13}C.

^{13}C - ^{14}C relationship

The Telele samples (Fig. 12.4.3.3.1) indicate a rather scattered distribution of points but the overall trend is that ^{13}C concentrations are more depleted for lower ^{14}C concentrations, which yields no evidence of carbon exchange or losses. It rather indicates that the groundwater below the Kalahari Red beds stems from recharge which occurred during the last glacial. Not only is the quality of the water better, indicating higher recharge, but during this period winter-type rainfall prevailed in the area (Vogel, 1993 and Talma, 1999 in press). This could explain why the ^{13}C concentrations are more depleted, namely because photosynthesis by vegetation at lower temperatures and in typical winter-rainfall conditions differs from that in summer rainfall, assimilating higher concentrations of ^{12}C and less ^{13}C.

Fig. 12.4.3.3.2 (^{13}C vs. alkalinity) shows no pattern between ^{13}C and ^{14}C values, but Fig. 12.4.3.3.3 clearly shows a trend of ^{13}C concentrations declining with depth. There is however no evidence of $\delta^{13}C$ concentrations becoming less negative (heavier) which should have occurred as the depth (age) increases, if exchange has occurred.

^{18}O concentrations

The plot of ^{18}O (Fig. 12.4.3.3.4) vaguely suggests that the ^{18}O composition of the water tends to be lighter (more negative) for lower ^{14}C concentrations. The ^{18}O concentrations therefore also do not indicate that carbon-exchange has occurred in the aquifer with time. This is also evidenced by Fig. 12.4.3.3.5 indicating ^{18}O plotted against the depth of the water strikes.

12.5 Estimation of recharge based on increase in age with depth

12.5.1 Age-depth relationship

A relationship between recharge and the age of groundwater in relation to depth has been derived by Bredenkamp and Vogel (1970), yielding the following equation for an aquifer of which the storativity decreases logarithmically with depth:

$$RE = S_o.d/t \quad \text{eq. 12.5.1.1}$$

where
RE = recharge
S_o = aquifer porosity at the water level
d = depth of water strike
t = age of water at depth d.

Based on an assumed porosity of 0,20 for the Telele aquifer, Eq. 12.5.1.1 yielded a value of recharge of 1,6 mm per annum (see Fig. 12.5.1.1).

However, it is more likely that the porosity of the aquifer does not decline with depth, in which case the following relationship should be used to derive the recharge:

$$RE = S_o.H/t.\ln(H/H\text{-}d) \quad \text{eq. 12.5.1.2}$$

where H = the total thickness of the aquifer and other terms as defined in eq. 12.5.1.1.

Assuming a total aquifer thickness as 150 m and aquifer storativity of 0,2 an average recharge of 1,6 mm per annum is obtained from this equation (see Fig. 12.5.1.2).

These recharge estimates are fairly consistent but seem lower than those derived from the depth of tritium penetration in the unsaturated zone which indicate recharge values between 5,5 to 17% of rainfall (16 - 50 mm) in the area (Verhagen et al., 1979). Verhagen stated that these estimates could be too high depending on water losses which could still occur from the unsaturated profile at depth. It would be realistic for recharge below the red clay to be very small.

12.6 ^{14}C and ^{13}C relationship in dolomitic aquifers based on time series monitoring of springs and of large abstractions in the Sishen mining area

12.6.1 Introduction

Isotopic measurements of the outflow of different dolomitic springs carried out over many years, have shown that the most consistent relationship appears to be that between ^{14}C concentrations and alkalinity, namely higher ^{14}C concentrations for higher bicarbonate concentrations (see Fig. 12.4.3.1.1). However, a similar pattern has not been revealed by the ^{14}C and ^{13}C concentrations of groundwater derived from boreholes. This is probably due to the fact that the composition of the water depends on the extent of mixing of waters with different characteristics due to differential contributions during abstraction and recharge, especially in heterogeneous aquifers such as dolomite. Hence the characteristics of groundwater abstracted from the Sishen open-pit mining site, previously studied by Verhagen et al. (1979), have been re-examined.

12.6.2 Sishen groundwater

According to Fig. 12.6.2.1 higher ^{14}C concentrations clearly accord with the rate of abstraction 8 months later (correlation coefficient = 0,86). This is rather surprising as it would have been more logical for the higher ^{14}C concentrations to have appeared directly in response to higher abstraction, as it could then have been attributed to a diversion of larger proportion of recent water to the pumping area. If that had been the case then the ^{14}C response would have been similar to that of the Kuruman eye (see Fig. 12.6.2.2) indicating higher ^{14}C concentrations for the initial period of higher flow, with only a small lag. As was indicated the higher flow was caused by the good recharge during the period 1974-77 with higher ^{14}C concentration (see Fig. 12.1.1). The observed ^{14}C pulse resulted from an imbalance in the relative contributions of flow from the two catchments feeding the spring. A larger quantity of recent recharge (high ^{14}C) was contributed by the western aquifer (Kuruman Hills) during the period of high flow, but as the flow receded to its normal discharge rate, the ^{14}C concentration declined to its usual average value, being a mix of constant ratio between recharge from the two catchments.

The flooding of the Sishen mine occurred as a result of rising groundwater levels following the high recharge during the period 1974 - 1977, the effect of which on water levels could be simulated by means of the CRD relationship with no lag (see Fig. 12.6.2.2). According to Fig. 12.6.2.3 the rate of abstraction closely follows the ${}_{36}^{1}CRD$ series but with a lag of 39 months. A lag of 39 months also occurred between the ^{14}C concentrations and the ${}_{36}^{1}CRD$ series (see Fig. 12.6.2.4). Similarly the average rainfall over a preceding period also corresponds to the water level fluctuations and was also compared with the isotopic responses observed in the Sishen mine abstractions. The average rainfall over the preceding 12 months, which also mimics the groundwater level fluctuations quite well, followed the response of the ^{14}C concentration but with a lag of 30 months (see Fig. 12.6.2.5). This leads to the conclusion that there was a delay of about 30 - 39 months before groundwater recharged during the period 1974 - 1977 started to appear in the pumped water. As is evident from the various graphs, interpretation of the isotopic responses is not straightforward, and variable pumping from different areas could have complicated the picture. Initially pumping took place from the Hill 2 area but was shifted to new production boreholes which have been drilled in the mining area. The best correlation (r = 0,86 - see Fig. 12.6.2.1) was established between the ^{14}C concentrations and the average 2 monthly abstractions in the Southern mine, eight months later. However, according to Fig. 12.6.2.2, the rise in groundwater level represented by the CRD series, occurred about 39 months earlier. Hence the belated increase in abstraction in the mining area only commenced after a long lag. It seems that the higher abstraction for practical reasons must have followed a pattern similar to that of the recharge contributions in order to lower the groundwater levels to their previous heights. The pumped water therefore merely reflected the isotopic characteristics of the groundwater in the aquifer which corresponded to that of the recharge but with a delay of about 39 months (see Fig. 12.6.2.4). The ^{13}C and bicarbonate concentrations measured in the pumped water revealed essentially the same pattern as the ^{14}C (see Fig. 12.6.2.6 and Fig. 12.6.2.7). Unfortunately only an incomplete set of bicarbonate analyses was available. The ^{14}C concentrations show some correspondence with the bicarbonate concentrations (see Fig. 12.6.2.8), clearly indicating interdependence between ^{14}C, ^{13}C and bicarbonate concentrations. The delay of about 39 months in the reappearance of the groundwater labelled with higher ^{14}C, is an indication of the turnover time of the recent component of water in the aquifer.

The apparent relationships between ^{14}C, ^{13}C and bicarbonate, indicating higher ^{14}C concentrations for higher alkalinity, which has been observed in various springs (Fig.12.4.3.1.1), is confirmed by the groundwater pumped from the Sishen mining area. The differences in the lags between the response of ^{14}C, ^{13}C and bicarbonate concentrations, are still unresolved.

In the case of groundwater obtained from Sishen Hill 2 the ^{14}C concentrations do not show a tendency similar to the one observed for springs in a more semi-arid environment, namely that higher ^{14}C concentrations are associated with higher alkalinity (see Fig. 12.6.2.9). There could be two groups which do not confirm the postulation that rapid recharge, bypassing the soil zone, is of lower alkalinity. Hence the indications are that recharge is taking place in the normal way. However, some preferential recharge via the Gamagara River and which has had little interaction with biogenic CO_2, could be present. This groundwater would have attained its final characteristic as it flows to the point of abstraction, and hence the general characteristic of the groundwater is a ^{14}C concentration of less than 83% mc, in spite of having been recharged very recently. Due to the short travelling time to the point of abstraction (39 months), the effect of subsequent exchange of carbon cannot be significant. The best correlation between ^{14}C and bicarbonate concentrations indicates that higher bicarbonate concentrations actually precede the ^{14}C response by one month, which is hard to explain. There is also reasonable correspondence between the ^{13}C concentrations and abstraction, and between ^{13}C and bicarbonate concentrations. This clearly indicates that the bicarbonate concentration is an important parameter in groundwater but the interpretation of carbon isotopes, especially in estimating turn-over times by ^{14}C age determinations, in carbonate aquifers is not yet fully understood.

The ^{14}C - ^{13}C relationship of the Sishen mining groundwater shows no evidence of carbon-exchange in the aquifer but instead indicates admixing of groundwater of a different characteristic (lower ^{13}C concentrations) which could have originated from recharge during a colder, more Mediterranean climate (see Fig. 12.6.2.10). This will be examined in the next section dealing with the ^{14}C - ^{13}C relationship of different dolomitic springs.

12.6.3 ^{14}C - ^{13}C relationship of dolomitic spring waters

The characteristics of several dolomitic springs have been examined in X-Y plots of the ^{14}C against ^{13}C to determine the end members of a mixing model, as is indicated in Fig. 12.6.2.10. However the reliability of the ^{13}C determinations is not sufficiently high for such a comparison to be made, due to errors caused by variable and incomplete recovery of ^{13}C from the water samples.

12.7 Summary of results and conclusions

The following picture emerges regarding the behaviour of ^{14}C:

1) Higher recharge and more direct infiltration which short-circuits much of the soil zone that generates biogenic CO_2, could yield groundwater with a lower alkalinity. This water will not attain an initial ^{14}C concentration of about 85% of the atmospheric ratio due to incomplete uptake of CO_2 in the soil zone, which limits the saturation level of bicarbonate that normally would be attained.

2) When water unsaturated in bicarbonate flows through the aquifer, limestone

with ^{14}C = 0% mc is dissolved. However, exchange between ^{14}C in the water and carbon within the aquifer would also occur. Whereas the dissolution of limestone continually exposes new surfaces of carbonate (with ^{14}C = 0% mc), reverse exchange causes a build-up of bomb ^{14}C on the interface of the carbonate matrix of the aquifer. This results in a higher build-up of ^{14}C concentrations during the exchange-equilibrium the more saturated the water is in bicarbonates and the smaller the turn-over time of the aquifer.

3) ^{14}C concentrations in groundwater, especially in dolomitic aquifers or springs which are recharged partially from non-dolomitic aquifers, are subject to an additional degree of ^{14}C modification. This seriously affects determination of the turn-over time ("age") of the water and the derivation of recharge according to the ^{14}C–age relationship with depth.

A re-examination and re-interpretation of ^{14}C concentrations of dolomitic springs needs to be carried out within the context of the new hypothesis using other tracers such as CFC's. Although the isotopic and hydrological interpretations are mutually supportive, a better understanding of the isotopic results has been obtained from the hydrological analysis of the data than the other way round.

13. Groundwater and climate fluctuations

13.1 Overview

The flow of major springs and the levels of coastal lakes fluctuate according to the status of groundwater levels which reflect the long-term oscillations of rainfall integrated over time and space. They also provide a means of assessing the hydrological impact of climate change. It is important that the selected series represent the natural responses and that they have not been affected by anthropogenic influences. By comparing the measurements to the cumulative departures of rainfall from the mean, or the moving average rainfall, it can be ascertained whether a series is reliable and what corrections, if any, have to be applied.

The following hydrological records in the RSA have been selected to assess possible impacts of climate change on the hydrological cycle.

Springflows: Corrected flow of the Uitenhage springs (Fig. 13.1.1) checked against the moving average (MA) rainfall.

Hydrographs: The fluctuations of groundwater levels of the Wondergat-sinkhole (Bo Molopo) provide a continuous series of natural groundwater level fluctuations from 1922 to 1980 whereafter the water levels have been affected by abstraction and have had to be corrected according to the CRD and MA rainfall series (see Fig. 13.1.2).

Lakes, estuary systems and runoff

Lake St Lucia (see Fig. 13.1.3.): - The natural hydrological balance of the system is reflected by the salinity fluctuation which responds inversely to the CRD and MA rainfall series (Bredenkamp et al., 1995). The anomaly caused by the Demoina cyclone is clearly seen.

Lake Mzingazi : - Although the lake level depends on both surface and groundwater, the latter appears to be the dominant component. However, the lake level has been affected by afforestation and by abstraction from the lake to supplement the water supply of Richards Bay. (Bredenkamp, 1995). Filtering out the effect of afforestation by means of the CRD relationship provides a record of the natural response of the lake which could be extrapolated back in time by means of historical rainfall records (see Fig. 13.1.4).

Although not included in the guide, the gross runoff from major catchments such as the Vaal Dam would also reflect long-term oscillations of above and below average rainfall. These long-term oscillations of rainfall are probably caused by periods of accumulation of energy somewhere in the ocean, coinciding with periods of sub-average rainfall, followed by periods of redistribution of the excess accumulated energy as higher rainfall. This could be accompanied by changes of the normal atmospheric circulation patterns causing droughts to occur in some areas and flooding in other areas. Major droughts are usually followed by a spell of higher rainfall and floods. The triggering of high rainfall after droughts is clearly evident from the hydrograph of the Wondergat which always recovers once it has reached a certain low level (see Fig. 13.1.2).

An interesting illustration of the long-term impact of rainfall has been reported by Gieske (1996) in a study of runoff in the Okavango Delta. Contrary to previous runoff models which incorporated certain discrepancies in the flow as having been caused by a change in the runoff response at a certain level of flow, the anomalous flows were proven to have resulted from antecedent rainfall fluctuations with a long memory. Excellent simulations have been obtained incorporating the cumulative rainfall departures from the mean as an index in the rainfall/runoff relationship. This tallies with findings that runoff in other areas are also linked to the status of groundwater. A typical example is that of the runoff from the Jonkershoek mountainous catchments most of which is delayed because it emanates from groundwater. In classical runoff models this response is incorporated as drainage from a reservoir with delayed outflow. According to the CRD analogy the base flow is clearly contributed by groundwater, varying according to the general piezometric levels of the catchment. So-called 'intermediate' flow could also be contributed by groundwater e.g. as the flow fed by larger fractures, responding faster than the matrix flow which contributes the more steady base-flow component of runoff. Any development of groundwater would therefore have repercussions on the surface flow. It is not impossible that the status of groundwater could also have an impact on climate, especially in the more humid areas, by inducing atmospheric conditions which are more conducive to triggering rainfall. This could be the reason why a major break-out from a severe drought marks the start of a new wet cycle.

For this reason a CRD-type runoff-model could also be successfully employed in simulating runoff, and would be worth investigating.

13.2 Impact of climate change

In all cases the collection of reliable rainfall records is as important as good records of the other hydrological variables which are monitored. Without reliable rainfall data it would be impossible to identify and assess the hydrological responses and impacts resulting from man's intervention, and those which have been caused by climate change.

The effect of climate change is possibly indicated by the corrected flows of the Uitenhage Springs which, since the start of the record, show increasing amplitudes of the major oscillations which appears to be 132 months. The spring flow is related to the average rainfall over 132 months (see Fig. 13.1.1).

In the case of Lake Mzimgazi (Fig. 13.1.4) the minor oscillations of wetter and drier cycles are not as prominent as those of Lake St Lucia (Fig. 13.1.3) and the Uitenhage springs. However the fluctuations of Lake Mzingazi (linked to the average rainfall) show evidence of a major drying cycle extending from 1925 to 1954 and a wetter than average rainfall cycle (rising lake level) over the period 1954 to 1990. It should be borne in mind that the hydrological impact of dry periods only occurs when conditions of wetness decline below the average, and similarly the hydrological impact of wet conditions prevails over periods of higher than average rainfall. The hydrological impact would therefore follow more directly in the case of runoff as it responds more quickly to rainfall, than does groundwater which reacts to the moving average rainfall over several preceding years.

The groundwater levels in the Wondergat also show the long-term rainfall cycles in the interior of the country (see Fig. 13.1.2). This also indicates higher amplitude peaking in 1980 which has been followed by some of the worst droughts especially in 1983 whereafter rainfall conditions improved slightly before drought conditions up to 1992 resumed.

Fig. 13.2.2 shows a comparison of major periods of above and below average hydrological response and their durations.

13.3 Conclusion

It should be realized that the hydrological cycle is a complex, interactive system of highly variable outputs of runoff and groundwater recharge. However, if the hydrological outputs are examined over larger time intervals they correspond to more simplistic relationships with rainfall, which are best illustrated by surface runoff impounded by dams, or by springs fed from underground reservoirs. The present study will hopefully lead to an appreciation of the role of groundwater not only as a supplementary water supply but as an important component of effective management of water resources.

Studies of drought and the assessment of its impact, based on the CRD or moving average rainfall series, could be linked directly to the hydrological elements of runoff, groundwater and soil moisture. Both the CRD and MA-rainfall methods provide a means of assessing and comparing the impacts of drought on a relative basis, and thus would provide an equitable basis of awarding compensation for losses due to the drought.

14. Establishment of an effective groundwater monitoring system.

14.1 General

The establishment of monitoring stations cannot be prescribed but depends to a large extent on the purpose for which the measurements are required, which should be clearly defined at the start. The study of specific aquifers which is usually part of an investigative programme, usually requires that several boreholes be drilled and used for pumping tests and as monitoring boreholes for several years to assess the exploitation potential of the aquifer. However only some of the boreholes need to be identified for long-term monitoring, or for purposes of modelling and management of the aquifer.

In both instances a good areal distribution of monitoring points is preferable but the density required would obviously have to be higher for a productive aquifer than to monitor the regional piezometric response. As a general rule approximately 6 to 9 well-spread monitoring points per aquifer or per selected area would provide a representative picture of the groundwater fluctuations in such an area. In such a case the production boreholes would serve as additional monitoring points in reliable hydrodynamic modelling of the aquifer. Naturally, for individual aquifers which are exploited, project managers would have to determine the number of boreholes to be monitored, with due consideration of the high cost of drilling and maintenance of stations.

For many aquifers a reduction in the number of monitoring points would not seriously affect the reliability of assessments of the groundwater exploitation potential, and the management of the aquifer. As has been indicated, the spatial variability of groundwater levels (recharge) is evened out by large fractures. The variability of rainfall is probably the least reliable factor, and monitored monthly total rainfall at each monitoring station would increase the reliability of the regressions between rainfall and the piezometric levels. The chemical analysis of the rainwater collected each month over a period, would improve the reliability of estimates of recharge derived by the chloride method. This would improve estimates of aquifer storativity and its spatial variation. The chloride concentrations could help to establish if a general relationship between rainfall and chloride concentrations exists and if it does, it would reduce the need to continue with the chloride analysis of rainwater.

In all cases monitoring could be done on a monthly basis as this is not only logical and practical, but would conform to the measurement of groundwater levels and of the input rainfall. Additional recording of daily rainfall would merely provide information on the statistical variability and stochastic nature of rainfall. For all the hydrographs which have been analysed for points in close proximity to reliable rainfall stations, good simulations of the groundwater levels have been obtained based only on monthly rainfall. There is therefore no need for rainfall and groundwater levels to be measured at shorter time intervals but the closer the rainfall is measured to the monitoring point the lower would be the uncertainty induced by spatial variability or unreliable rainfall.

Flow measurements and water quality analyses of major perennial springs would have to be continued as they represent groundwater fluctuations over larger areas. Apart from information on the groundwater situation the springs also provide an integrated picture of rainfall fluctuations and climatic variations, and could be one of the most reliable indicators of climatic change and its hydrological impacts.

14.2 Effective monitoring and data evaluation

The following guidelines have been prepared to assist in the establishment of an effective monitoring network and in evaluating existing stations according to certain criteria.

The decision to establish a monitoring station has to be based on:

1) the importance of groundwater as a primary/secondary water supply;
2) the existence of exploitable aquifers for irrigation or as urban water supplies, which is generally indicated by the occurrence of boreholes with high sustainable yields;
3) the sustainability of groundwater exploitation based on the average annual rainfall - areas of high rainfall being of greater importance, and because groundwater exploitation affects the base flow of streams and the ecology. However, exploitation of groundwater on a large scale also occurs in drier areas, where groundwater has been replenished over many years or during pluvial periods of the

past;

4) monitoring points already in operation and rainfall stations in the area.

Selection of additional monitoring points:

Before a new monitoring station is opened

1) all groundwater records in the area should first be evaluated according to the CRD and MA method; abandoned stations should be scrutinized and checked for reliability;
2) The objectives and criteria of monitoring should be clearly defined, e.g.
   - whether the station would serve a purpose other than merely measuring the groundwater levels;
   - whether rainfall records which date back several years could be utilized;
   - whether it aids surface or groundwater impact studies or aquifer management by means of simple CRD models or a hydrodynamic model;.
   - whether spring flows integrate rainfall variability over a larger area;
   - what additional information would be forthcoming for the assessment of the hydrological impact of drought intensities and duration;
   - the accessibility of the points;
   - whether the selected point would be sufficiently representative of a typical aquifer in a specific climatic environment.
3) With regard to the frequency of measurements
   - monthly measurements would be adequate, but
   - if more frequent monitoring is considered necessary, the objectives should be defined clearly and the cost should be assessed.

14.3 Closure of stations

A decision on whether or not to close a station, should be arrived at in a similar way, using the same criteria mentioned above. Any stations which are closed could later be reopened and examined by the CRD and MA methods.

15. Benefits and research opportunities derived from monitoring

The study on monitoring (Bredenkamp, 1999) has opened up several opportunities for extending the CRD and MA rainfall methods to other parts of the world, especially to areas of Africa where rainfall measurements but only limited groundwater level data are available. Sparse data is no longer a constraint in simulating the trend in water level fluctuations as it could be inferred from the CRD method or the MA rainfall series. Reasonably reliable estimates of recharge and of aquifer potential could be derived, based on short series of water level measurements and estimates of recharge derived by means of the chloride method. Reliable hydrodynamic models of aquifers could then be compiled and applied effectively for aquifer management.

In areas where much data have been accumulated, but the impact of abstraction has affected the natural hydrological responses of groundwater, the CRD and MA rainfall methods could simulate the natural responses and even obtain an estimate of the average abstraction. The approach and the methods set out in the present guide could be used to analyse groundwater in a consistent way in all parts of the world. It would also help to characterize aquifers in a more meaningful and effective way than in the past.

The CRD/MA method should be studied as a means of assessing the impact of abstraction and recharge on individual boreholes drilled in fractured and dolomitic aquifers. In this way an effective model to simulate water level responses in fractured aquifers which are the most dominant in may parts of the world, could be established.

Studies of runoff analogous to the CRD method should be considered, although it could be argued that they merely provide additional runoff models. However the emphasis should be shifted to surface runoff yields on a monthly basis due to the averaging effect of storage dams on the variability of surface flow. The CRD method and average rainfall method provide the same possibility of incorporating the fast response runoff, and intermediate response as well as base flow contributions. Hence the CRD method would be the component used in sophisticated hydrological models. Even though the simulation of runoff might not be as good over smaller time intervals, estimates of volumes would be reliable over longer intervals, which in water resources management is very important. The relationship between surface flow and groundwater should be studied in greater depth, especially in assessing the impact of development such as afforestation and in relation to groundwater exploitation

It is proposed that detailed studies of mountainous catchments be carried out according to the approach set out in the present study, utilizing particularly the chloride measurements which have been collected previously by the Department of Water Affairs and Forestry. This data is likely to be lost if it is not used in a purposeful way. The response of recharge and runoff to rainfall could be revealed, and the significance of groundwater, leading to a clearer understanding of hydrological interactions. In this way the gap between surface and groundwater hydrology could be reduced.

The present network of monitoring points in the RSA should be examined according to the directives given in the study but only after all the monitoring data worth analysing have been studied. In many areas the density of stations could be reduced without seriously affecting the assessment of the natural response of groundwater or the impact of abstraction on different aquifers. In some areas monitoring stations would have to be established mainly to provide a representative coverage of the different groundwater areas. This should be done according to the importance of groundwater and an assessment of the impact which large scale abstraction would have on groundwater reserves. Fewer groundwater monitoring points backed by monthly rainfall measurements, would help to fill in gaps in areas where groundwater records are sparse.

The impact of exploitation of groundwater and determination of the "reserve" is a critical determinant in the new water law in South Africa. The CRD and MA approach is probably the most effective method to assess the reserve in a simplistic way. In the case of groundwater the impact could be quantified in a comparative way and the effect on wet-lands and base flow could be dealt with in a consistent manner without complicated modelling of the hydrological cycle.

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Fig. 4.2.1.4.4 Flow of different springs in the Bo Molopo area in relation to the groundwater level of the Wondergat.

Fig. 4.2.1.4.5 Flow of different springs in the Pretoria area in relation to the groundwater level of the Wondergat.

Fig. 4.2.1.4.6 Relationship between 120-month average rainfall and the flow of the Pretoria fountains.

Fig. 5.2.1 Infilling of missing groundwater levels by means of CRD relationship of station D5N519 (Williston).

Fig. 5.3.1.1 Correlation coefficients for Slurry rainfall in relation to stations at different distances.

Fig. 5.3.2.1 Actual and simulated flow of the Buffelshoek eye in relation to the Wondergat levels.

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Fig 5.3.2.3 Period of abstraction indicated by k-values >1 (shown as negative values) for the CRD relationship in relation to the observed flows of the Molopo eye.

Fig. 5.3.2.4 Different k-values which have to be introduced to simulate the water levels of the Wondergat from the ${}_{36}^{1}CRD$ series.

Fig. 5.3.2.5 Correspondence between the measured water levels in monitoring borehole BB35, and those simulated with the CRD series.

Fig. 5.3.2.6 Correspondence between the measured water levels in monitoring borehole GN41, and those simulated with the CRD series.

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Fig. 5.4.5.1	Distribution of groundwater level monitoring boreholes stored in the National Groundwater Database (NGDB).
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Fig. 7.2.2.5 Correspondence between measured values and those simulated by the ${}^{1}_{av}CRD$ series lagged by 14 months. Subtracting the monthly pumping from the simulated series produced an exceptionally good match with the observed water levels for conditions of exploitation.

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Fig. 8.3.1.2 Recharge, in percent and millimetres, derived from equal volume periods applicable to the Pretoria Fountains.

Fig. 8.3.1.3 Recharge relationship for the Pretoria Springs based on the EV method, clearly indicating a discrepancy which is attributed to erroneous flow data.

Fig. 8.3.1.4 Flow of the Uitenhage springs expressed as a percentage of the average rainfall - applying the EV method.

Fig. 8.3.2.2.1 The Uitenhage Groundwater Control Area showing the location of the springs and major artesian boreholes in the Coega Ridge Aquifer.

Fig. 8.3.2.2.2 Correlation between the flow of the Uitenhage Springs and the ${}^{1}_{132}CRD$ series.

Fig. 8.3.2.2.3 Recharge as percentages and millimetres, derived by means of the EV method, for the Uitenhage Springs.

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Fig. 9.2.2.2 Average inflow over 60 months into the Rietvlei dam in relation to the flow of the Rietvlei springs.

Fig. 9.3.1 *Simulated groundwater levels at Danielskuil indicating that the levels* would have risen higher than the ground surface.

Fig. 10.2.1 Irrigation abstraction in the Grootfontein compartment based on rainfall over preceding 12 months. The higher abstraction after 1978 is due to larger areas being irrigated.

Fig. 10.3.1 Abstraction from the Grootfontein eye for two periods in relation to rainfall.

Fig. 11.3.3.1.1 Comparison of the chloride and sulphate concentrations of Bloemhof and Trompsburg springs.

Fig. 11.3.3.1.2	Chloride and sulphate concentrations of the Bloemhof eye indicating an improvement in water quality.
Fig. 11.3.3.1.3	Plot of chloride concentrations of the Bloemhof eye against time, and the cumulative plot indicating that the initial concentrations cannot be used to determine recharge by the chloride-ratio method.
Fig. 11.3.3.1.4	Comparison of the chloride and sulphate concentrations of Turffontein eye - normalized plot.
Fig. 11.3.3.2.1	Comparison of the chloride and sulphate concentrations of Turffontein, Maloney's and Gerhard Minnebron eyes.
Fig. 11.3.3.2.2	Comparison of the chloride and sulphate concentrations of Gerhard Minnebron.
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Fig. 11.3.3.3.1	Chloride and sulphate concentrations of Renosterfontein.
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Fig. 11.4.1.1	Chloride concentrations in rainfall derived from measurements in the Zululand area and in Botswana.
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Fig. 11.4.1.3	Chloride depositioning of rainfall in the Zululand area.
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Fig. 11.4.2.1	Recharge derived from chloride concentrations of different dolomitic springs - *assuming no contamination*.
Fig. 11.5.2.1	Temporal variation of water quality of production boreholes at Zuurbekom.
Fig. 11.5.2.2	Sulphate concentrations of Zuurbekom boreholes in relation to the cumulative rainfall departures, incorporating a lag of 2 years in the sulphate response.
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Fig. 11.5.3.3	Cumulative plot of the abstraction in relation to the chloride concentrations of Centurion boreholes.
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Fig.12.1.1	Natural variation in tritium and ^{14}C concentrations in the atmosphere, showing effect of bomb input.
Fig. 12.3.1	Relationship between deuterium and oxygen-18 in natural waters as well as the effect of evaporation.
Fig. 12.4.2.1	Variation of ^{14}C concentrations of the Grootfontein eye in the Rietvlei-Pretoria area.
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Fig. 12.4.2.2b	^{14}C concentrations of Maloney's eye in relation to the average rainfall over 60 months.
Fig. 12.4.2.3	^{14}C concentrations in relation to the average discharge rate of springs.
Fig. 12.4.2.4	^{14}C concentrations measured at main Kuruman eye.
Fig. 12.4.3.1.1	^{14}C concentrations of different dolomitic springs in relation to alkalinity.

Fig. 12.4.3.1.2	Recharge of different springs inferred by means of the chloride method in relation to alkalinity.
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Fig. 12.4.3.2.4	^{14}C concentrations in relation to alkalinity for groundwaters in the Sishen area.
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Fig. 12.4.3.3.2	^{13}C concentrations compared to alkalinity measurements of groundwater in the Telele area.
Fig. 12.4.3.3.3	Comparison of ^{14}C and ^{13}C concentrations of groundwater in the Telele profile area.
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Fig. 12.5.1.1	^{14}C ages of groundwater in relation to the depth of water strikes used to derive recharge from the linear increase in age with depth.
Fig. 12.5.1.2	^{14}C ages of groundwater in relation to the depth of water strikes used to derive recharge from the increase in age with depth.

Fig. 12.6.2.1	^{14}C concentrations of groundwater in the Sishen mining area in relation to the rate of abstraction.
Fig. 12.6.2.2	Rise in groundwater levels in the Sishen area in relation to the Cumulative Rainfall Departure series, with no lag.
Fig. 12.6.2.3	Rise in groundwater levels in relation to the increased abstraction required lowering the groundwater levels in the Sishen mining area.
Fig. 12.6.2.4	^{14}C concentration in the pumped water of the Sishen mine in relation to the CRD series representing the rise in groundwater levels.
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Fig. 12.6.2.6	^{13}C concentrations in relation to alkalinity of groundwater in the Sishen mining area.
Fig. 12.6.2.7	Alkalinity of groundwater in relation to ^{13}C concentration in the Sishen mining area.
Fig. 12.6.2.8	^{14}C concentration in the Sishen mining area in relation to alkalinity of the groundwater.
Fig. 12.6.2.9	Graph of ^{14}C concentrations against alkalinity for Sishen Hill 2 samples.
Fig. 12.6.2.10	Relationship between ^{14}C and ^{13}C concentrations.
Fig. 13.1.1	Inferred natural flow of the Uitenhage springs as an indication of climate change on hydrological components.
Fig. 13.1.2	Inferred natural water levels of the Wondergat sinkhole in relation to the moving average rainfall over 72 months.
Fig. 13.1.3	Salinity variations of Lake St Lucia measured at Charter's Creek in relation to the moving average rainfall over 48 months.
Fig. 13.1.4	Variation of the water level of Lake Mzingazi.
Fig. 13.2.2	Presentation of the major wet and dry cycles.

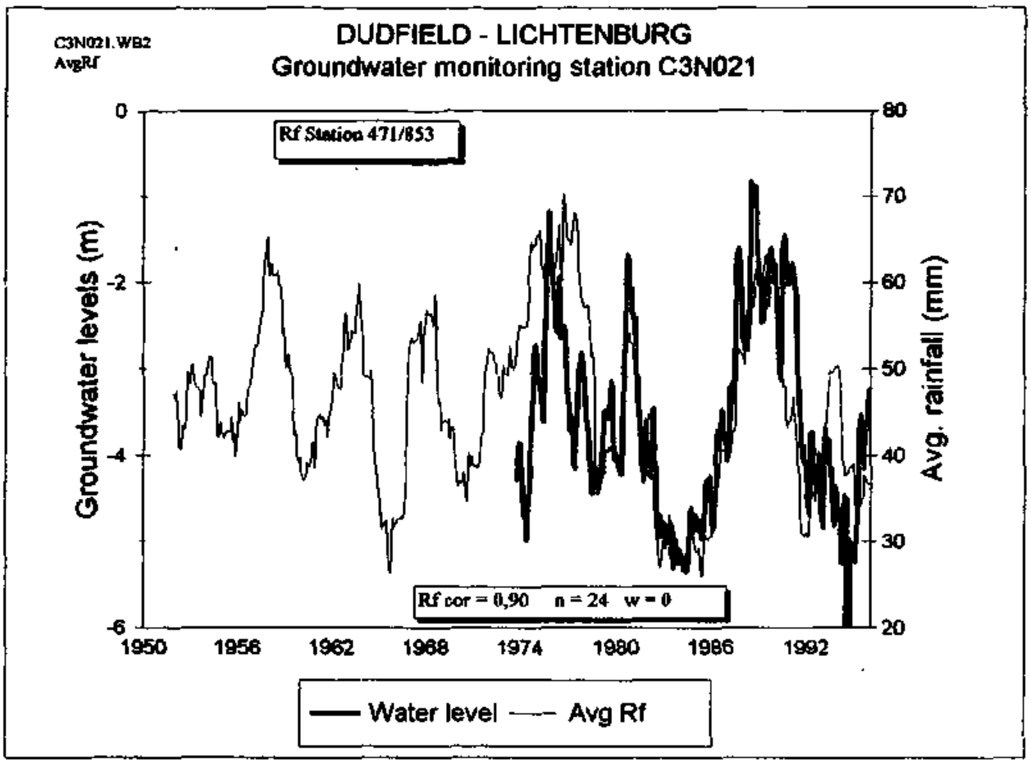


Fig. 3.2.1.1 Water levels in C3N021 in relation to the average rainfall over 24 months.

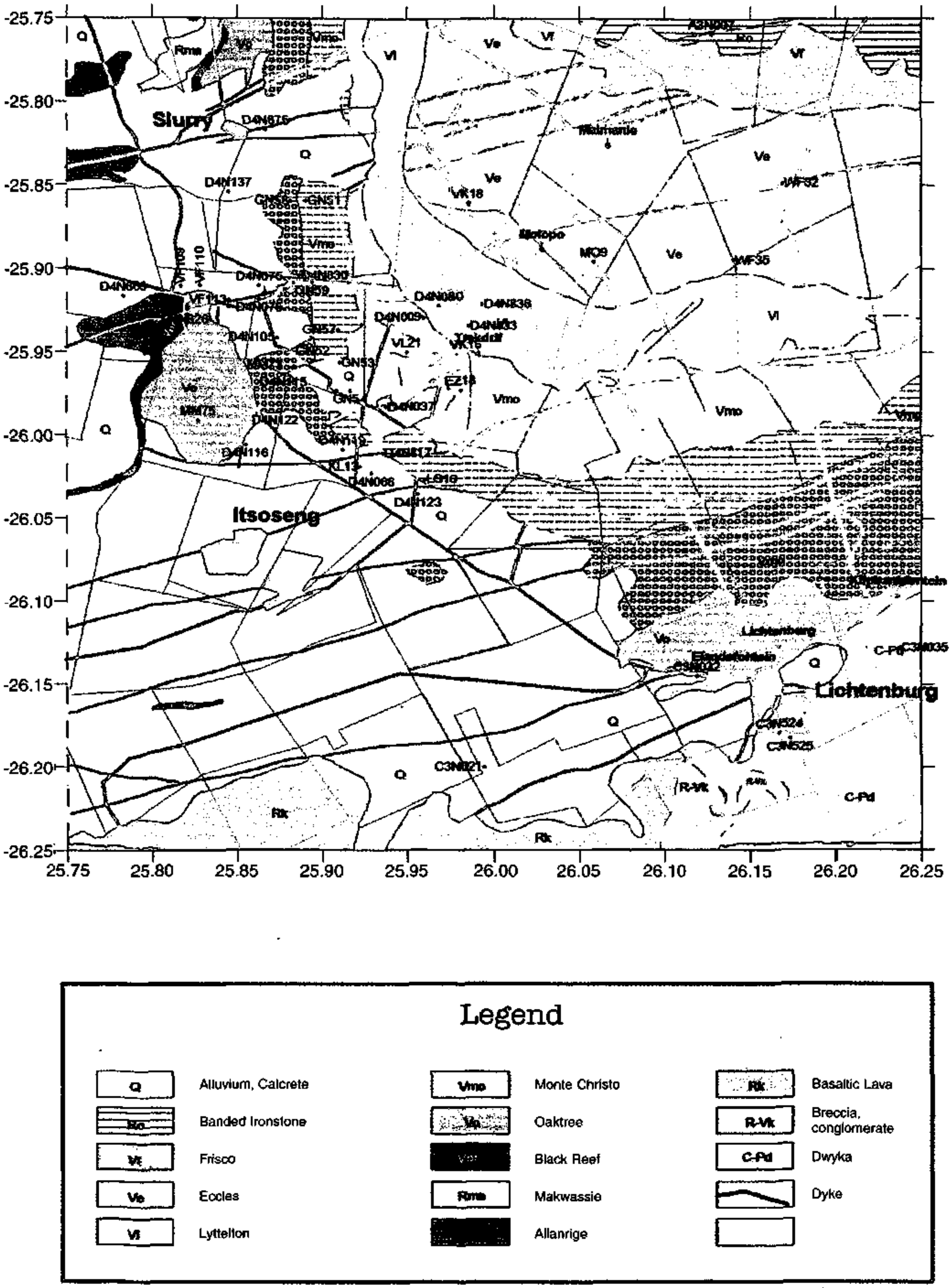


Fig. 3.2.1.2: Positions of monitoring boreholes in the Grootfontein area in relation to the geology.

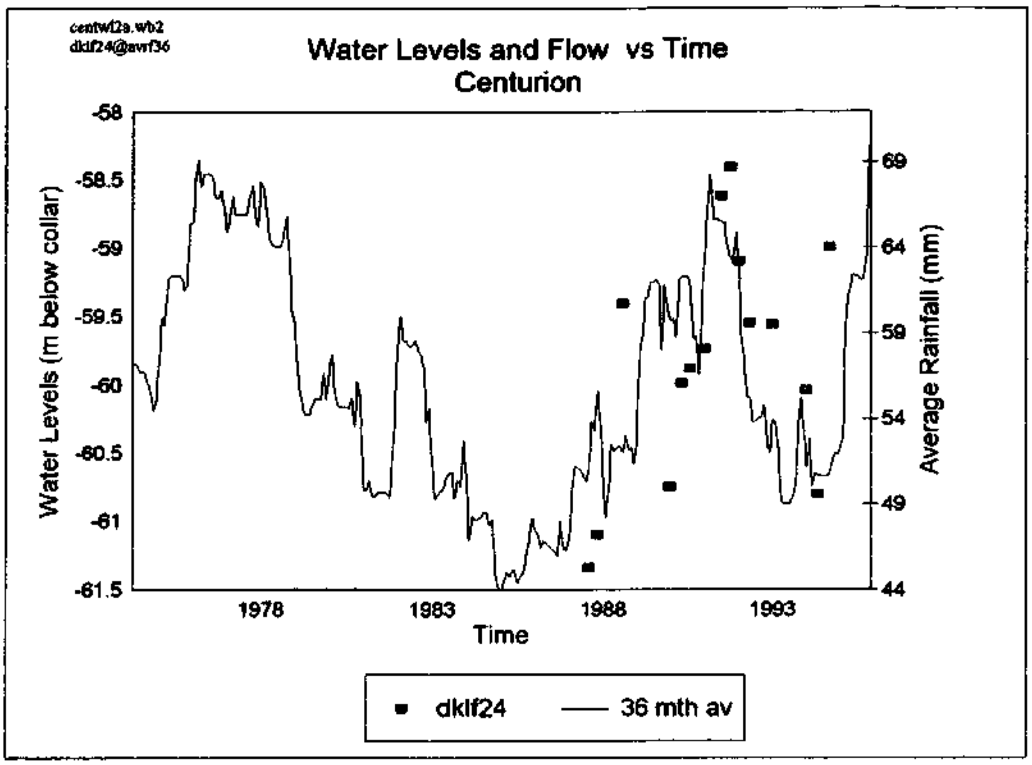


Fig. 3.2.1.3 Water levels in borehole Dklf24 in relation to the average rainfall over 36 months.

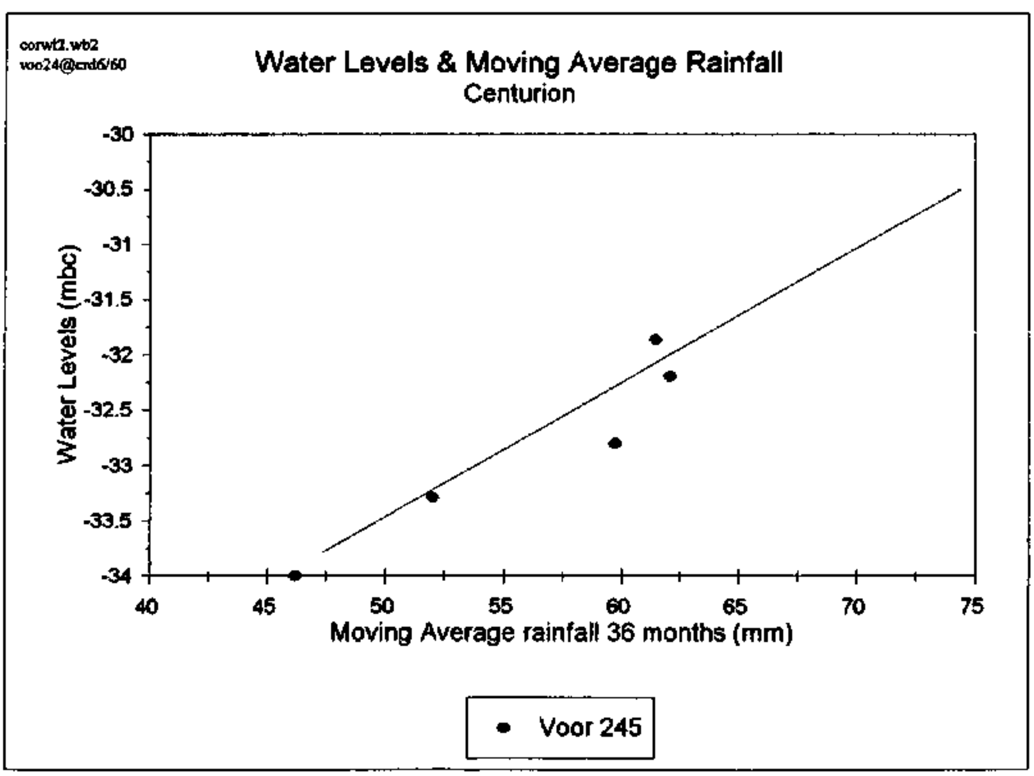
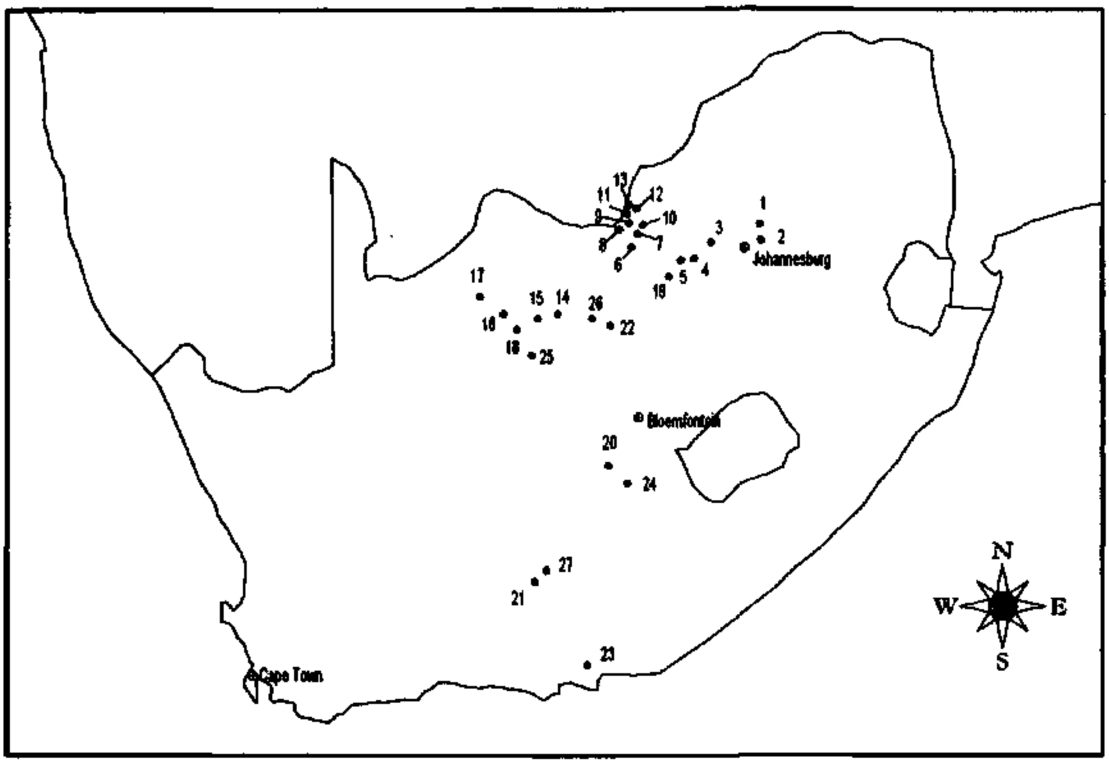


Fig. 3.2.1.4 Water levels in borehole Voor245 in relation to the average rainfall over 36 months.



Relative position of springs

1. Pretoria Fountains
2. Rietvlei and Sterkfontein
3. Maloney's Eye
4. Upper Mooirivier
5. Schoonspruit
6. Lichtenburg
7. Molopo Eye
8. Grootfontein
9. Olievendraai
10. Buffelshoek
11. Vergenoegd
12. Tweefontein
13. Dinokana
14. Thabasikwa
15. Vlakfontein
16. Kuruman
17. Tsineng
18. Groot Kono
19. Gerhard Minnebron
20. Trompsburg
21. Toorberg
22. Bloemhof
23. Uitenhage
24. Mosterdhoek
25. Boetsap
26. Kraalkop
27. Mackies

Fig. 4.1.1 Location of different springs in the RSA of which the flows and groundwater chemistry have been measured.

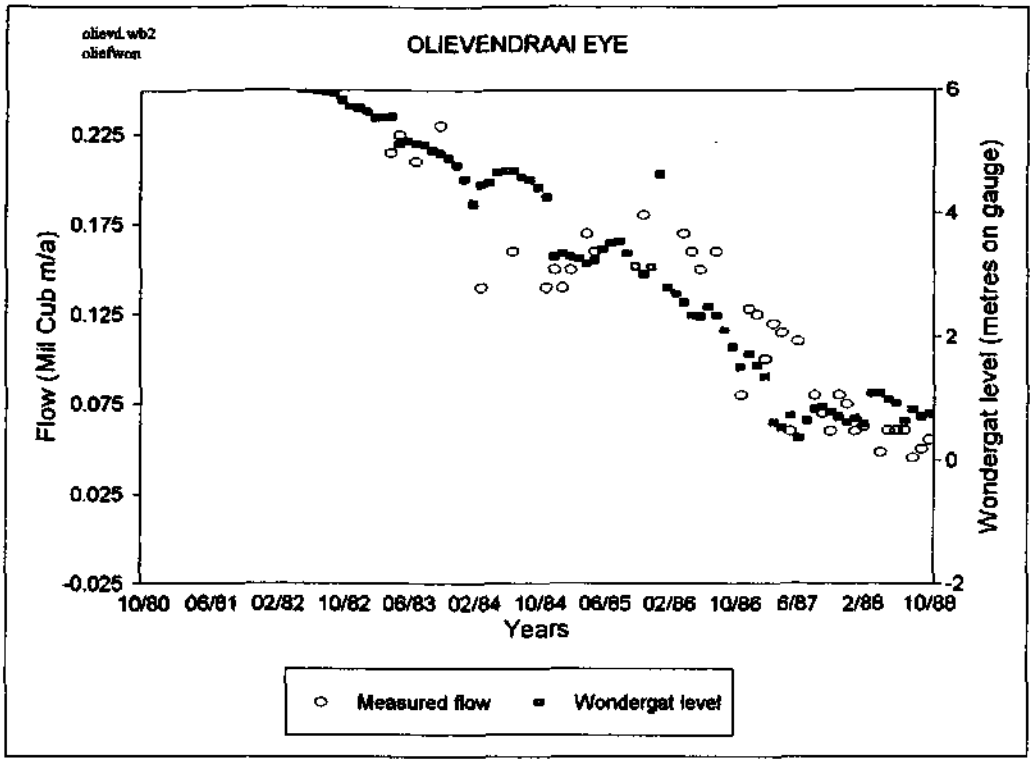


Fig. 4.2.1.1.1 Flow of Olievendraai eye in relation to the Wondergat water level.

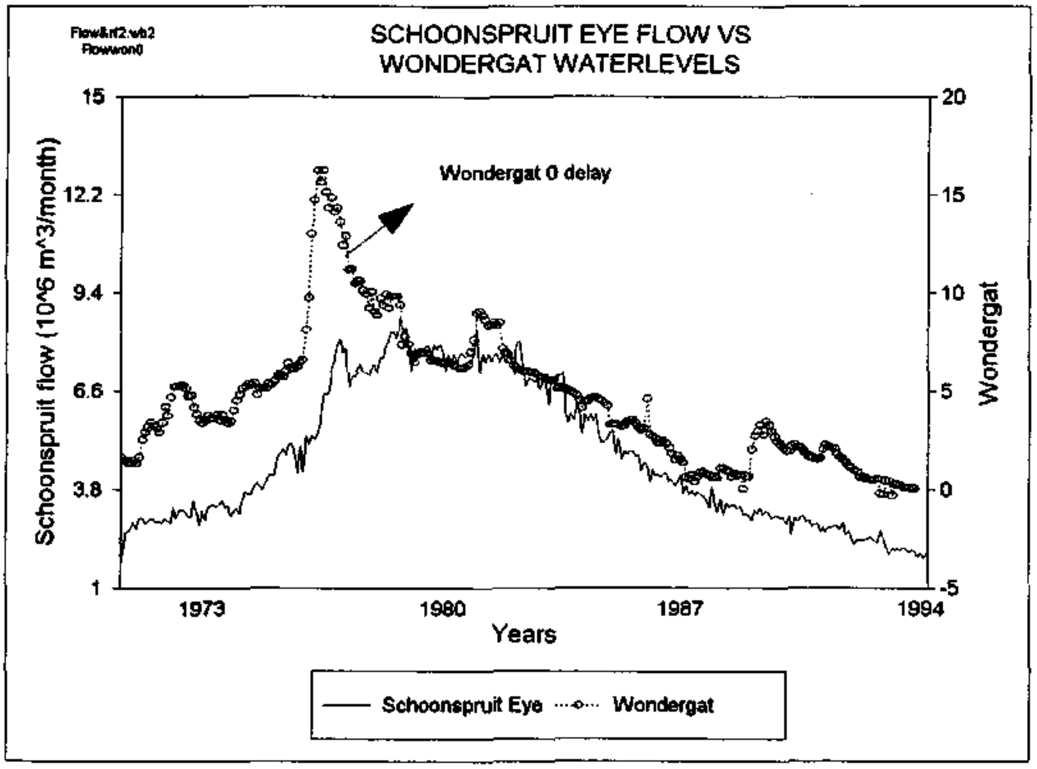


Fig. 4.2.1.1.2 Flow of Schoonspruit eye in relation to the Wondergat water level.

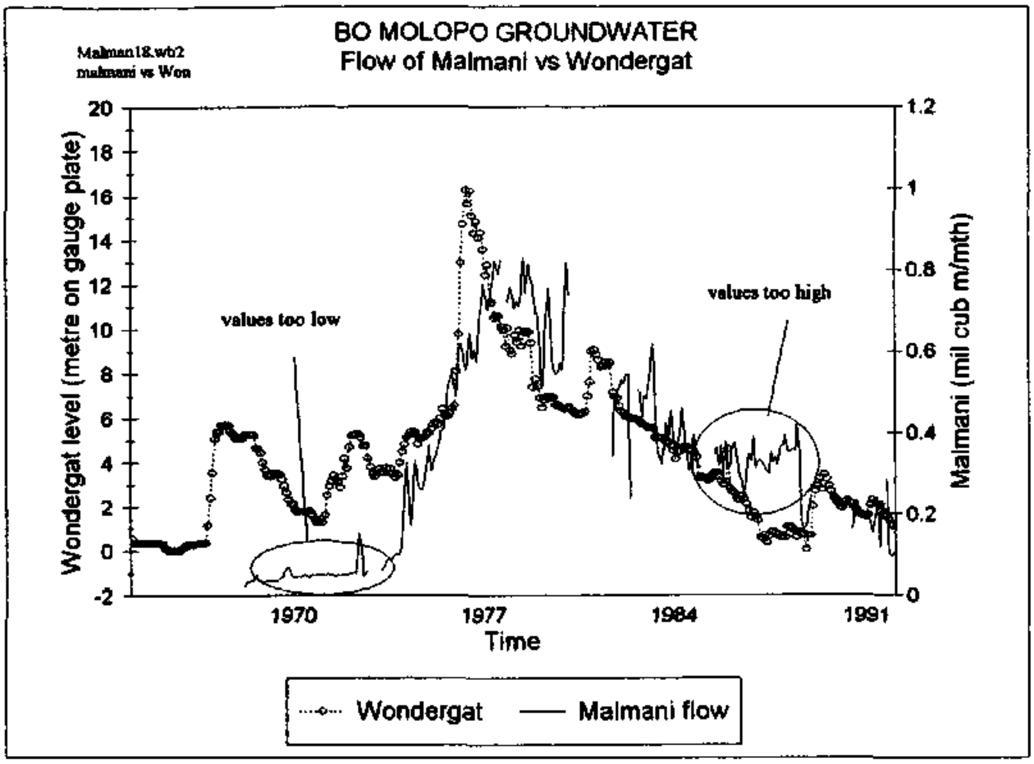


Fig. 4.2.1.1.3 Flow of Malmani eye in relation to the Wondergat water level.

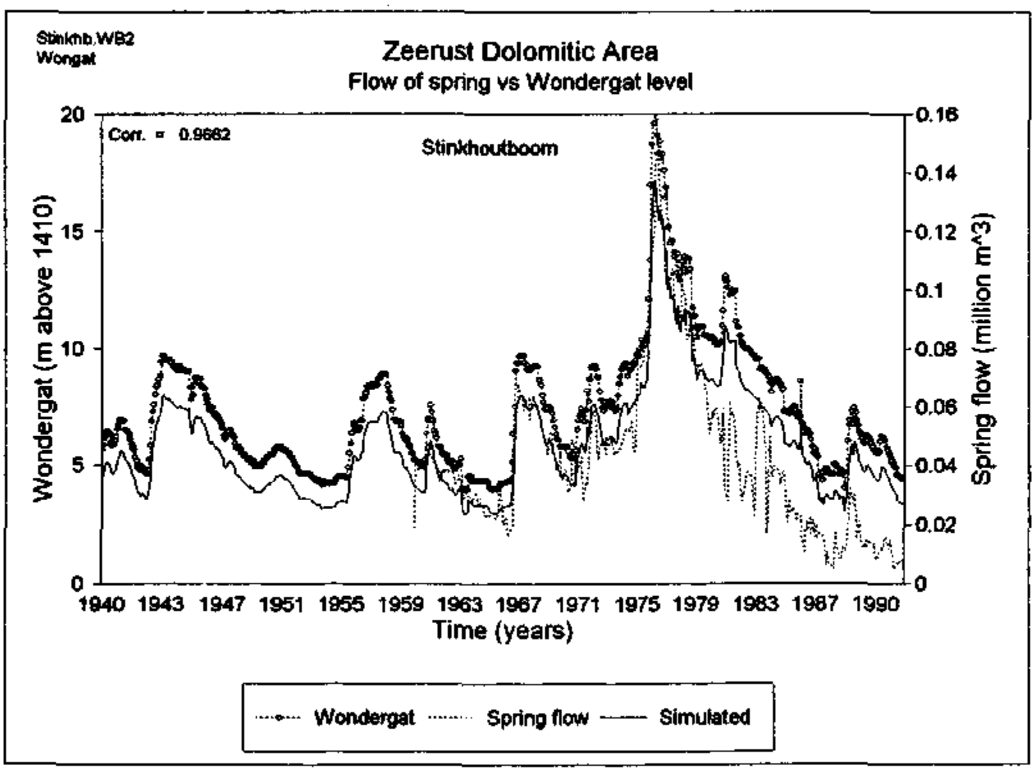


Fig. 4.2.1.1.4 Flow of Stinkhoutboom in relation to the Wondergat levels.

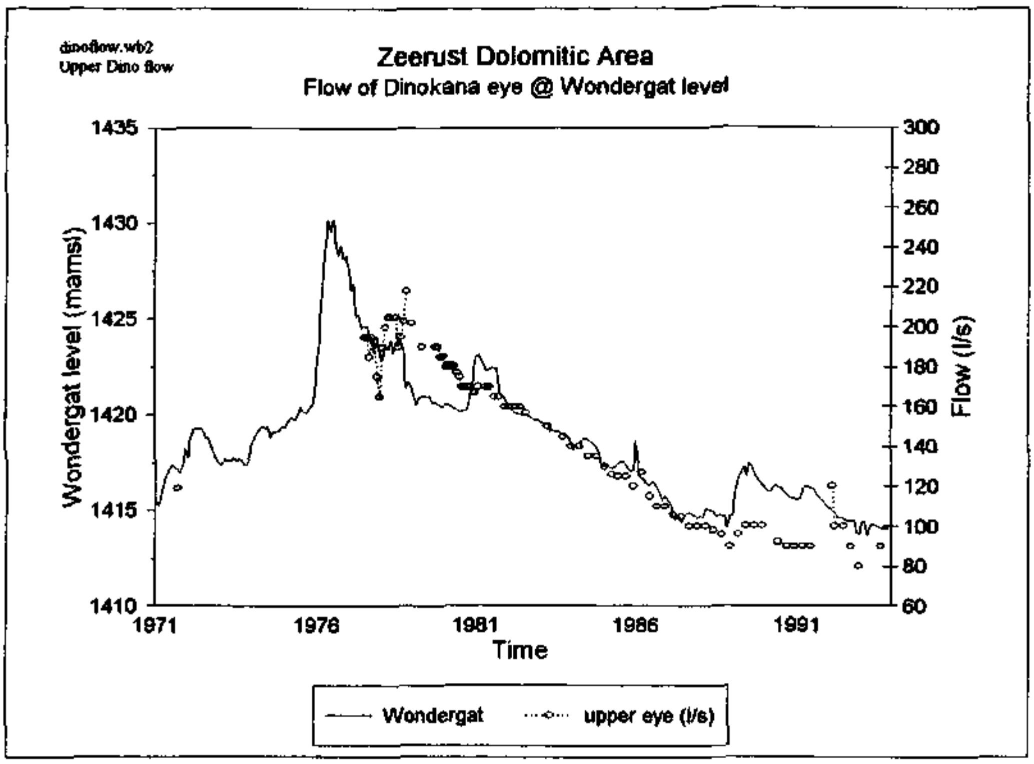


Fig. 4.2.1.1.5 Flow of Dinokana eye in relation to the water levels of the Wondergat.

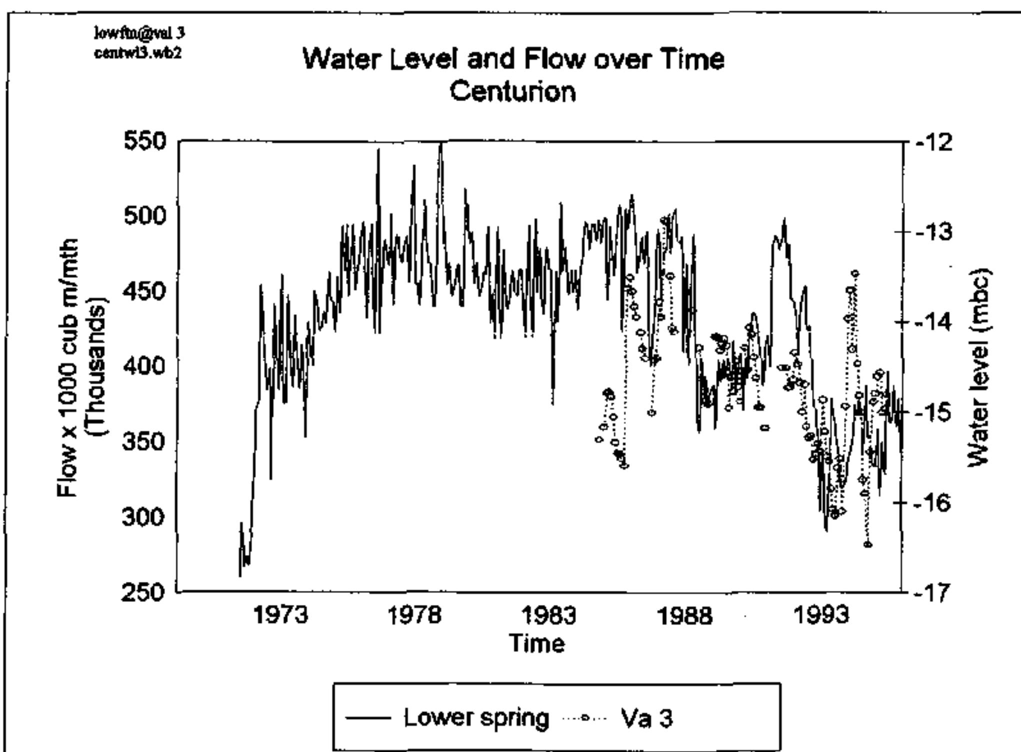


Fig. 4.2.1.2.1 Flow of the Lower Fountain in relation to the water level fluctuation of borehole Va3 at Valhalla.

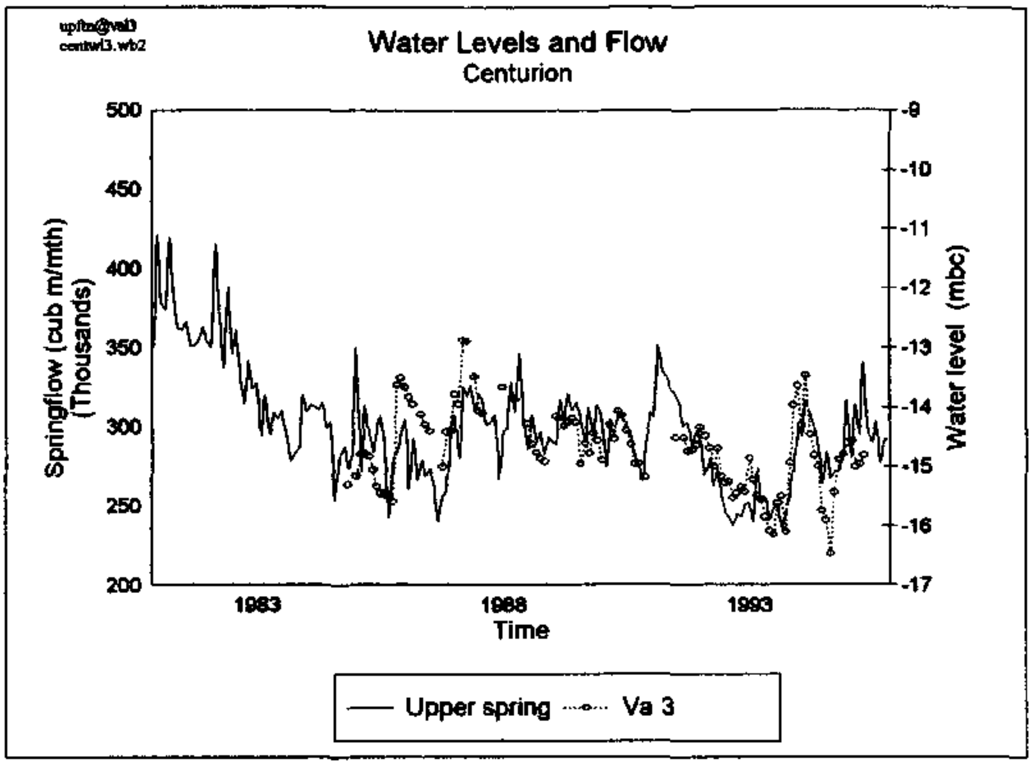


Fig. 4.2.1.2.2 Flow of the Upper Fountain in relation to the water level of borehole Va3 at Valhalla.

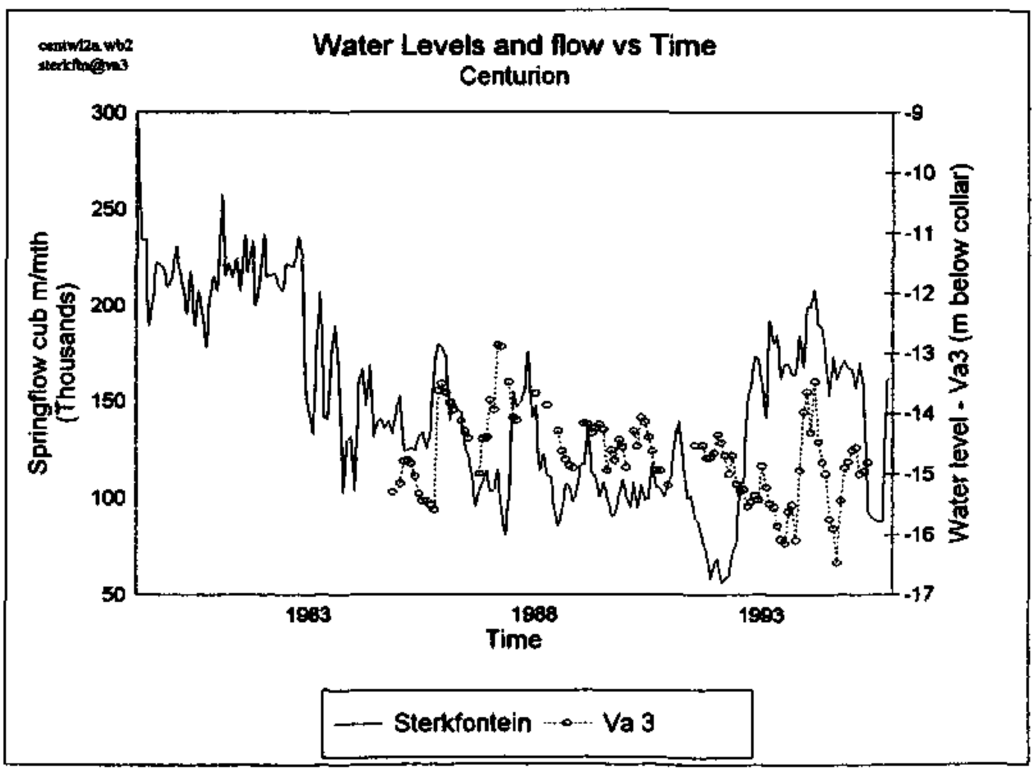


Fig. 4.2.1.2.3 Flow of the Sterkfontein eye in relation to the water level of borehole Va3 at Valhalla.

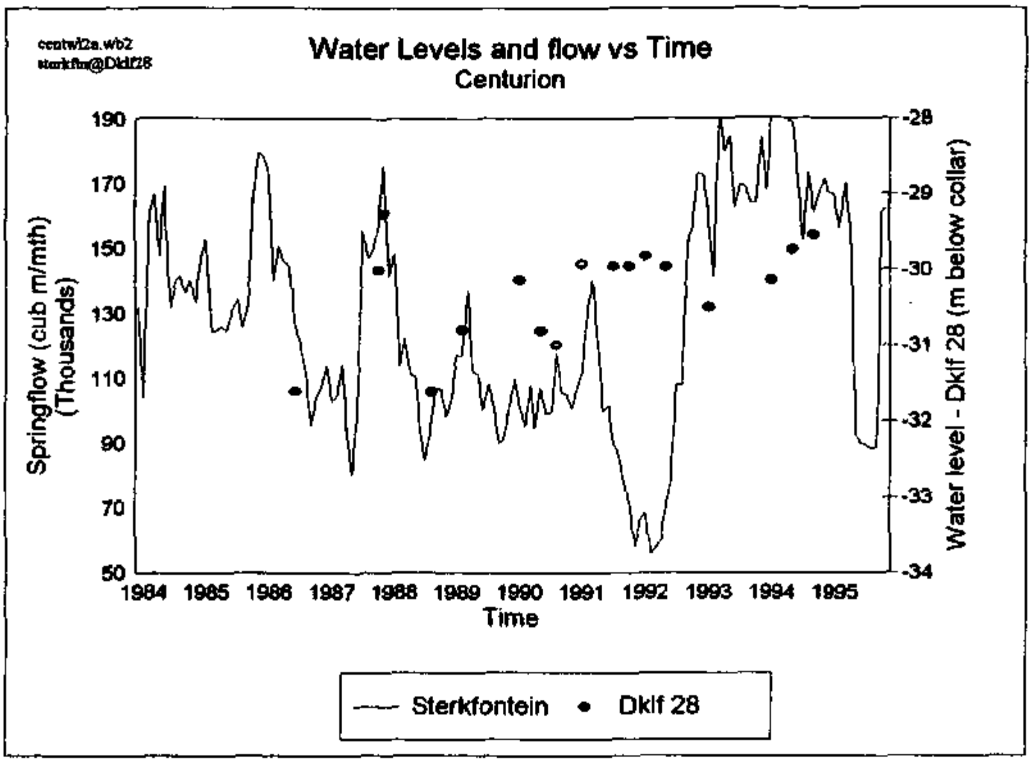


Fig. 4.2.1.2.4 Flow of the Sterkfontein eye in relation to the water levels of borehole Dklf28 (Doornkloof 28).

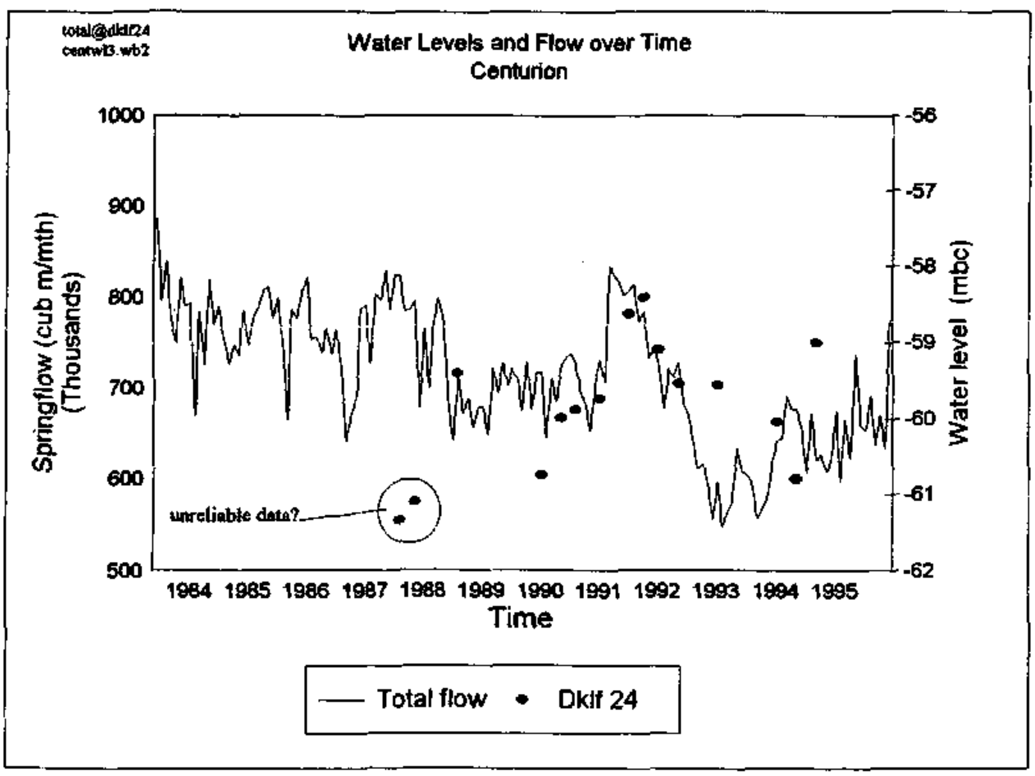


Fig. 4.2.1.2.5 Total flow of the Fountains in relation to the water levels of borehole DKlf24.

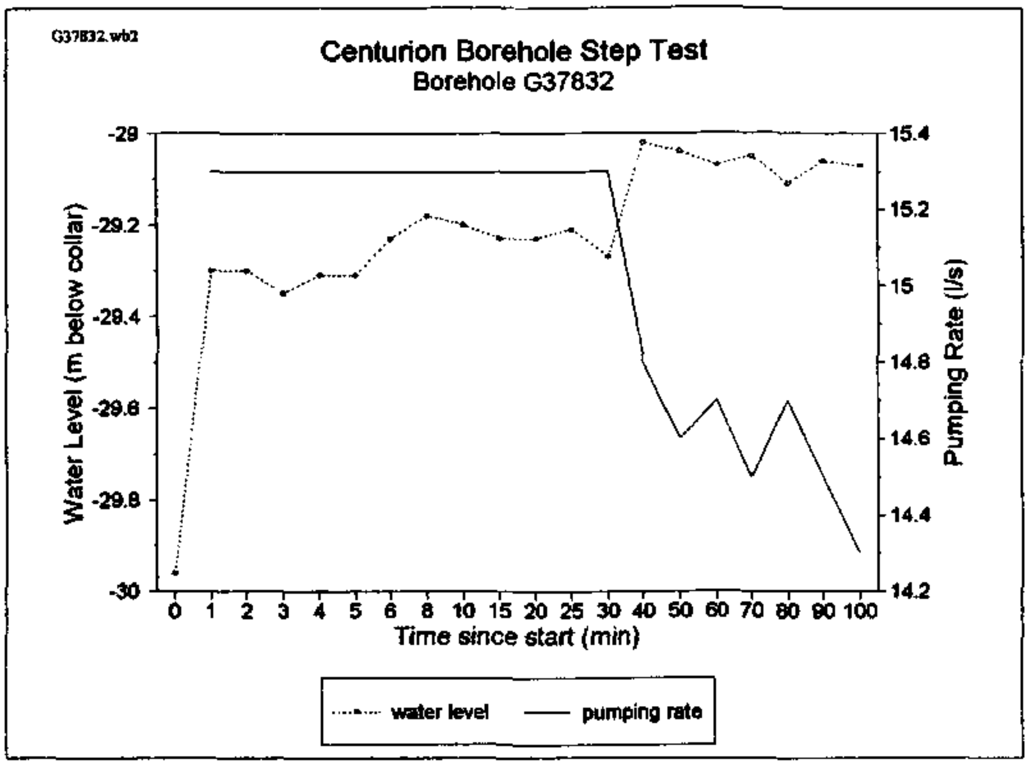


Fig. 4.2.1.2.6 Water level in abstraction borehole in Centurion area indicating a rise during the initial period of pumping.

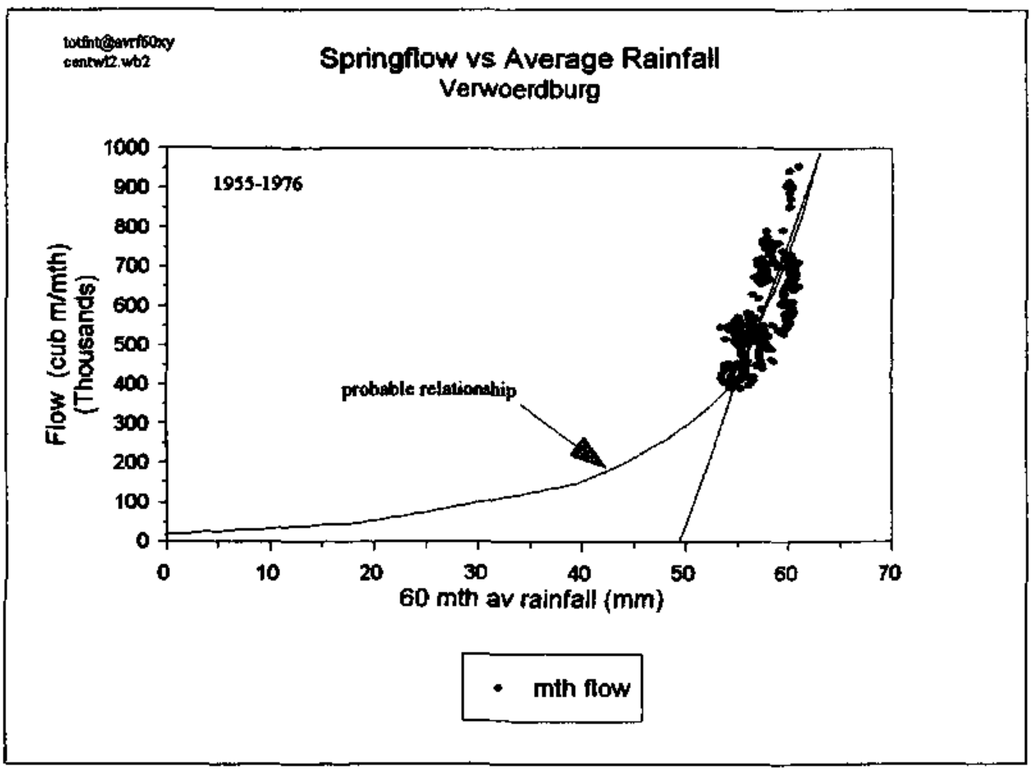


Fig. 4.2.1.4.1 Flow of the Sterkfontein eye in relation to the average rainfall over 60 preceding months.

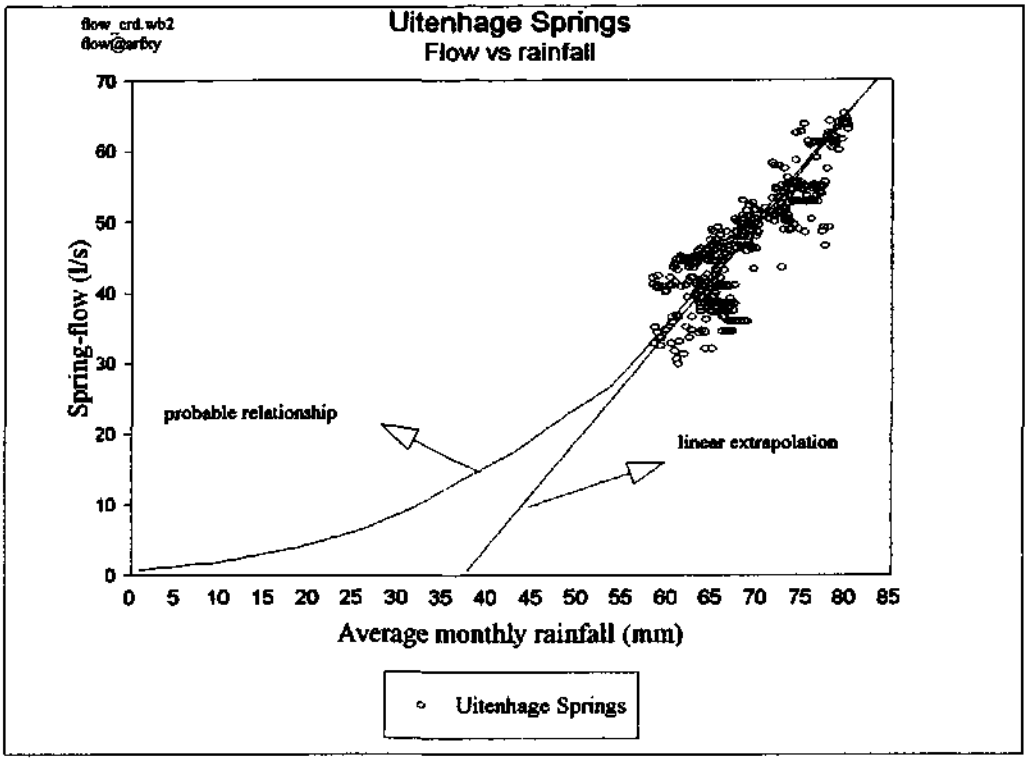


Fig. 4.2.1.4.2 Flow of the Uitenhage spring in relation to the average rainfall over 132 months preceding a specific month.

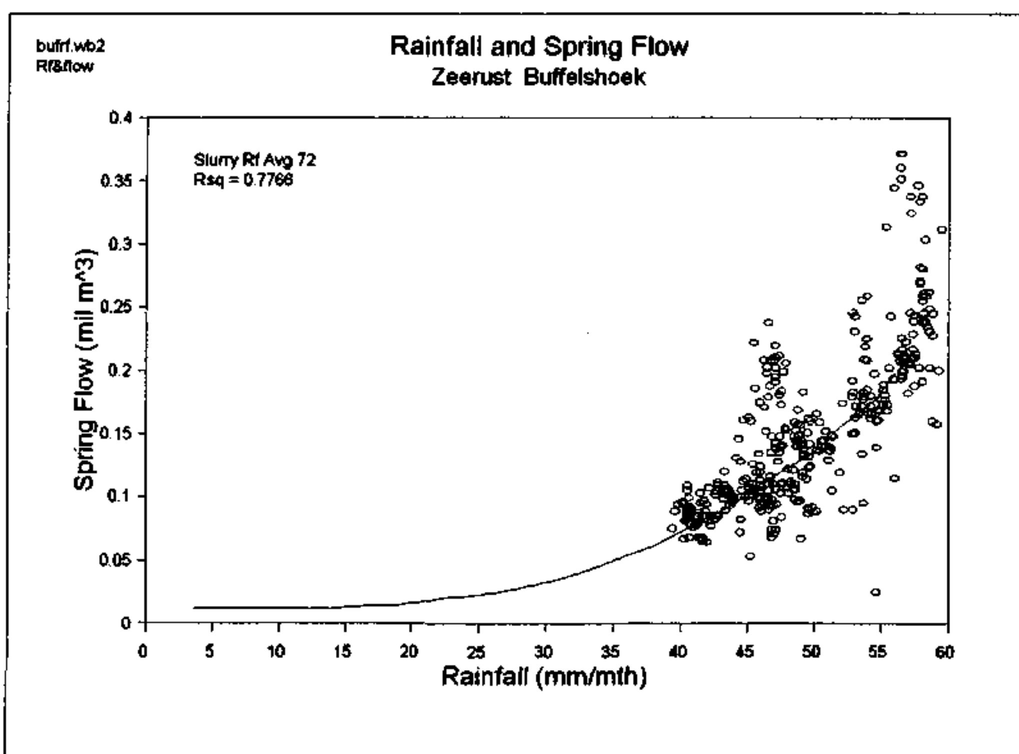


Fig. 4.2.1.4.3 Average monthly flow of the Buffelshoek eye in relation to rainfall experienced during period of equal volume.

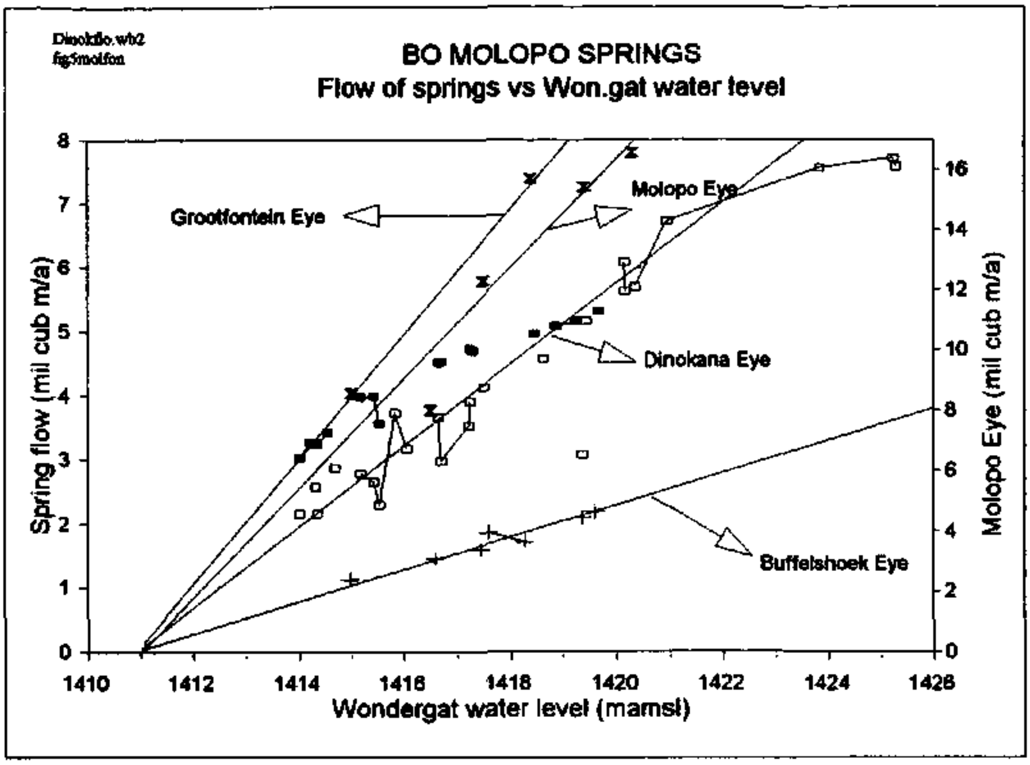


Fig. 4.2.1.4.4 Flow of different springs in the Bo Molopo area in relation to the groundwater level of the Wondergat.

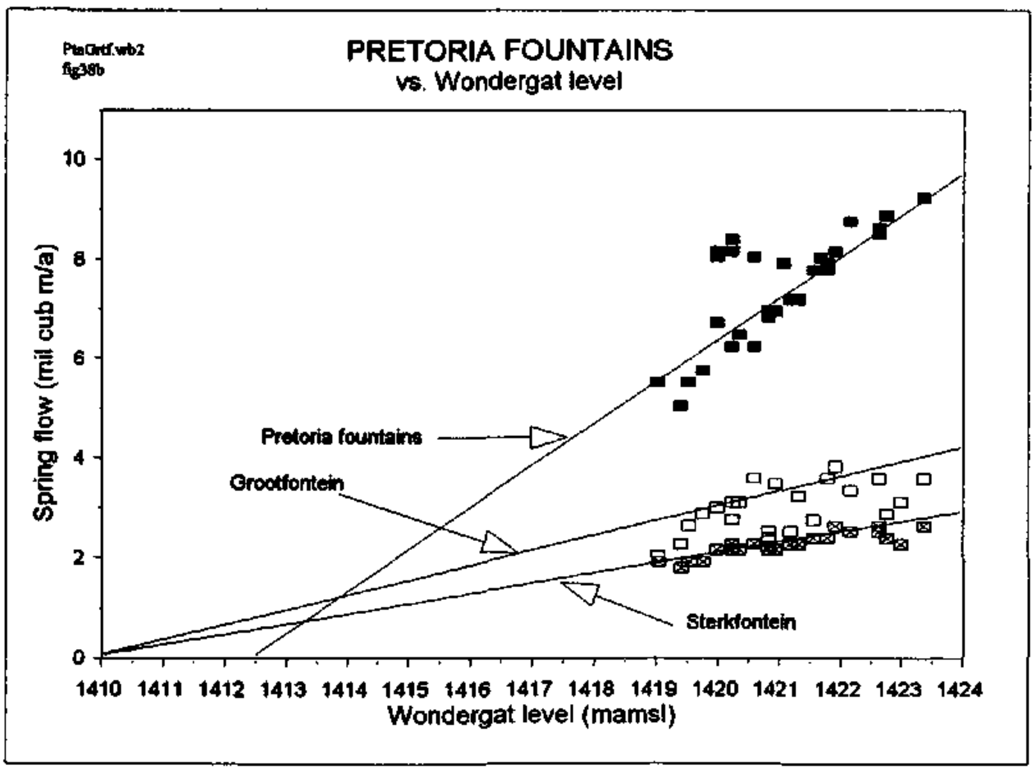


Fig. 4.2.1.4.5 Flow of different springs in the Pretoria area in relation to the groundwater level of the Wondergat.

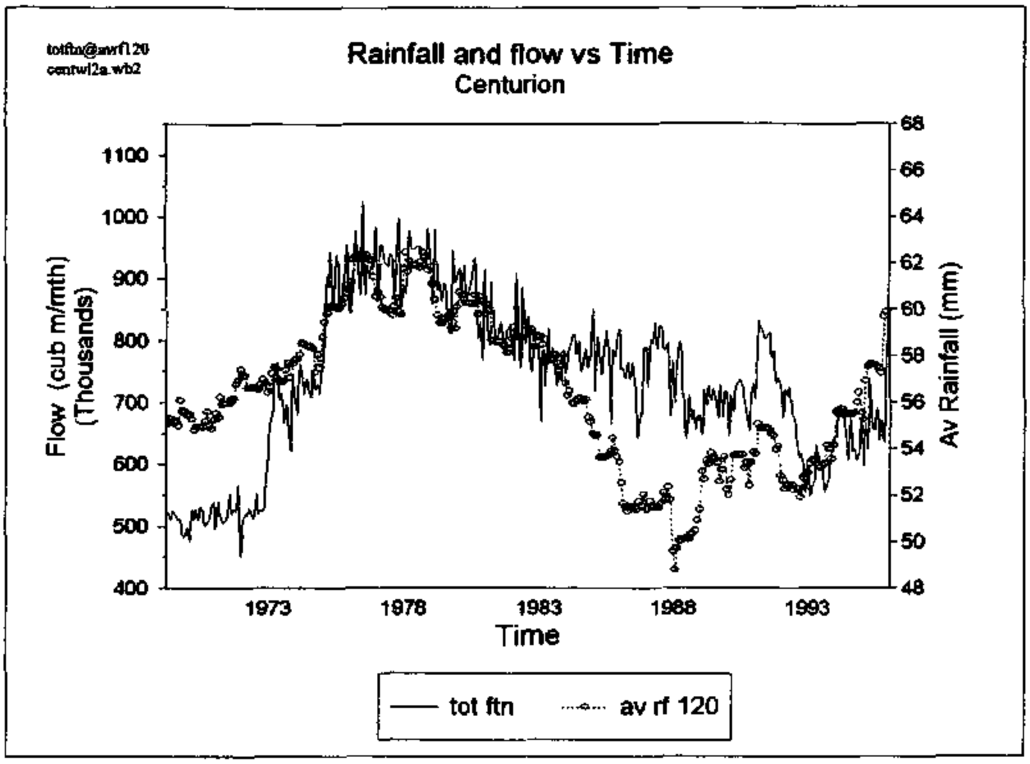
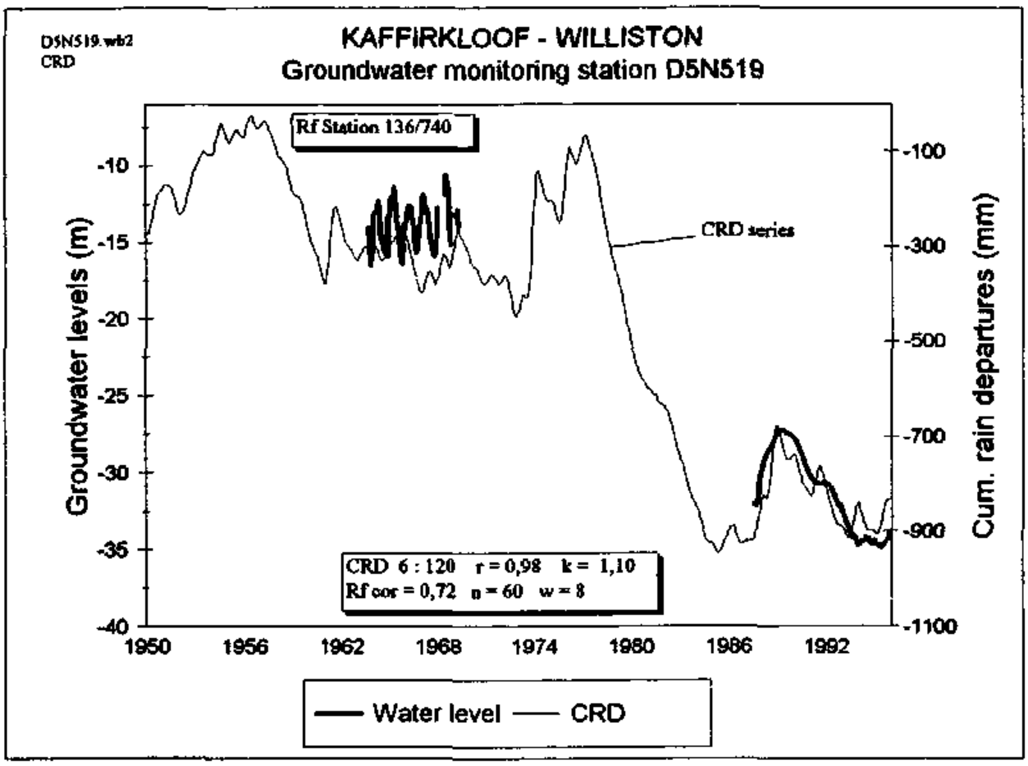


Fig. 4.2.1.4.6 Relationship between 120-month average rainfall and the flow of the Pretoria fountains.



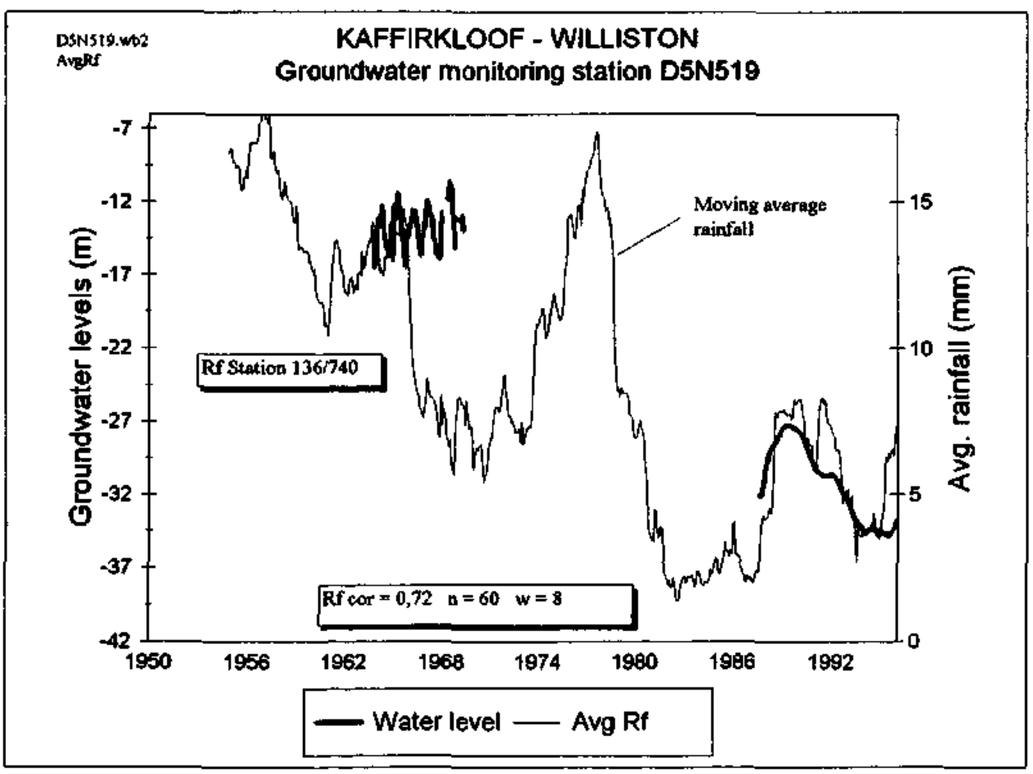


Fig. 5.2.1 Infilling of missing groundwater levels by means of CRD relationship of station D5N519 (Williston).

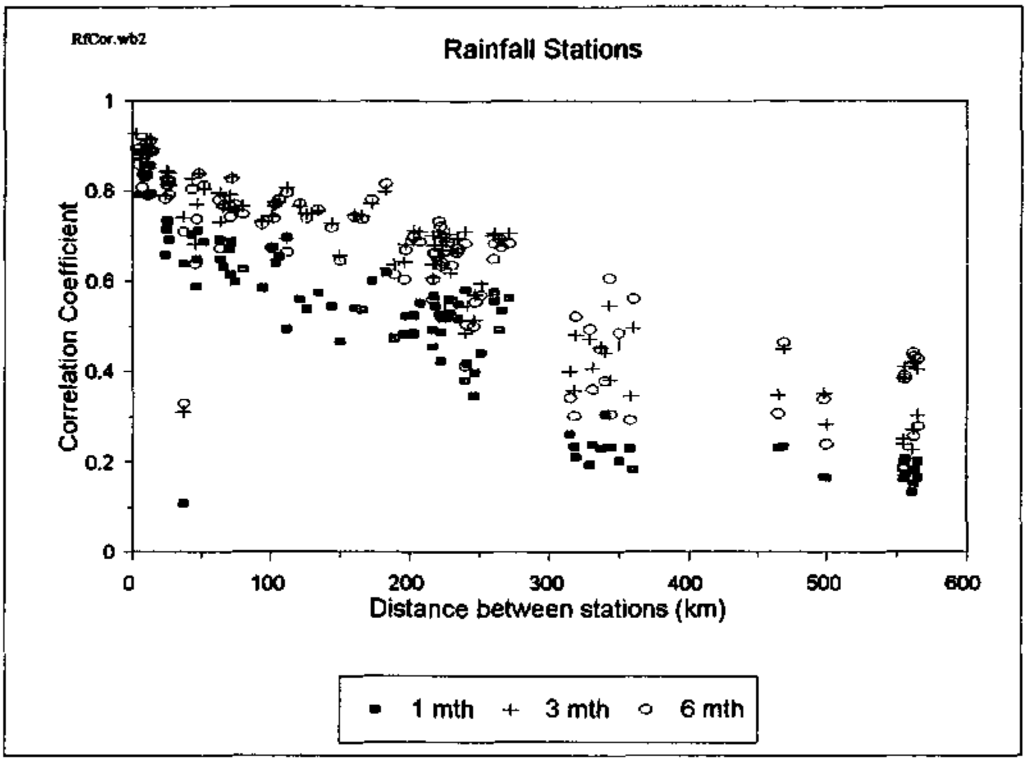


Fig. 5.3.1.1 Correlation coefficients for Slurry rainfall in relation to stations at different distances.

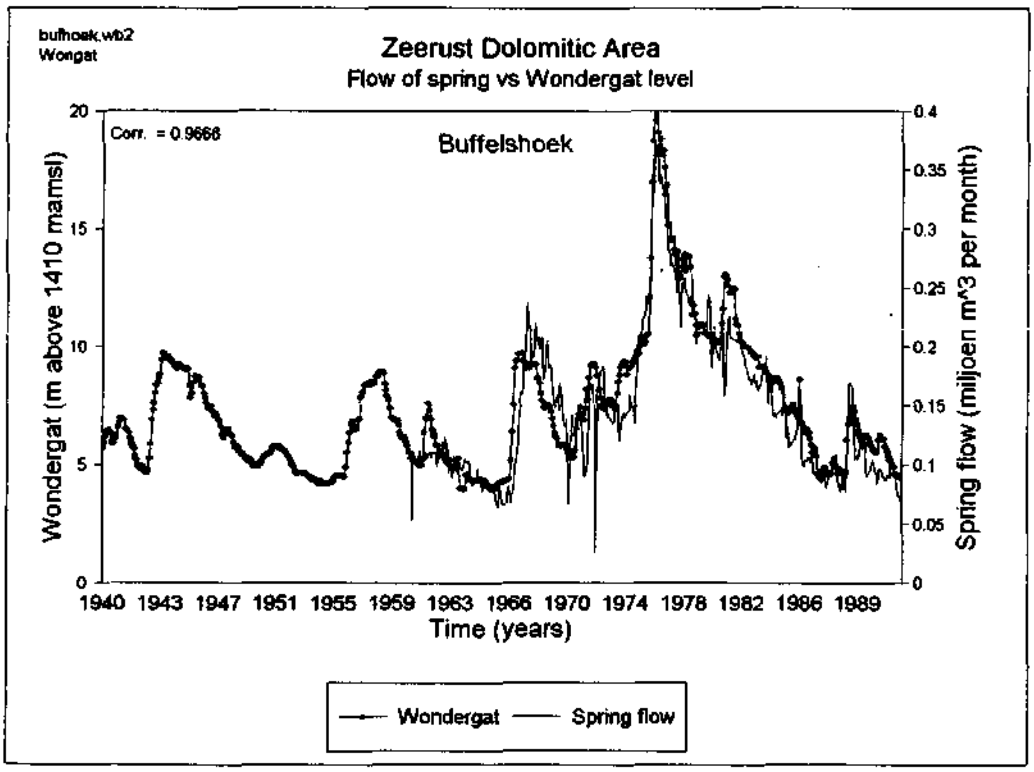


Fig. 5.3.2.1 Actual and simulated flow of the Buffelshoek eye in relation to the Wondergat levels.

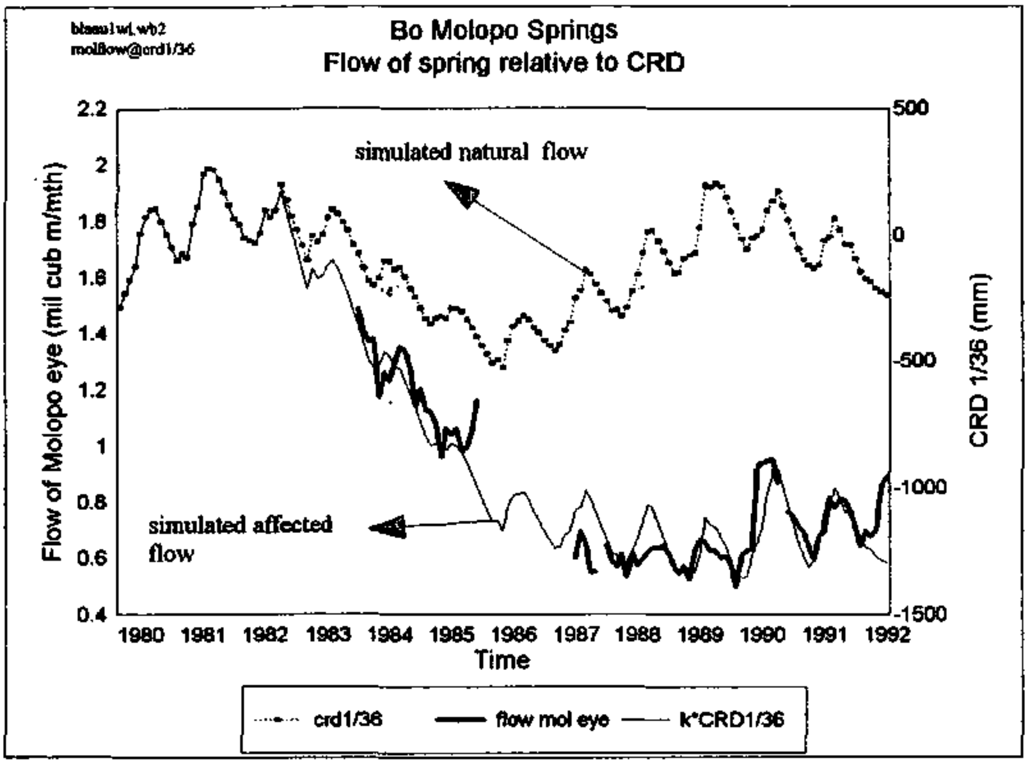


Fig. 5.3.2.2 Comparison between the flow of Molopo eye and that simulated by the $\frac{1}{36}CRD$ series having incorporated a k-factor representing the effect of abstraction.

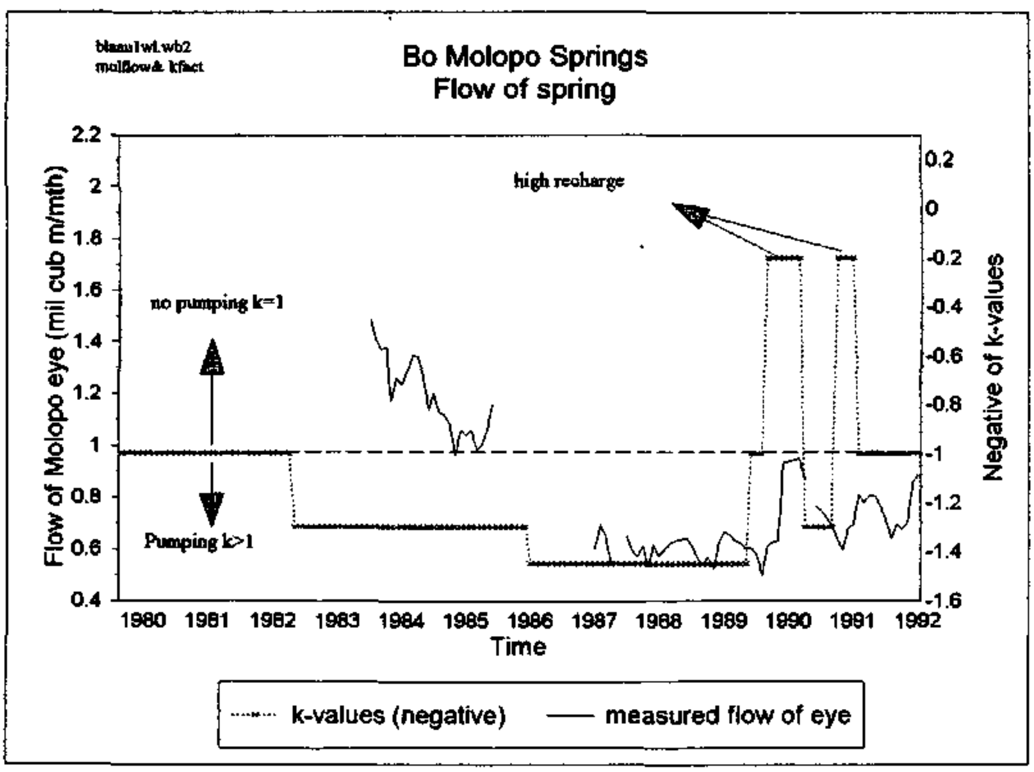


Fig 5.3.2.3 Period of abstraction indicated by k-values >1 (shown as negative values) for the CRD relationship in relation to the observed flows of the Molopo eye.

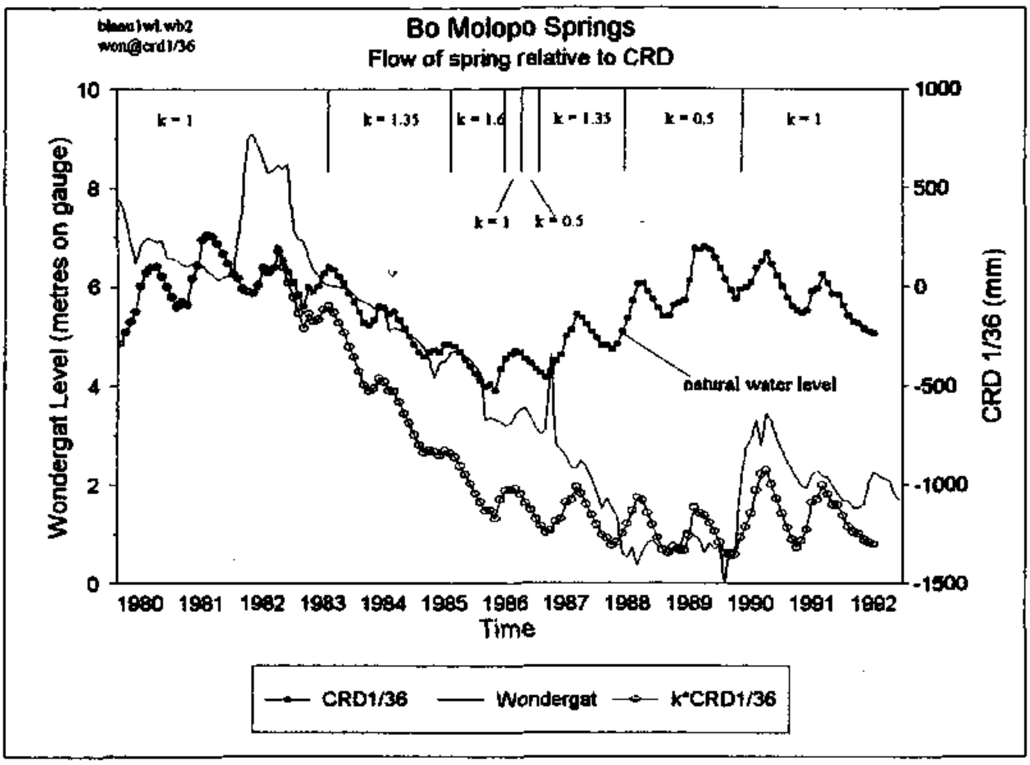


Fig. 5.3.2.4 Different k-values which have to be introduced to simulate the water levels of the Wondergat from the $_{36}^{1}CRD$ series.

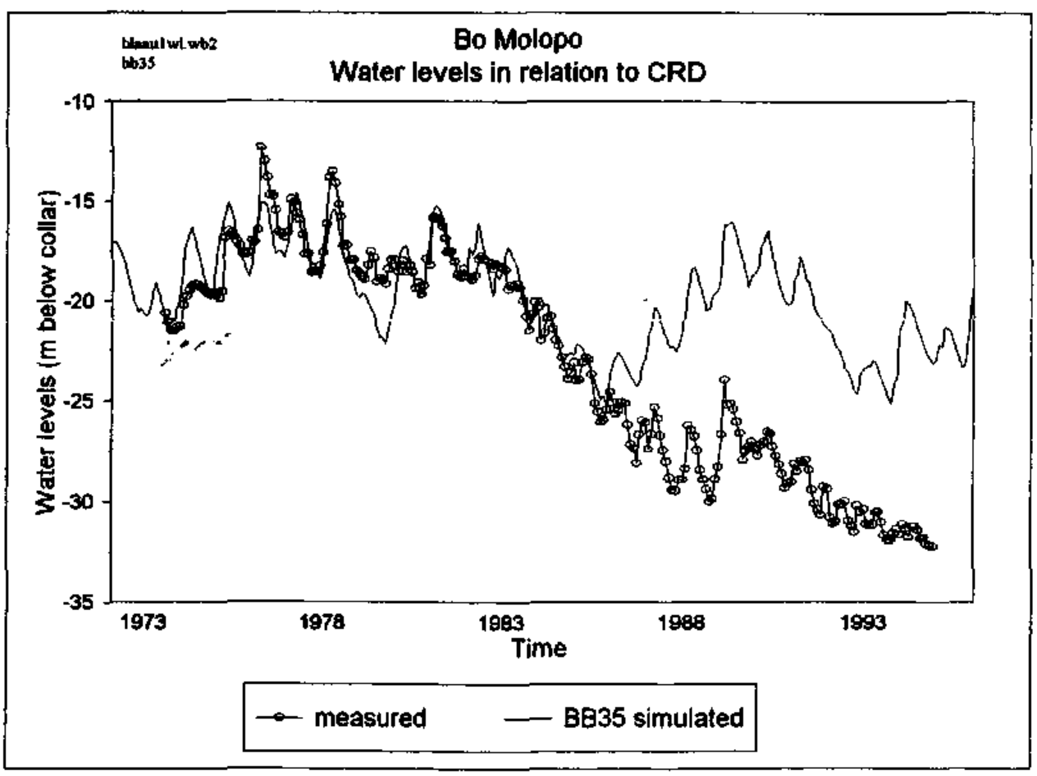


Fig. 5.3.2.5 Correspondence between the measured water levels in monitoring borehole BB35, and those simulated with the CRD series.

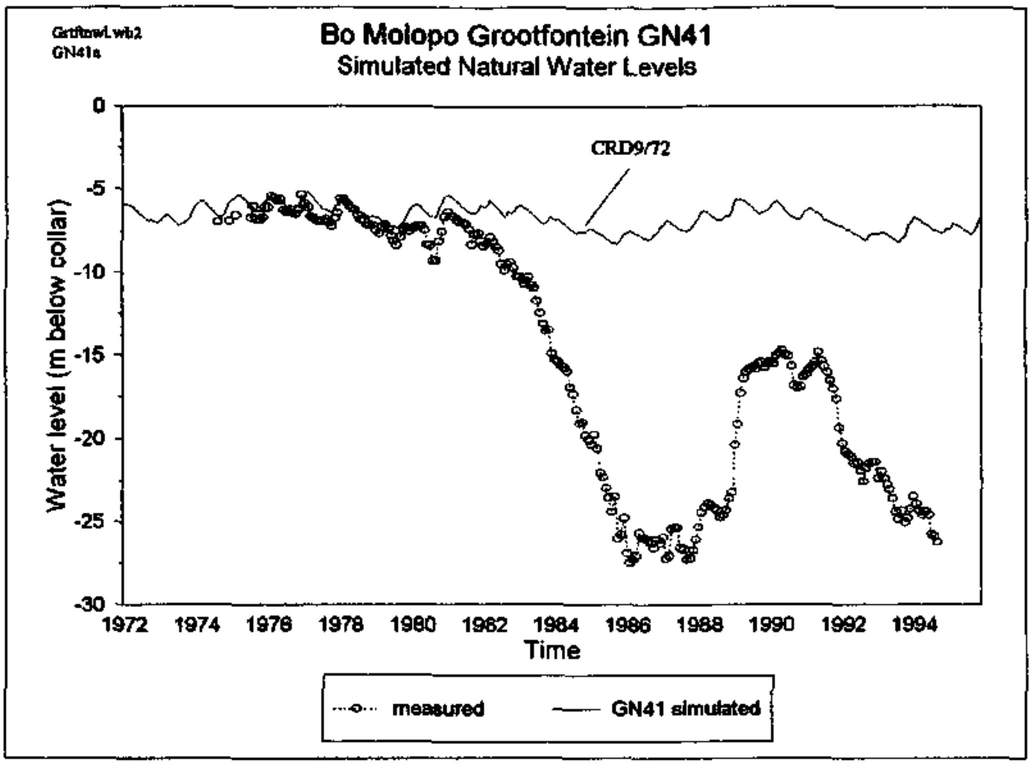


Fig. 5.3.2.6 Correspondence between the measured water levels in monitoring borehole GN41, and those simulated with the CRD series.

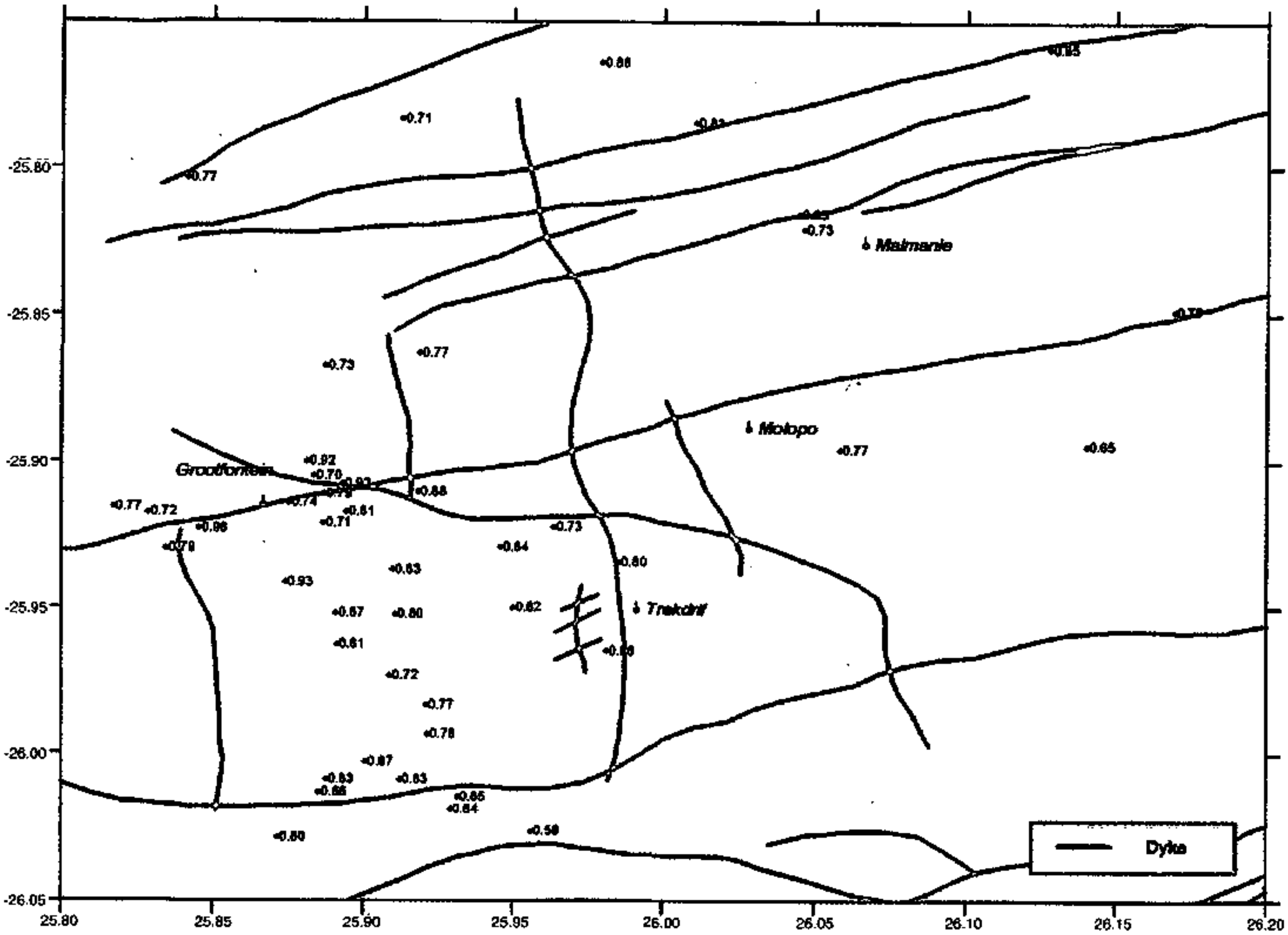


Fig. 5.3.3.1: The correlation coefficiens indicating the reliability of the CRD regression in simulating the groundwater levels and estimating the aquifer storativity (Bo-Molopo Area).

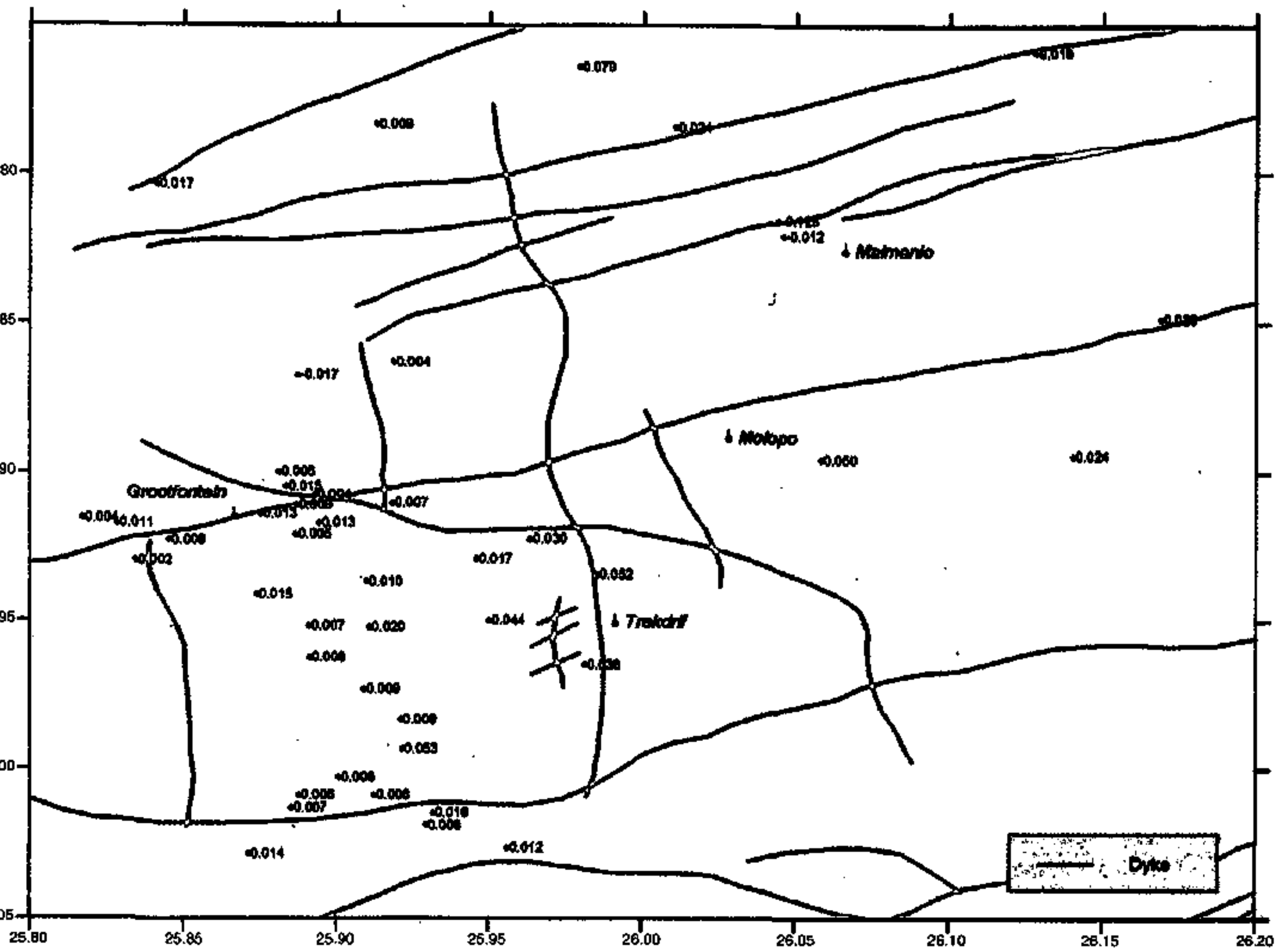


Fig. 5.3.3.2: Storativity values (S) derived by means of the CRD-relationship for different hydrographs in the Bo-Molopo Area.

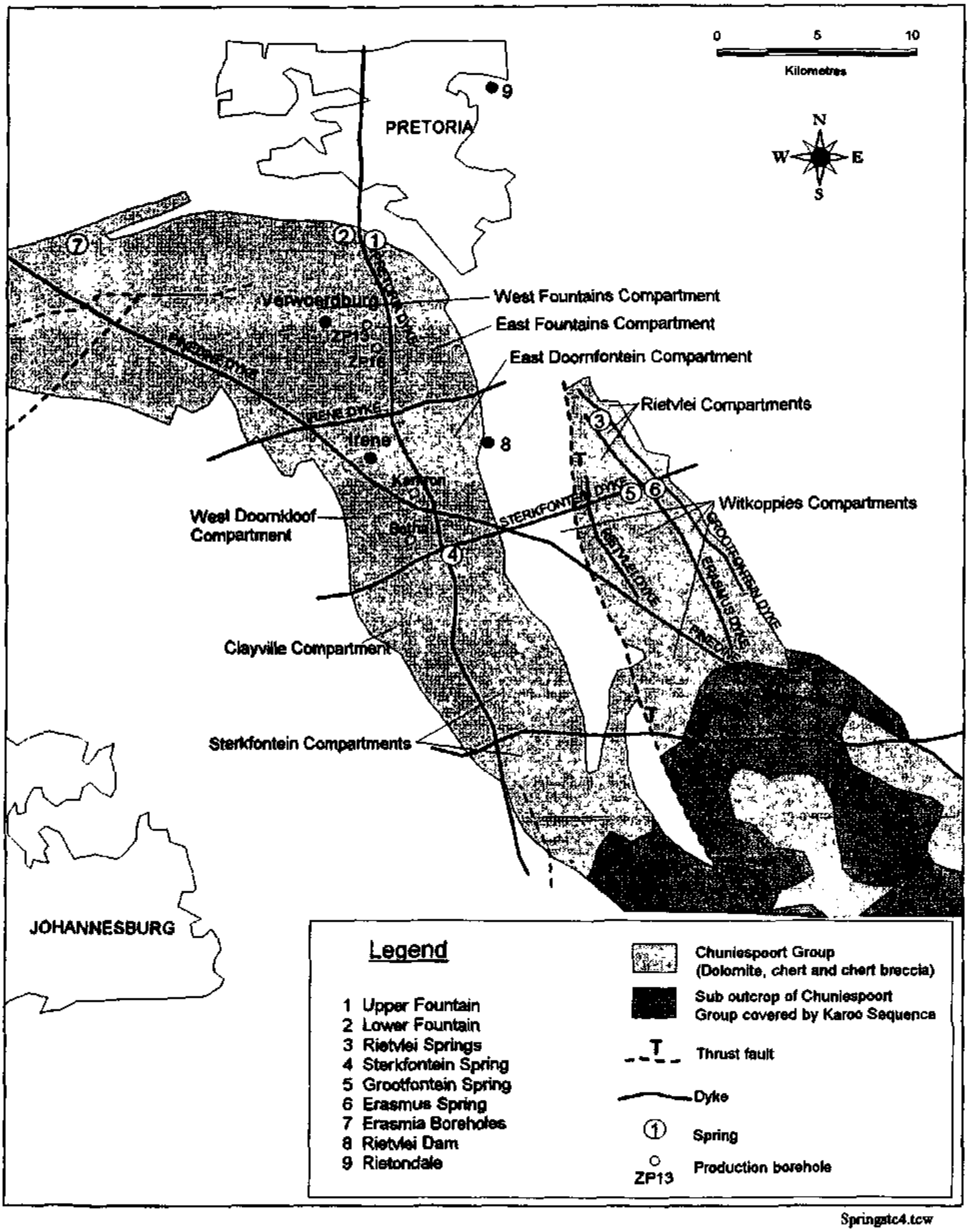


Fig. 5.4.1.1 **Groundwater level monitoring points in the Pretoria dolomitic area.**

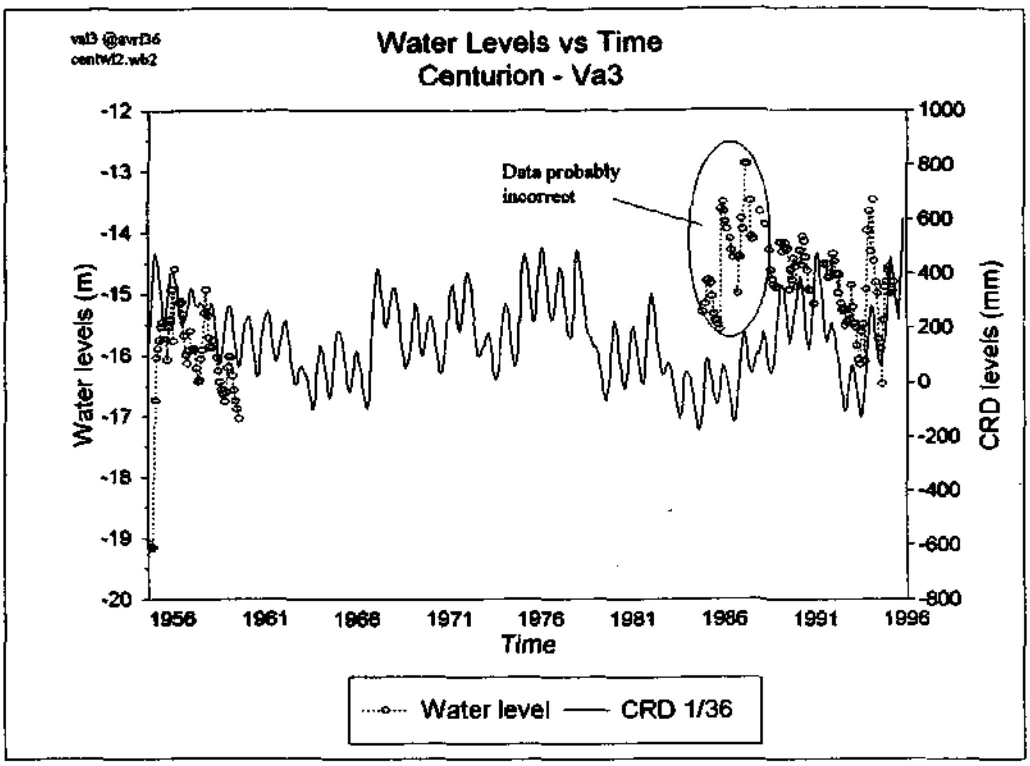


Fig. 5.4.1.2 Measured water levels in borehole Va3 relative to values simulated by the ${}^{1}_{36}CRD$ series.

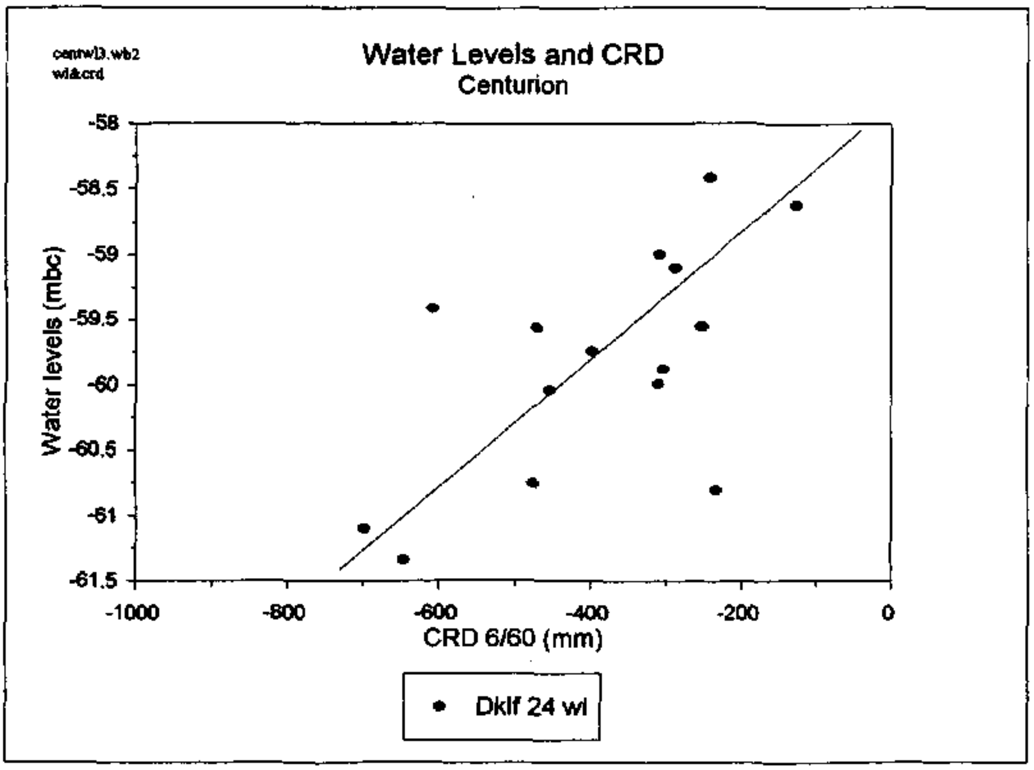


Fig. 5.4.1.3 Water levels in borehole Dklf24 in relation to the CRD series.

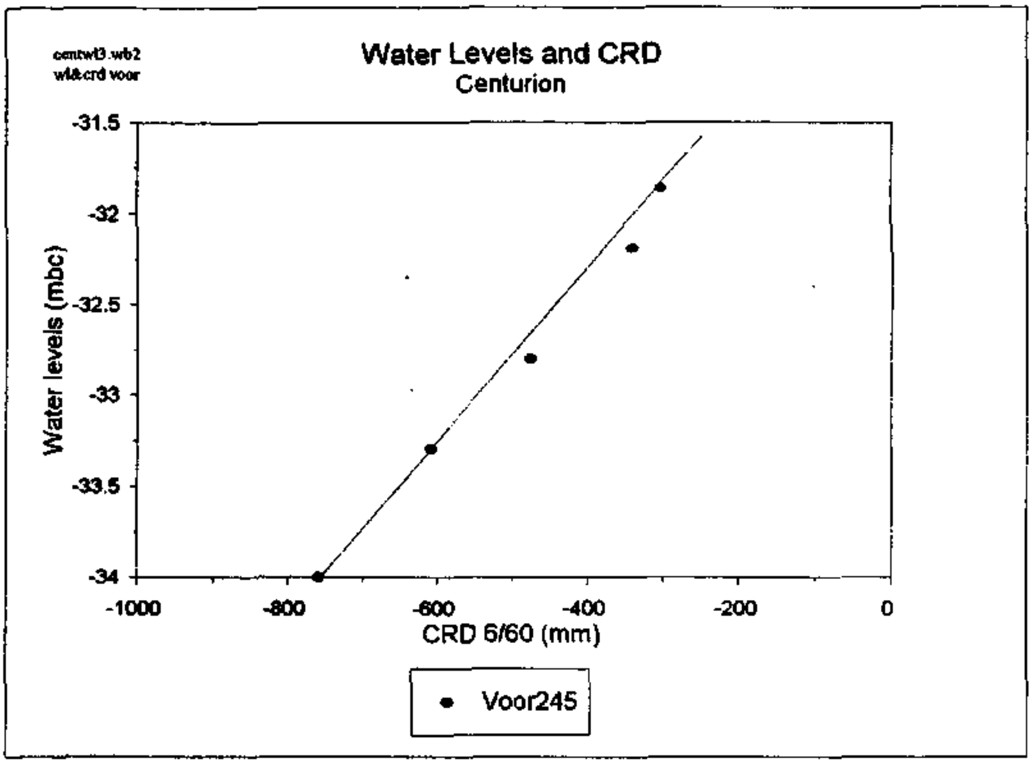


Fig. 5.4.1.4 Water levels in borehole Voor245 in relation to the CRD series.

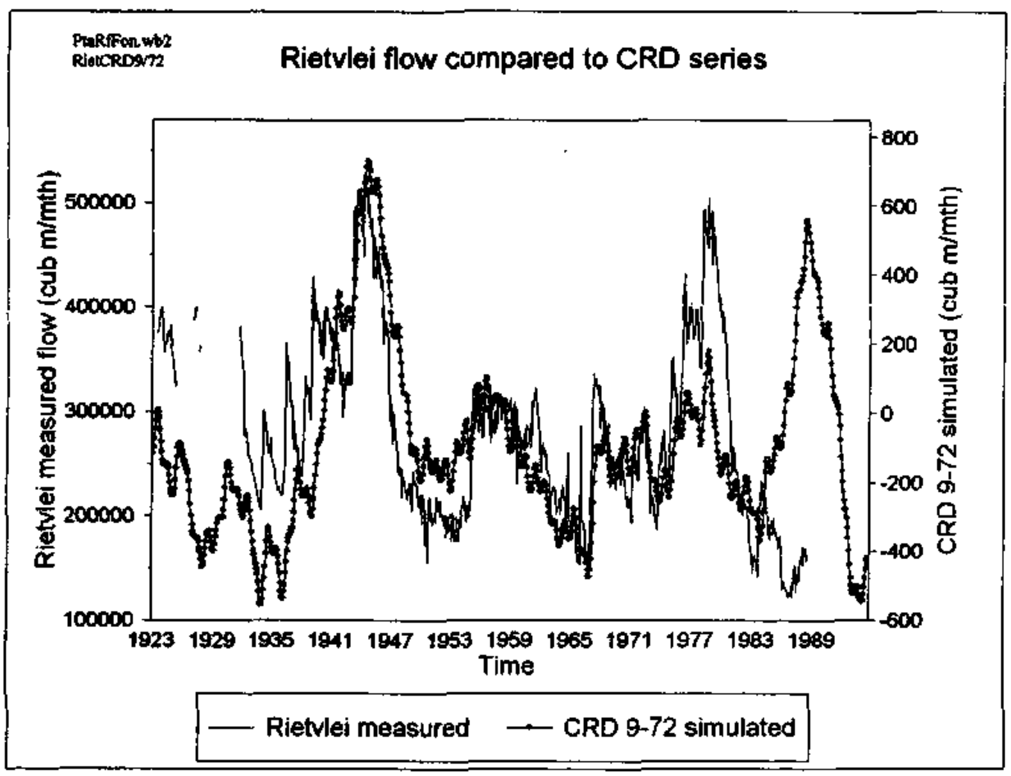


Fig. 5.4.2.1 Flow of the Rietvlei springs in relation to flow simulated by the ${}^{9}_{72}CRD$ series.

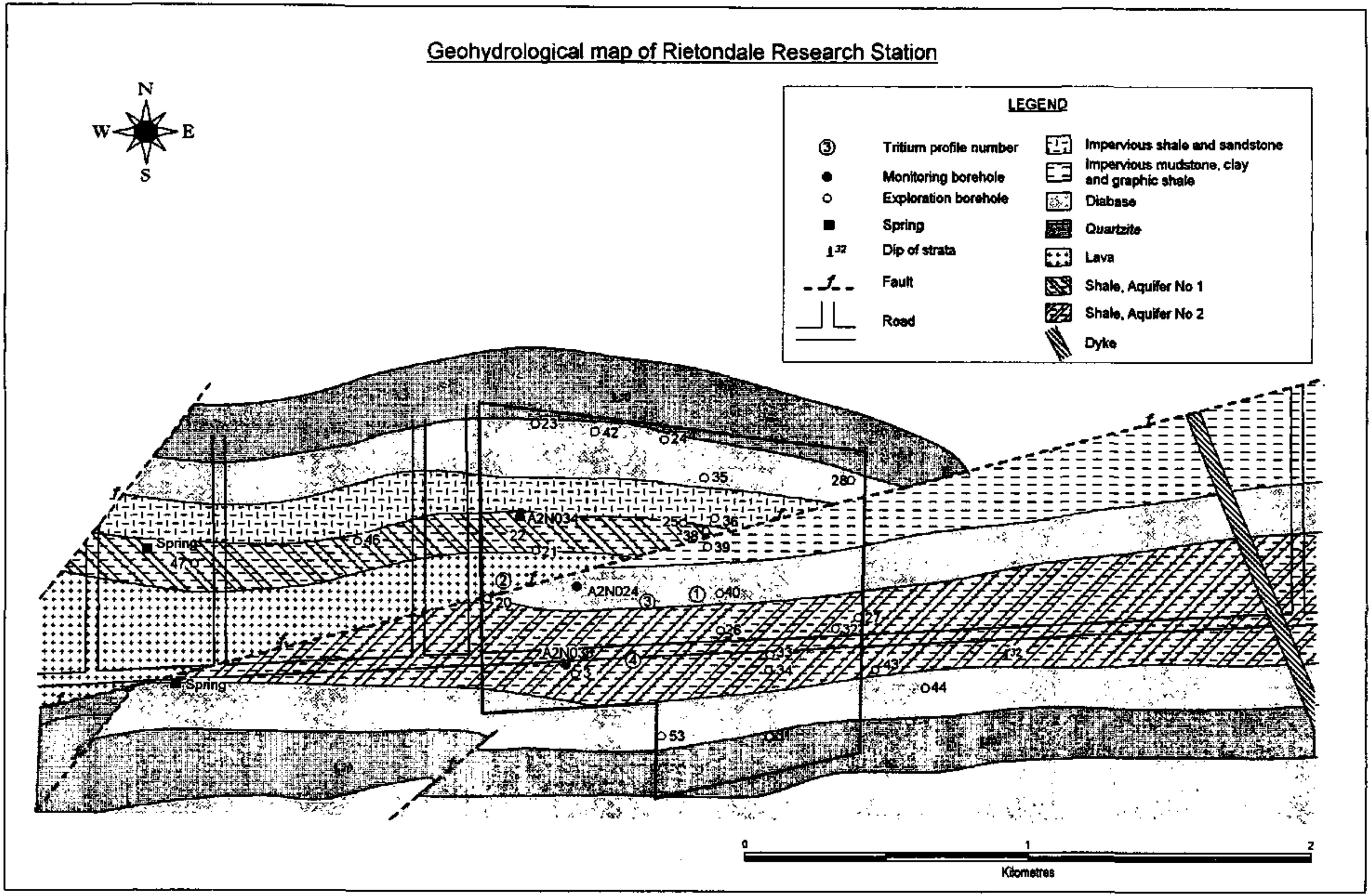


Fig. 5.4.3.1 Water level stations in the Rietondale aquifer, and monitoring points.

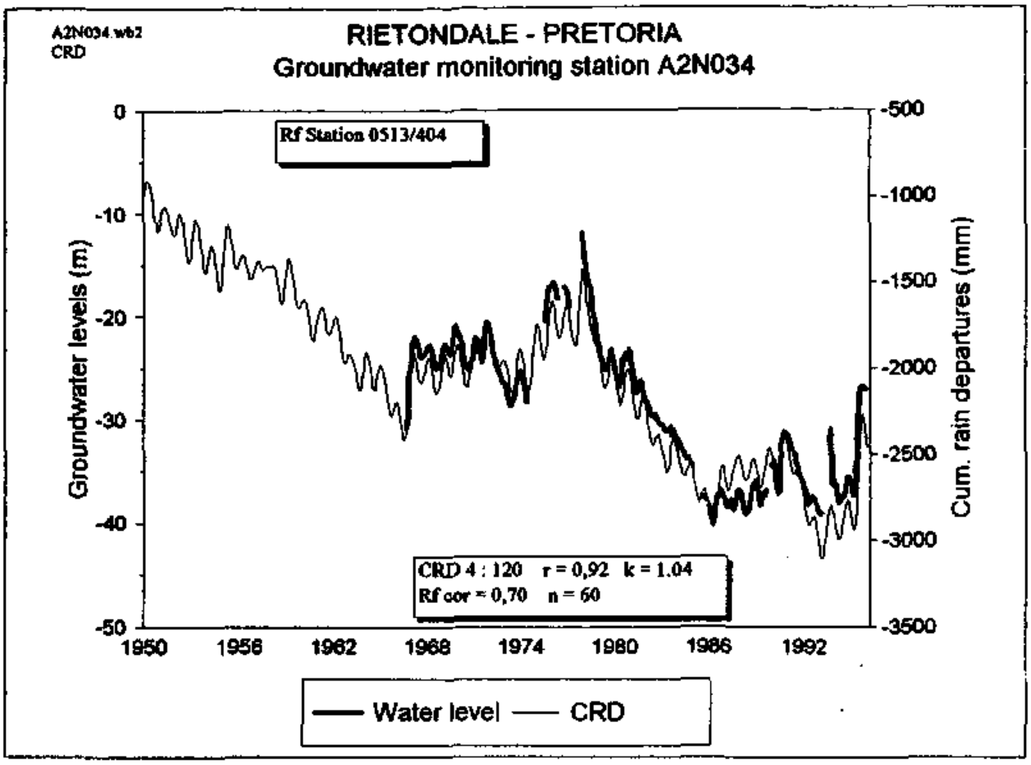


Fig. 5.4.3.2 Water levels in borehole A2N034 (Rietondale) in relation to the CRD series.

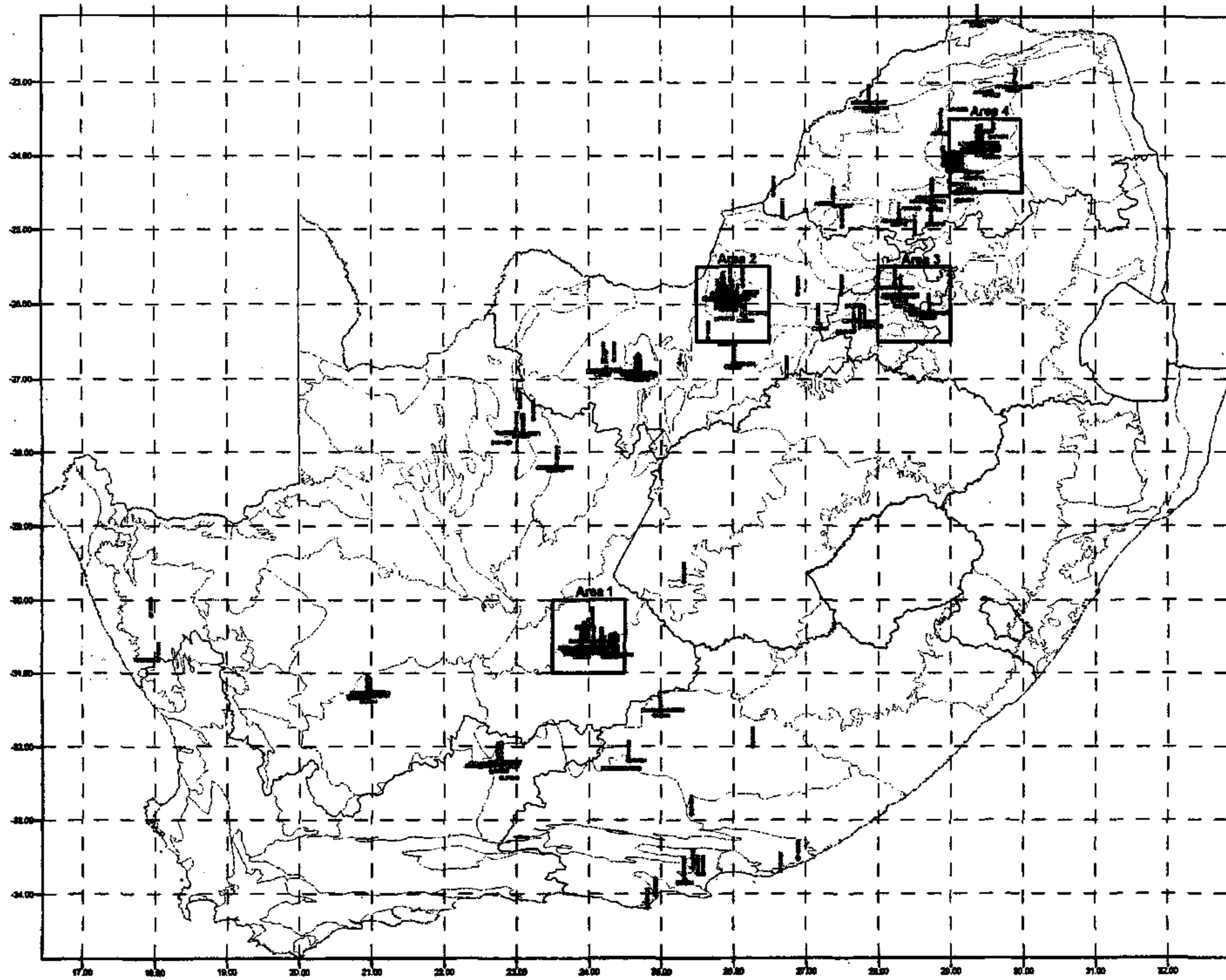


Fig. 5.4.5.1 Distribution of groundwater levels monitoring boreholes stored in the National Groundwater Database (NGDB).

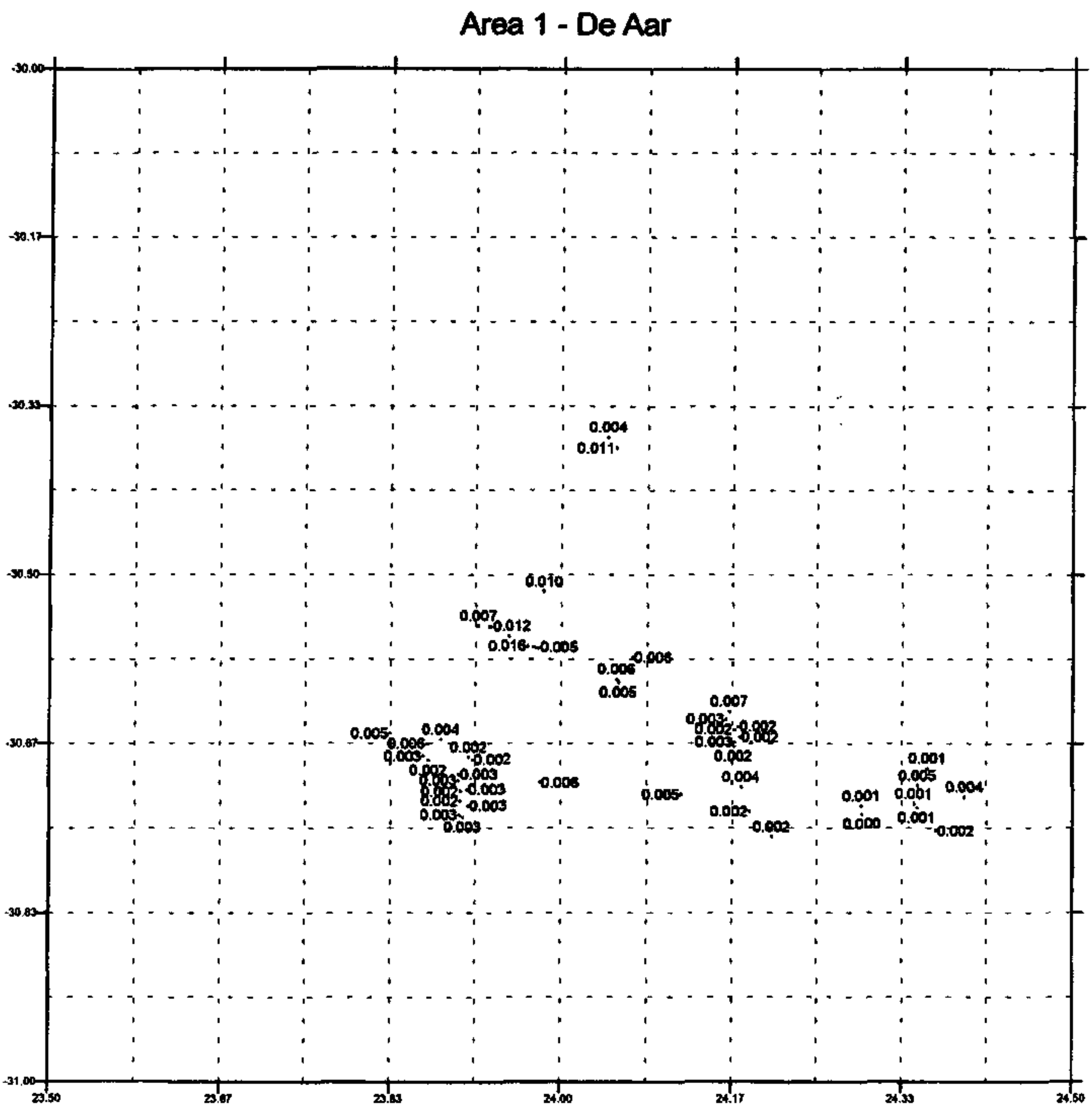


Fig. 6.2.1 : Storativity values derived from CRD relationship for area 1 – De Aar Karoo.

Area 2 - Bo Molopo

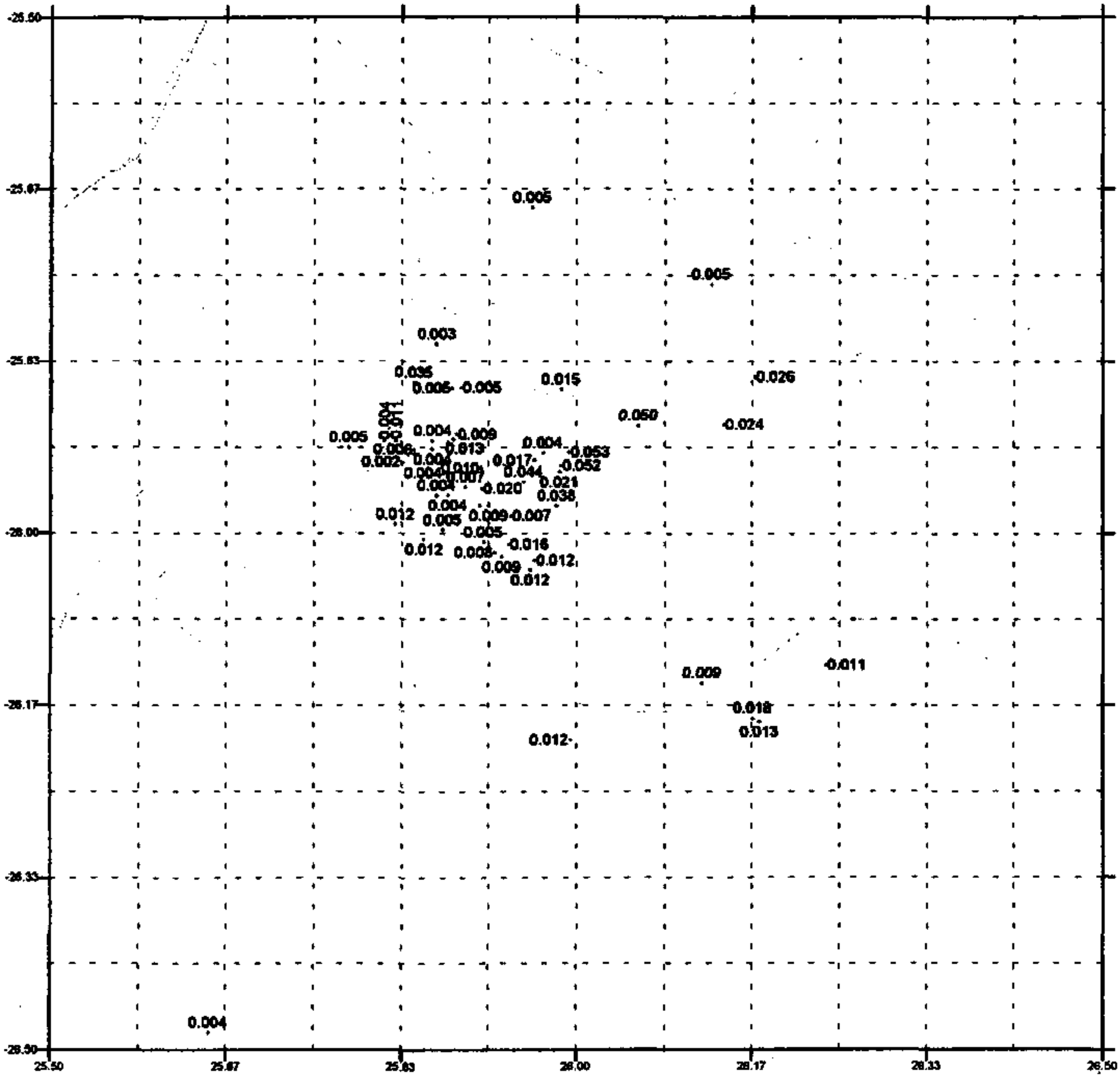


Fig. 6.2.2 : Storativity values derived from CRD relationship for area 2 – Bo Molopo.

Area 3 - Pretoria

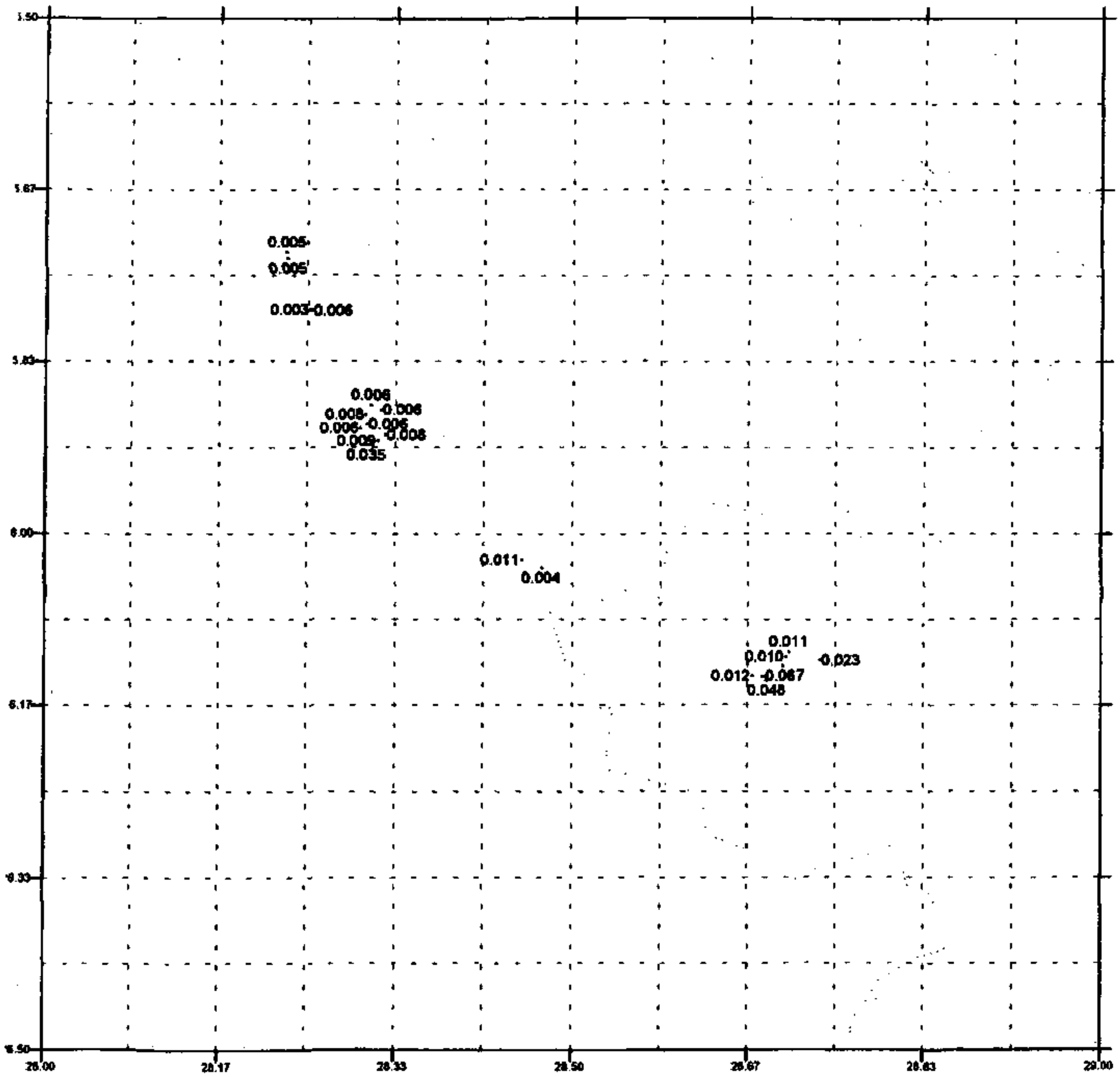


Fig. 6.2.3 : The aquifer storativity derived from the CRD relationship, of the Pretoria dolomitic aquifer.

Area 4 - Pietersburg

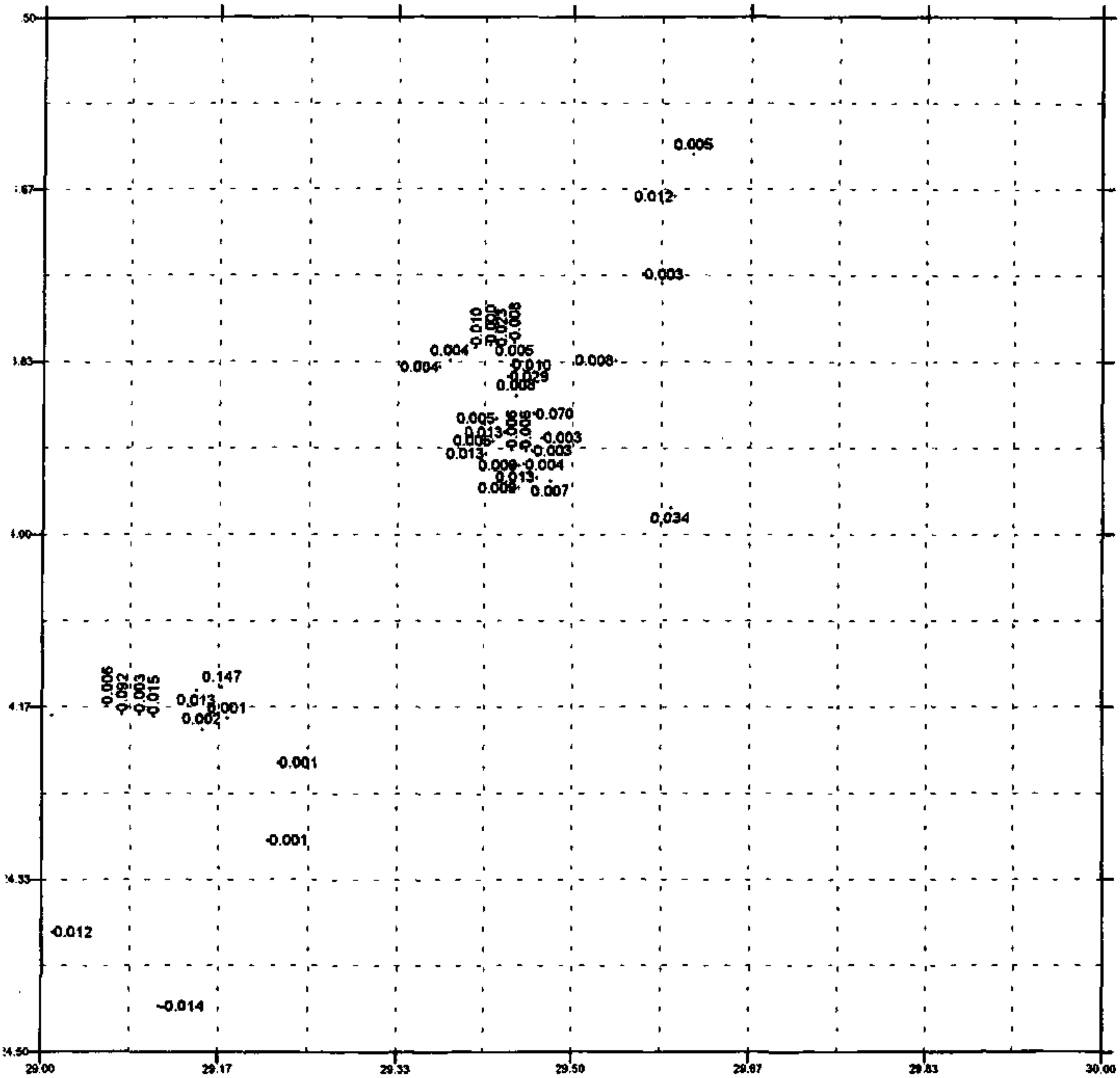


Fig. 6.2.4 : The aquifer storativity derived from the CRD relationship, of the Pietersburg granitic aquifer.

Area 1 - De Aar

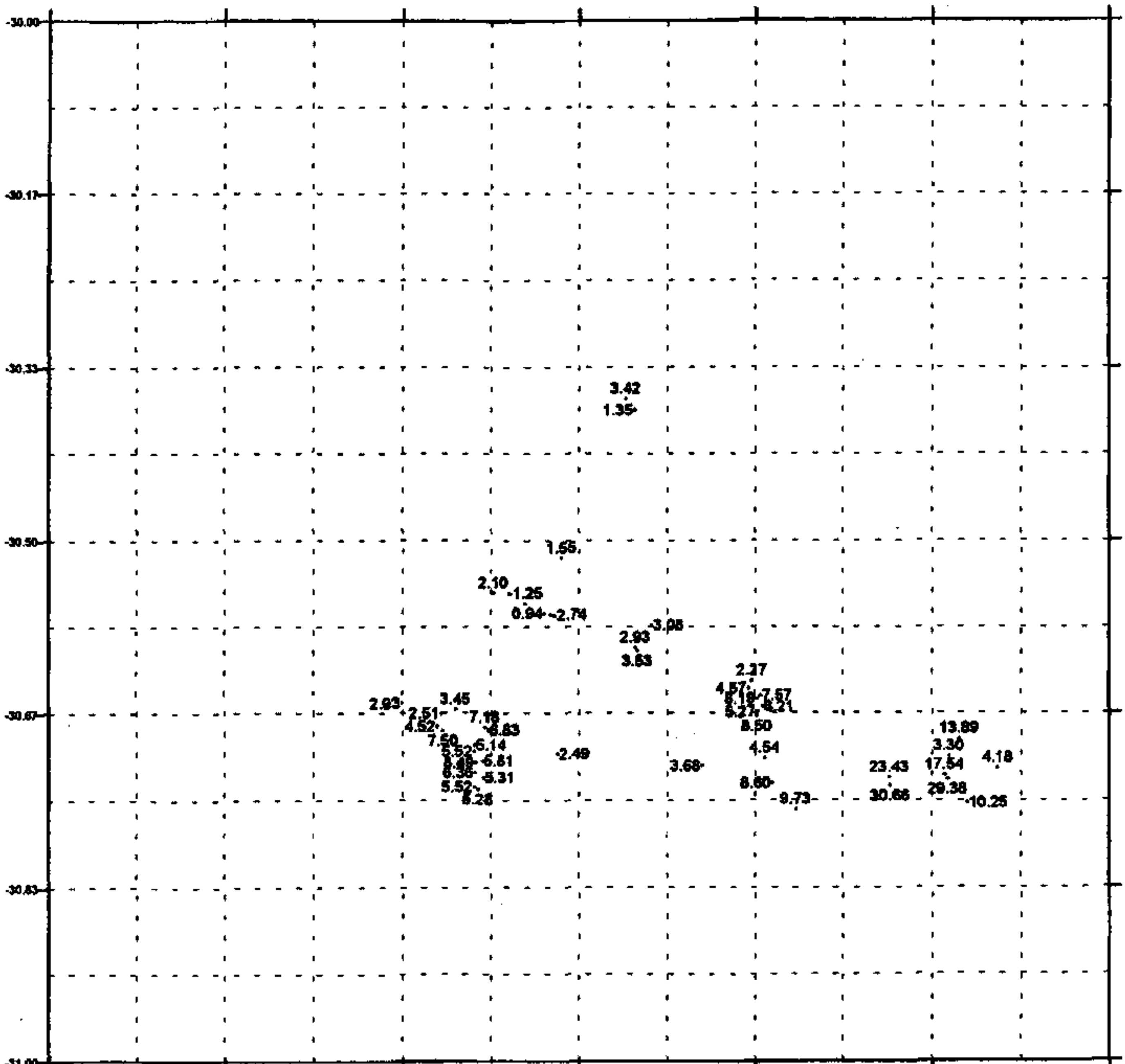


Fig. 6.3.1 : Unit response of groundwater levels to recharge - area 1: De Aar Karoo.

Area 2 - Bo Molopo

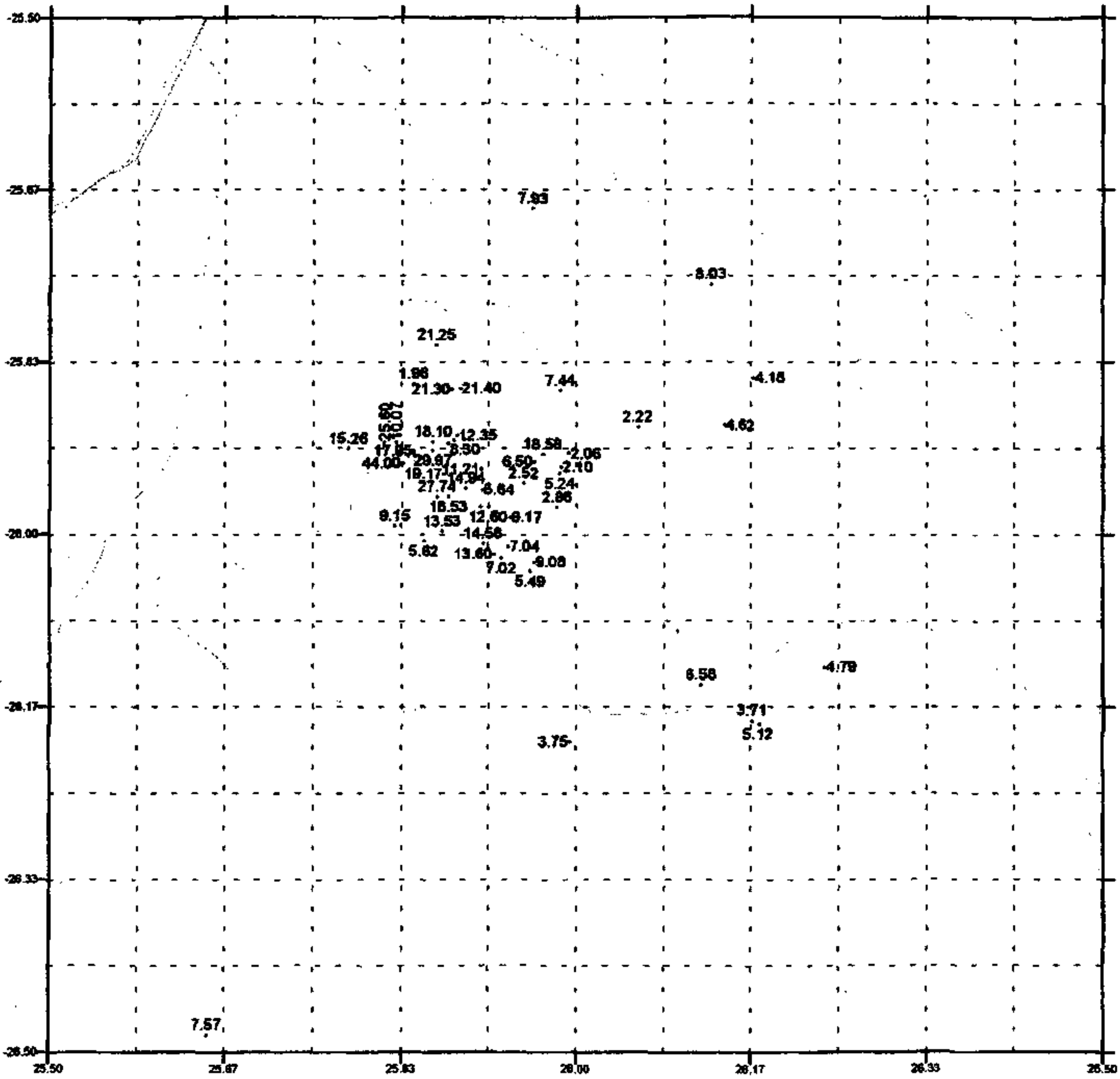


Fig. 6.3.2 : Unit response of groundwater levels to recharge - area 2: Bo Molopo area.

Area 3 - Pretoria

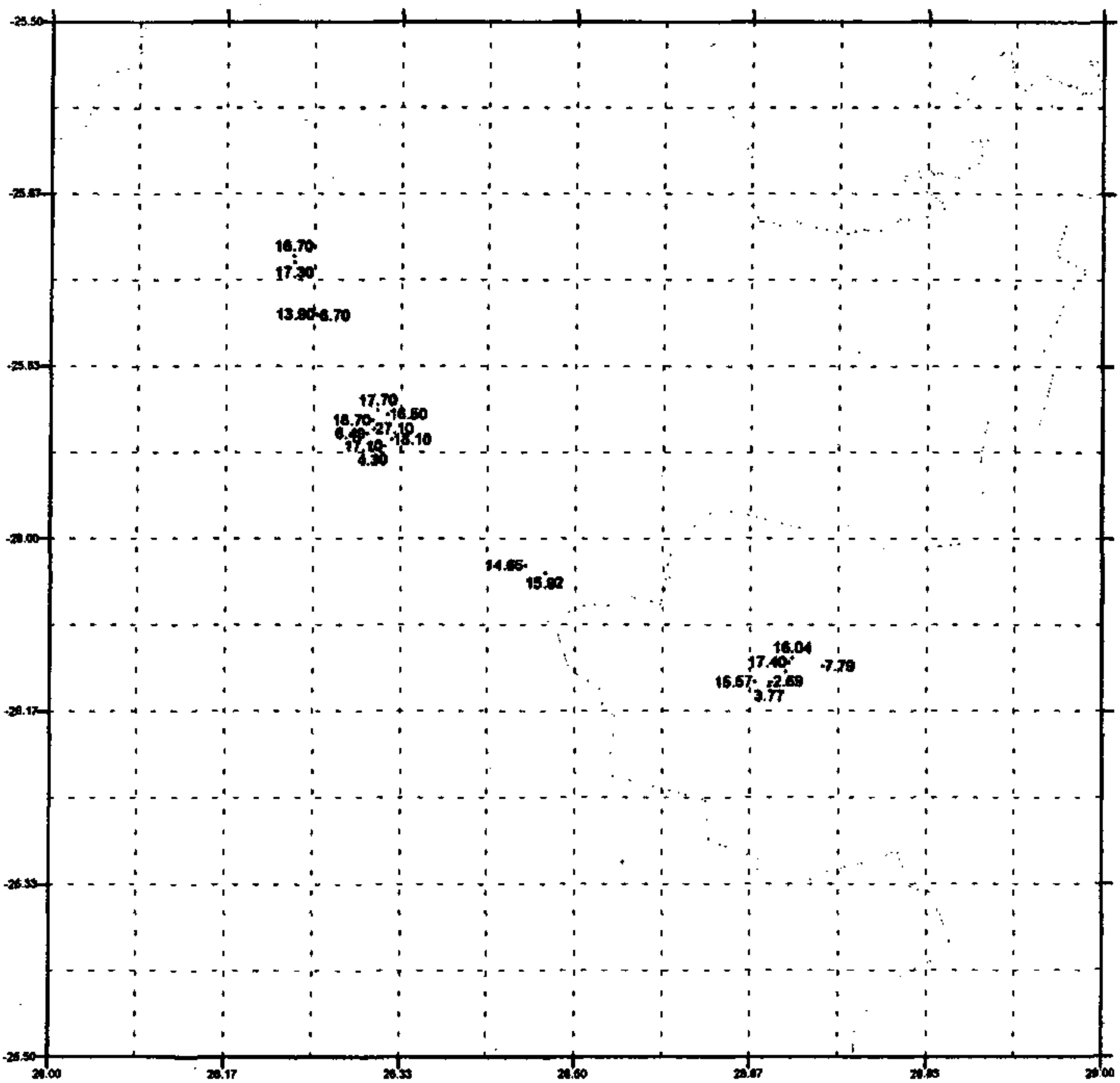


Fig. 6.3.3 : Unit response of groundwater levels to recharge - area 3: Pretoria area.

Area 4 - Pietersburg

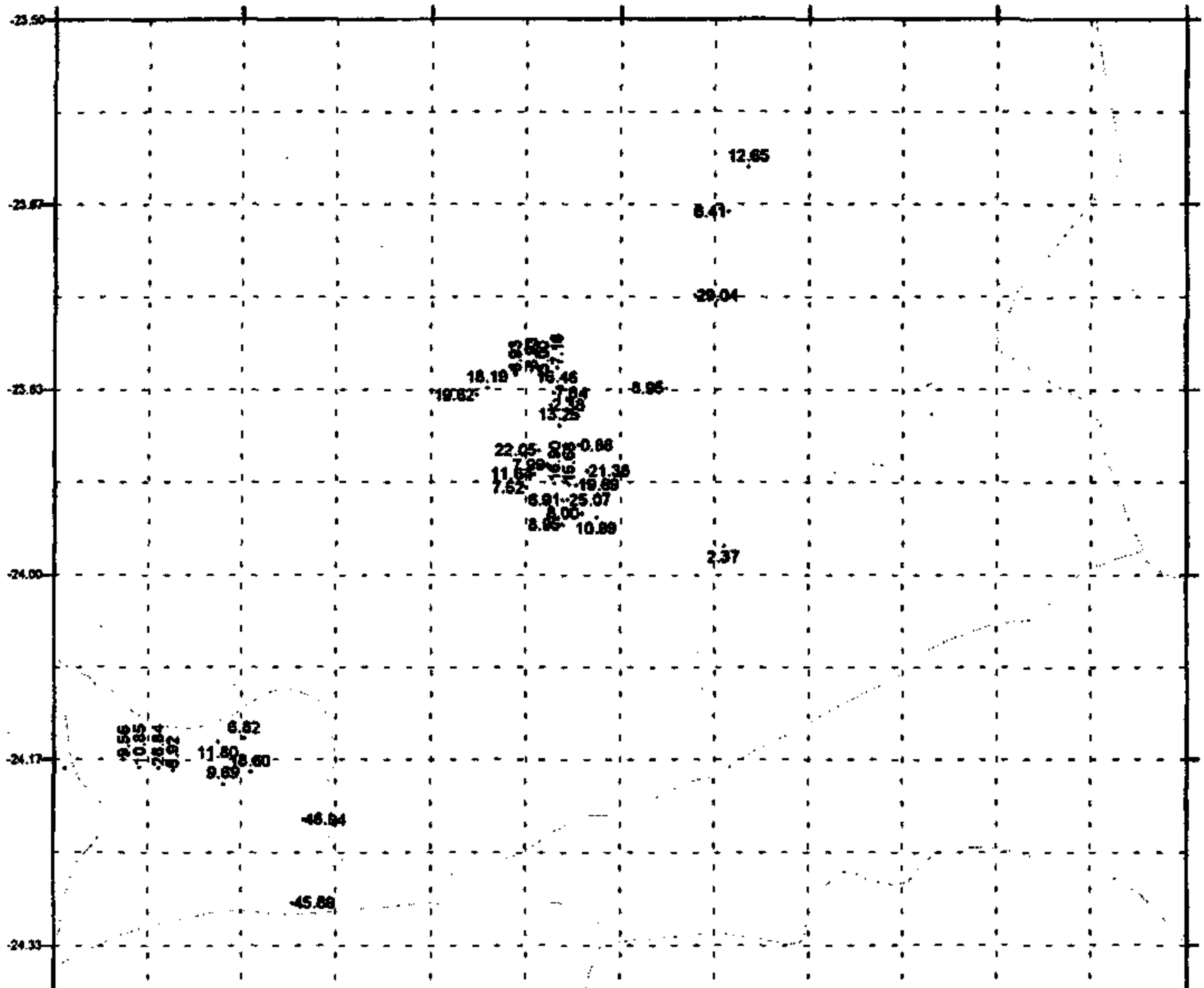


Fig. 6.3.4 : Unit response of groundwater levels to recharge - area 4: Pietersburg area.

Area 2 - Bo Molopo

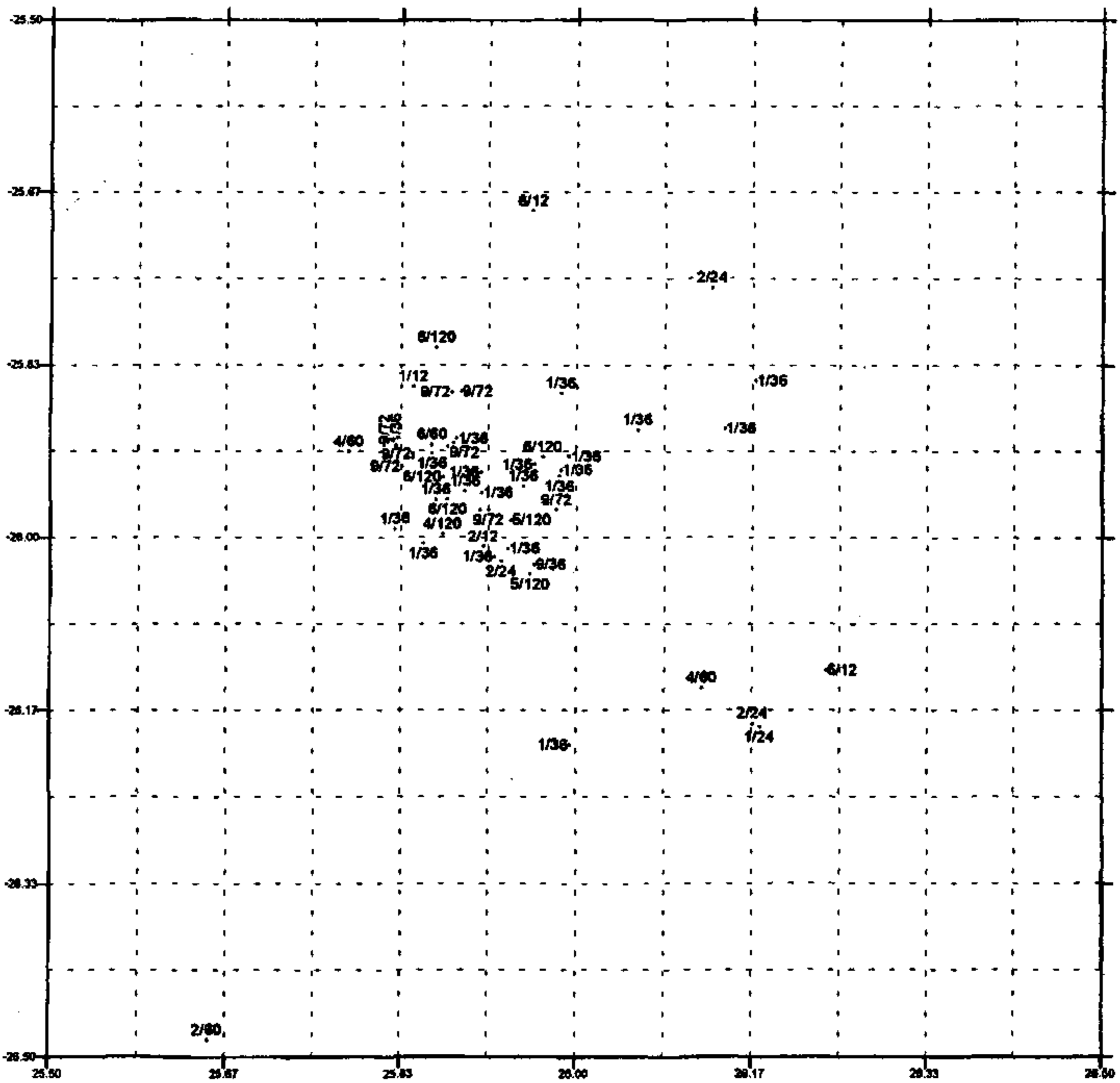


Fig. 6.4.1 : The m/n memories of the CRD analysis of the Bo-Molopo dolomitic aquifer.

Area 1 - De Aar

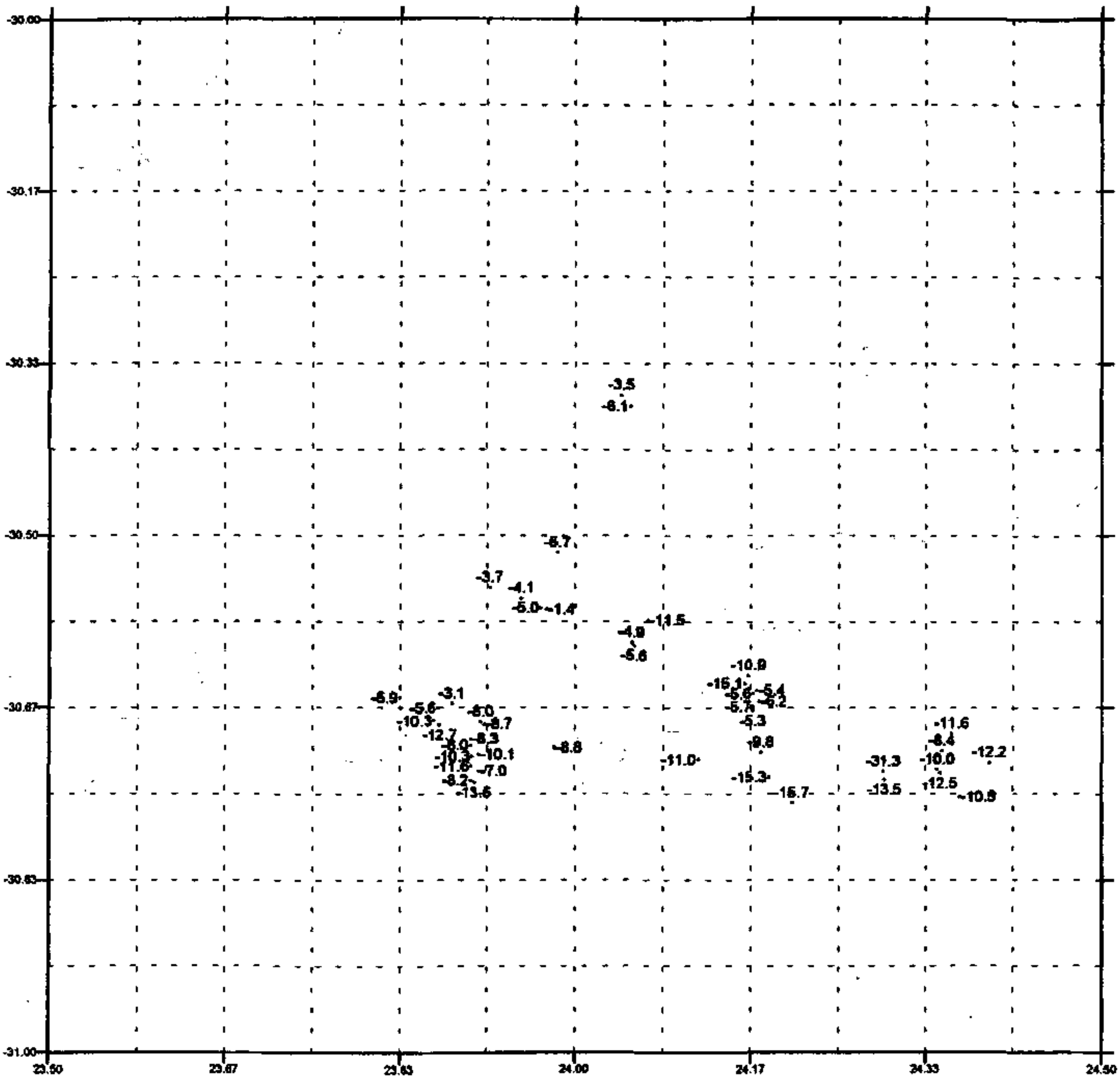


Fig. 6.5.1 : The inferred depth(m) of the aquifer derived from the MA regression of the Karoo aquifer in the De Aar region.

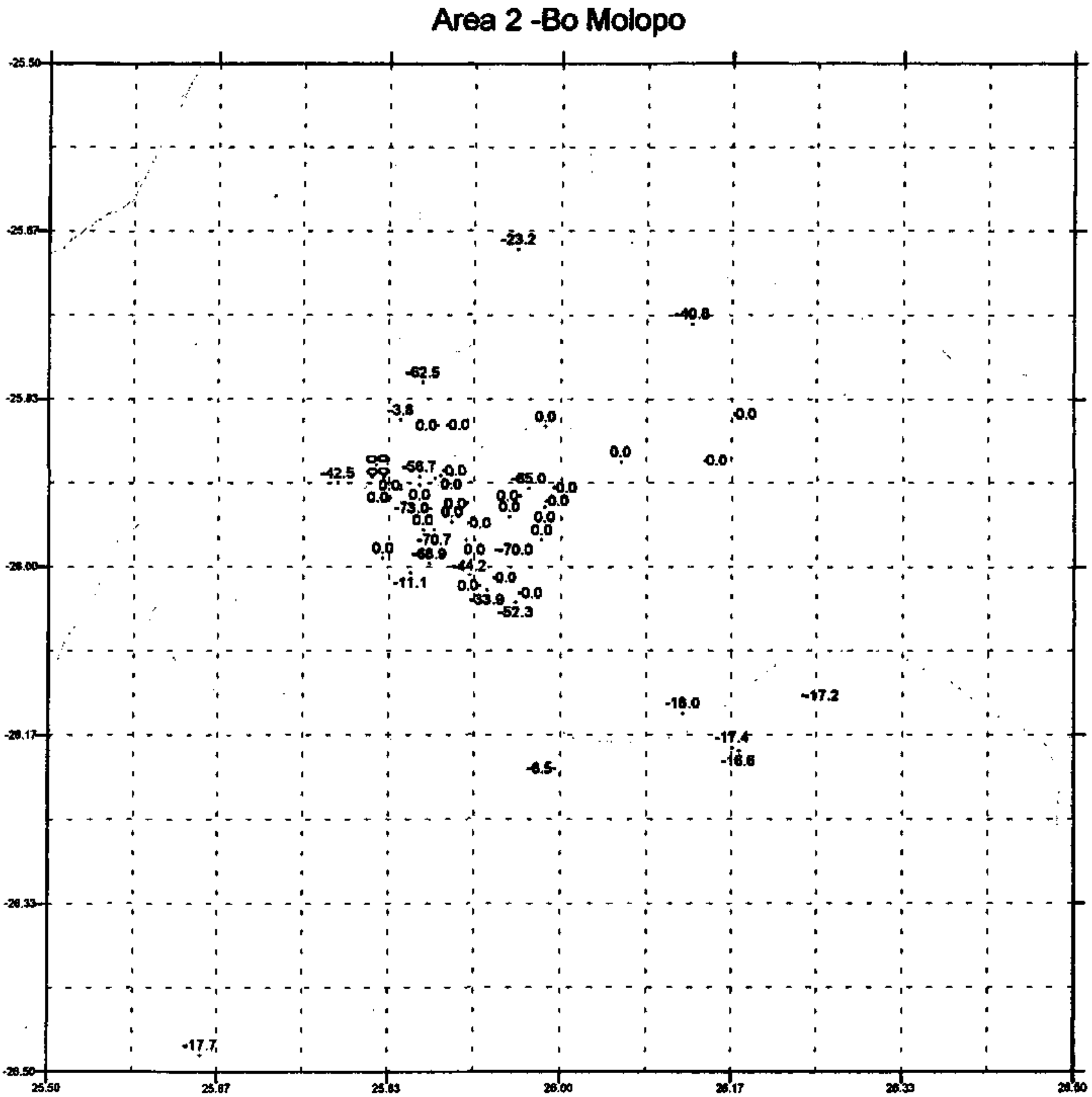


Fig. 6.5.2 : The inferred depth(m) of the aquifer derived from the MA regression of the Bo-Molopo dolomitic aquifer.

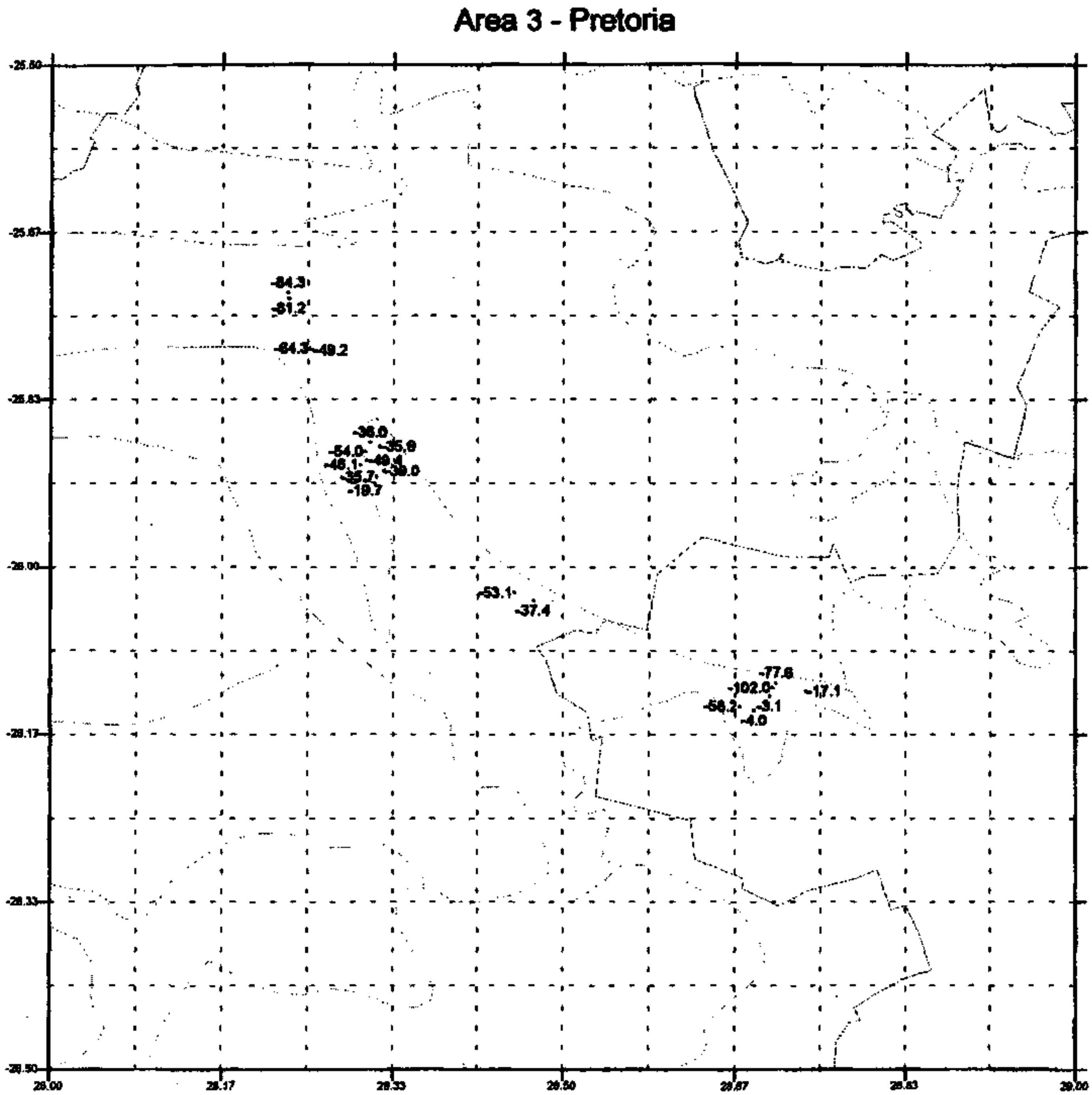


Fig. 6.5.3 : The inferred depth(m) of the aquifer derived from the MA regression of the Pretoria dolomitic aquifer.

Area 4 - Pietersburg

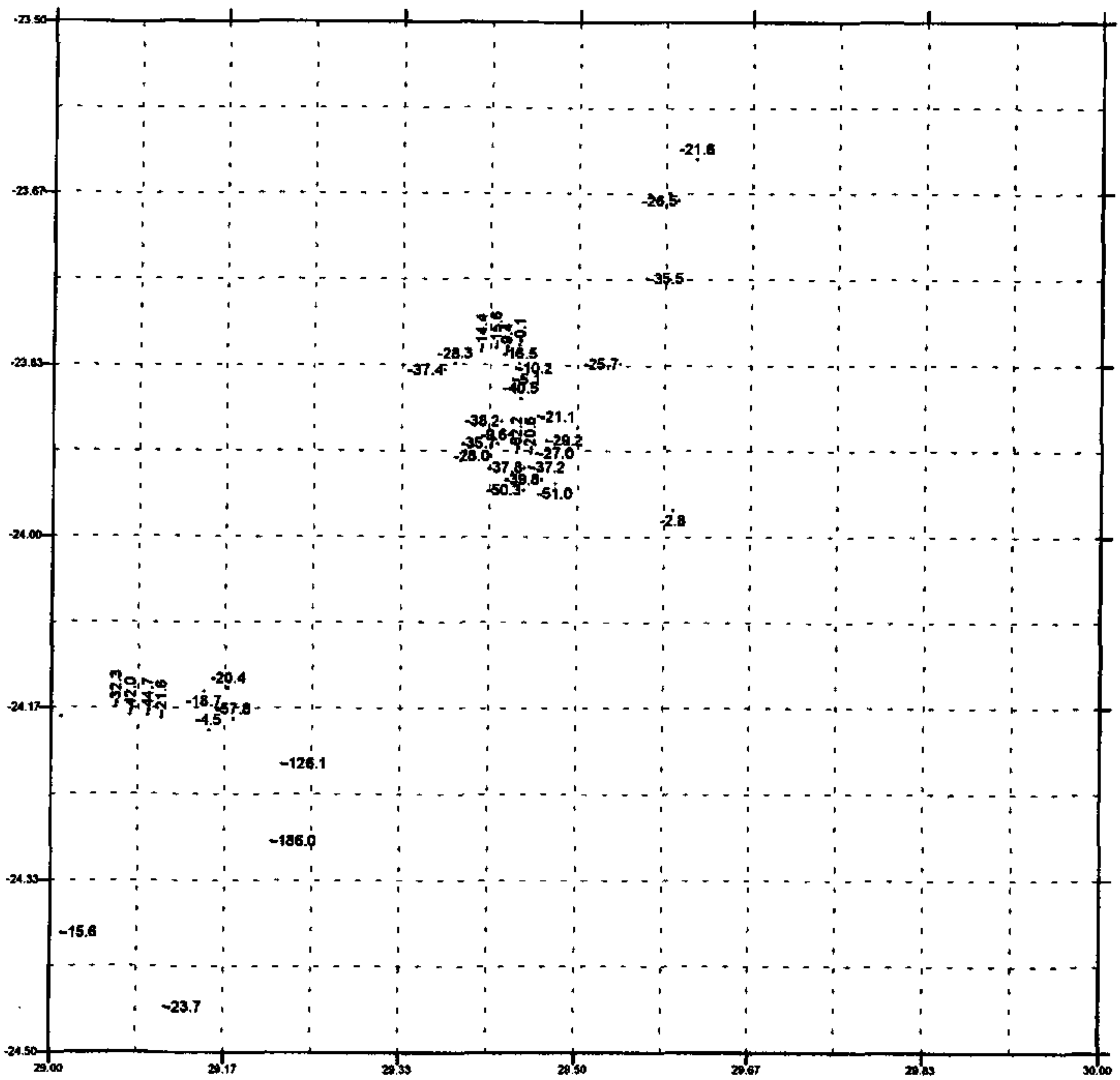


Fig. 6.5.4 : The inferred depth(m) of the aquifer derived from the MA regression of the Pietersburg granitic aquifer.

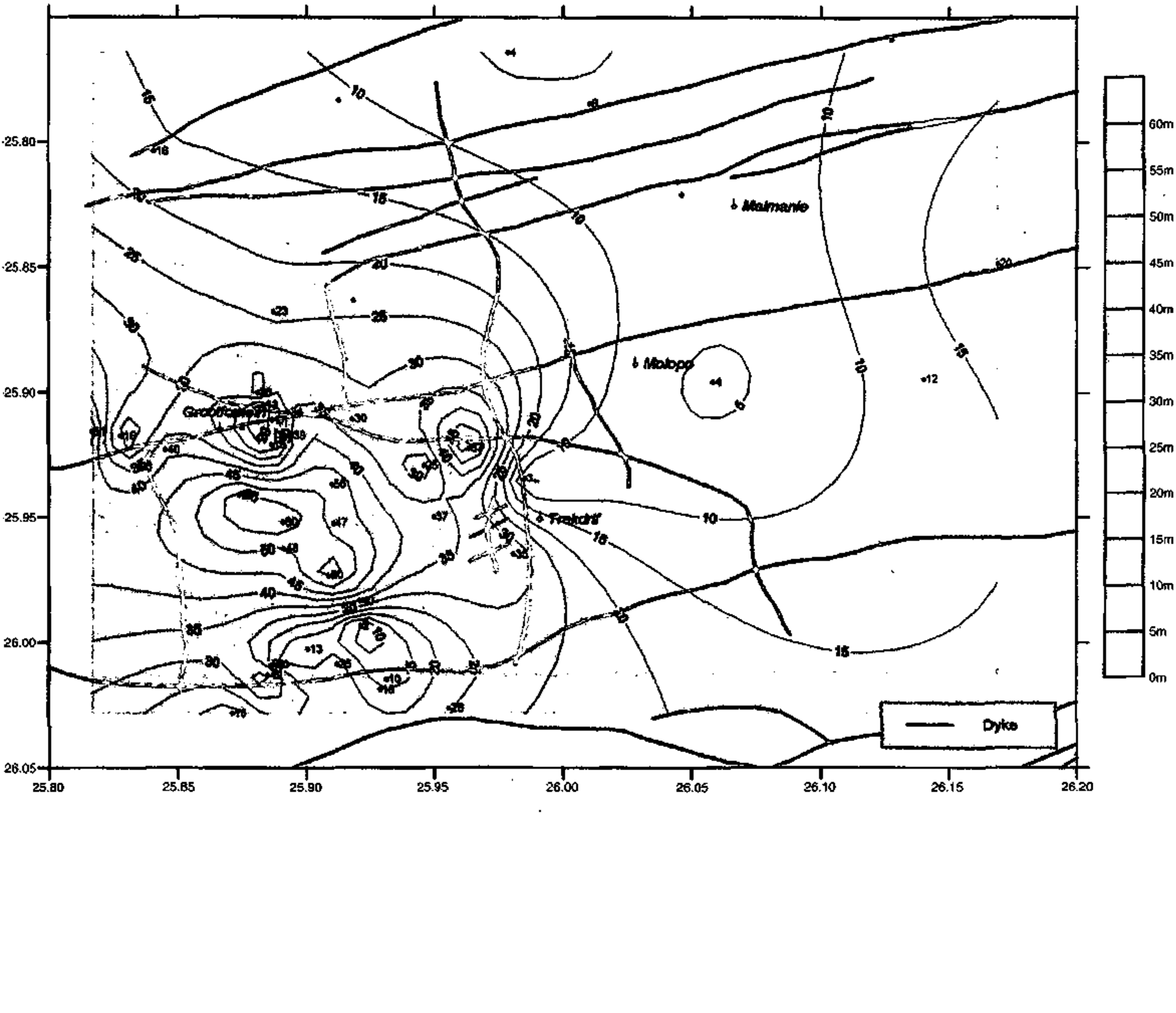
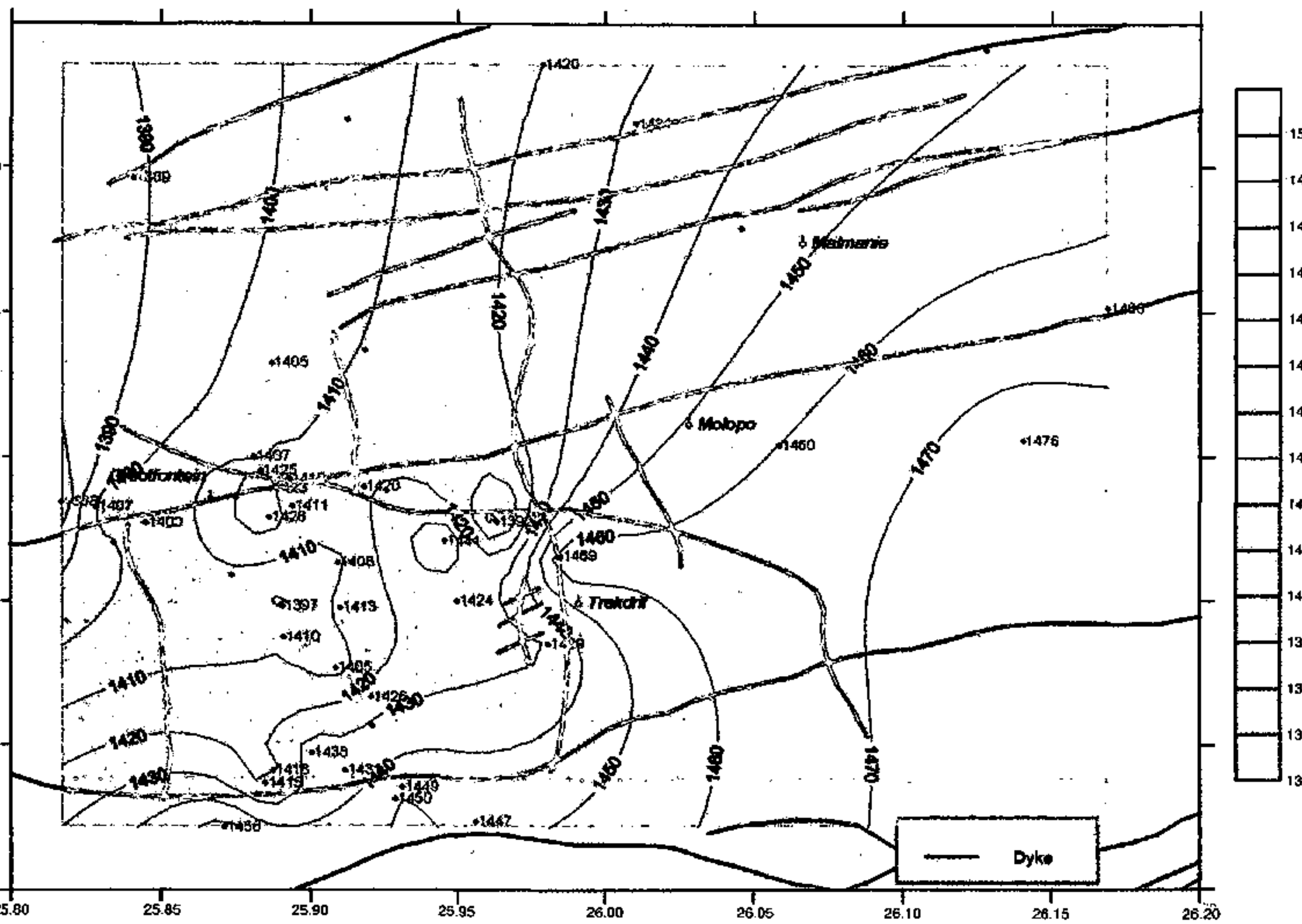


Fig. 6.5.5: Contour map of the thickness (m) of the saturated aquifer in the Bo-Molopo dolomitic aquifer.



Malmanie
Molopo
Trekdrif
Dyke
-25.80
-25.85
-25.90
-25.95
-26.00
-26.05
25.80
25.85
25.90
25.95
26.00
26.05
26.10
26.15
26.20
1500
1490
1480
1470
1460
1450
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1430
1420
1410
1400
1390
1380
1370
1360

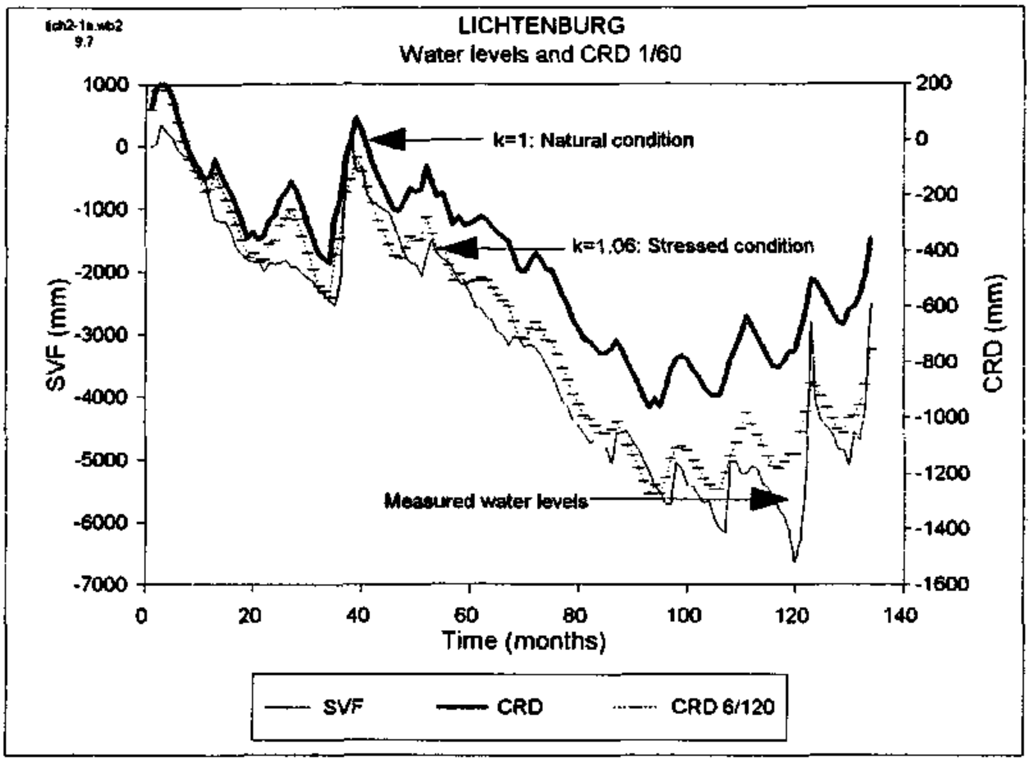


Fig. 7.2.1.1 Integrated water levels of the Lichtenburg aquifer in relation to abstraction.

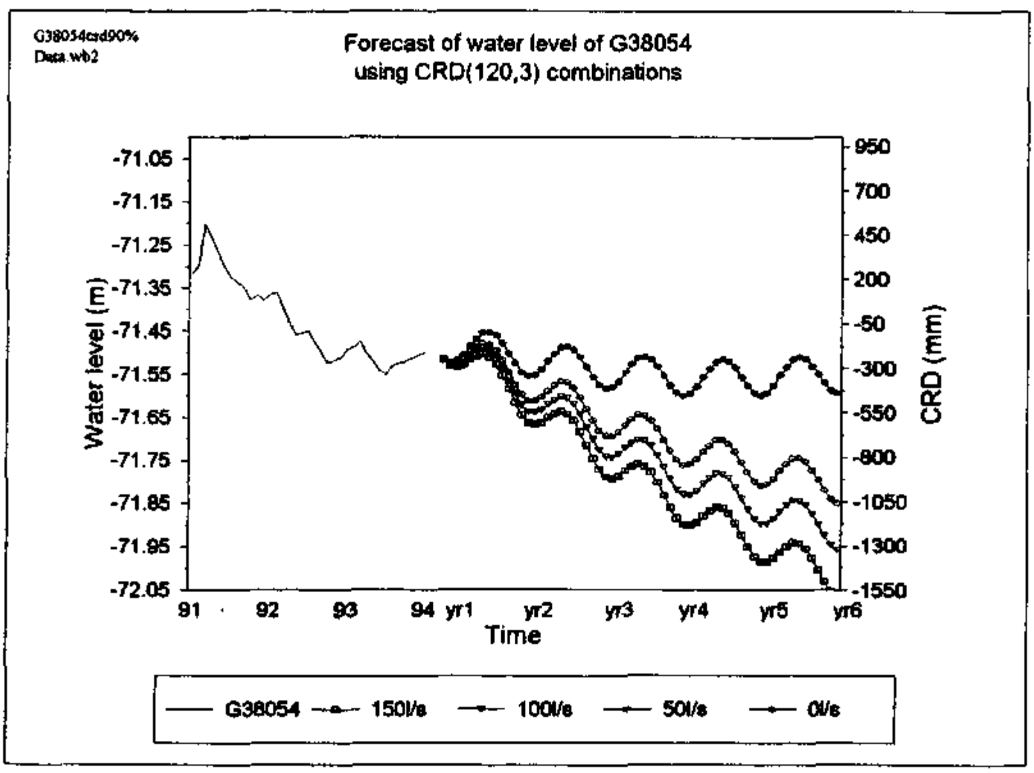


Fig. 7.2.1.2 Relationship between flow of Grootfontein spring, simulated by the CRD series for different rates of abstraction.

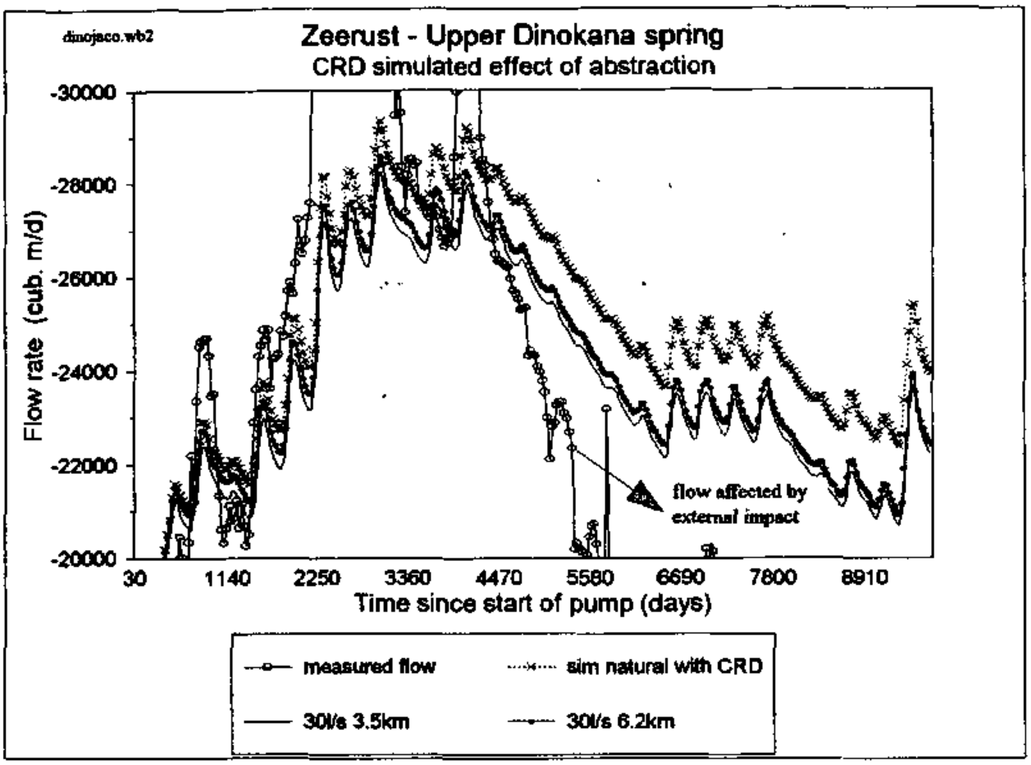


Fig. 7.2.1.3 Simulated flow of the Upper Dinokana spring for abstraction at different distances from the eye.

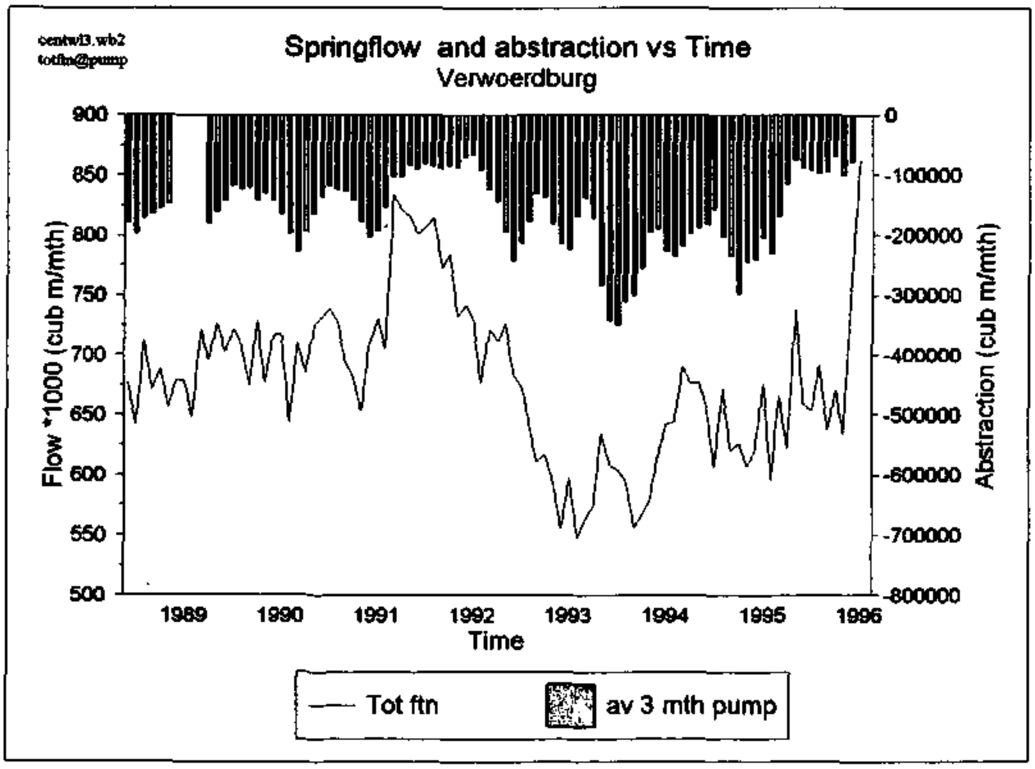


Fig. 7.2.1.4 Flow of the Pretoria Fountains relative to abstraction from the compartment.

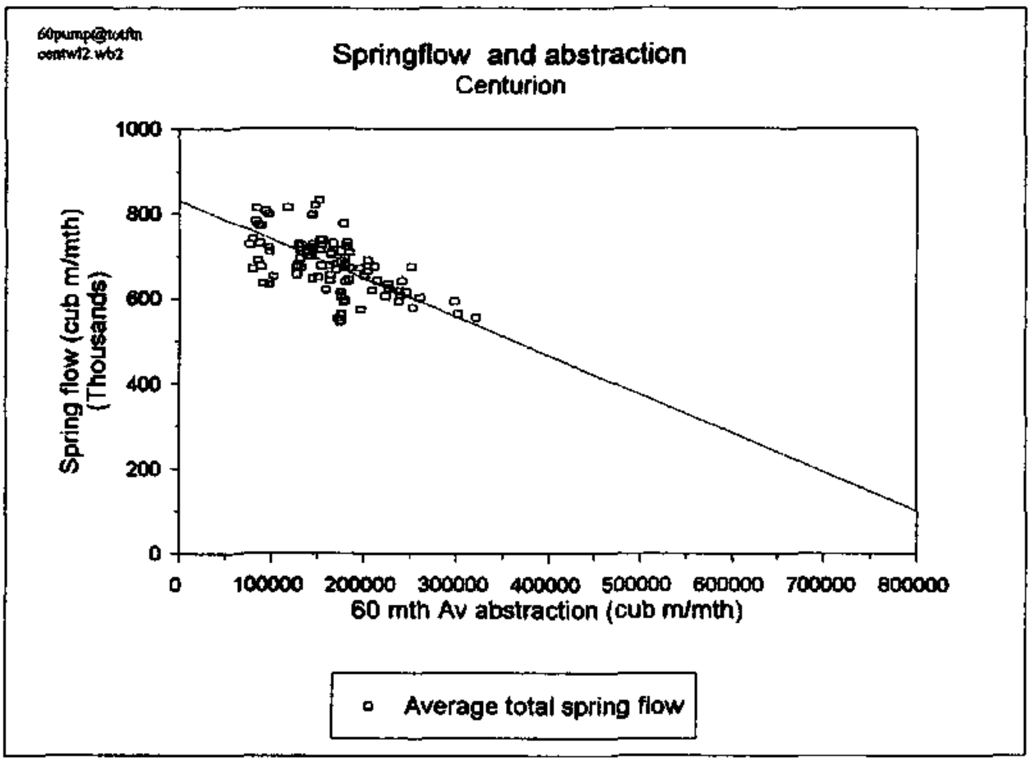


Fig. 7.2.1.5 X-Y plot of the discharge of the Pretoria Fountains in relation to the rates of abstraction in the recharge area.

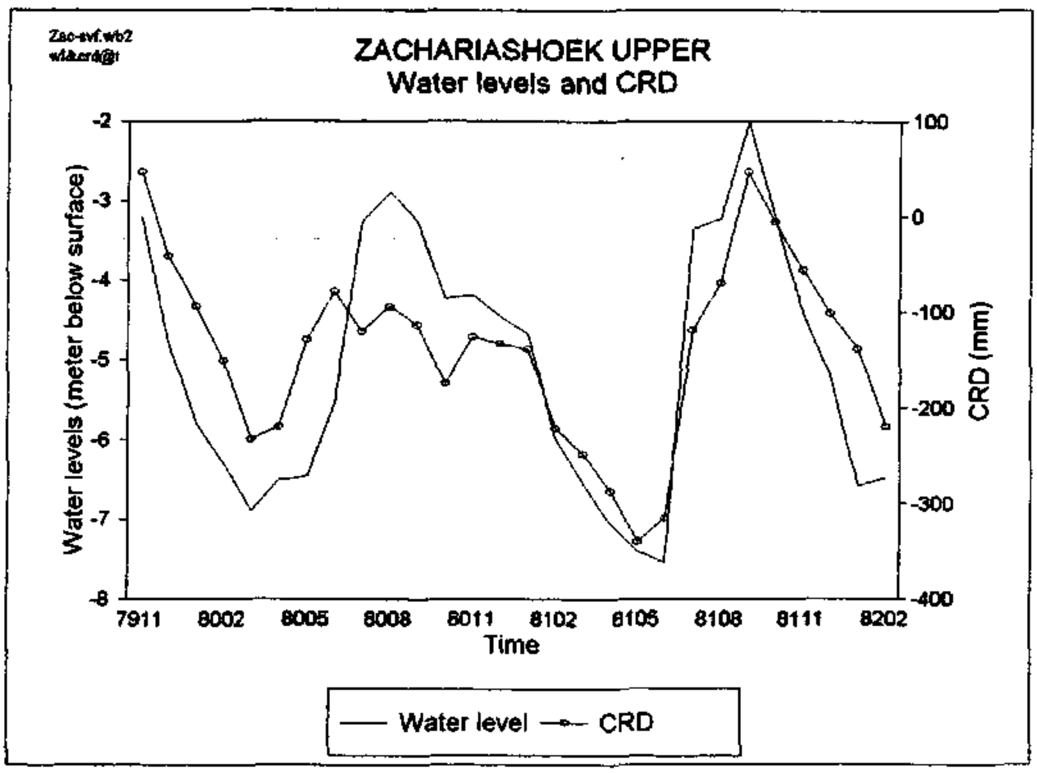


Fig. 7.2.2.1 Correspondence between the groundwater levels in the Jonkershoek mountains compared to the $^{1}_{av}CRD$ series.

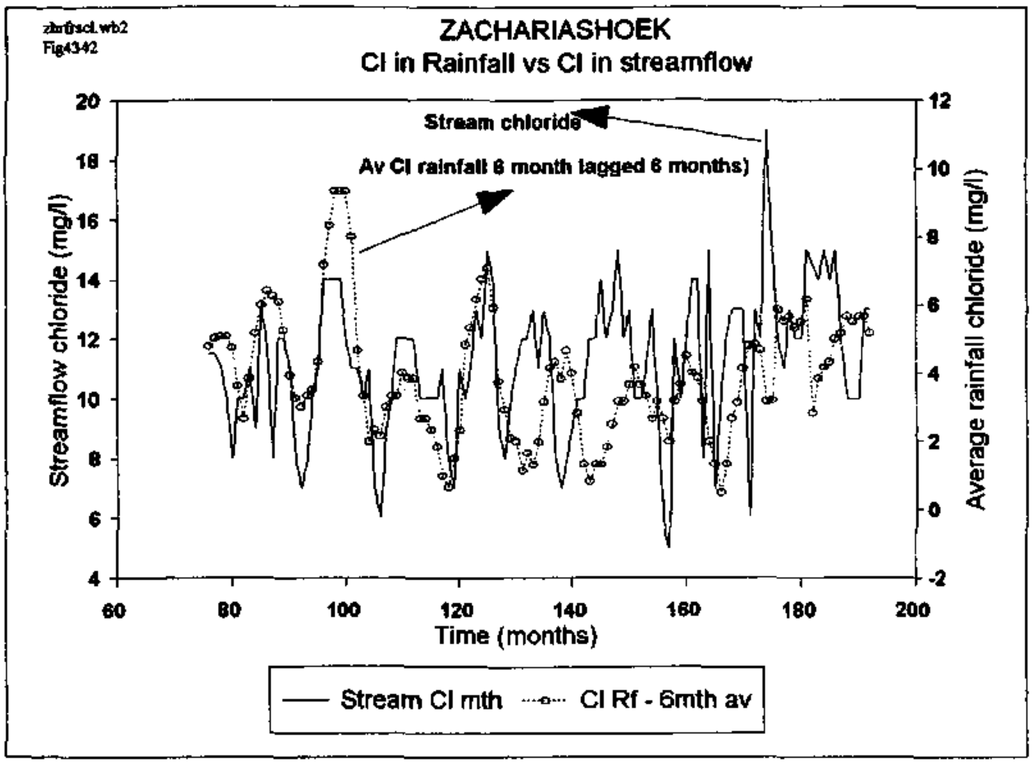


Fig. 7.2.2.2 Lag between rainfall and its appearance of stream flow, derived from the variation of chloride in the rainfall in comparison to that of the runoff.

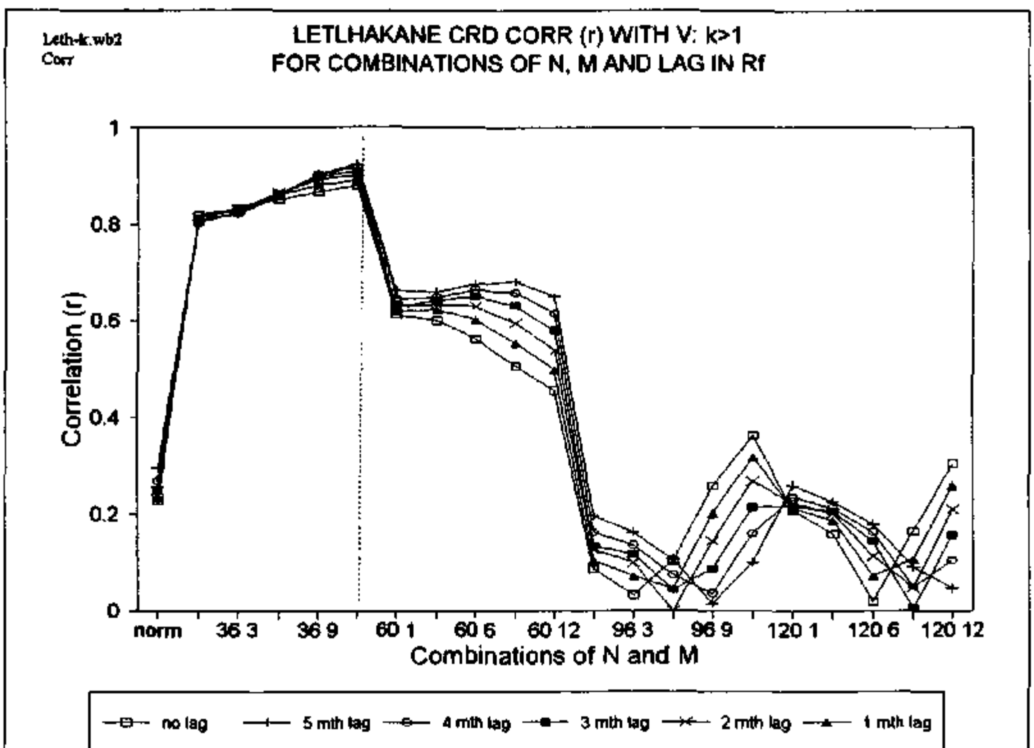


Fig. 7.2.2.3 Correlations between the groundwater levels and the CRD series for different long (n) and short memories (m) for the Letlhakane-Orapa area.

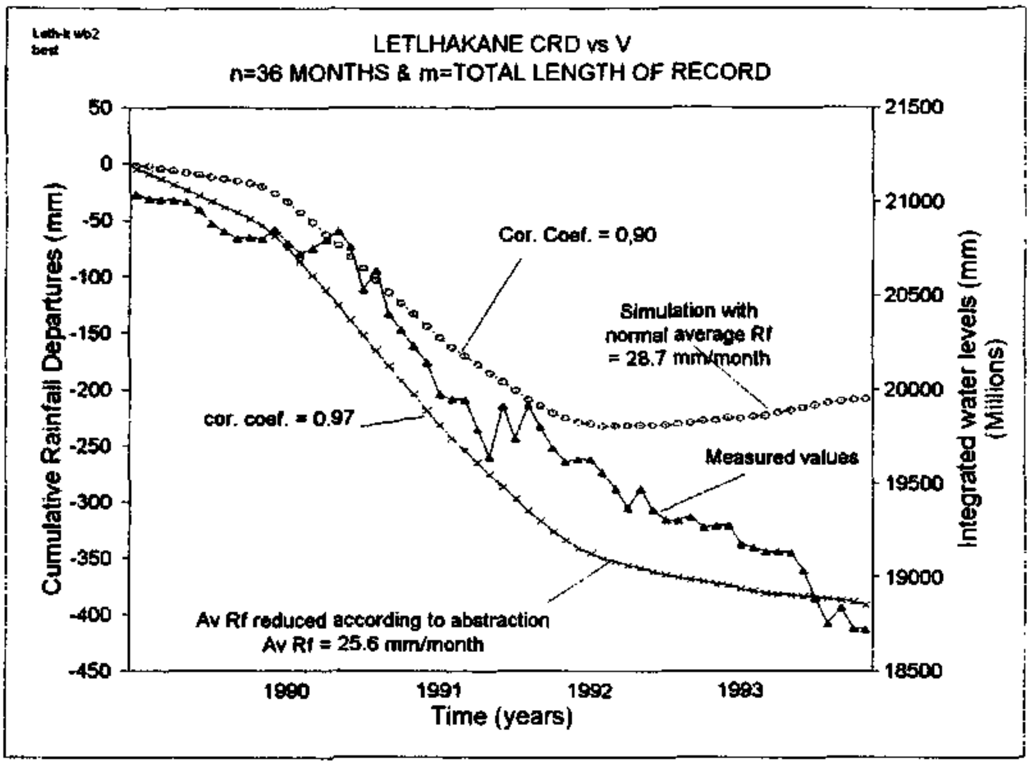


Fig. 7.2.2.4 Correspondence between the $_{av}^{1}CRD$ series and groundwater levels, assuming that recharge is determined by the average rainfall and that losses are controlled by rainfall during the short memory.

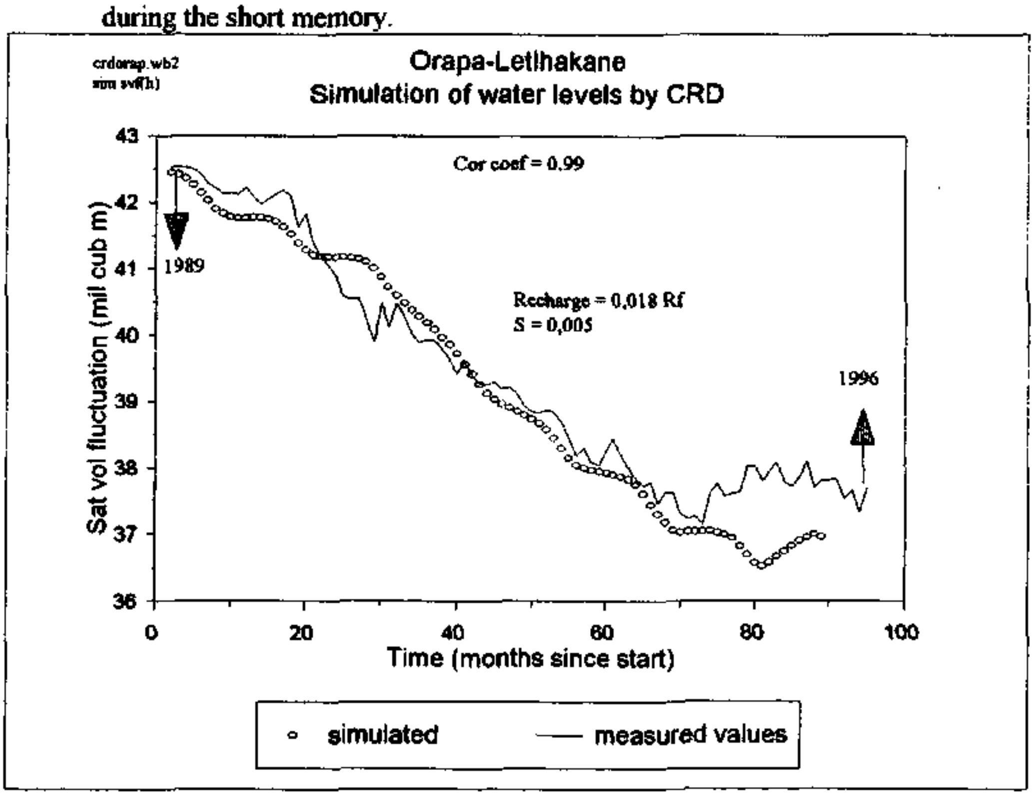


Fig. 7.2.2.5 Correspondence between measured values and those simulated by the $_{av}^{1}CRD$ series lagged by 14 months. Subtracting the monthly pumping from the simulated series produced an exceptionally good match with the observed water levels for conditions of exploitation.

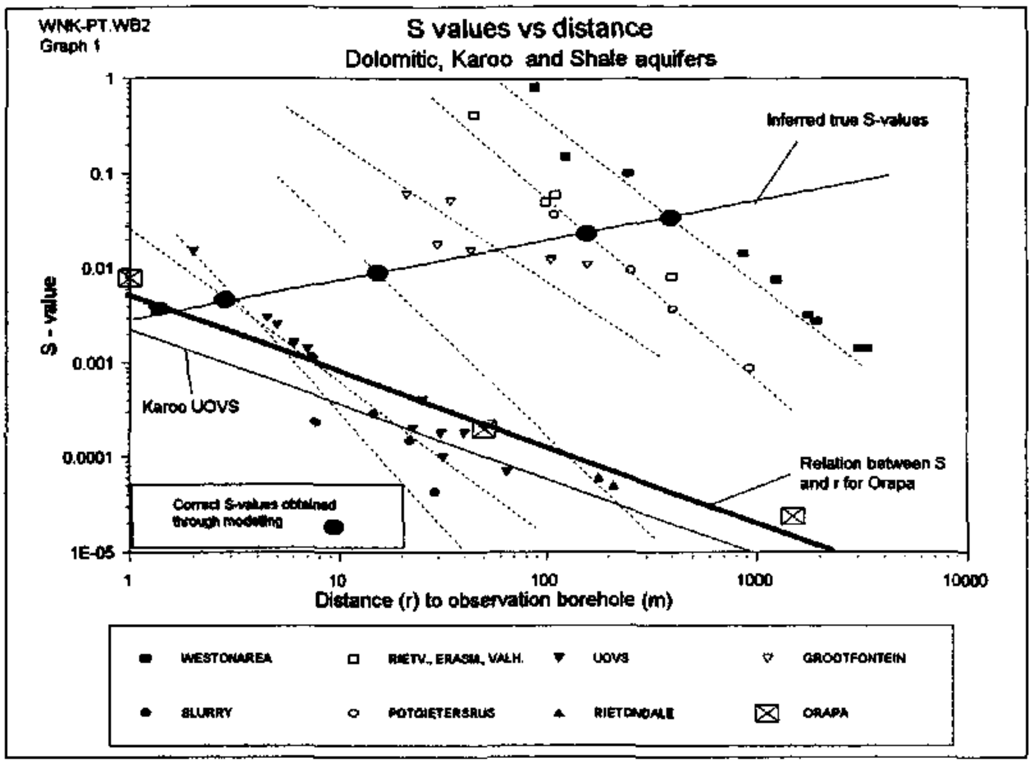


Fig. 7.2.2.6 Calibration curve of the real S-values in relation to the dependence of aquifer storativity on the distance (r) of observation boreholes from the pumping borehole.

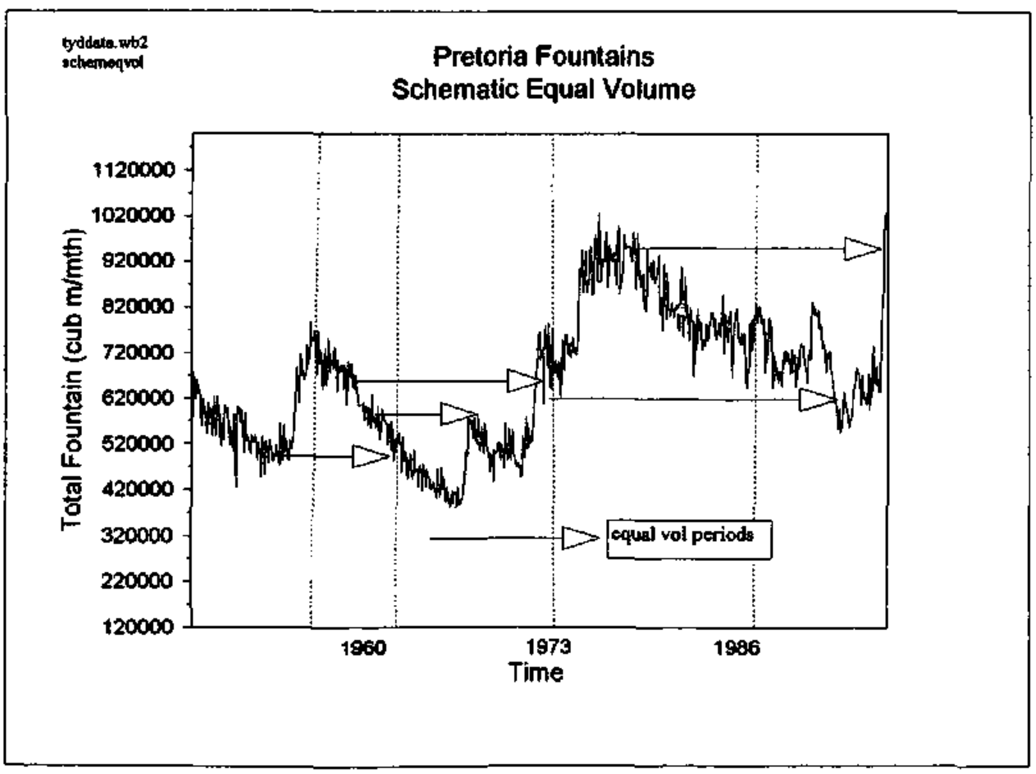


Fig. 8.3.1.1 Schematic diagram of the equal volume method (EV method) applied to spring flows.

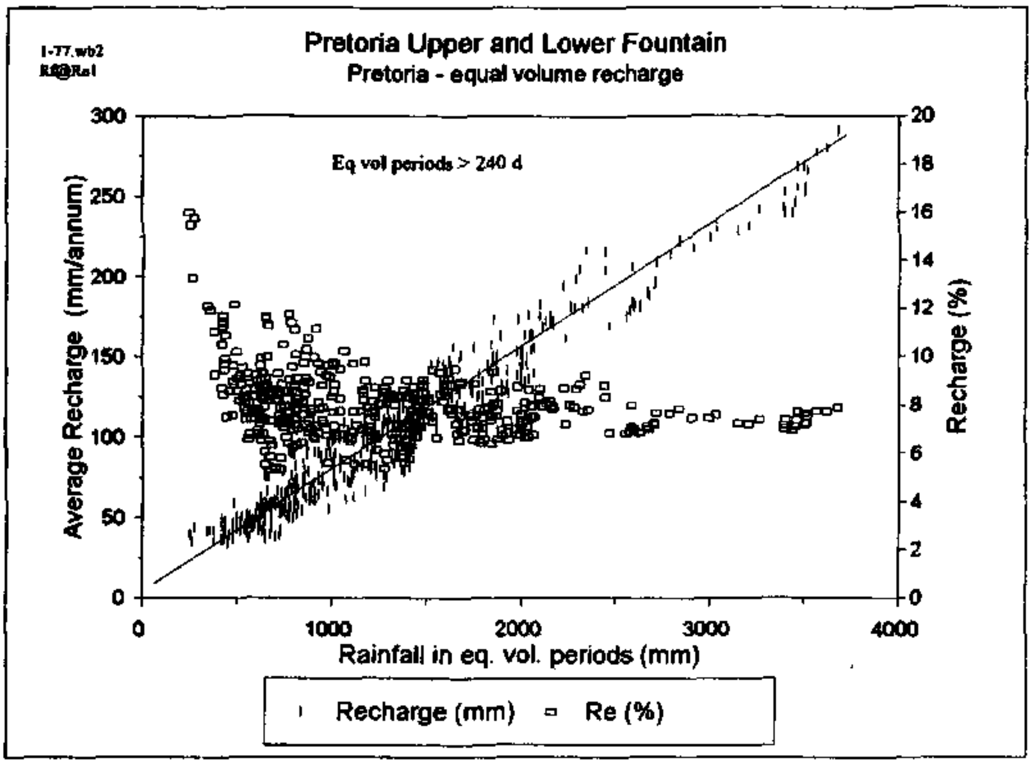


Fig. 8.3.1.2 Recharge, in percent and millimetres, derived from equal volume periods applicable to the Pretoria Fountains.

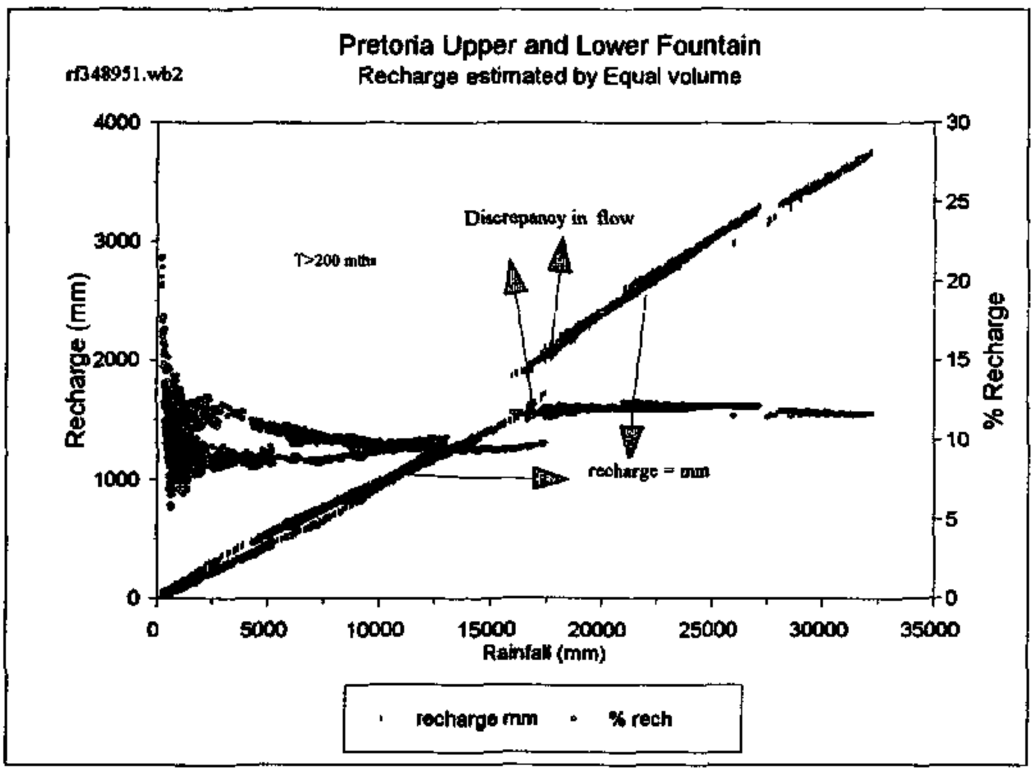


Fig. 8.3.1.3 Recharge relationship for the Pretoria Springs based on the EV method, clearly indicating a discrepancy which is attributed to erroneous flow data.

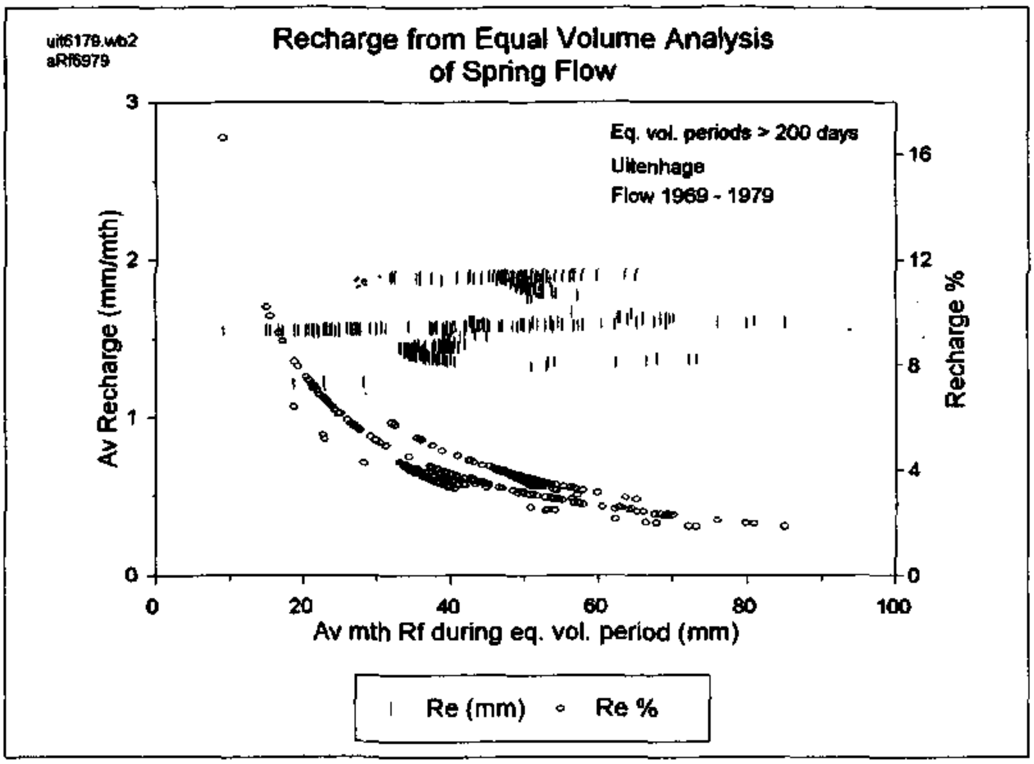


Fig. 8.3.1.4 Flow of the Uitenhage springs expressed as a percentage of the average rainfall - applying the EV method.

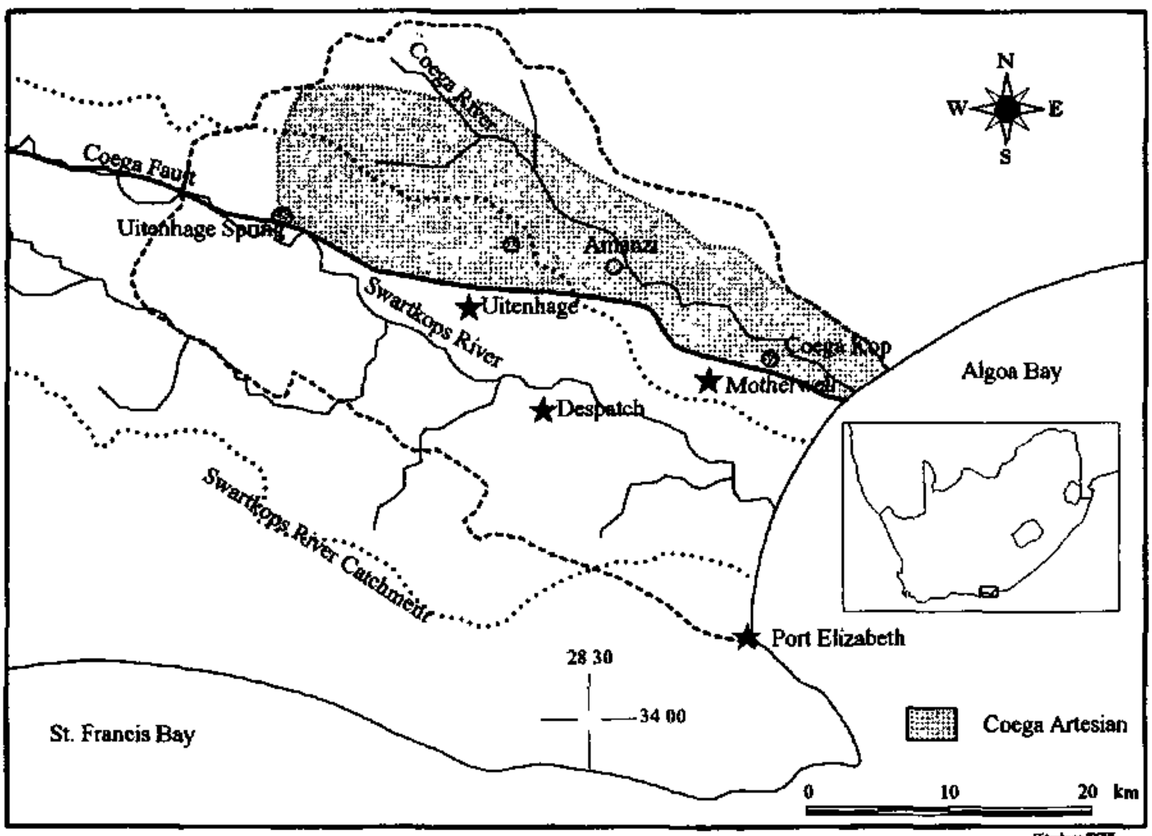


Fig. 8.3.2.2.1 The Uitenhage Groundwater Control Area showing the location of the springs and major artesian boreholes in the Coega Ridge Aquifer.

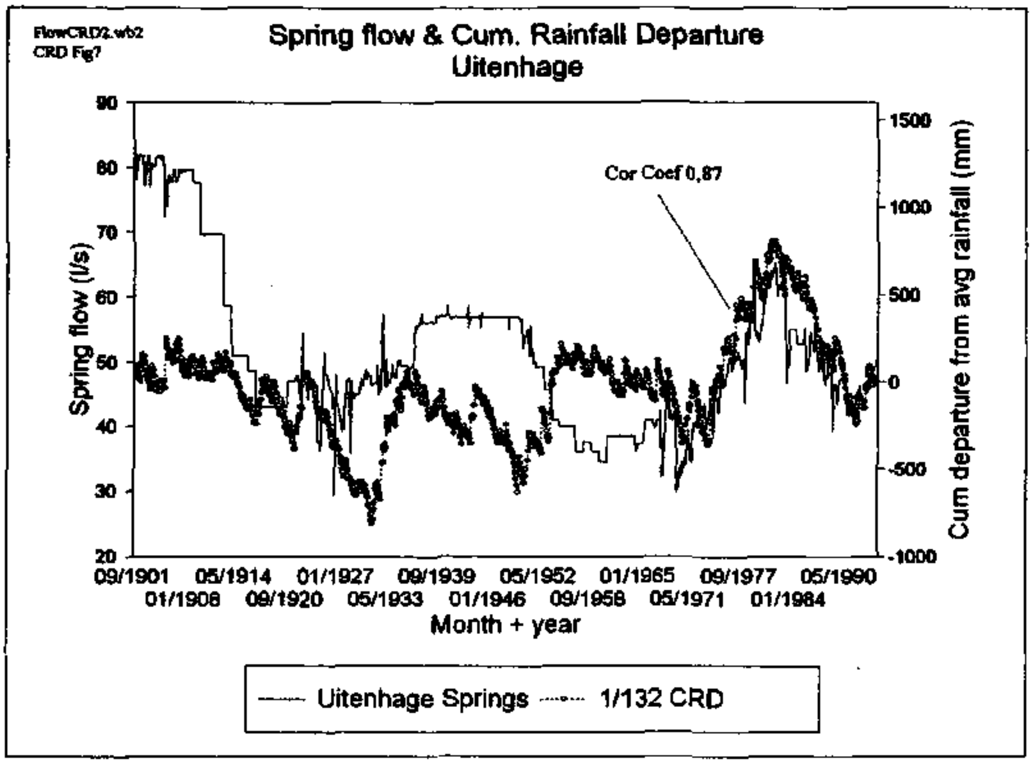


Fig. 8.3.2.2.2 Correlation between the flow of the Uitenhage Springs and the $_{132}^{1}CRD$ series.

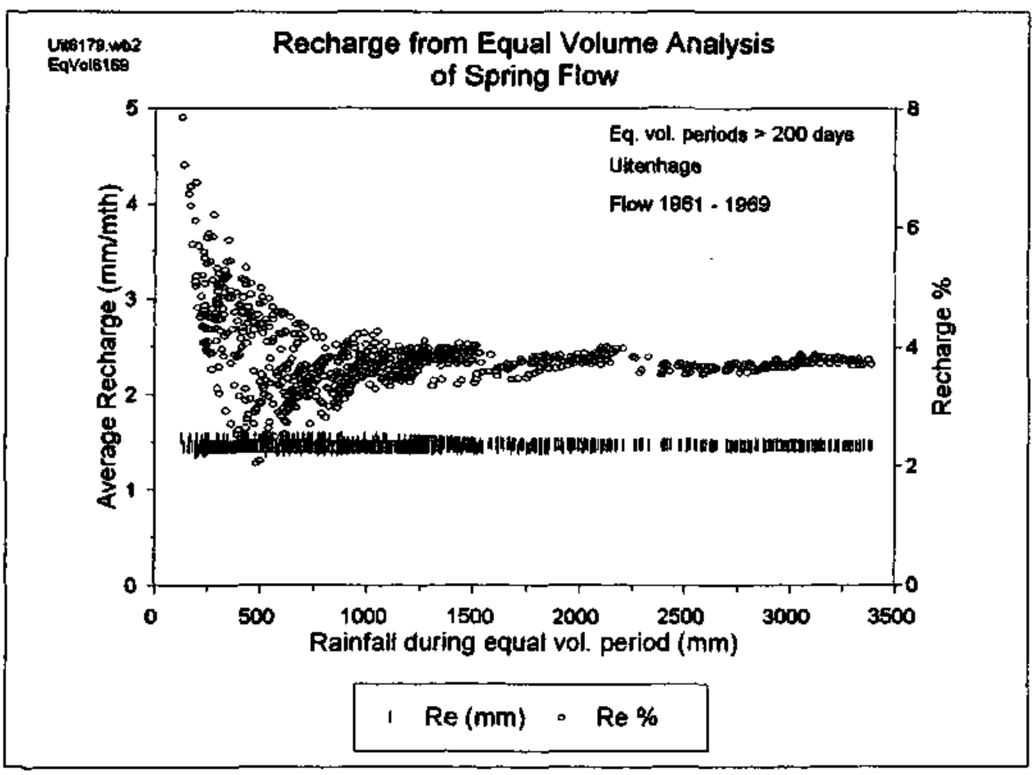


Fig. 8.3.2.2.3 Recharge as percentages and millimetres, derived by means of the EV method, for the Uitenhage Springs.

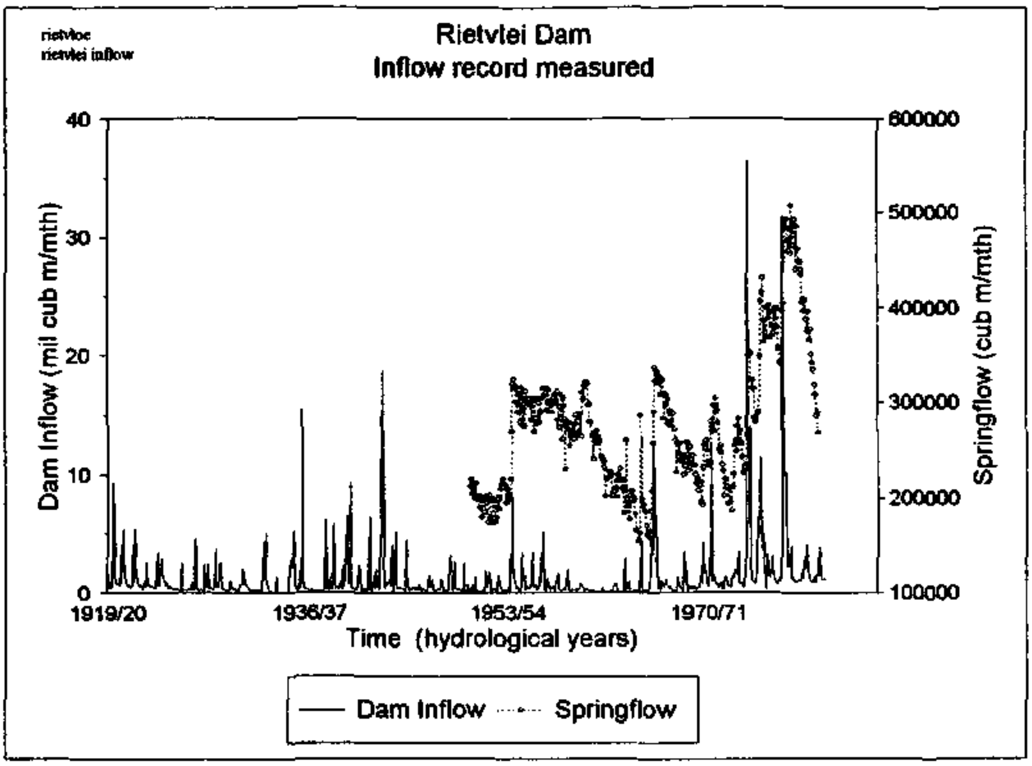


Fig. 9.2.2.1 Monthly inflow into the Rietvlei Dam in relation to the flow of the Rietvlei springs.

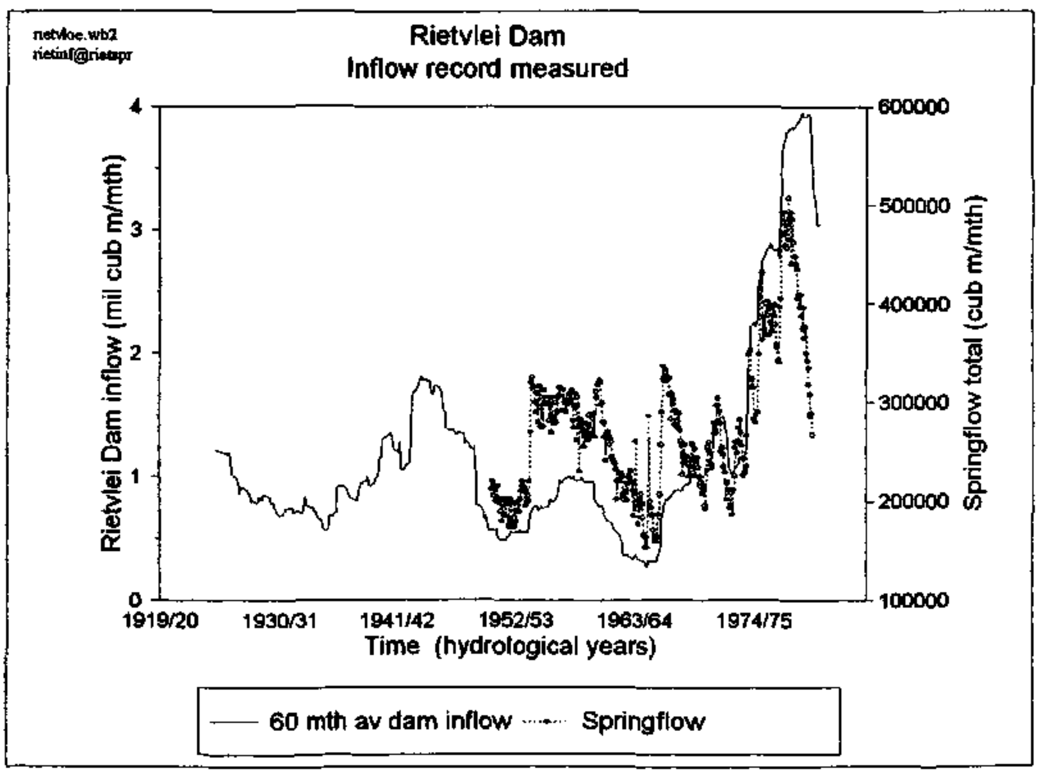
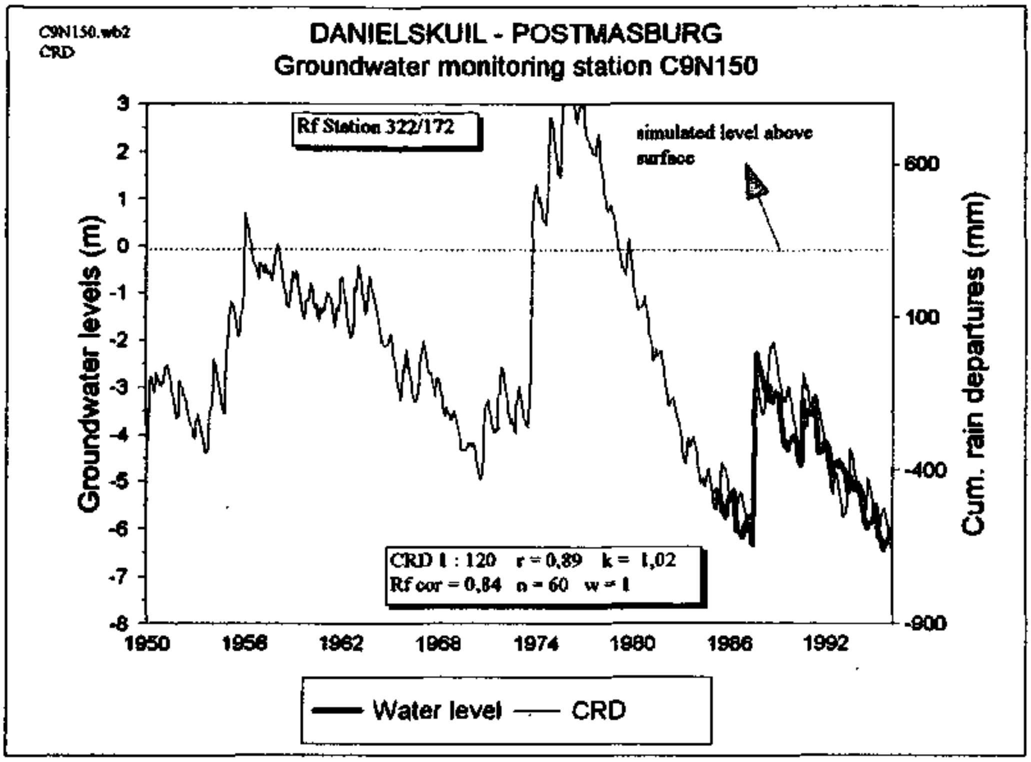


Fig. 9.2.2.2 Average inflow over 60 months into the Rietvlei dam in relation to the flow of the Rietvlei springs.



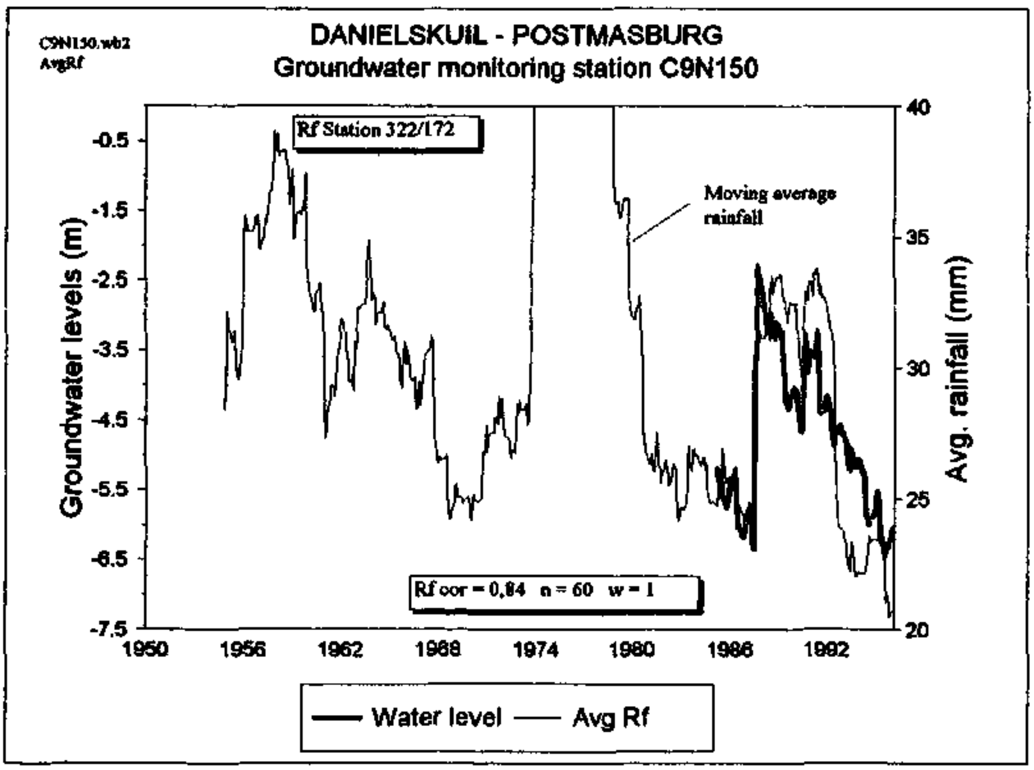


Fig. 9.3.1 Simulated groundwater levels at Danielskuil indicating that the levels would have risen higher than the ground surface.

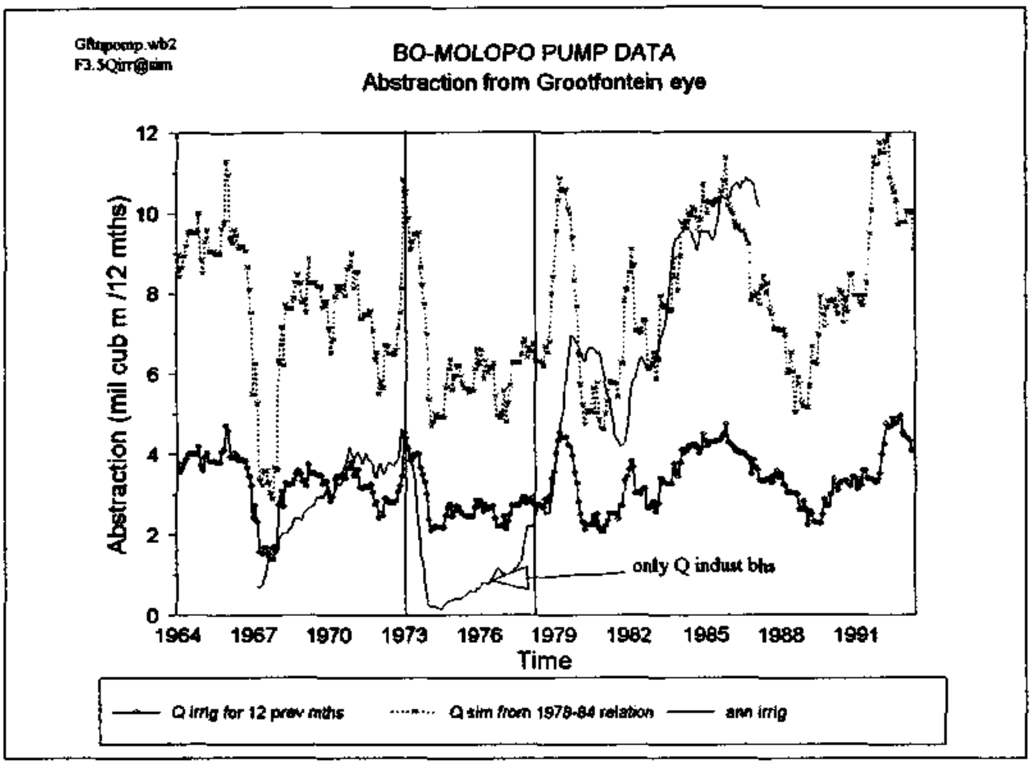


Fig. 10.2.1 Irrigation abstraction in the Grootfontein compartment based on rainfall over preceding 12 months. The higher abstraction after 1978 is due to larger areas being irrigated.

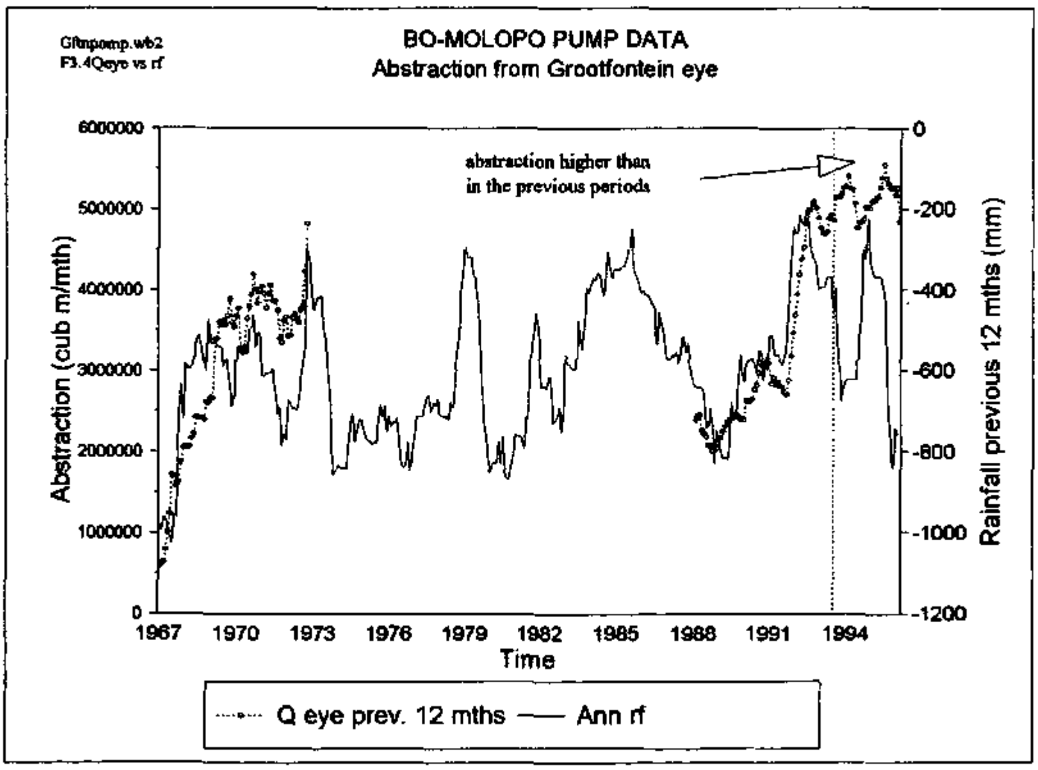


Fig. 10.3.1 Abstraction from the Grootfontein eye for two periods in relation to rainfall.

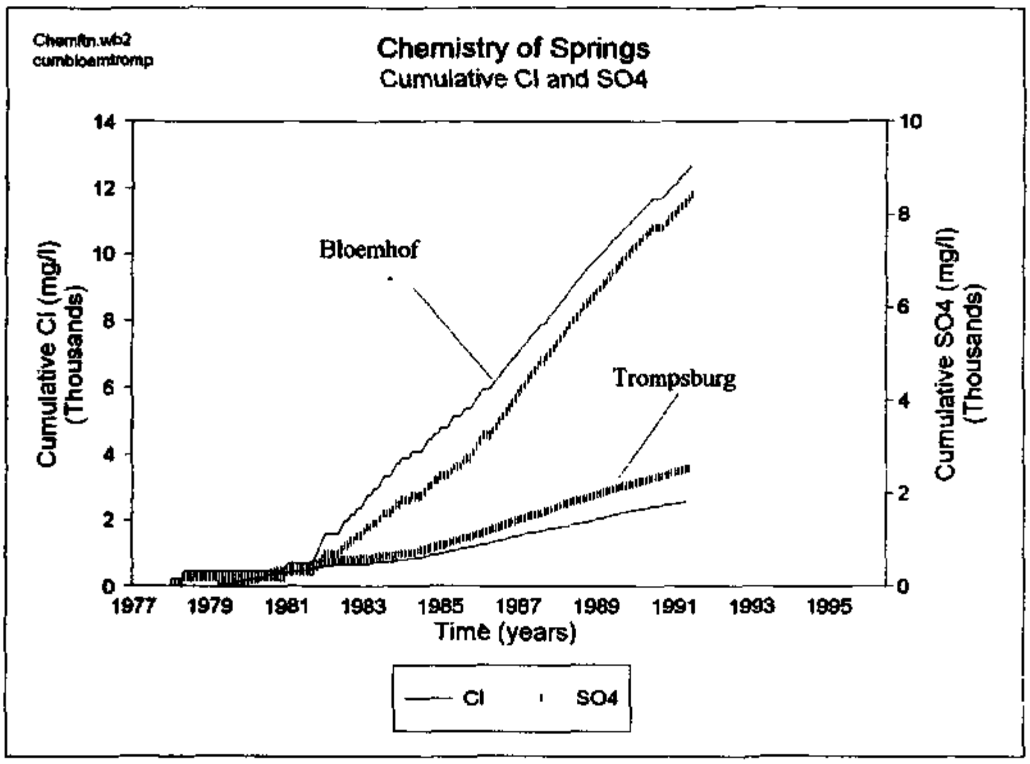


Fig. 11.3.3.1.1 Comparison of the chloride and sulphate concentrations of Bloemhof and Trompsburg springs.

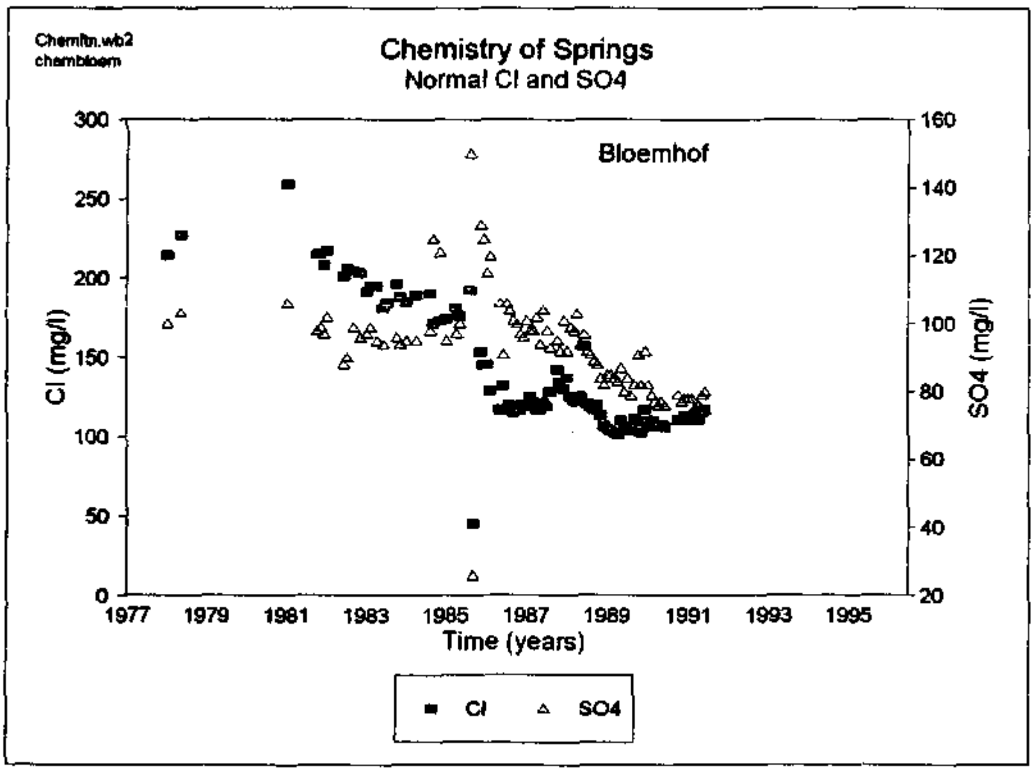


Fig. 11.3.3.1.2 Chloride and sulphate concentrations of the Bloemhof eye indicating an improvement in water quality.

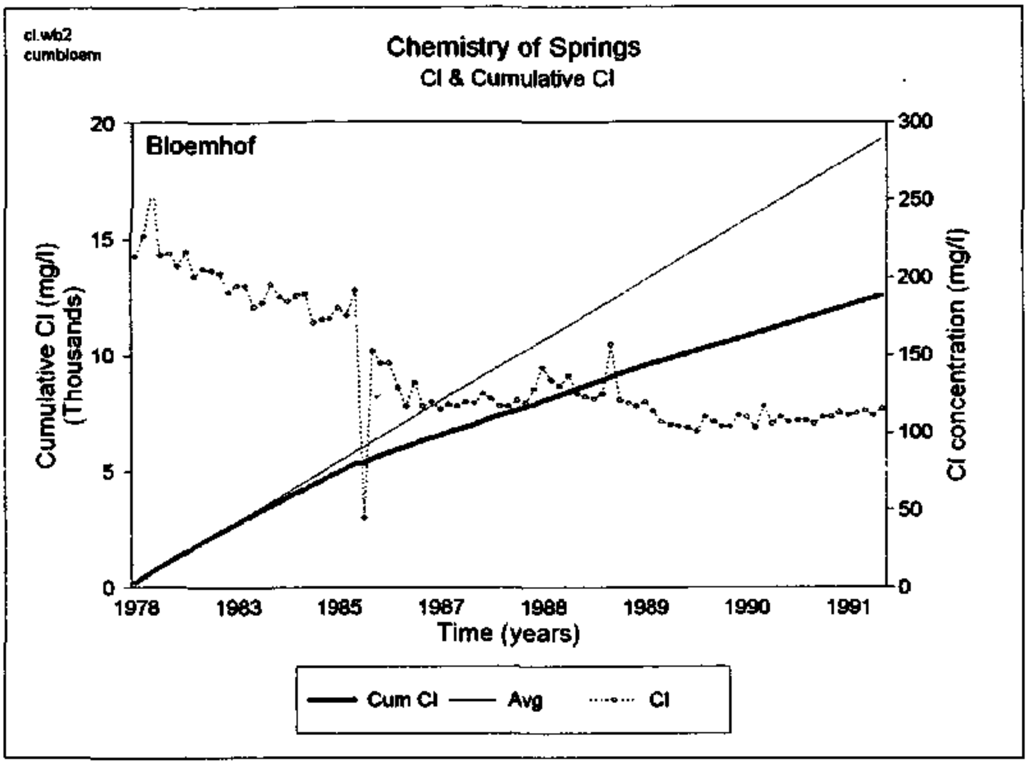


Fig. 11.3.3.1.3 Plot of chloride concentrations of the Bloemhof eye against time, and the cumulative plot indicating that the initial concentrations cannot be used to determine recharge by the chloride-ratio method.

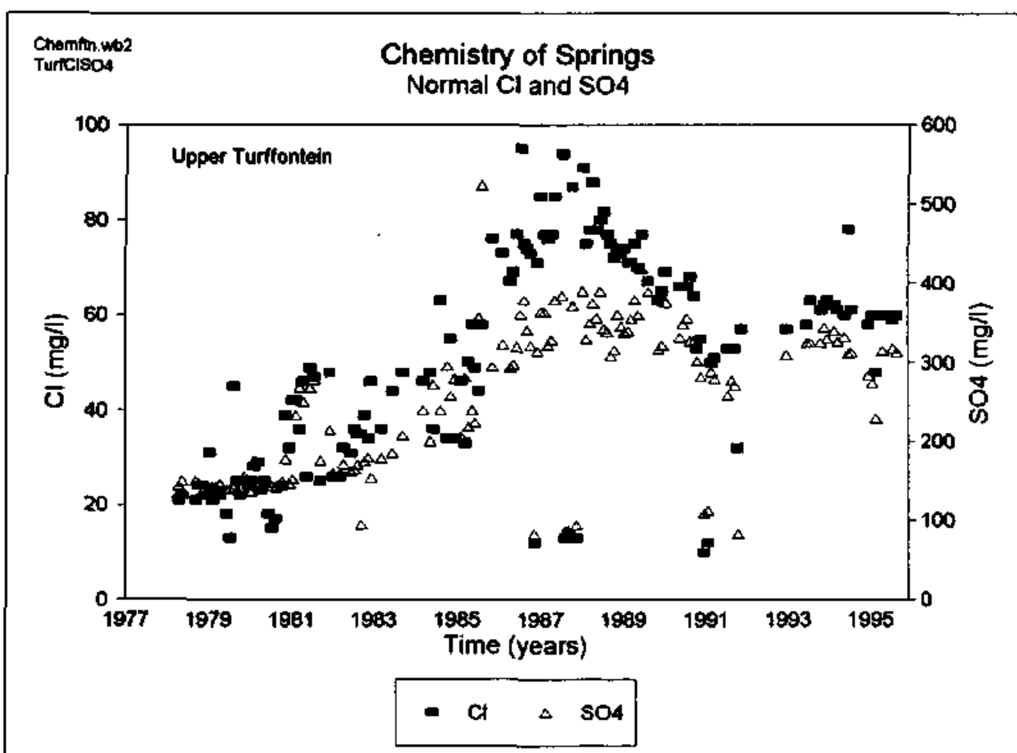


Fig. 11.3.3.1.4 Comparison of the chloride and sulphate concentrations of Turffontein eye - normalized plot.

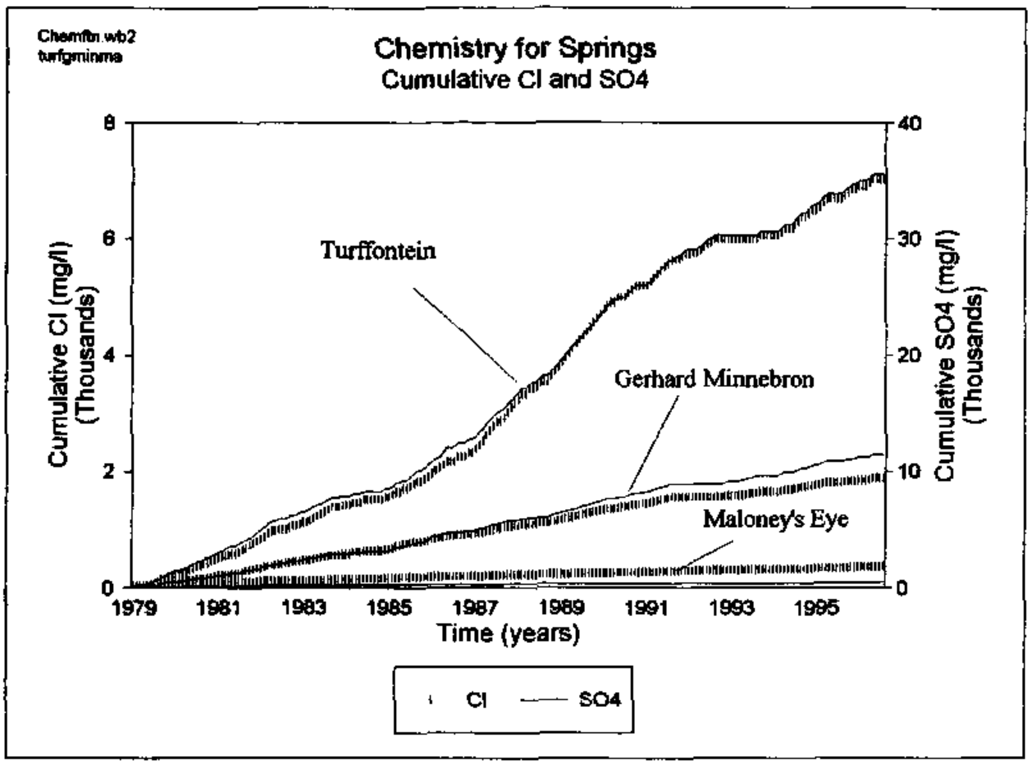


Fig. 11.3.3.2.1 Comparison of the chloride and sulphate concentrations of Turffontein, Maloney's and Gerhard Minnebron eyes.

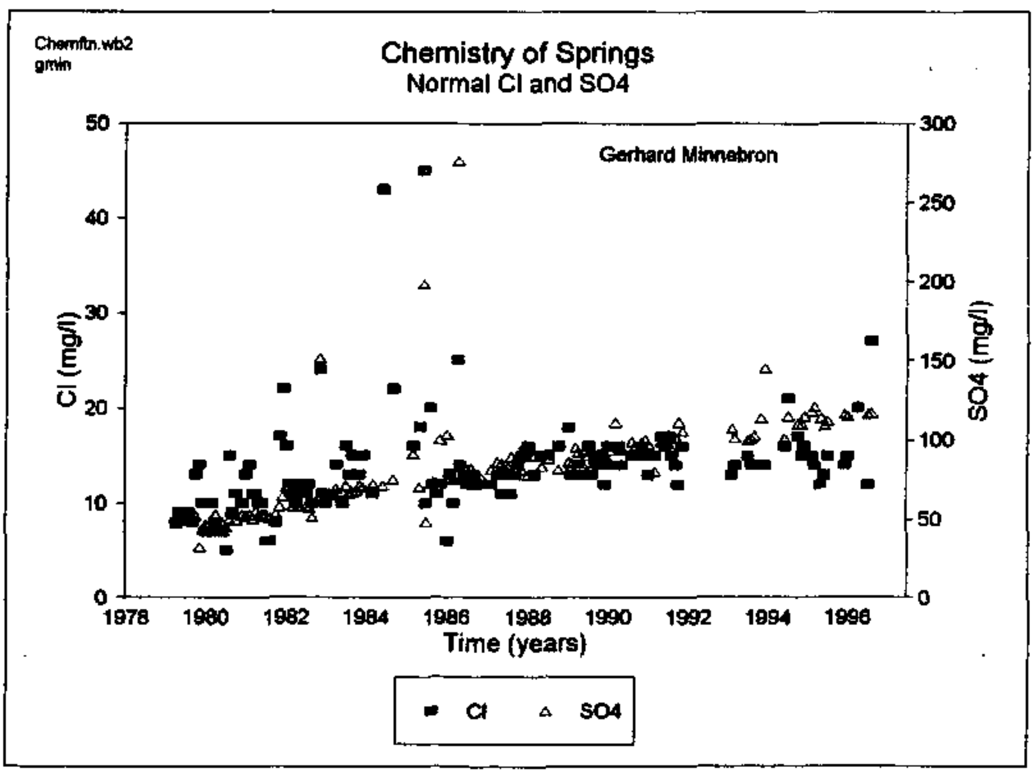


Fig. 11.3.3.2.2 Comparison of the chloride and sulphate concentrations of Gerhard Minnebron.

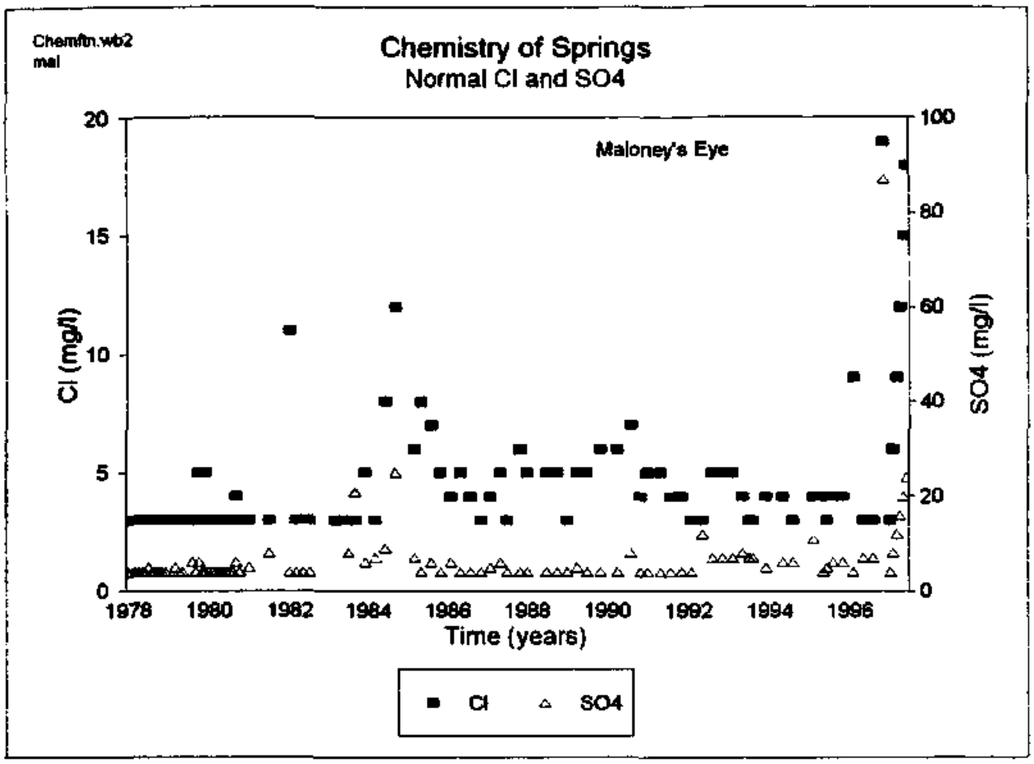


Fig. 11.3.3.2.3 Comparison of the chloride and sulphate concentrations of Maloney's eye.

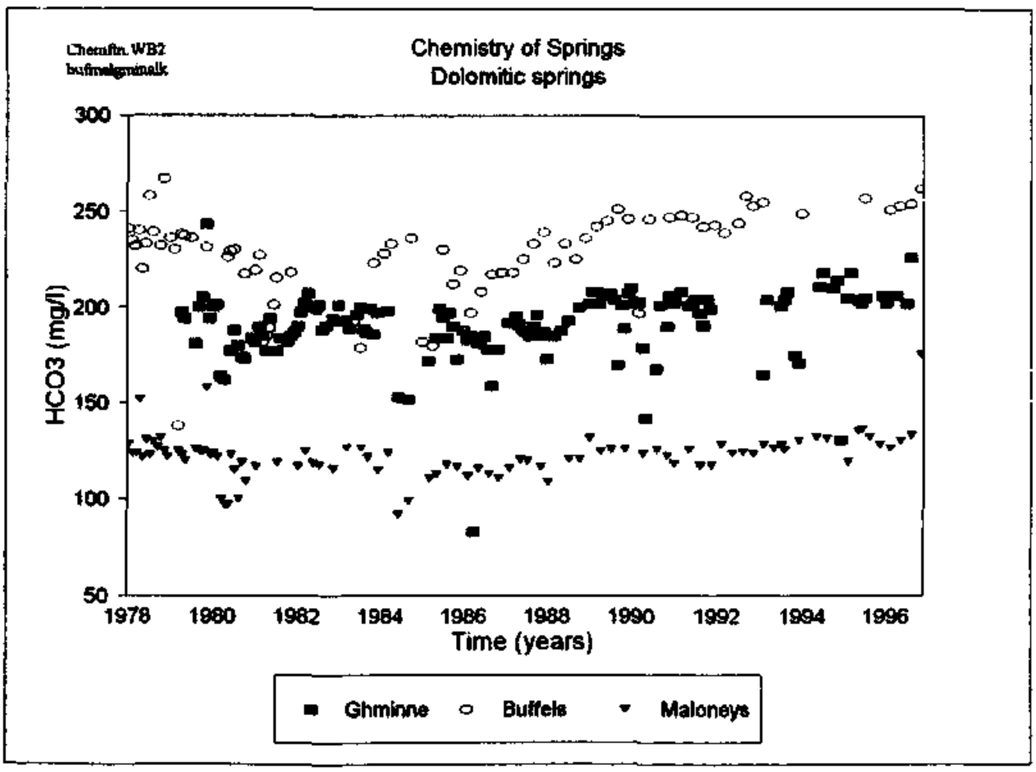


Fig. 11.3.3.2.4 Comparison of bicarbonate of the Buffelshoek, Maloney's and Gerhard Minnebron dolomitic springs.

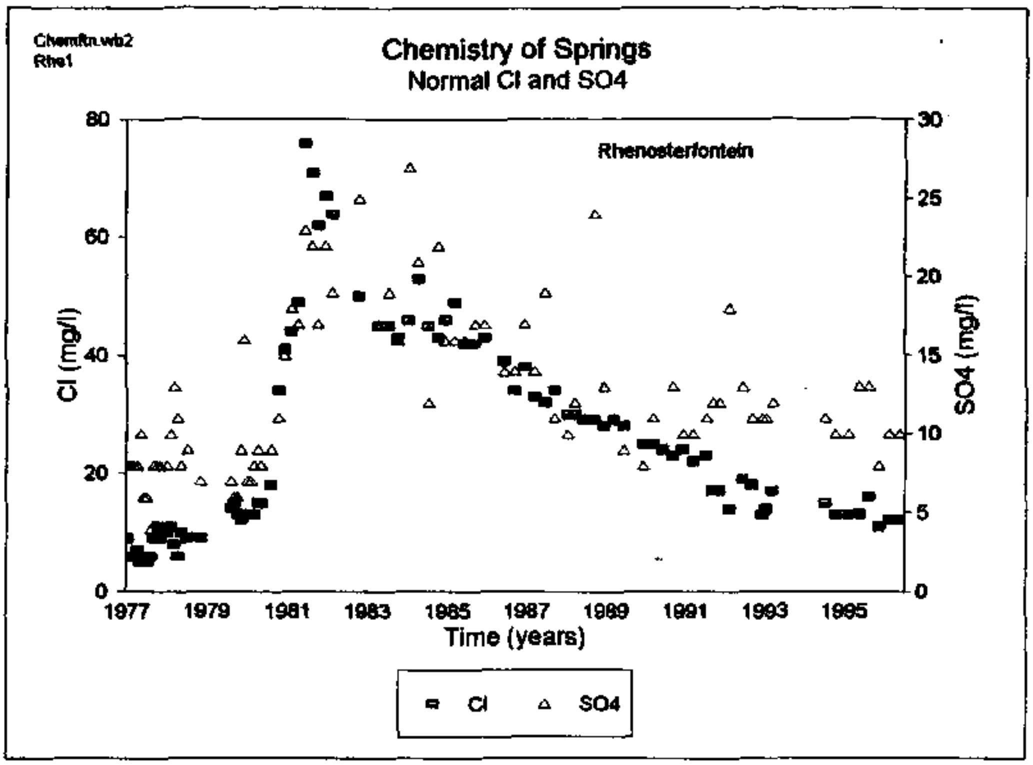


Fig. 11.3.3.3.1 Chloride and sulphate concentrations of Renosterfontein.

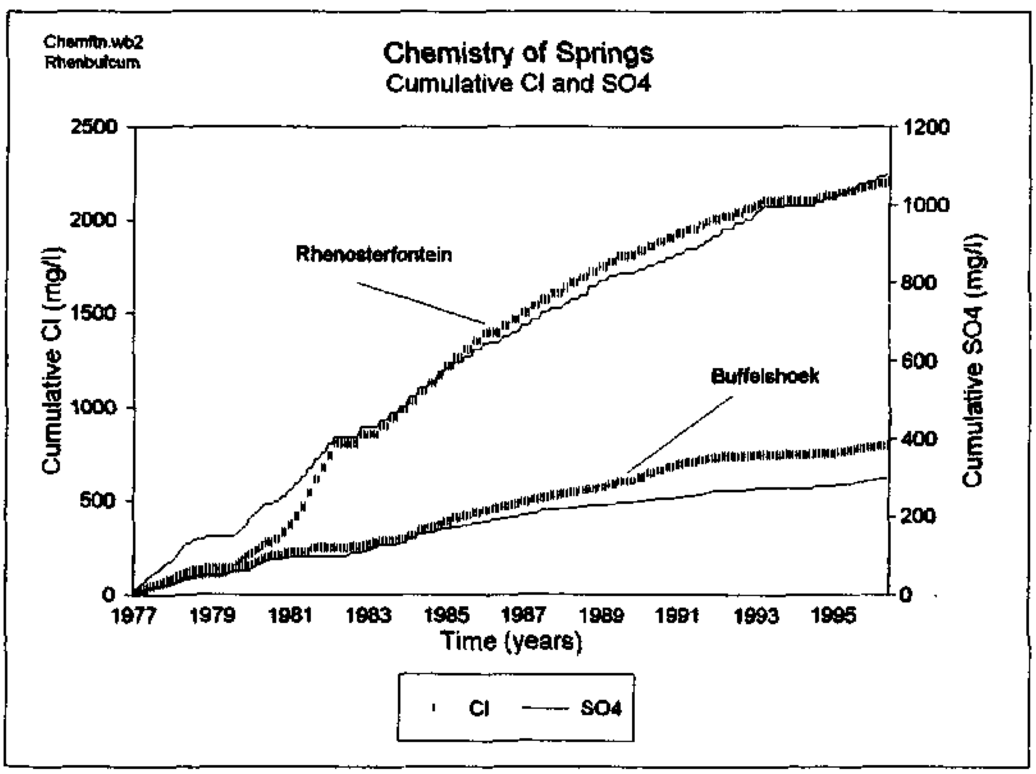


Fig. 11.3.3.3.2 Chloride and sulphate concentrations of Buffelshoek eye compared to those of Rhenosterfontein eye.

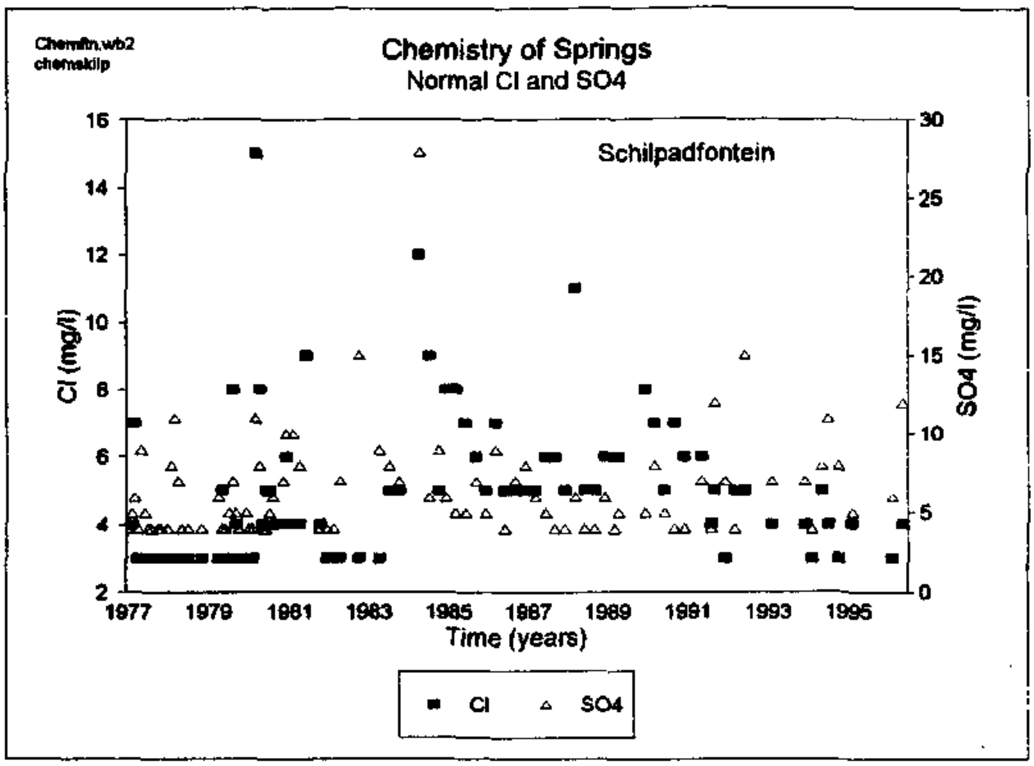


Fig. 11.3.3.3.3 Chloride and sulphate concentrations of Schilpadfontein eye.

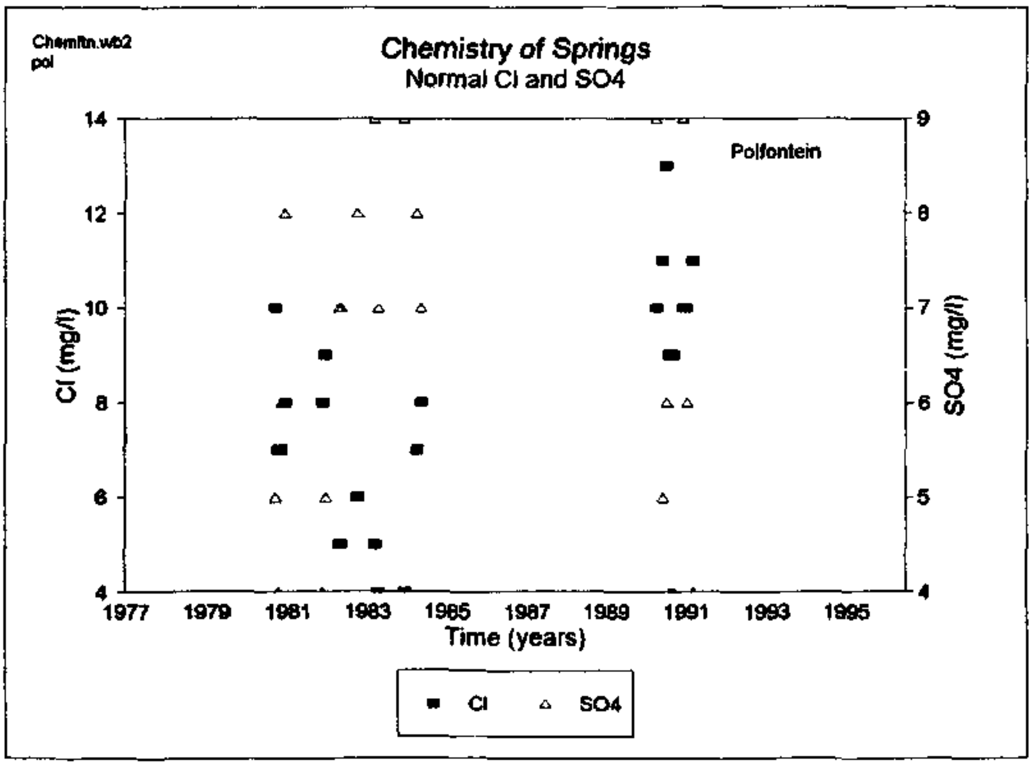


Fig. 11.3.3.3.4 Water quality of the Polfontein eye indicating little evidence of contamination.

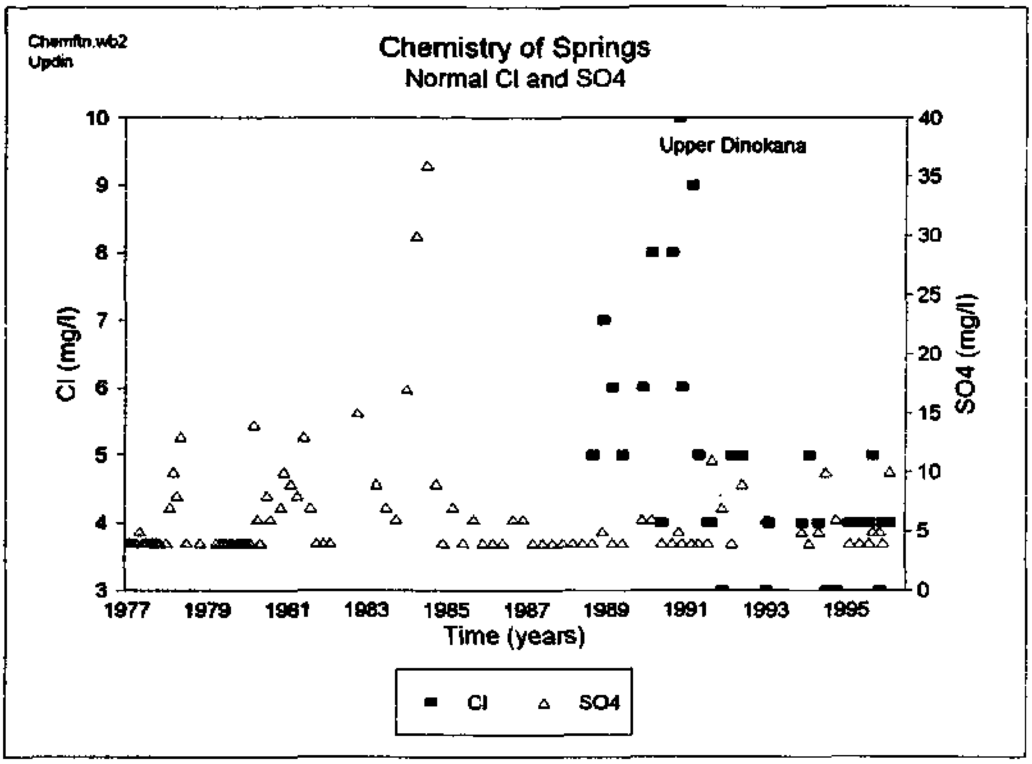


Fig. 11.3.3.3.5 Low concentrations of chloride and sulphate of the Upper Dinokana eye showing no clear evidence of pollution.

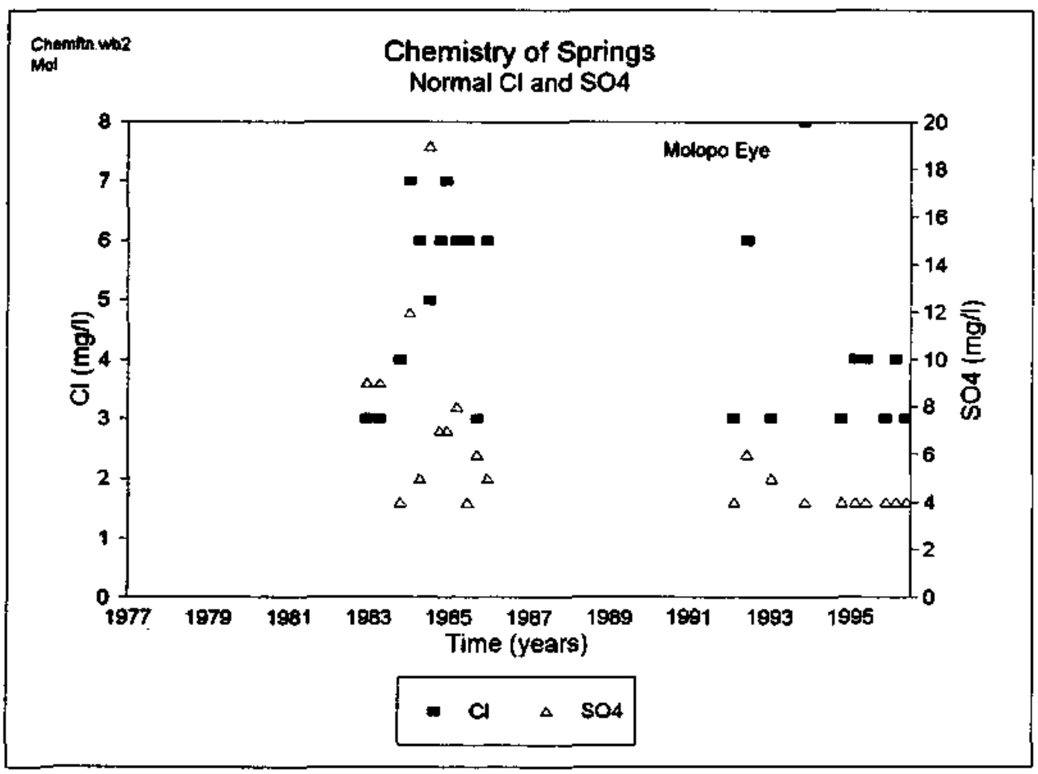


Fig. 11.3.3.3.6 Water quality of the Molopo eye showing no clear evidence of pollution.

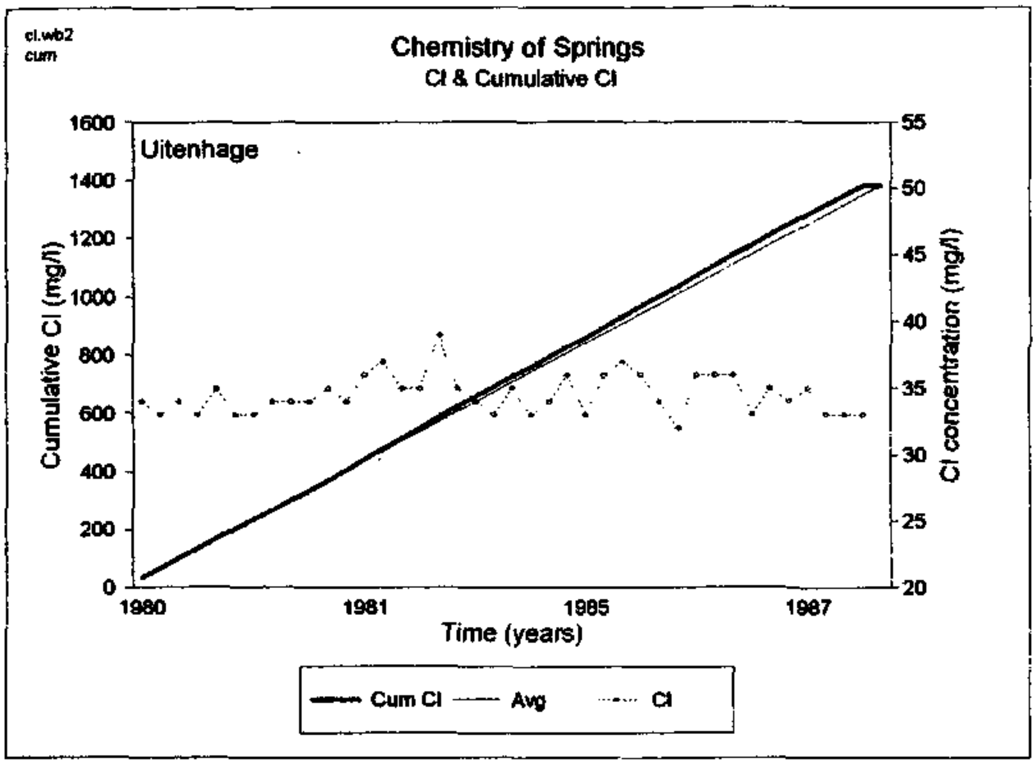


Fig. 11.3.3.4.1 Modified cumulative plot of chloride concentrations of the Uitenhage springs.

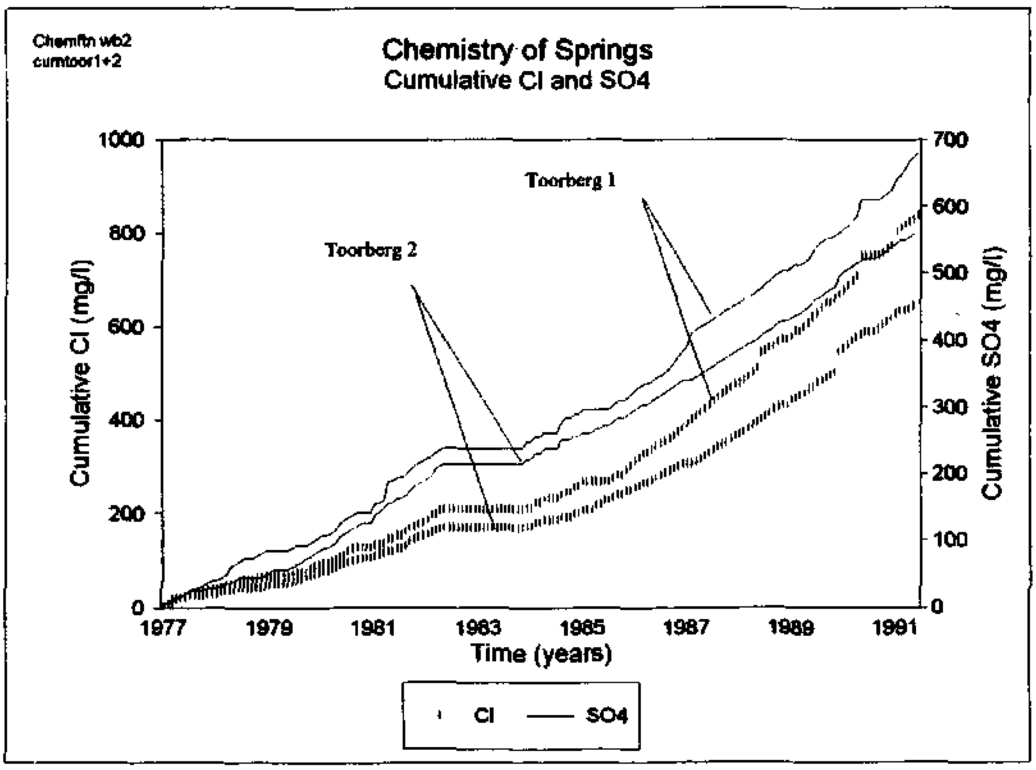


Fig. 11.3.3.4.2 Water quality of the Toorberg springs.

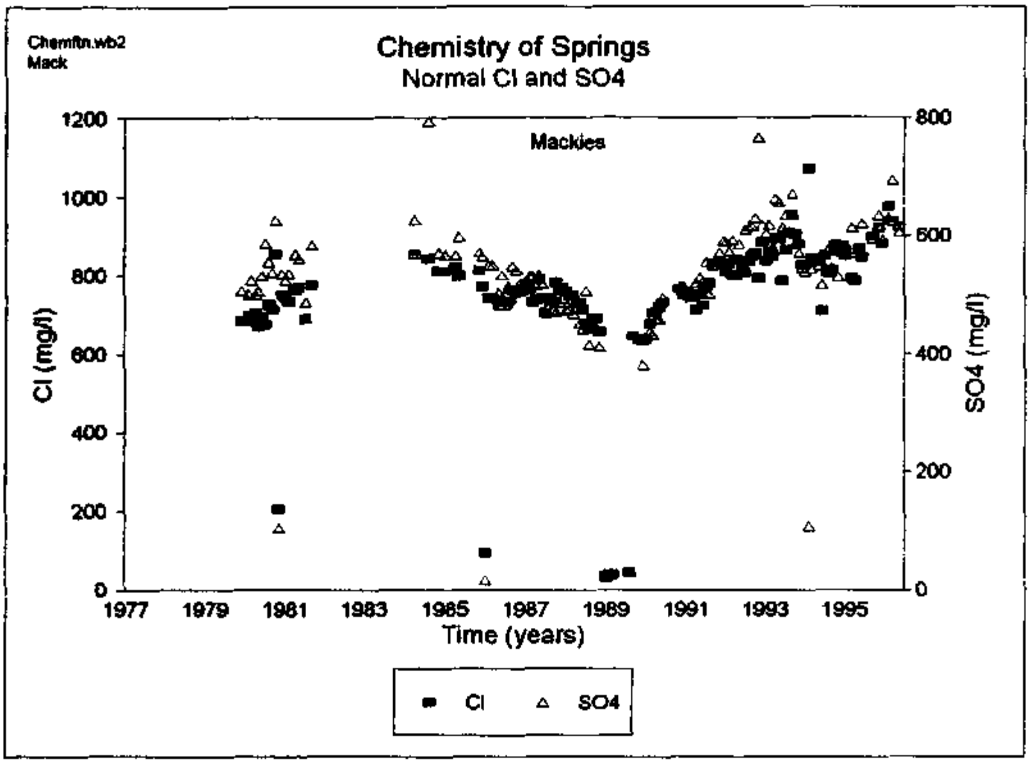


Fig. 11.3.3.4.3 Water quality of Mackies spring.

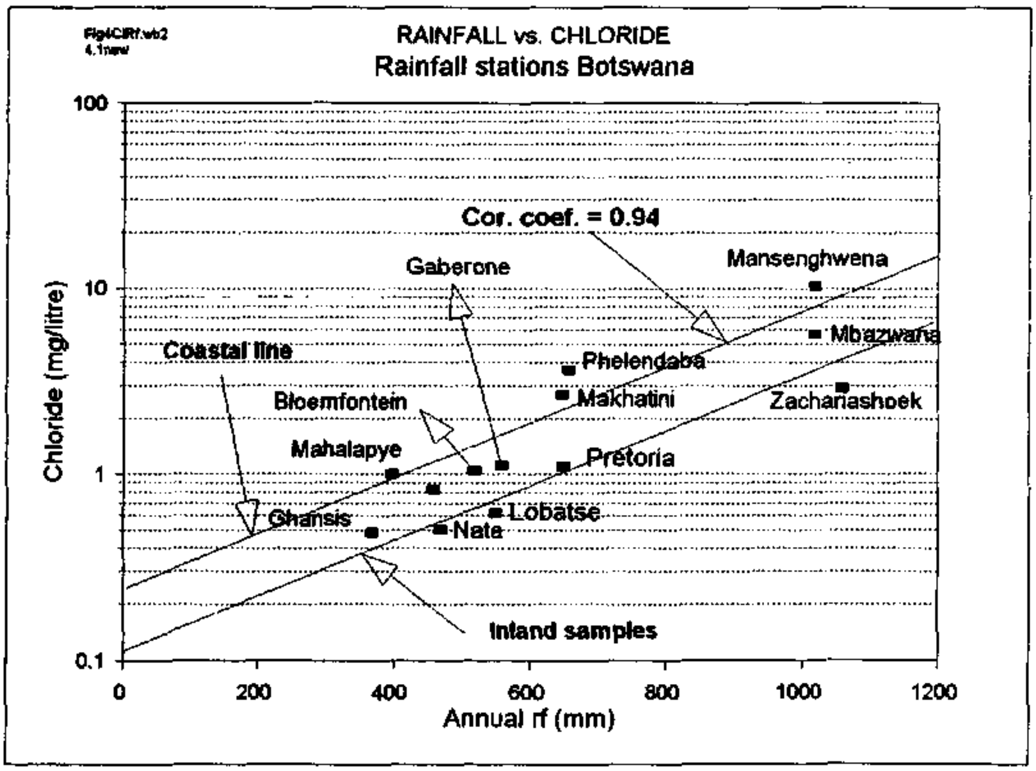


Fig. 11.4.1.1 Chloride concentrations in rainfall derived from measurements in the Zululand area and in Botswana.

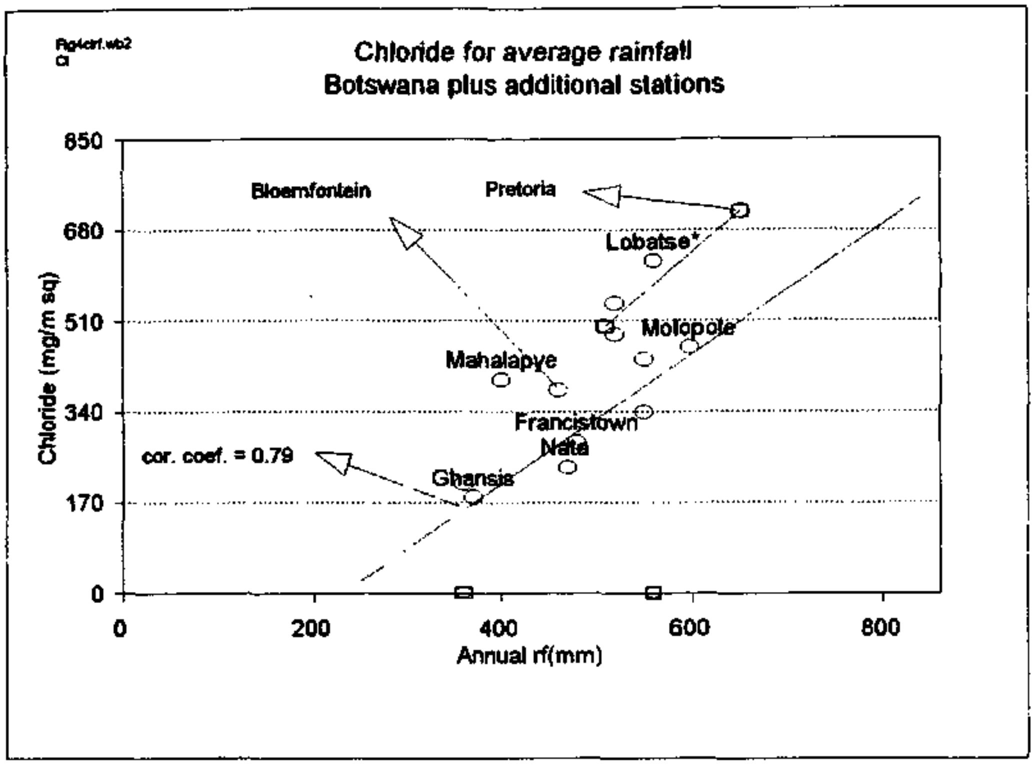


Fig. 11.4.1.2 Chloride in Botswana rainfall.

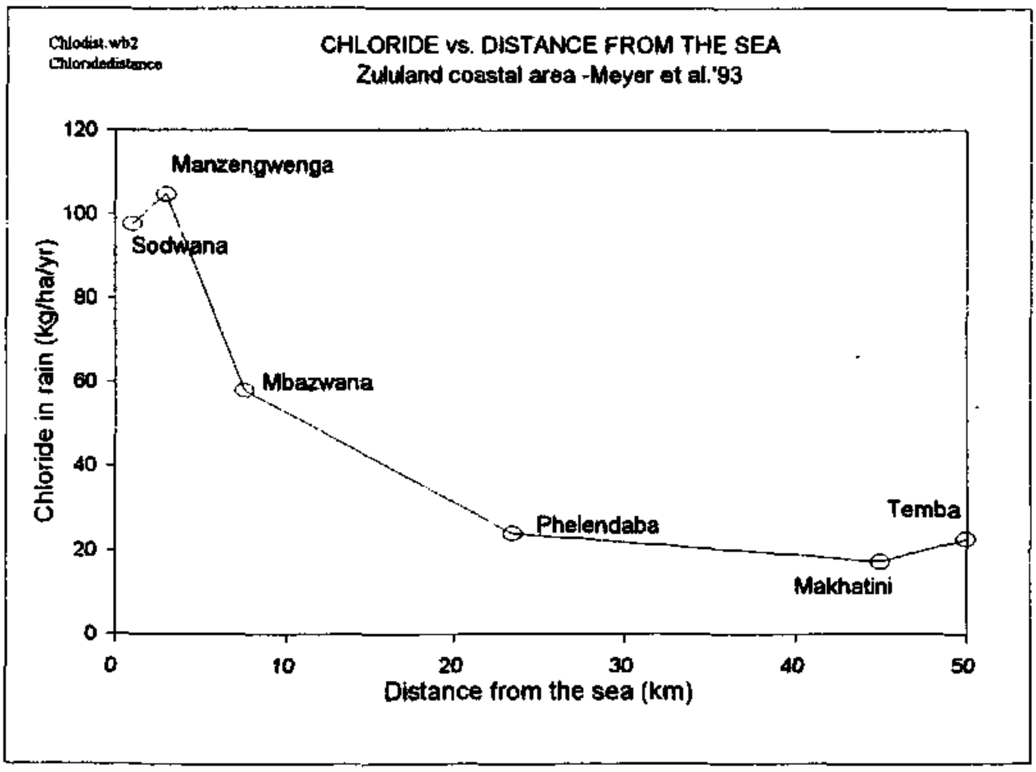


Fig. 11.4.1.3 Chloride depositioning of rainfall in the Zululand area.

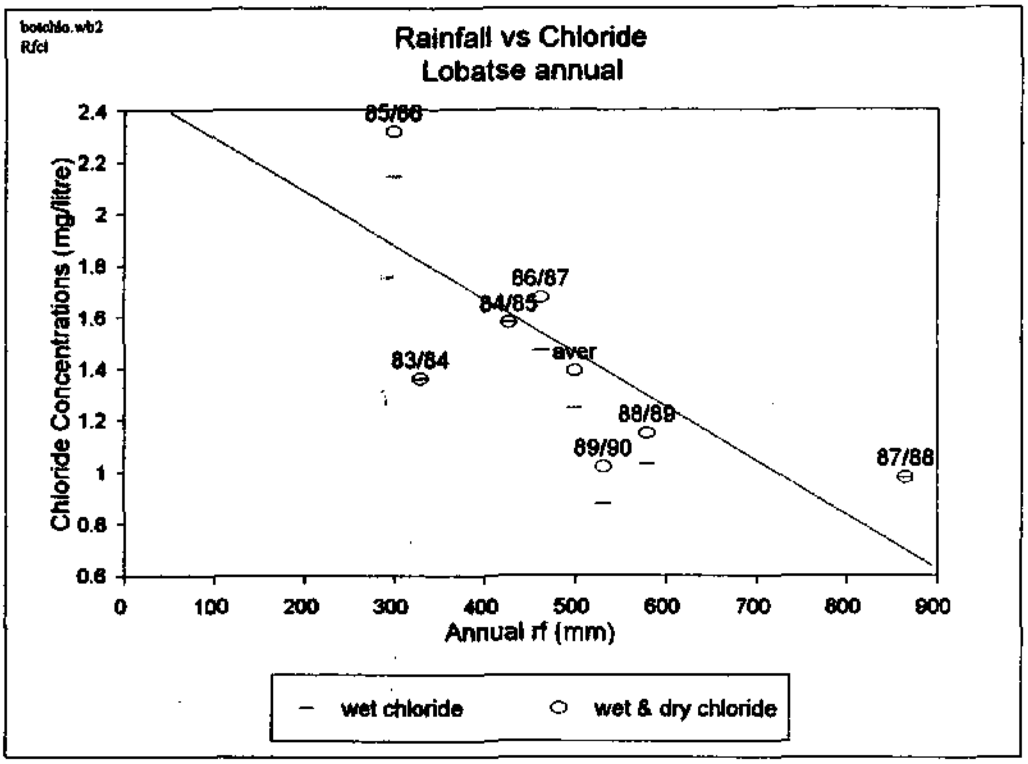


Fig. 11.4.1.4 Average annual chloride concentrations of rainfall measurements at Lobatse.

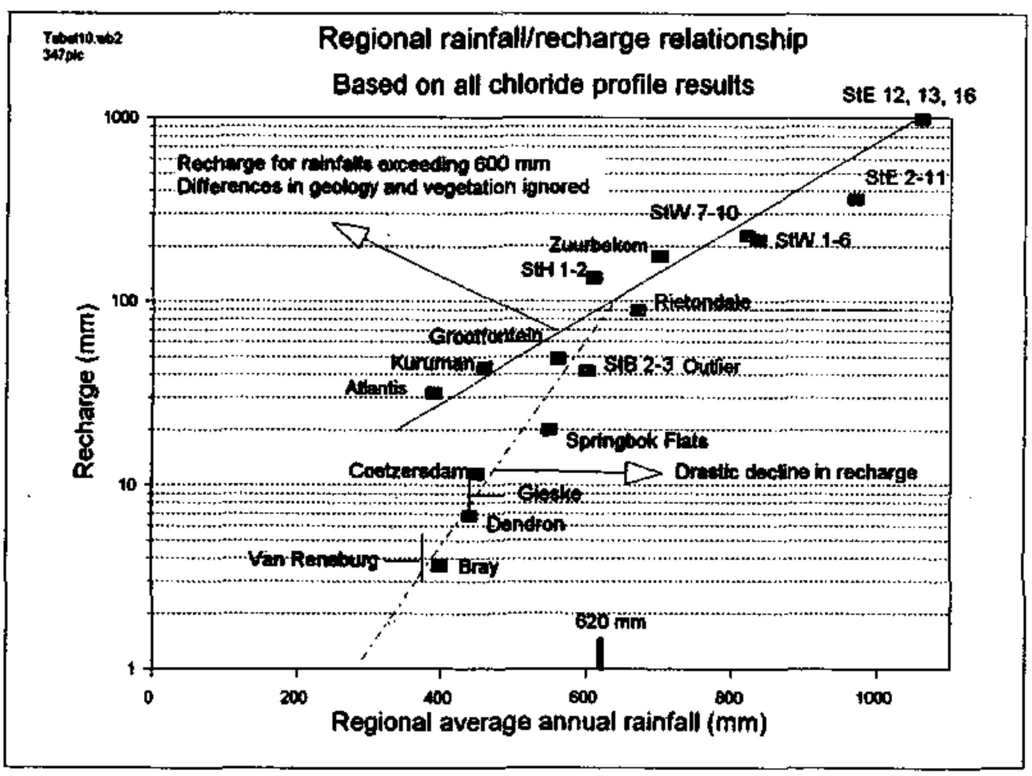


Fig. 11.4.1.5 Rainfall/recharge relationship derived by the chloride ratio method.

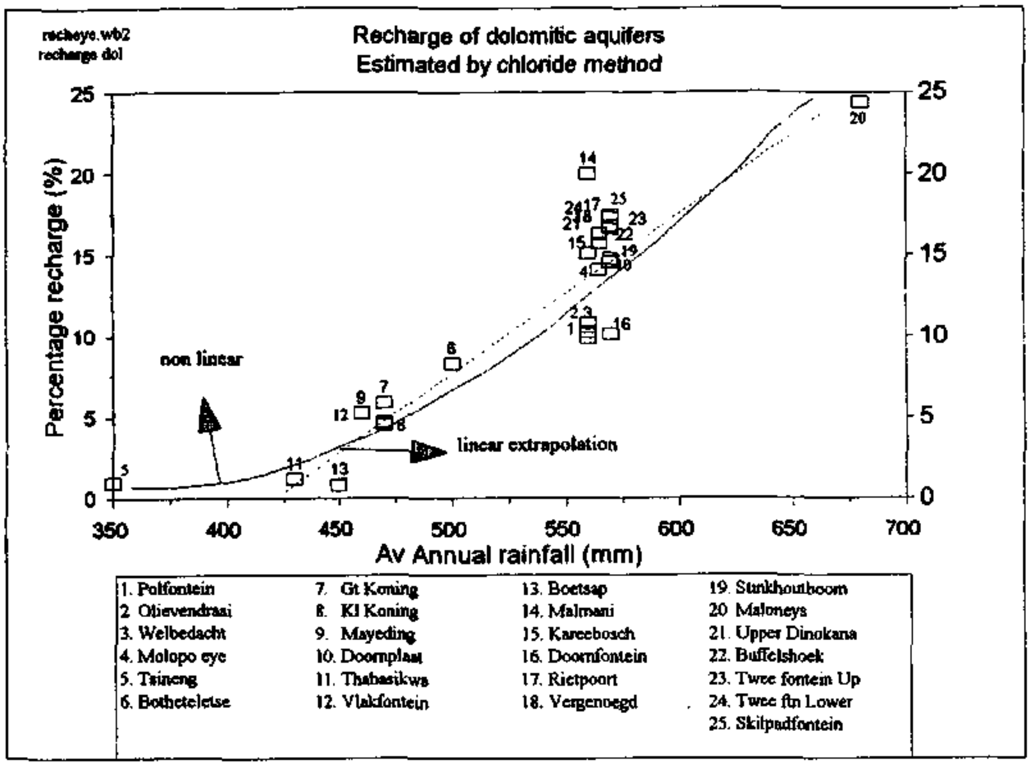


Fig. 11.4.2.1 Recharge derived from chloride concentrations of different dolomitic springs - assuming no contamination.

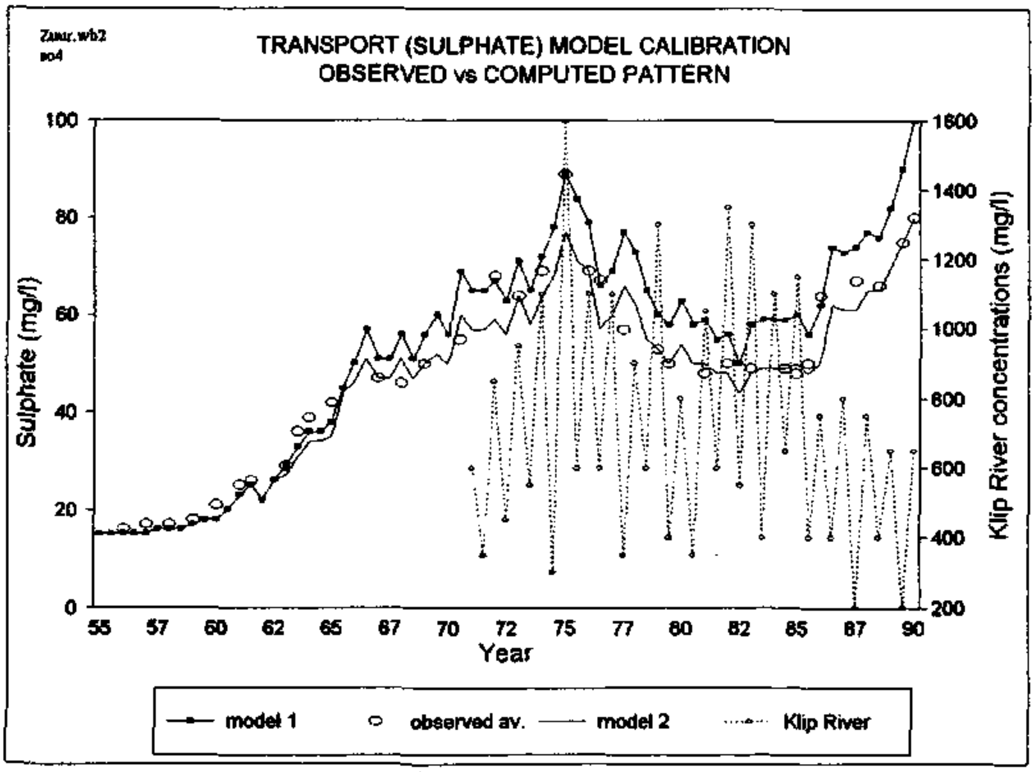


Fig. 11.5.2.1 Temporal variation of water quality of production boreholes at Zuurbekom.

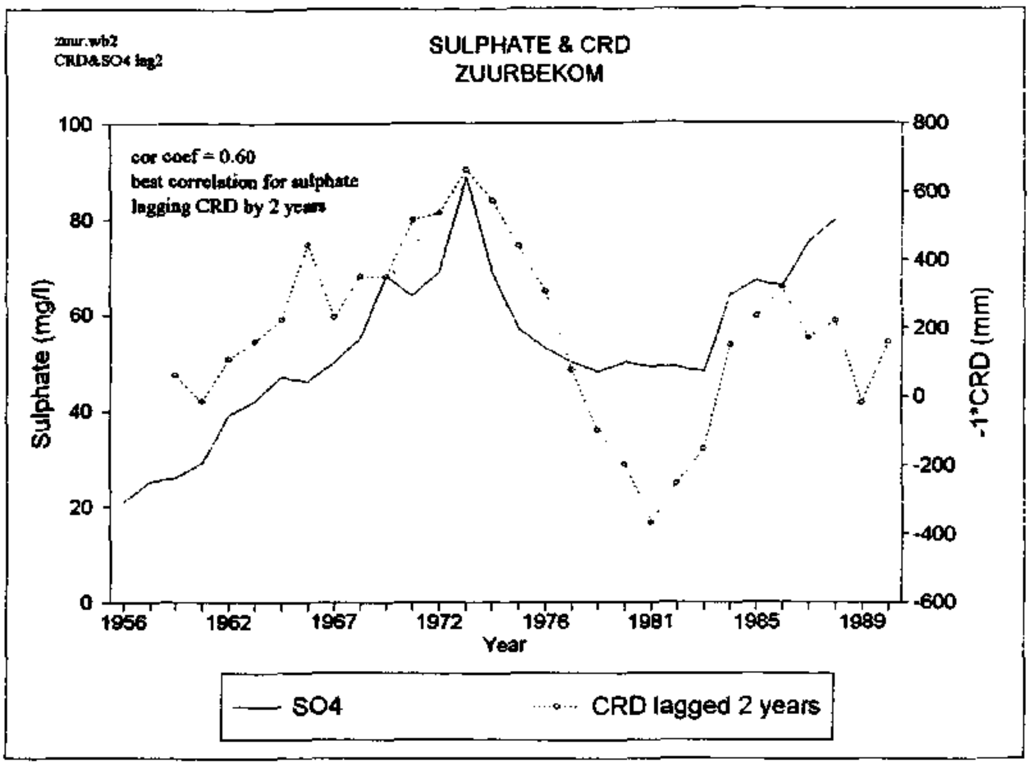


Fig. 11.5.2.2 Sulphate concentrations of Zuurbekom boreholes in relation to the cumulative rainfall departures, incorporating a lag of 2 years in the sulphate response.

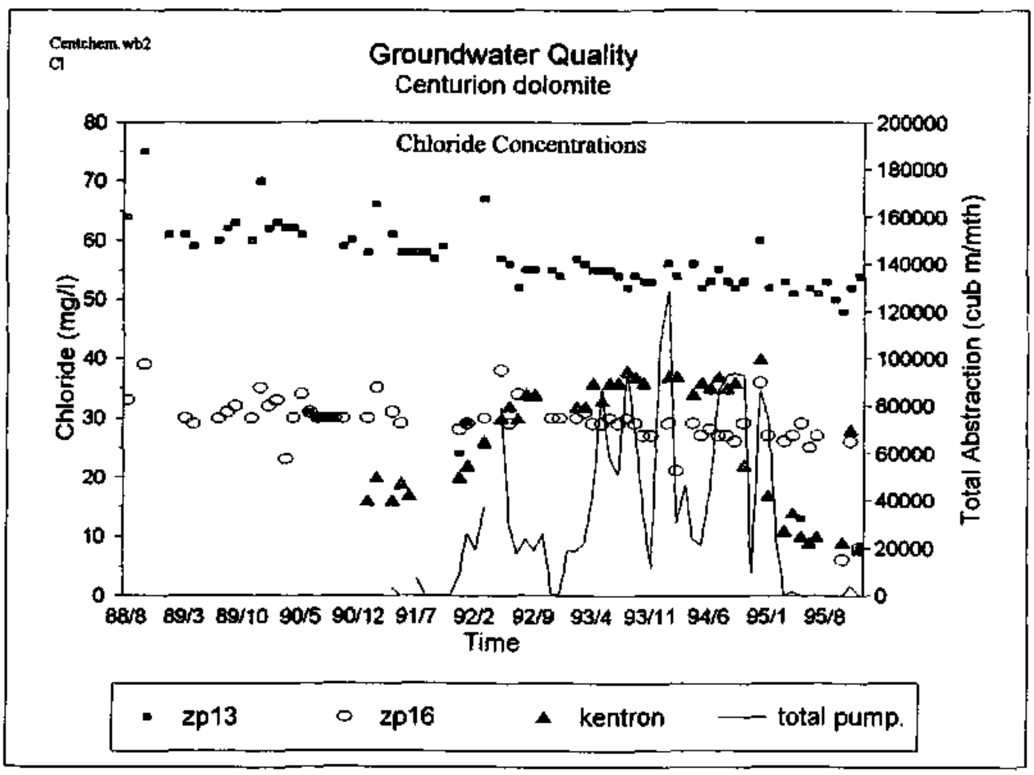


Fig. 11.5.3.1 Comparison of the chloride concentrations of groundwater obtained from the Centurion production boreholes.

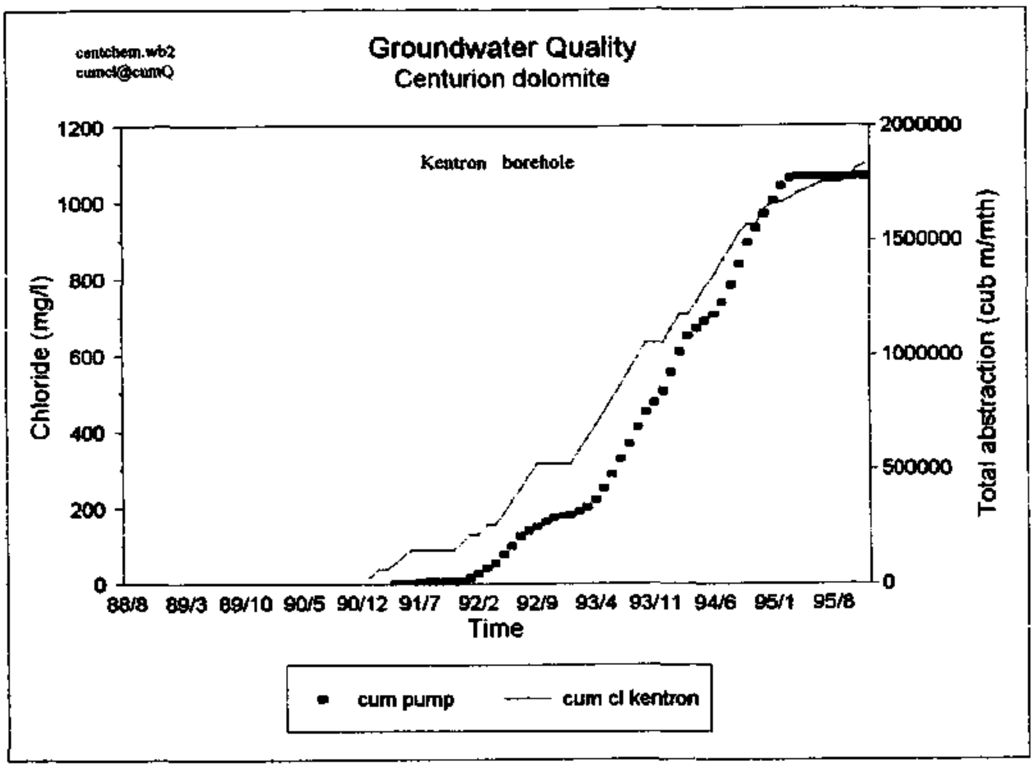


Fig. 11.5.3.2 Comparison of sulphate and chloride concentrations of the Pretoria Fountains.

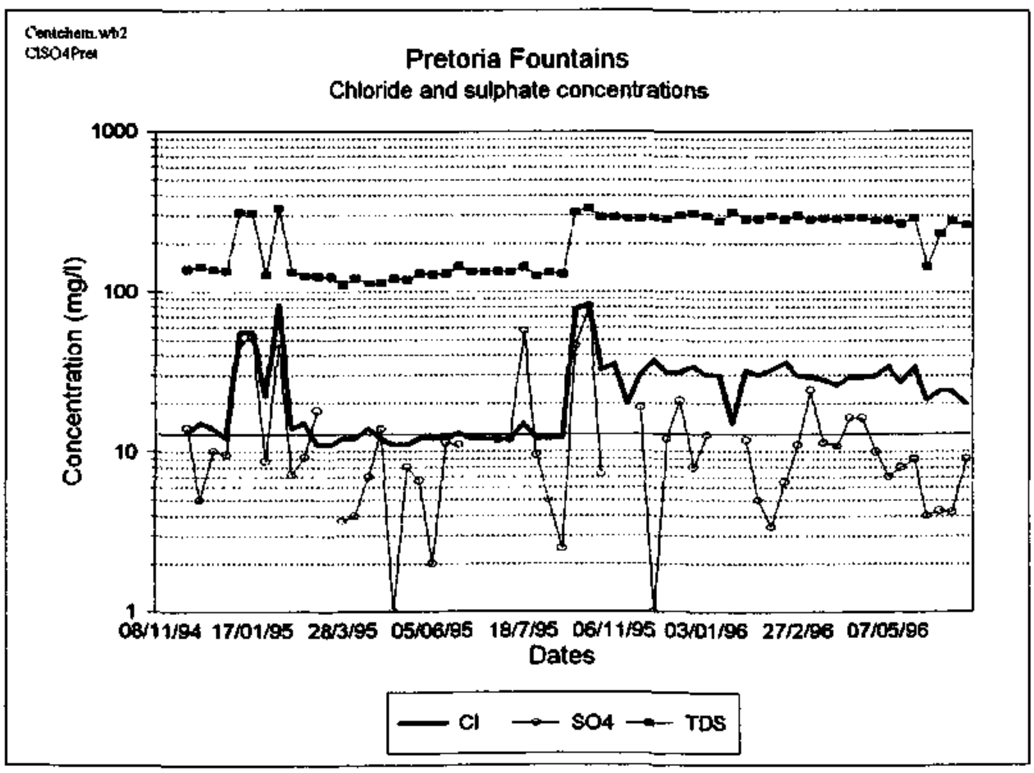


Fig. 11.5.3.3 Cumulative plot of the abstraction in relation to the chloride concentrations of Centurion boreholes.

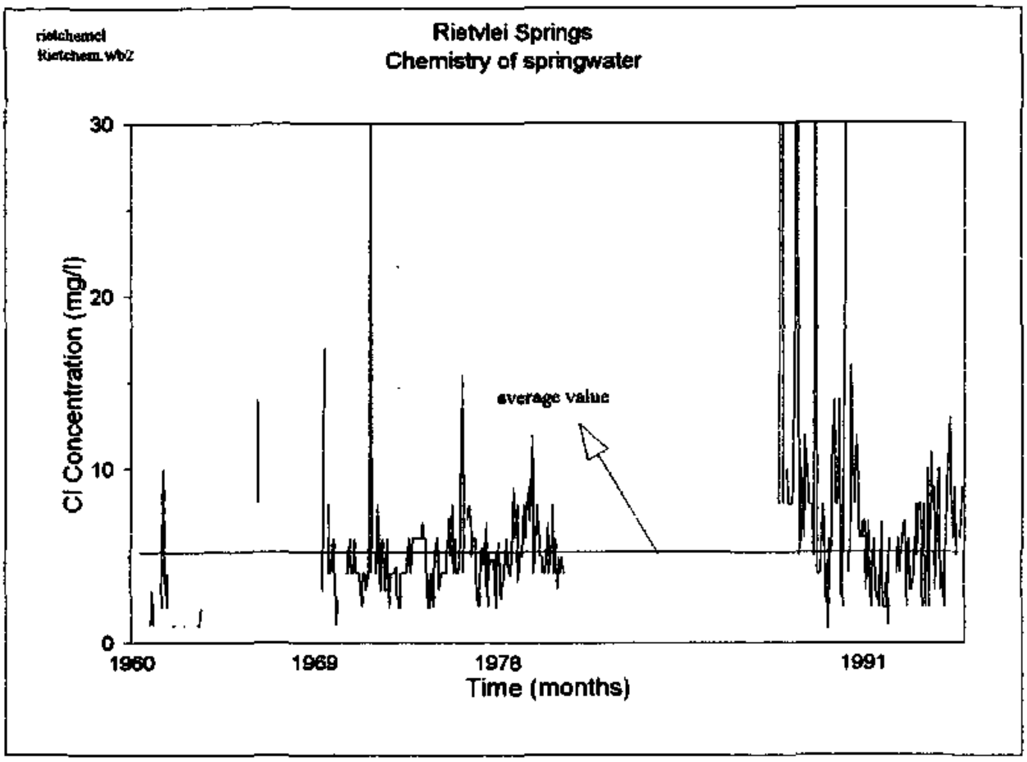


Fig. 11.5.3.4 Comparison of chloride concentrations of the Rietvlei springs in Pretoria.

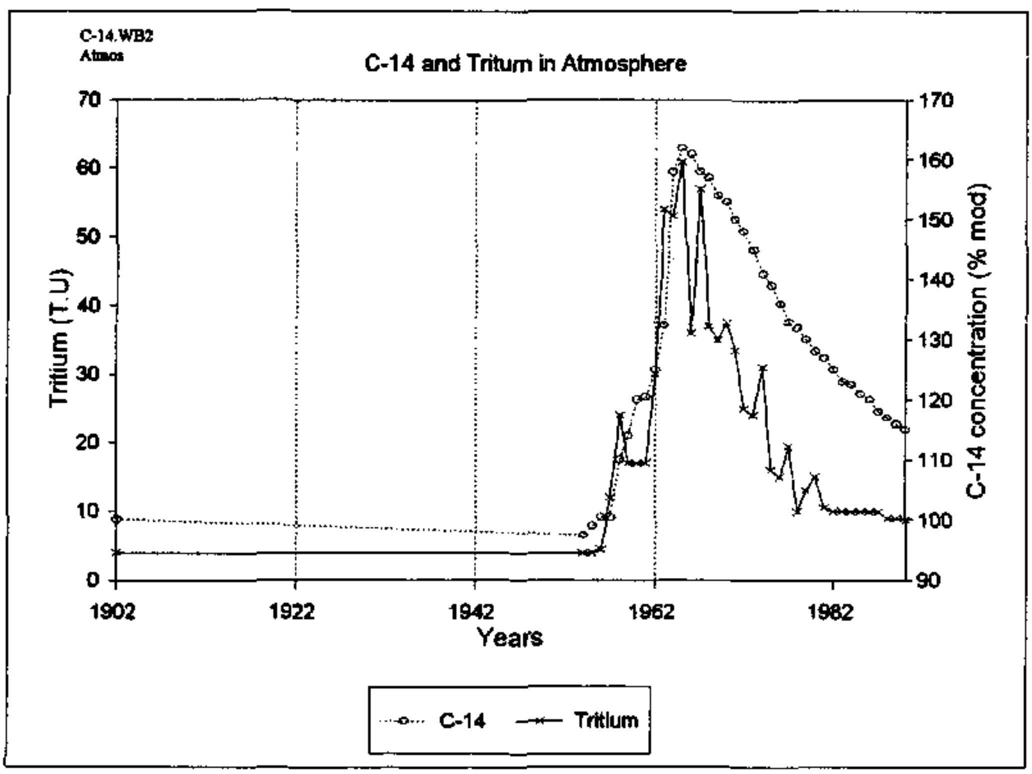


Fig.12.1.1 Natural variation in tritium and ^{14}C concentrations in the atmosphere, showing effect of bomb input.

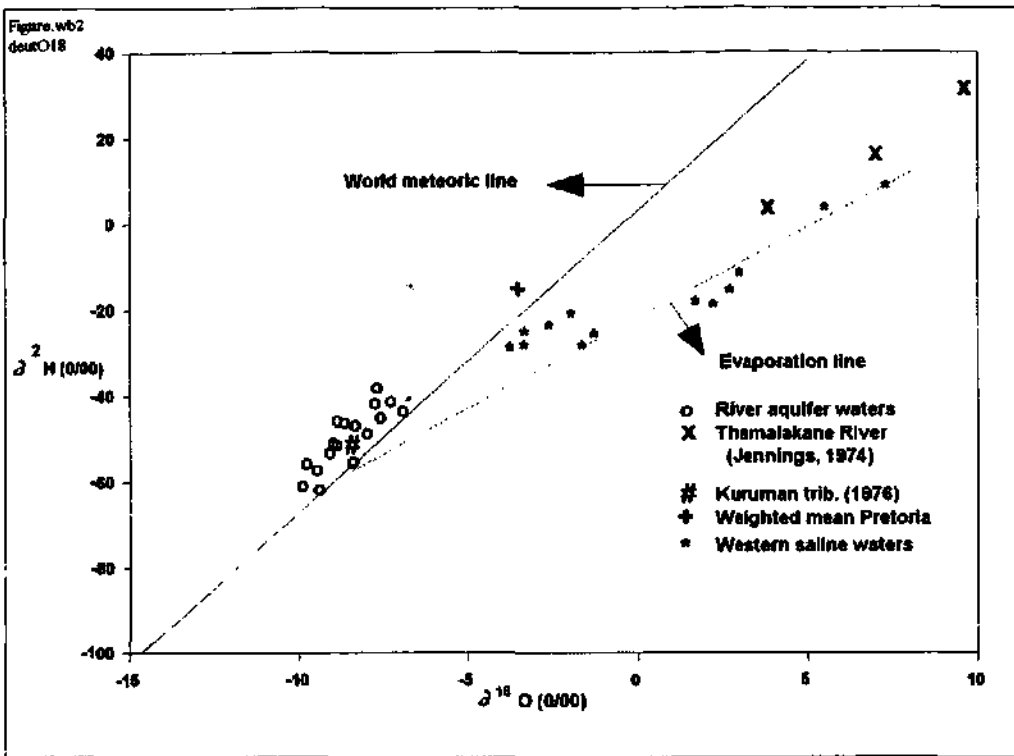


Fig. 12.3.1 Relationship between deuterium and oxygen-18 in natural waters as well as the effect of evaporation.

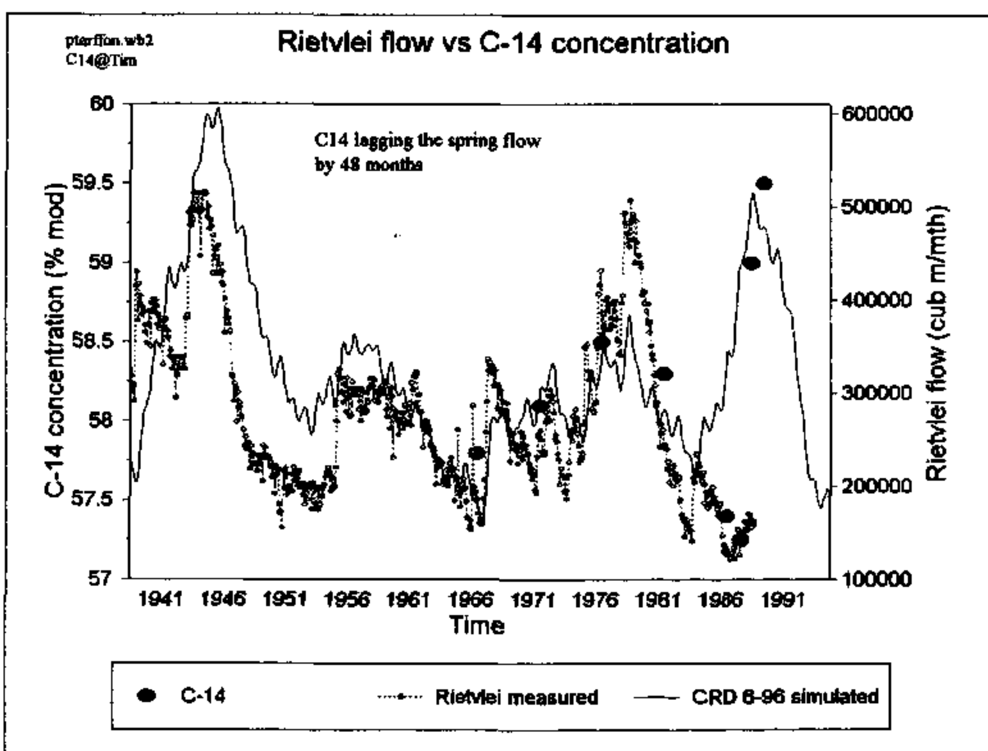


Fig. 12.4.2.1 Variation of ^{14}C concentrations of the Grootfontein eye in the Rietvlei-Pretoria area.

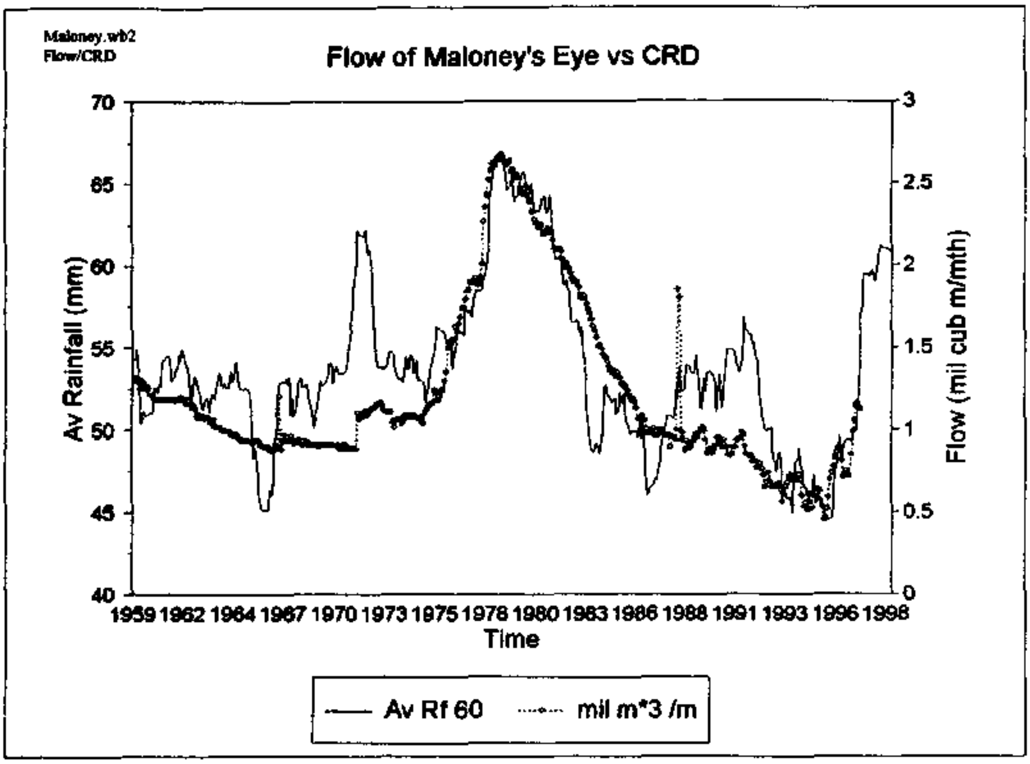


Fig. 12.4.2.2a Flow of Maloney's Eye in relation to the moving average rainfall over 60 months.

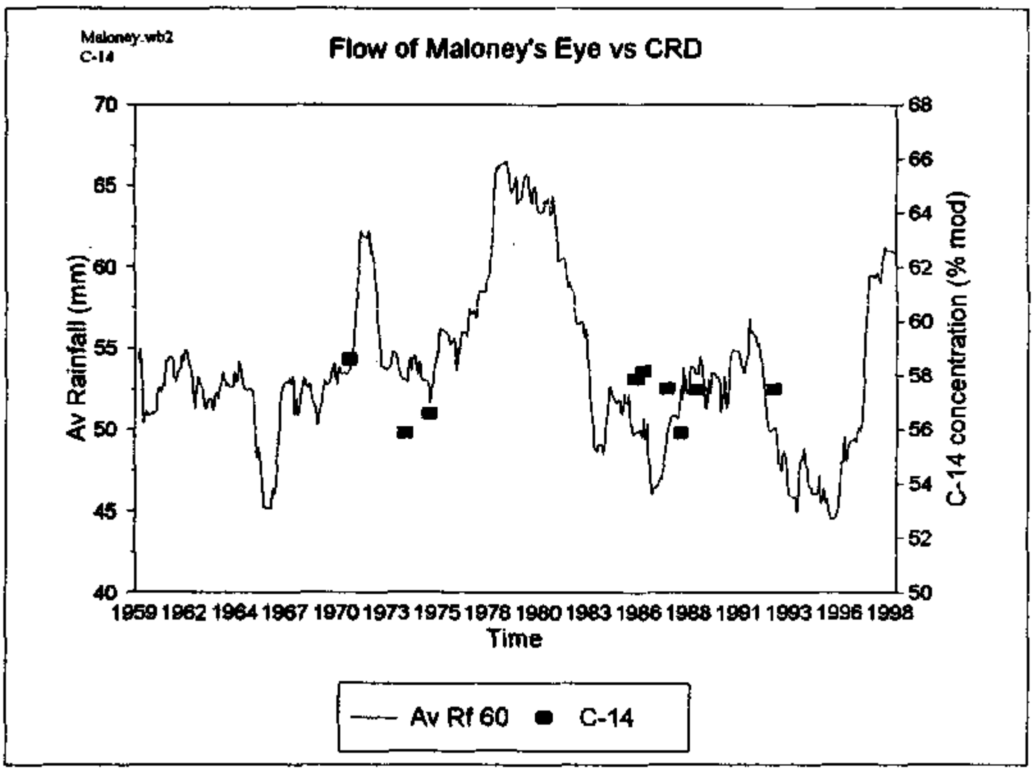


Fig. 12.4.2.2b ^{14}C concentrations of Maloney's eye in relation to the average rainfall over 60 months.

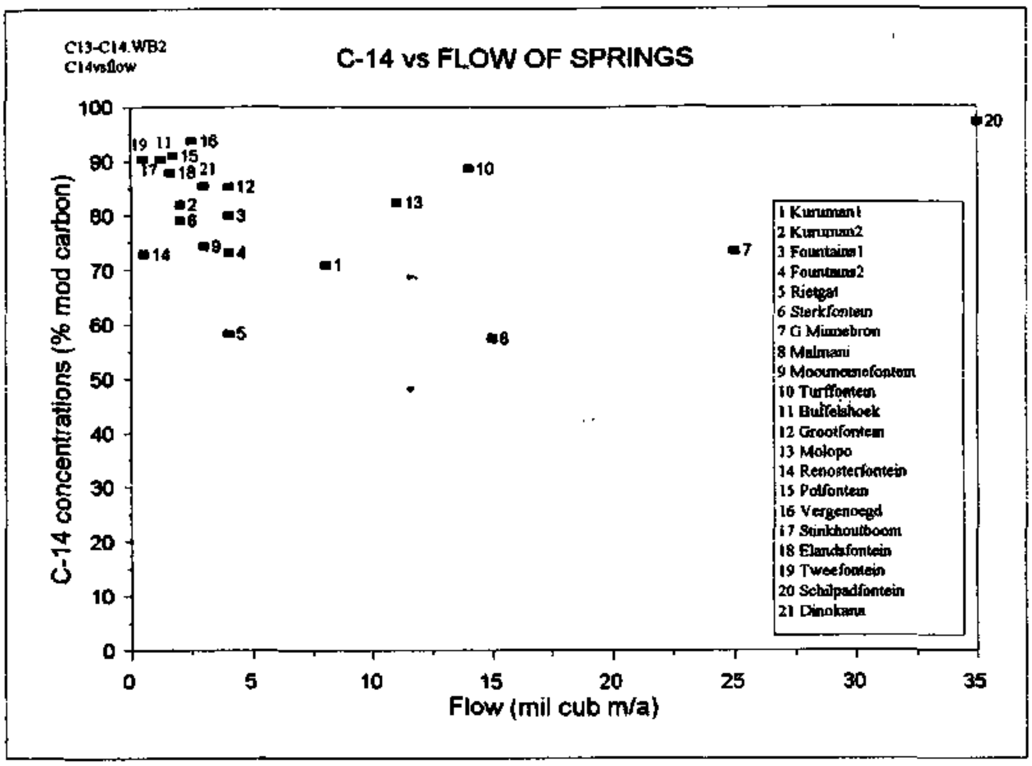


Fig. 12.4.2.3 ^{14}C concentrations in relation to the average discharge rate of springs.

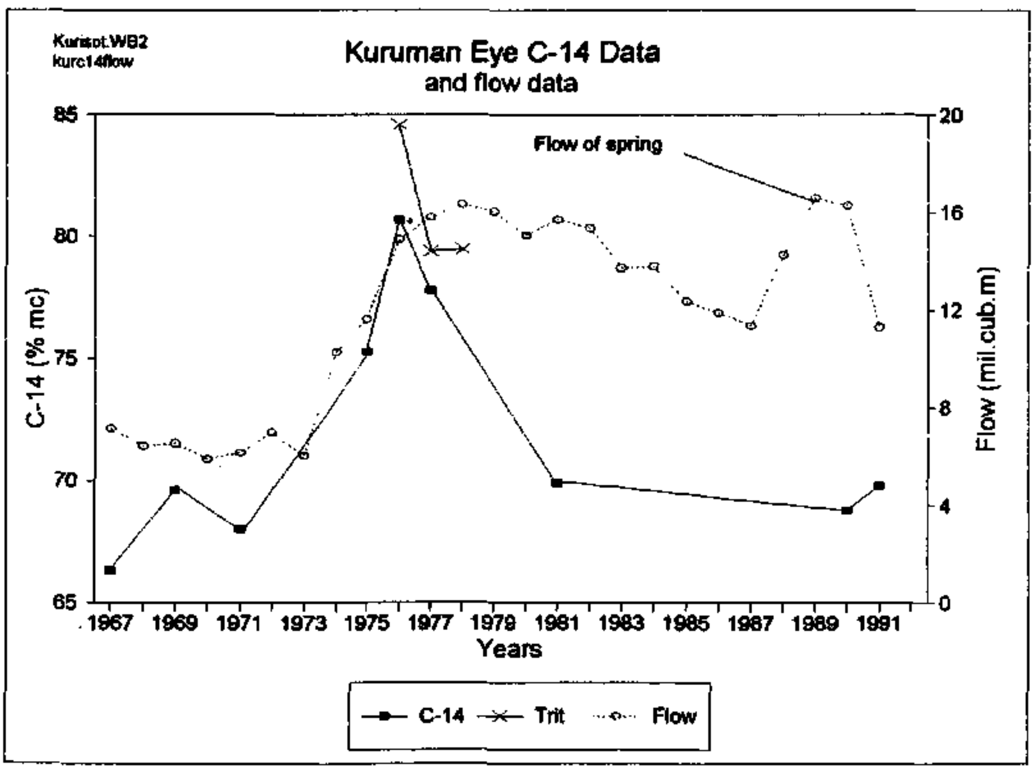


Fig. 12.4.2.4 ^{14}C concentrations measured at main Kuruman eye.

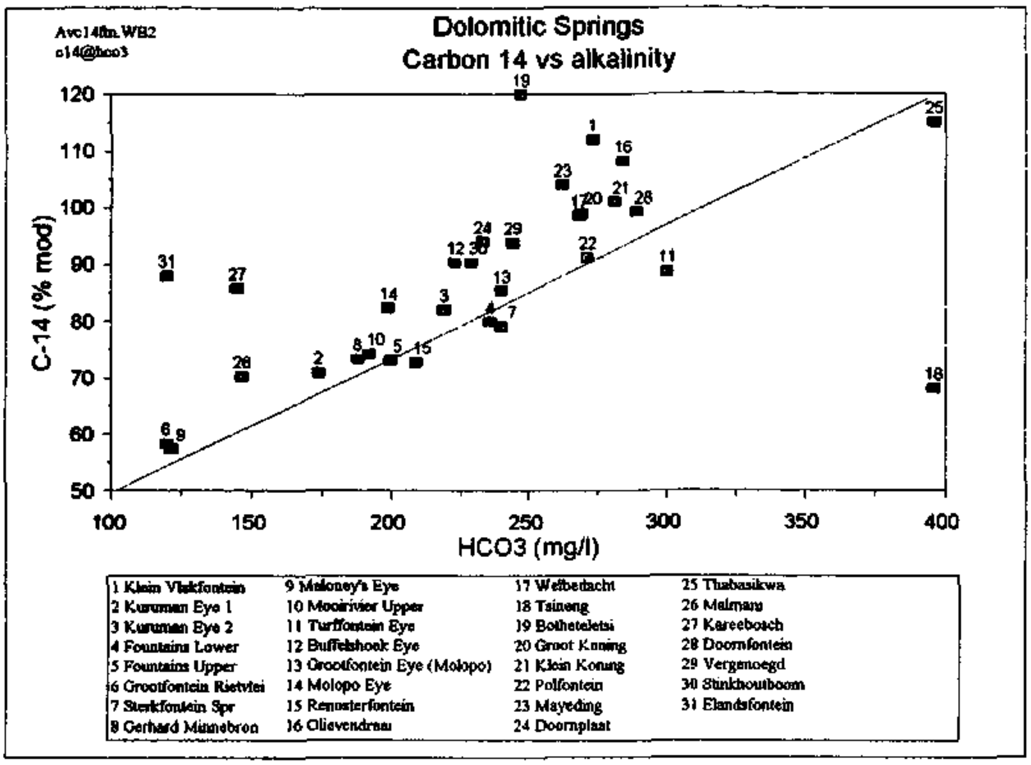


Fig. 12.4.3.1.1 ^{14}C concentrations of different dolomitic springs in relation to alkalinity.

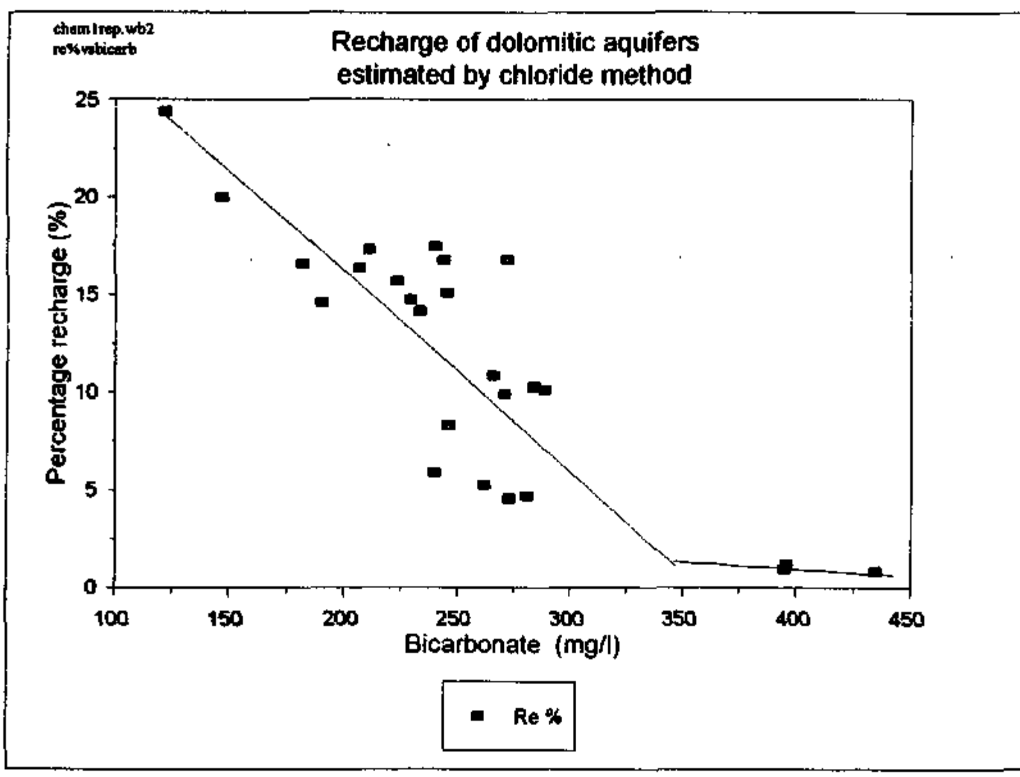


Fig. 12.4.3.1.2 Recharge of different springs inferred by means of the chloride method in relation to alkalinity.

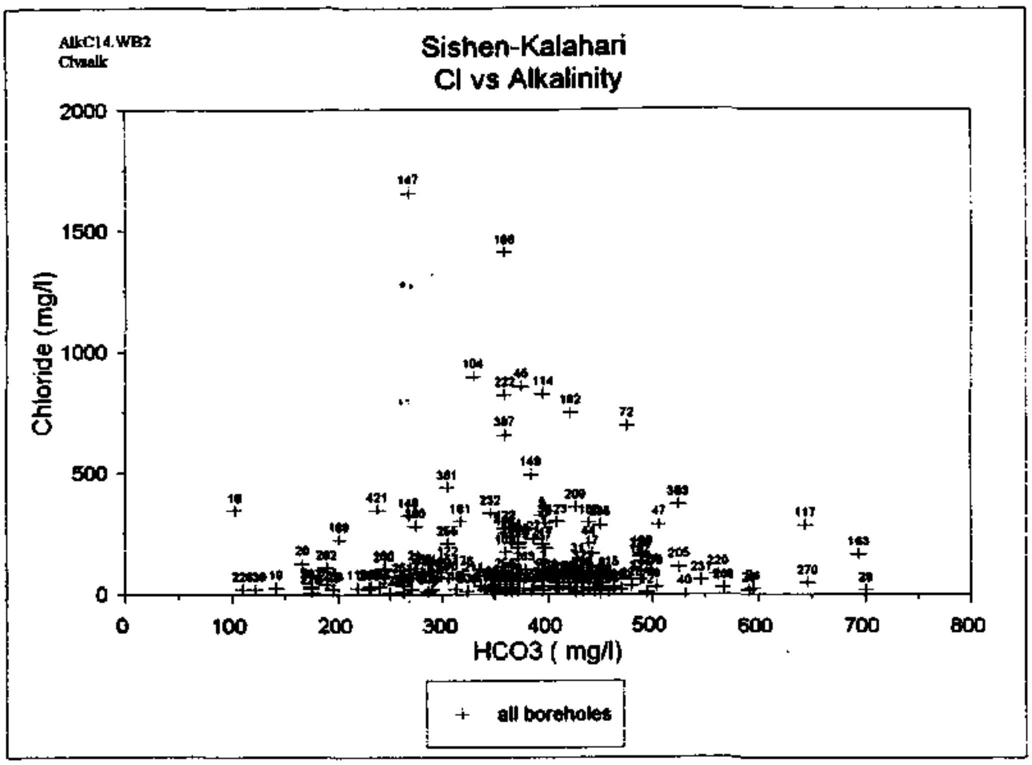


Fig. 12.4.3.2.2 Chloride concentrations in relation to the alkalinity of groundwaters in the Sishen area.

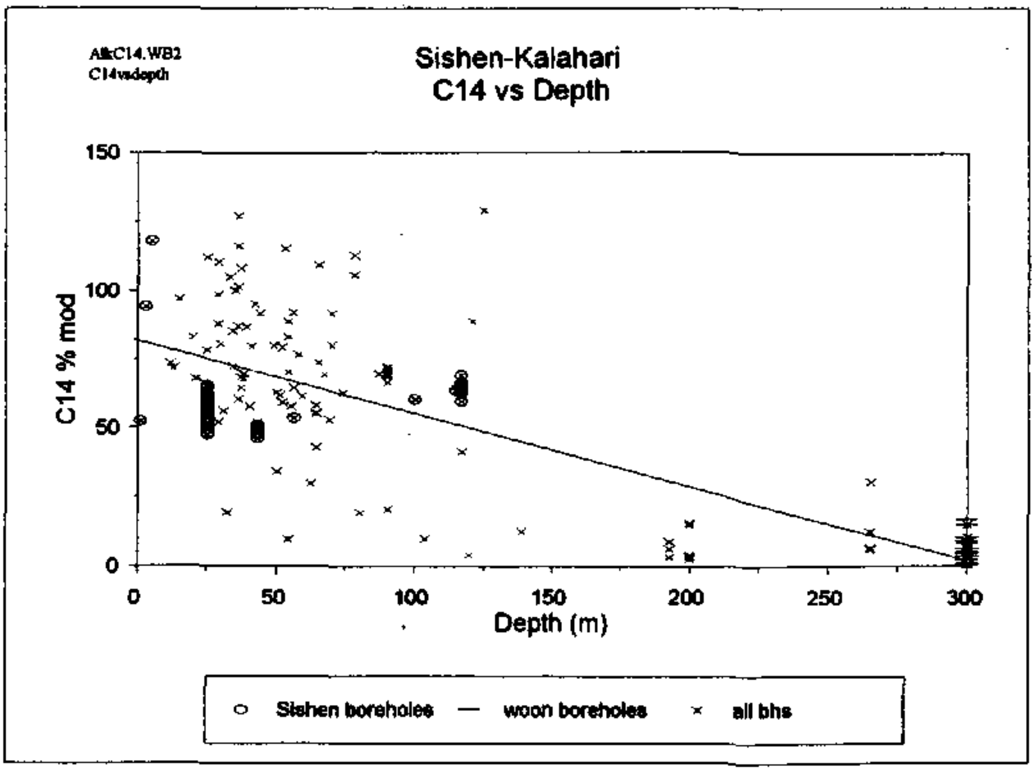


Fig. 12.4.3.2.3 ^{14}C concentrations in relation to the depth of water strike.

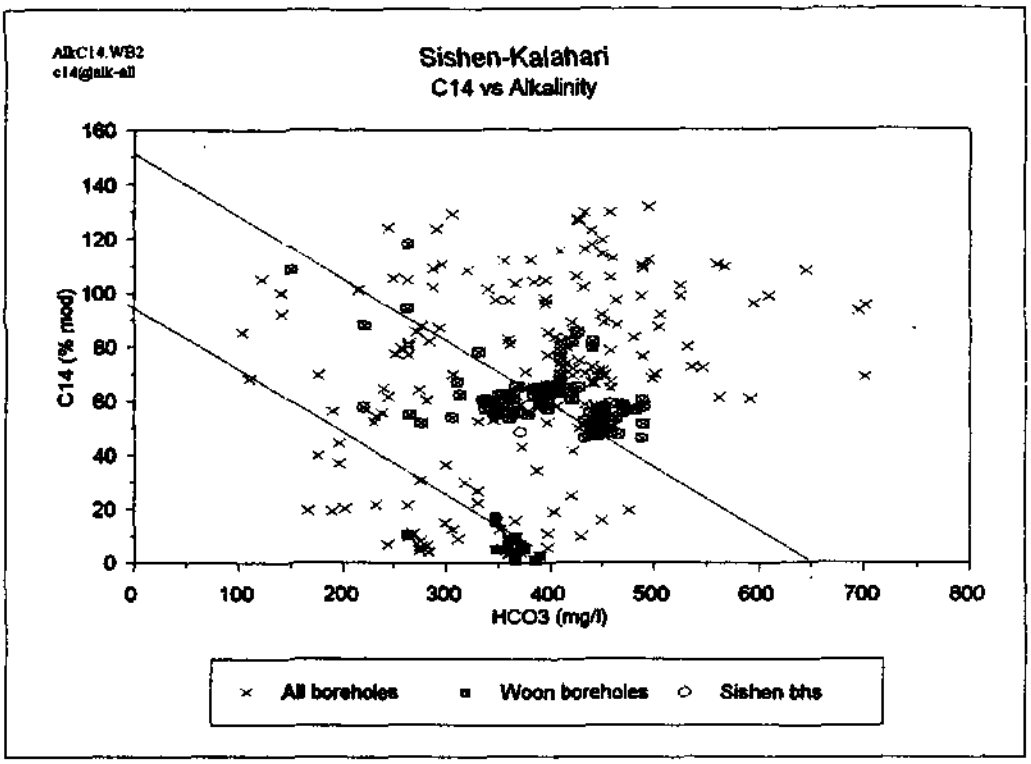


Fig. 12.4.3.2.4 ^{14}C concentrations in relation to alkalinity for groundwaters in the Sishen area.

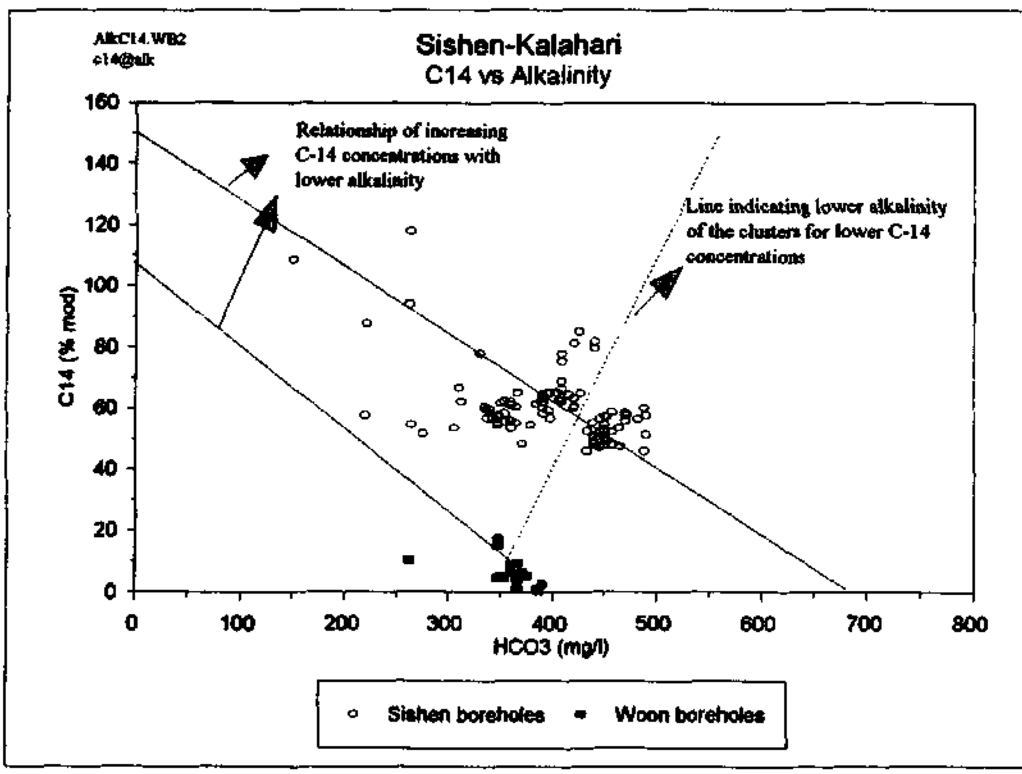


Fig. 12.4.3.2.5 Relationship between ^{14}C and alkalinity of groundwater in the Sishen mining area.

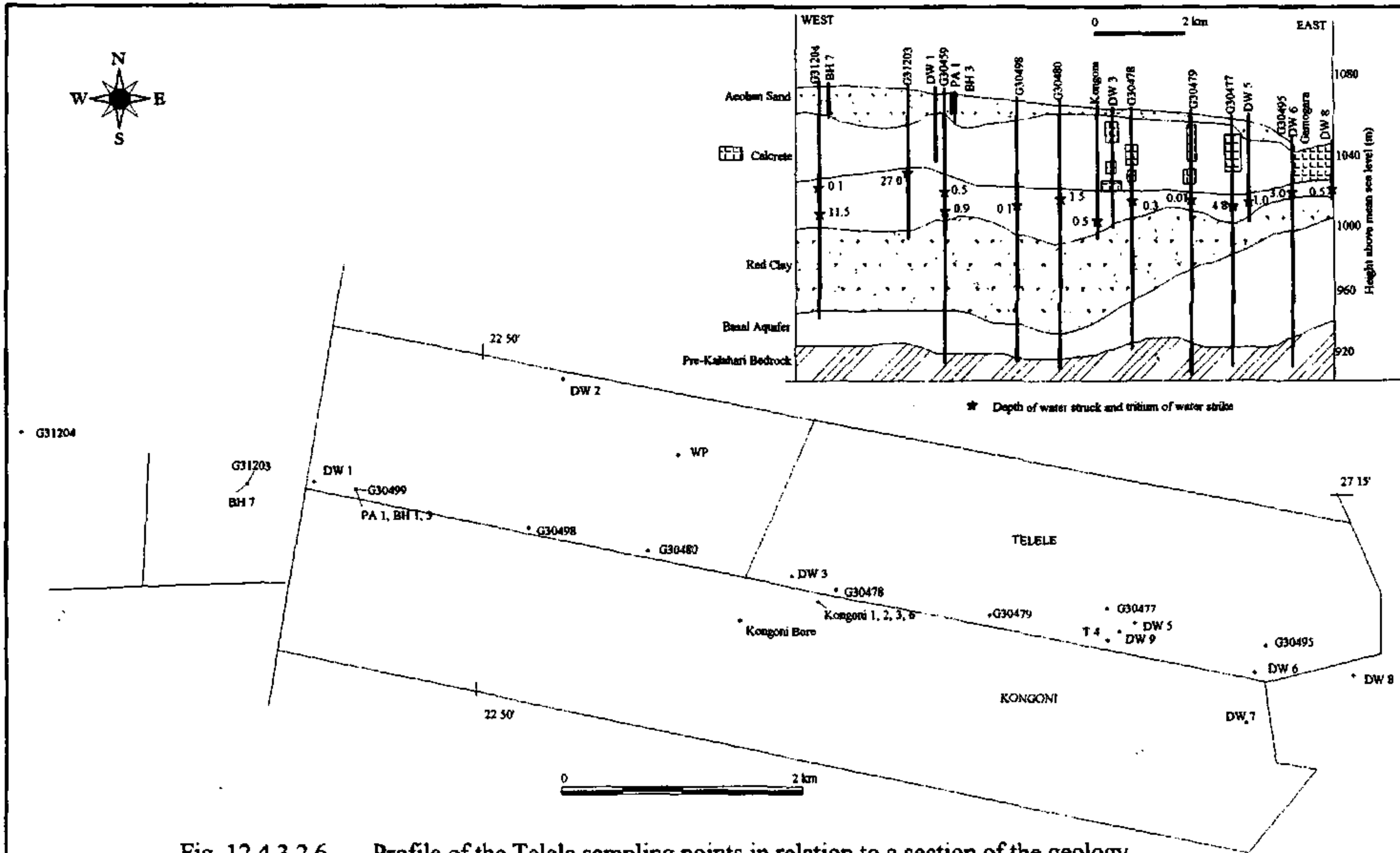


Fig. 12.4.3.2.6 Profile of the Telele sampling points in relation to a section of the geology.

Telele.tcw

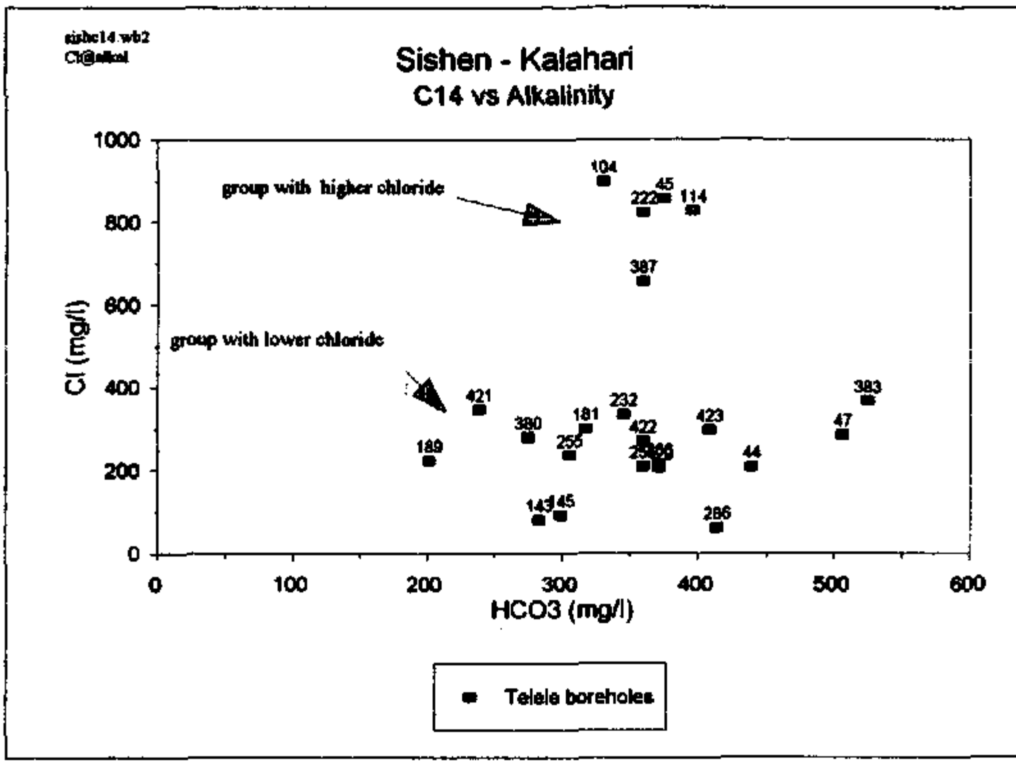


Fig. 12.4.3.2.7 Chloride concentrations in relation to alkalinity (HCO_3^-) indicating two groupings.

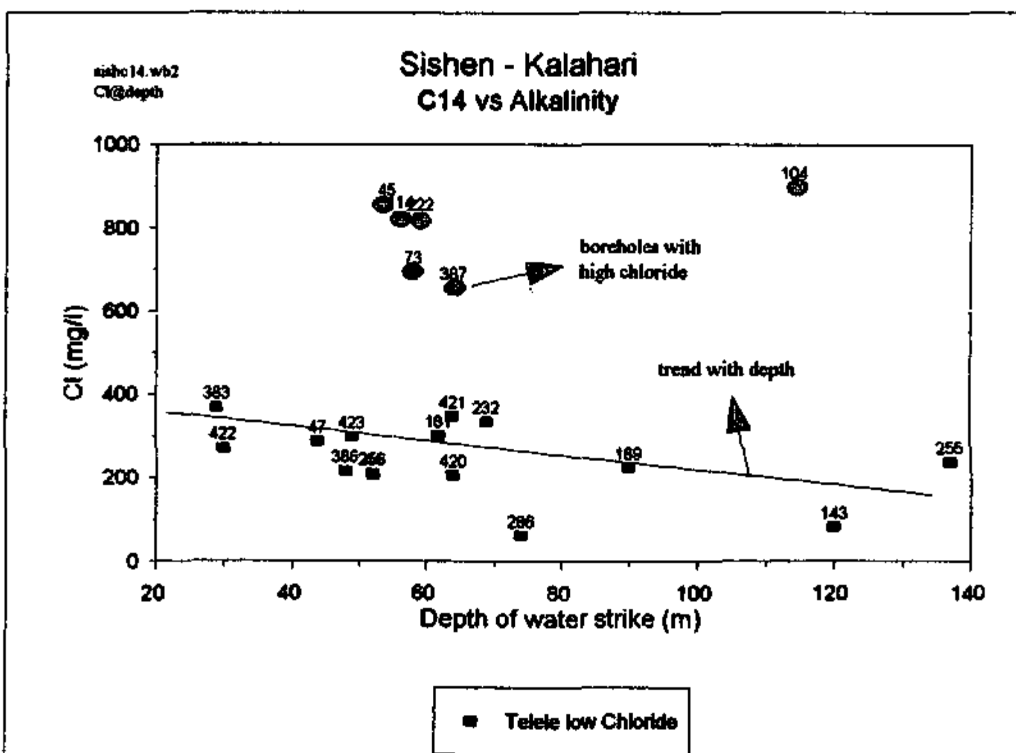


Fig. 12.4.3.2.8 Chloride concentrations relative to the depth of water strikes.

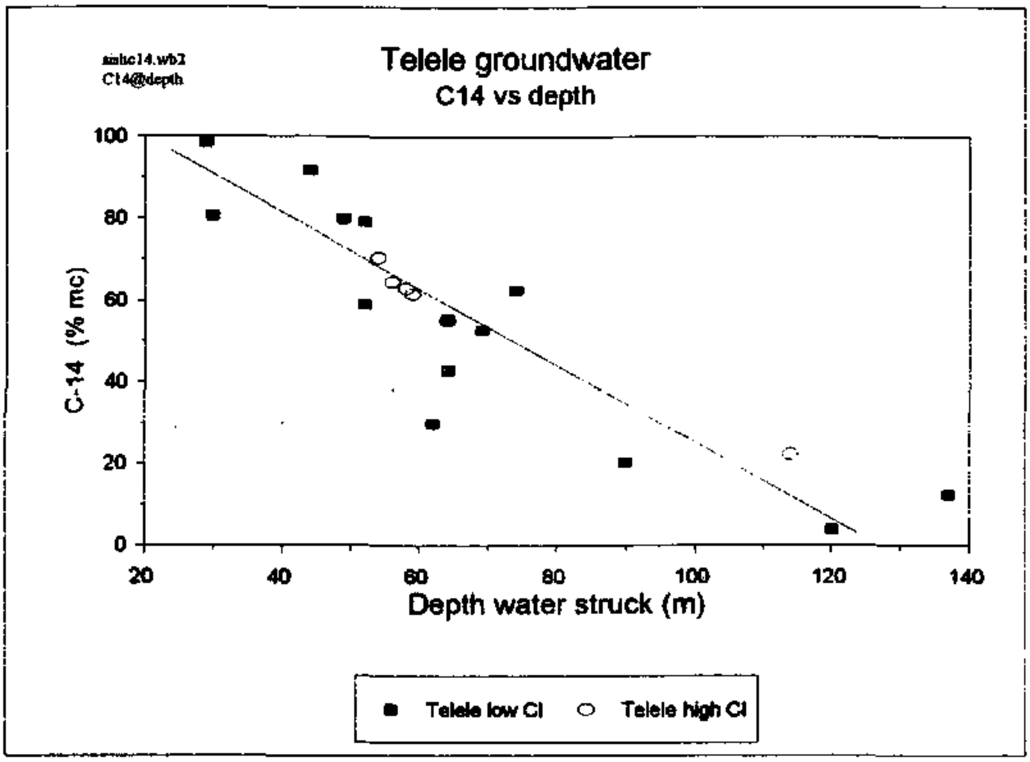


Fig. 12.4.3.2.9 ^{14}C concentrations in relation to the depth of water strikes.

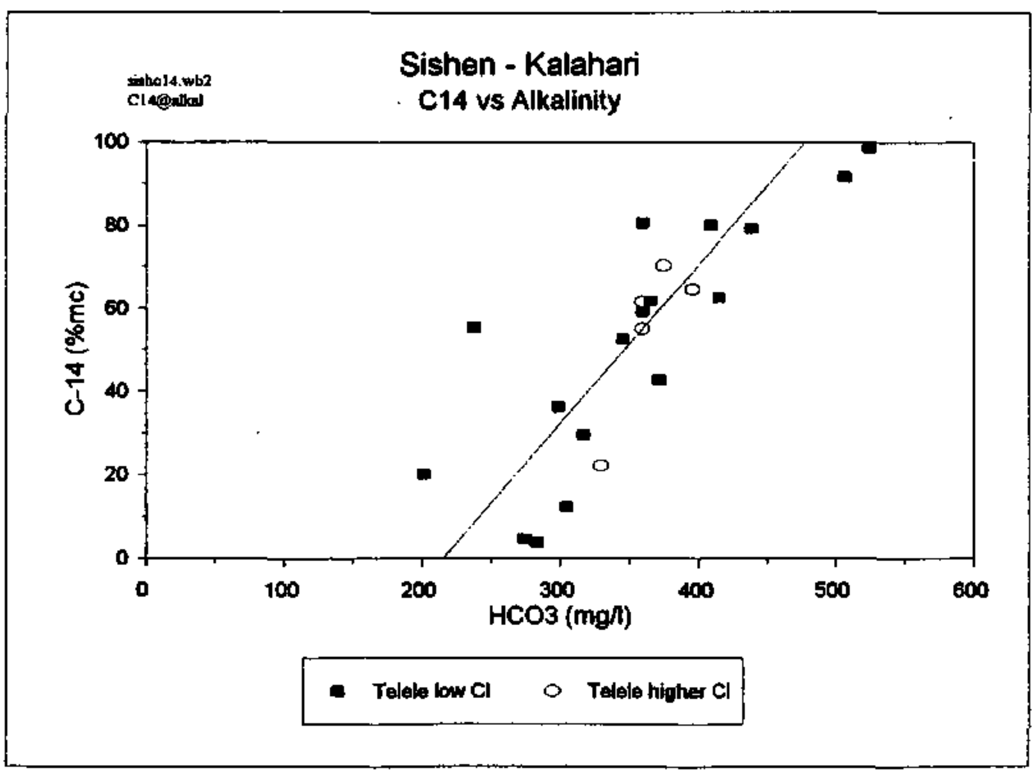


Fig. 12.4.3.2.10 ^{14}C concentrations of all Telele samples relative to the alkalinity of the water samples.

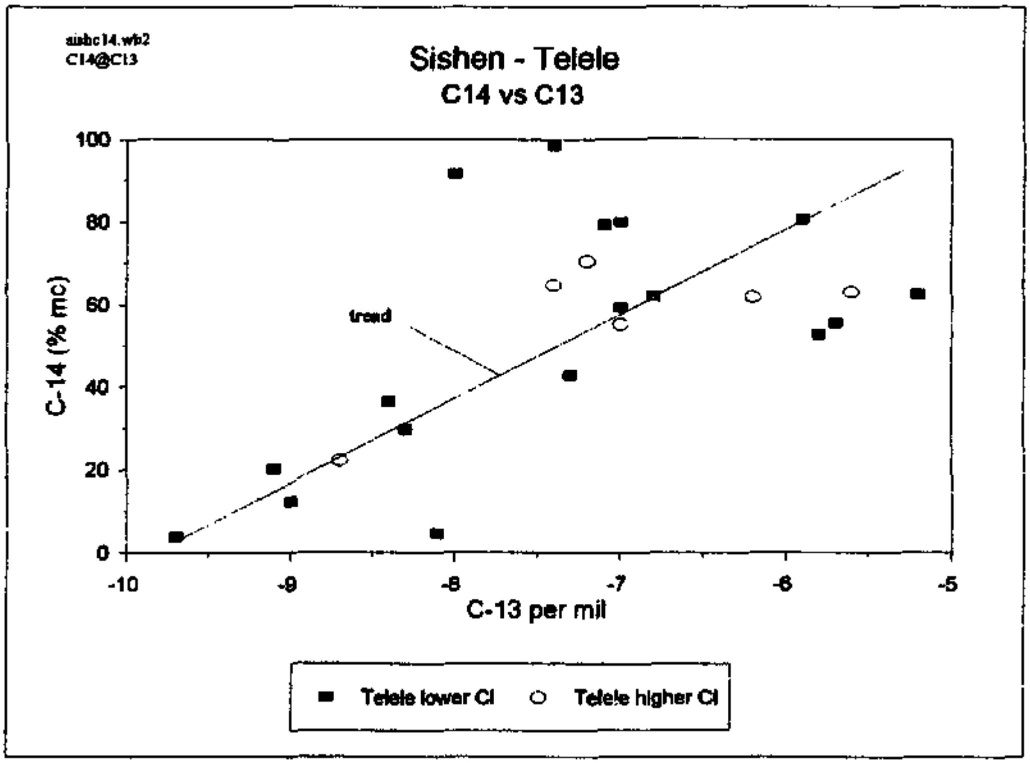


Fig. 12.4.3.3.1 ^{14}C vs ^{13}C relationship of groundwater in the Telele Profile area.

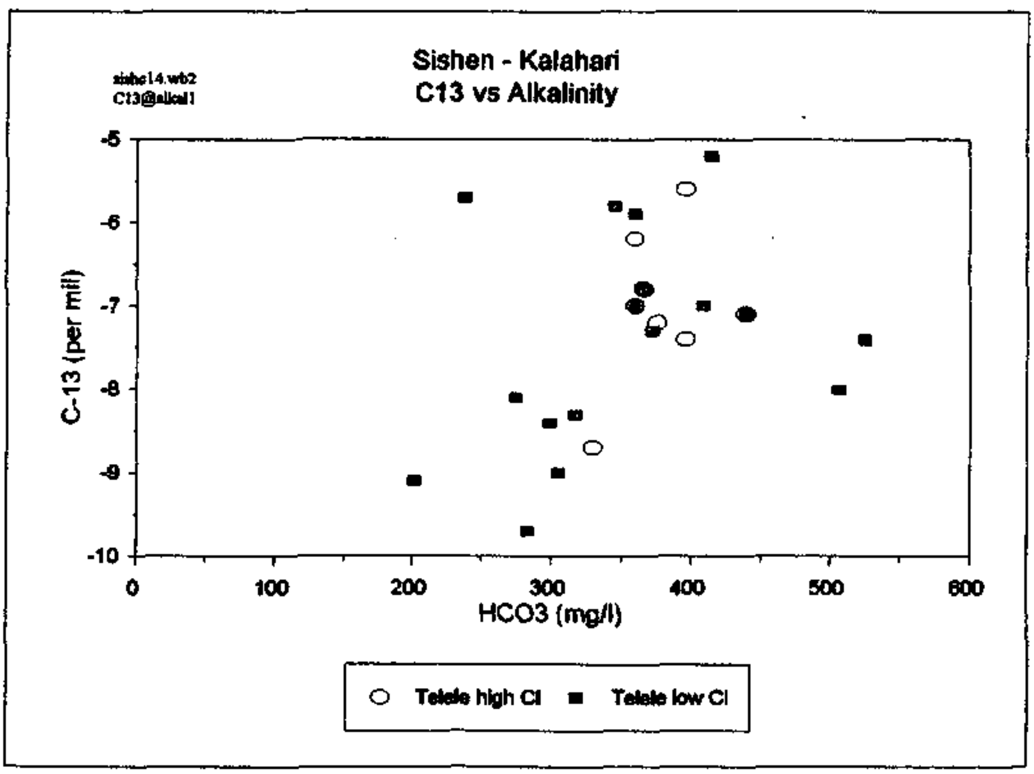


Fig. 12.4.3.3.2 ^{13}C concentrations compared to alkalinity measurements of groundwater in the Telele area.

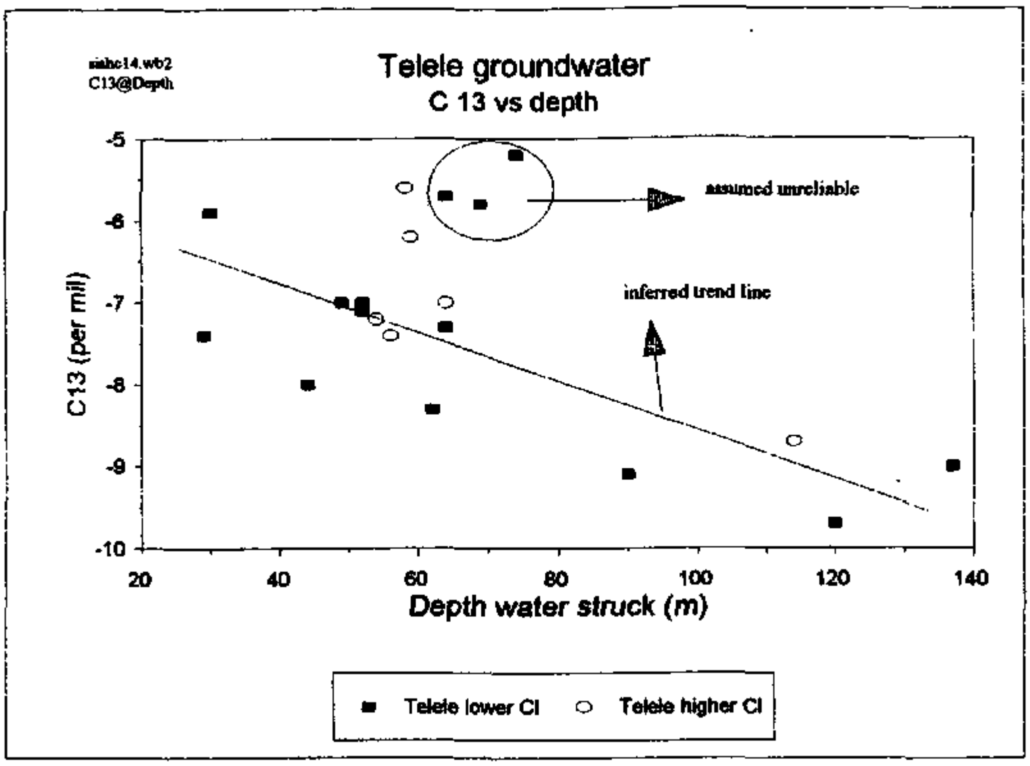


Fig. 12.4.3.3.3 ^{13}C concentrations in relation to the depth of water strikes.

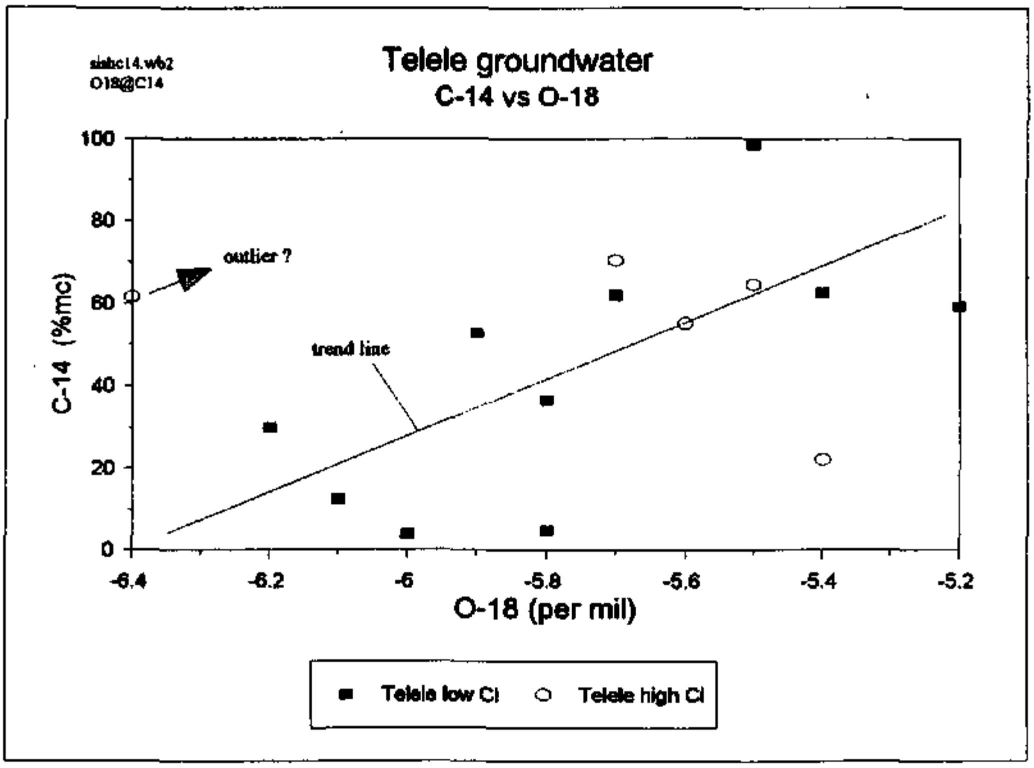


Fig. 12.4.3.3.4 Comparison of ^{18}O and ^{14}C concentrations with depth of water strikes.

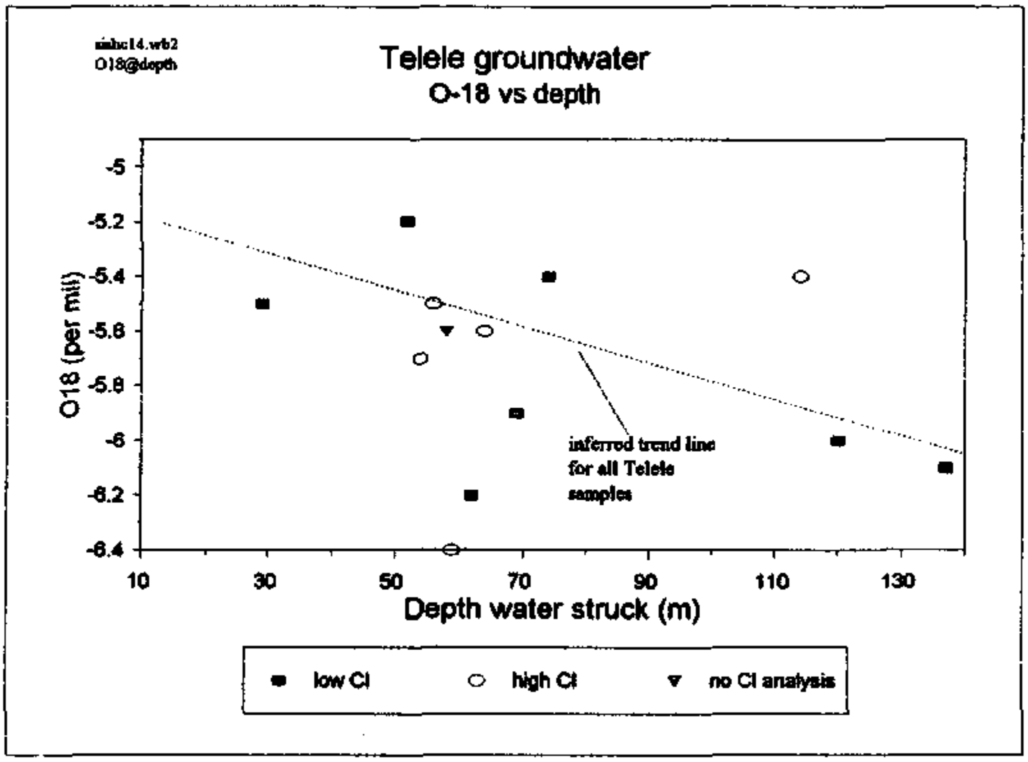


Fig. 12.4.3.3.5 ^{18}O concentrations in relation to the depth of water strikes.

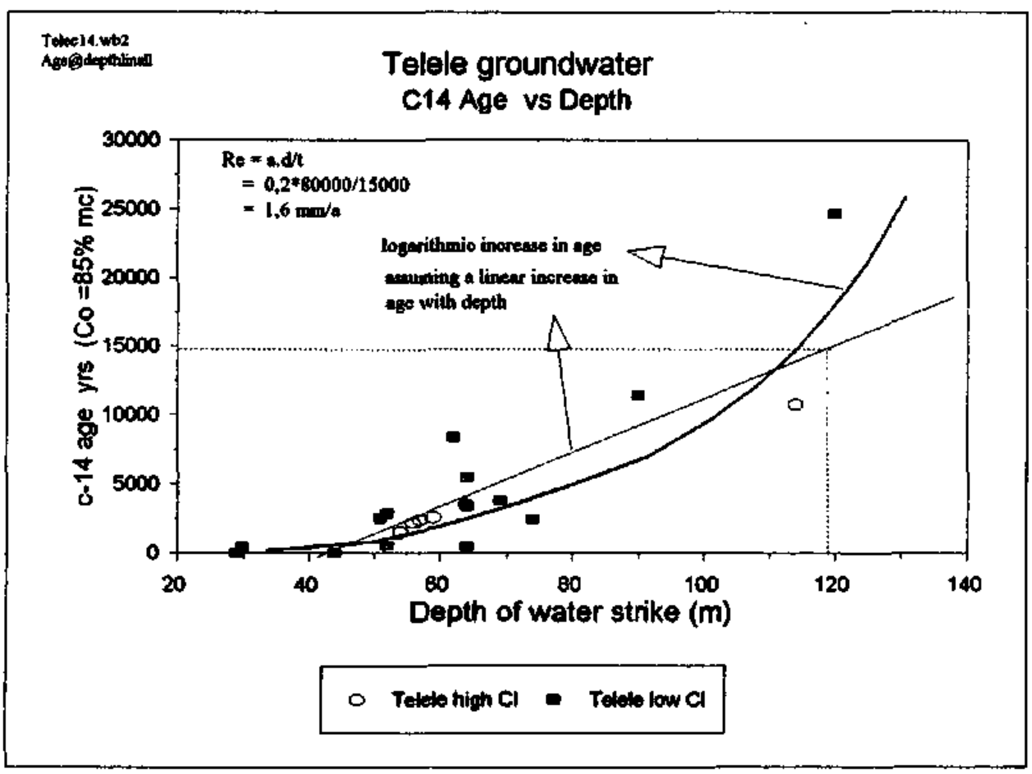


Fig. 12.5.1.1 ^{14}C ages of groundwater in relation to the depth of water strikes used to derive recharge from the linear increase in age with depth.

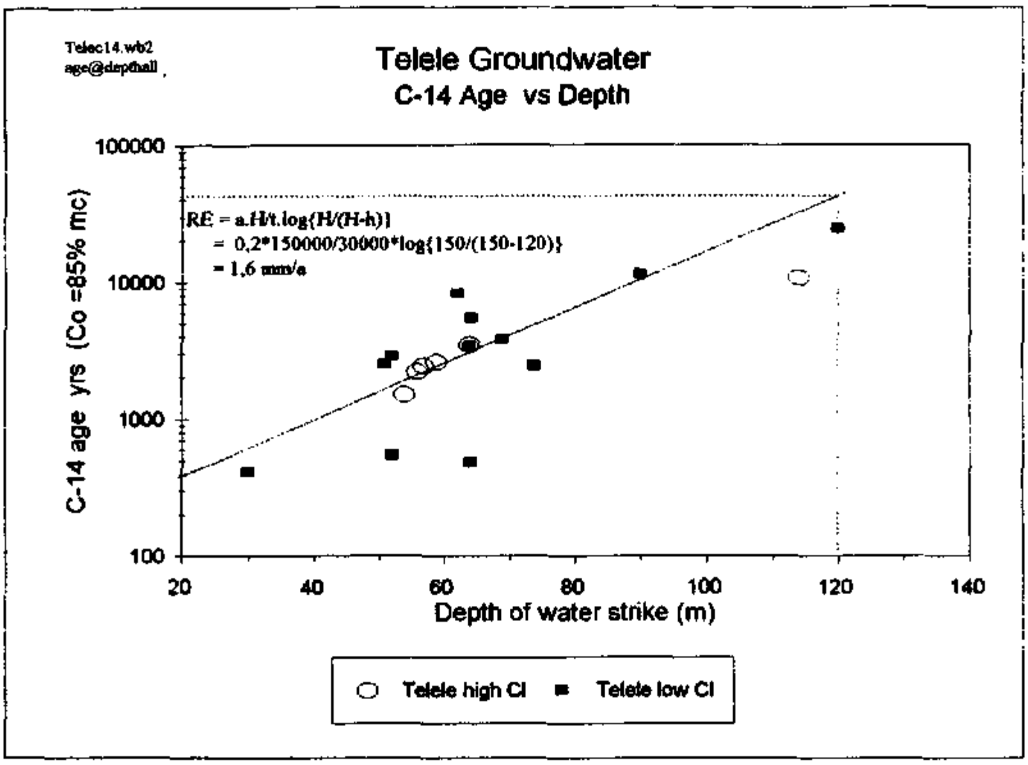


Fig. 12.5.1.2 ^{14}C ages of groundwater in relation to the depth of water strikes used to derive recharge from the increase in age with depth.

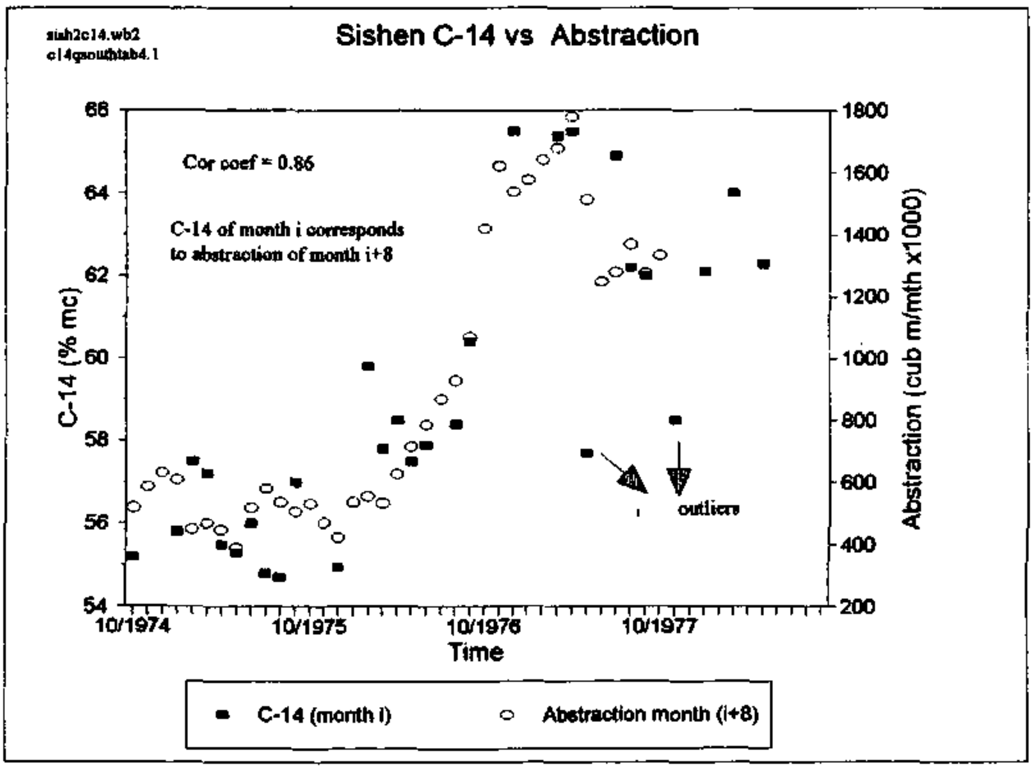


Fig. 12.6.2.1 ^{14}C concentrations of groundwater in the Sishen mining area in relation to the rate of abstraction.

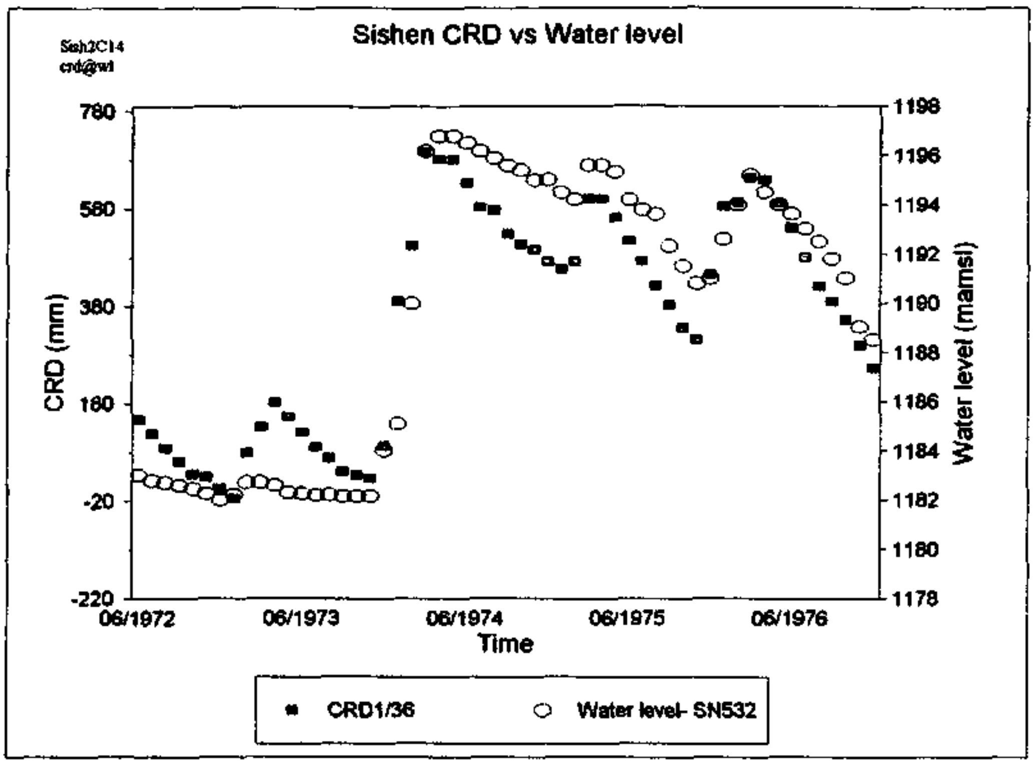


Fig. 12.6.2.2 Rise in groundwater levels in the Sishen area in relation to the Cumulative Rainfall Departure series, with no lag.

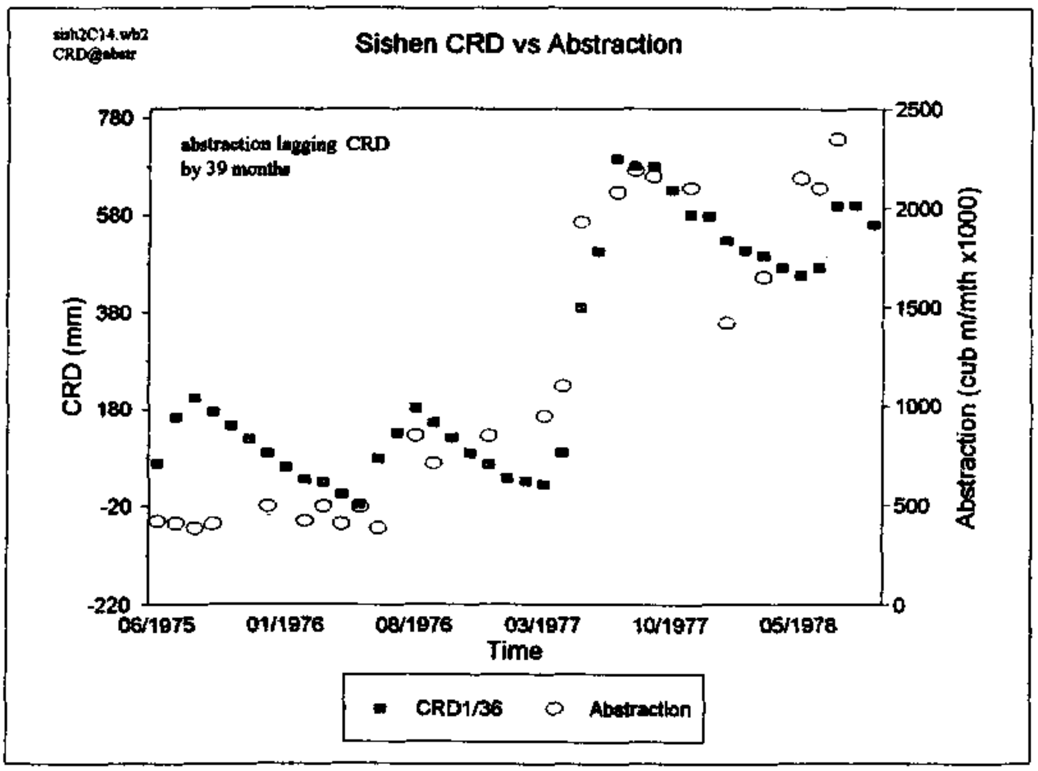


Fig. 12.6.2.3 Rise in groundwater levels in relation to the increased abstraction required lowering the groundwater levels in the Sishen mining area.

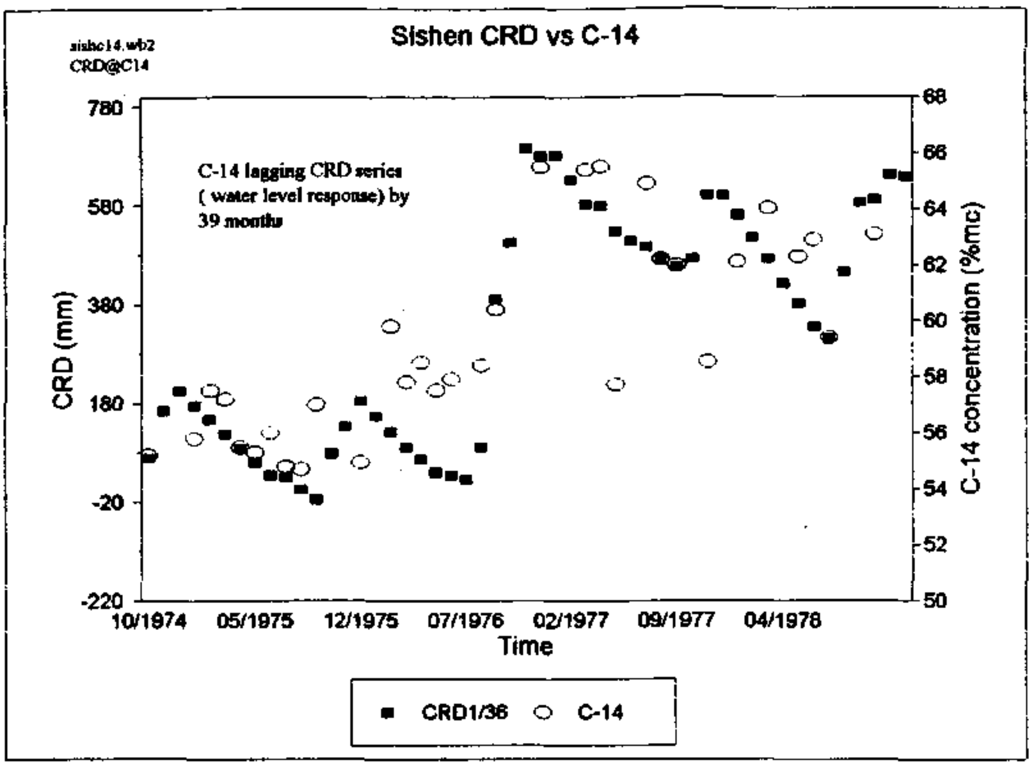


Fig. 12.6.2.4 ^{14}C concentration in the pumped water of the Sishen mine in relation to the CRD series representing the rise in groundwater levels.

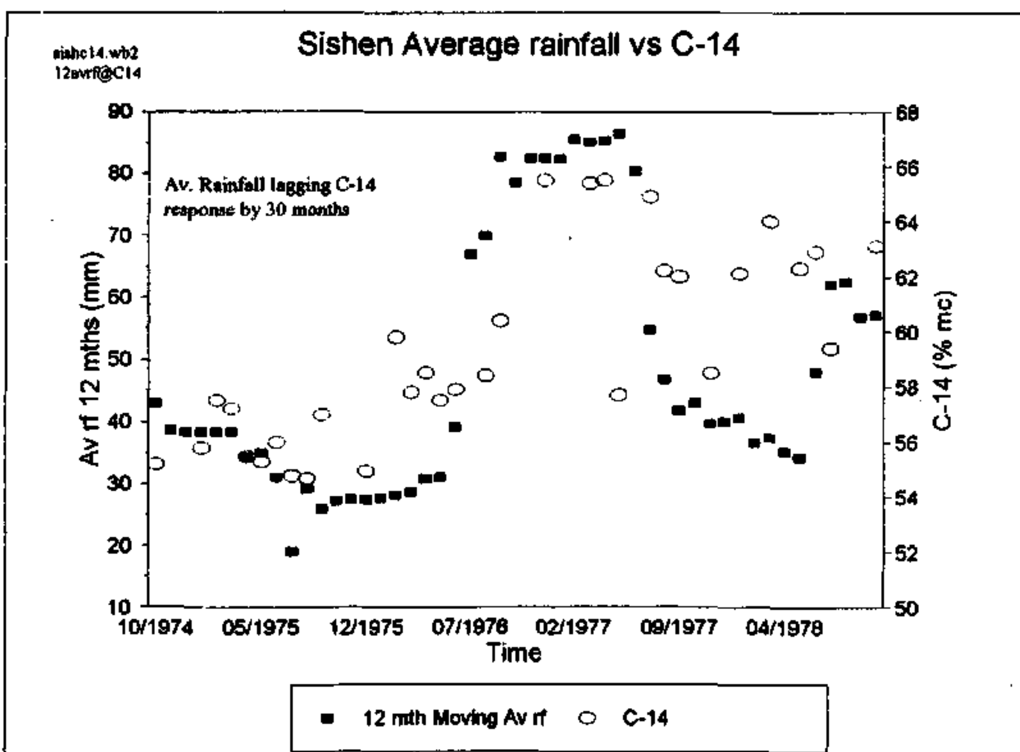


Fig. 12.6.2.5 ^{14}C concentrations in relation to the average rainfall over the preceding 12 months which also corresponds to the rise in groundwater levels.

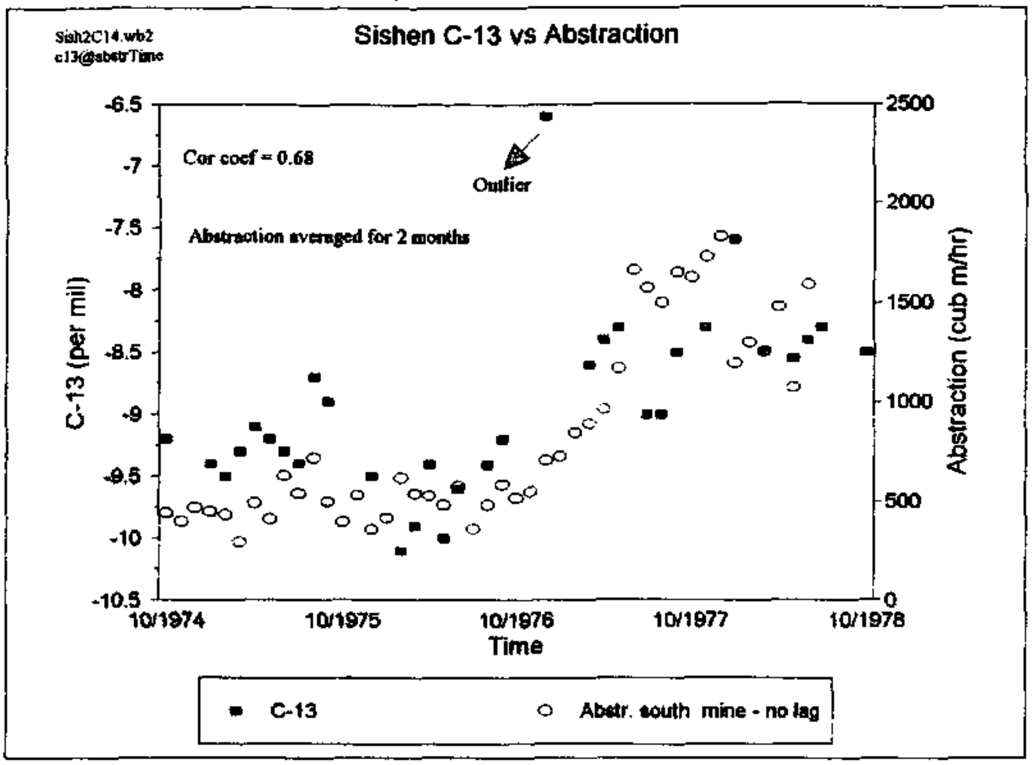


Fig. 12.6.2.6 ^{13}C concentrations in relation to alkalinity of groundwater in the Sishen mining area.

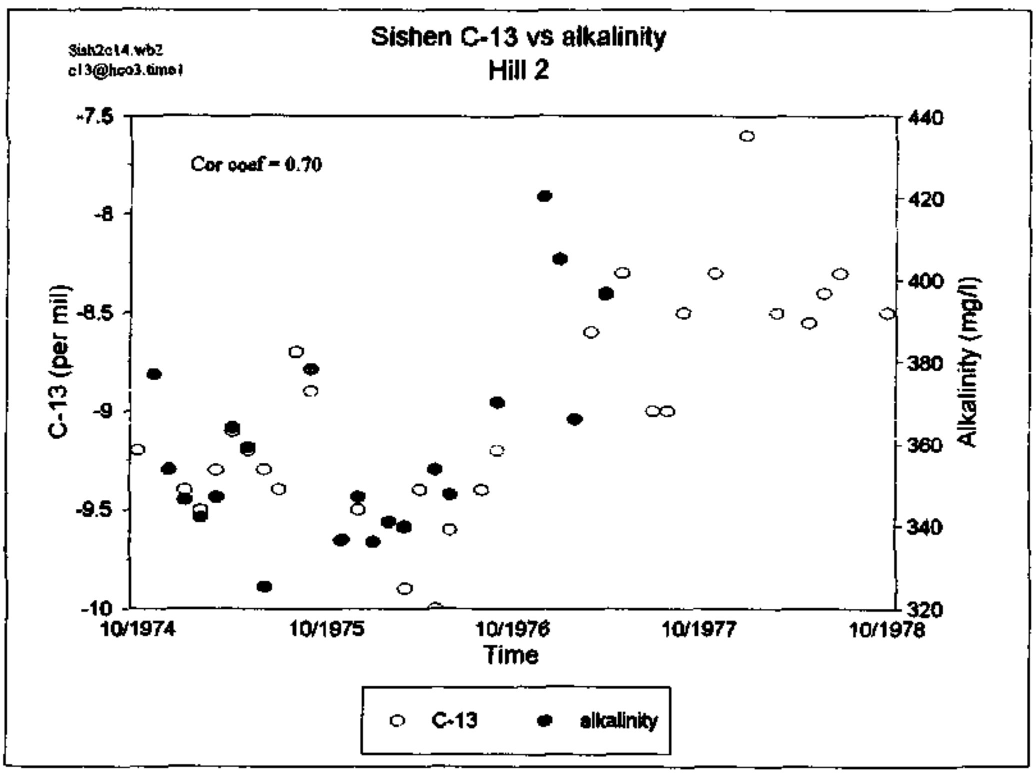


Fig. 12.6.2.7 Alkalinity of groundwater in relation to ^{13}C concentration in the Sishen mining area.

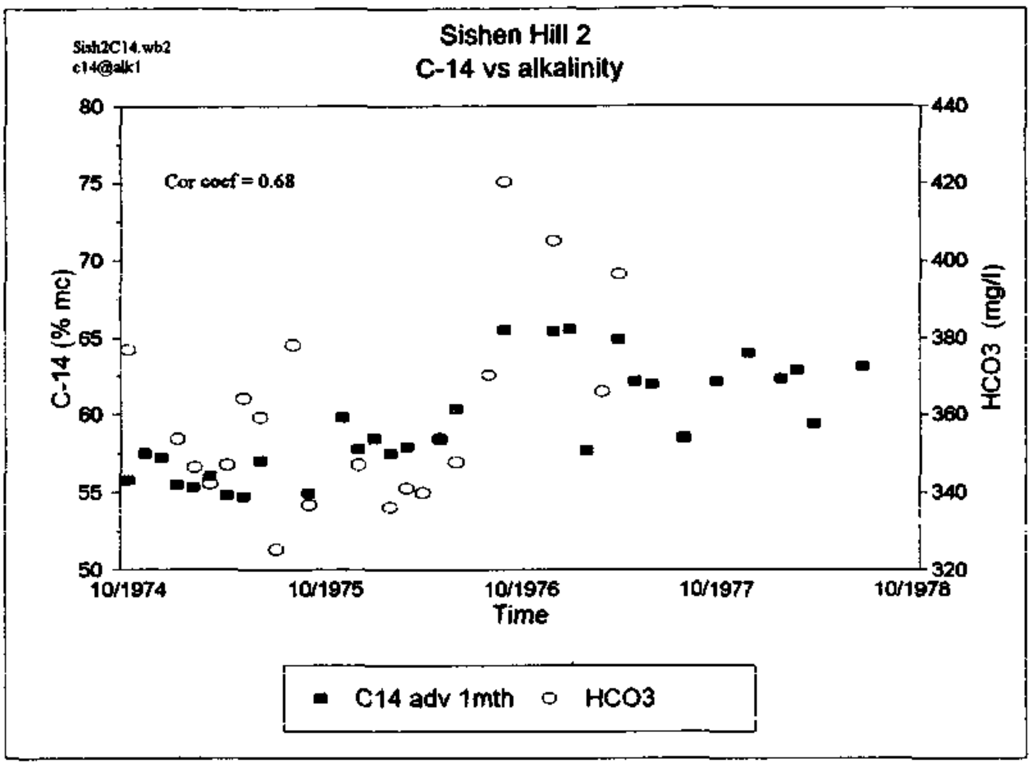


Fig. 12.6.2.8 ^{14}C concentration in the Sishen mining area in relation to alkalinity of the groundwater.

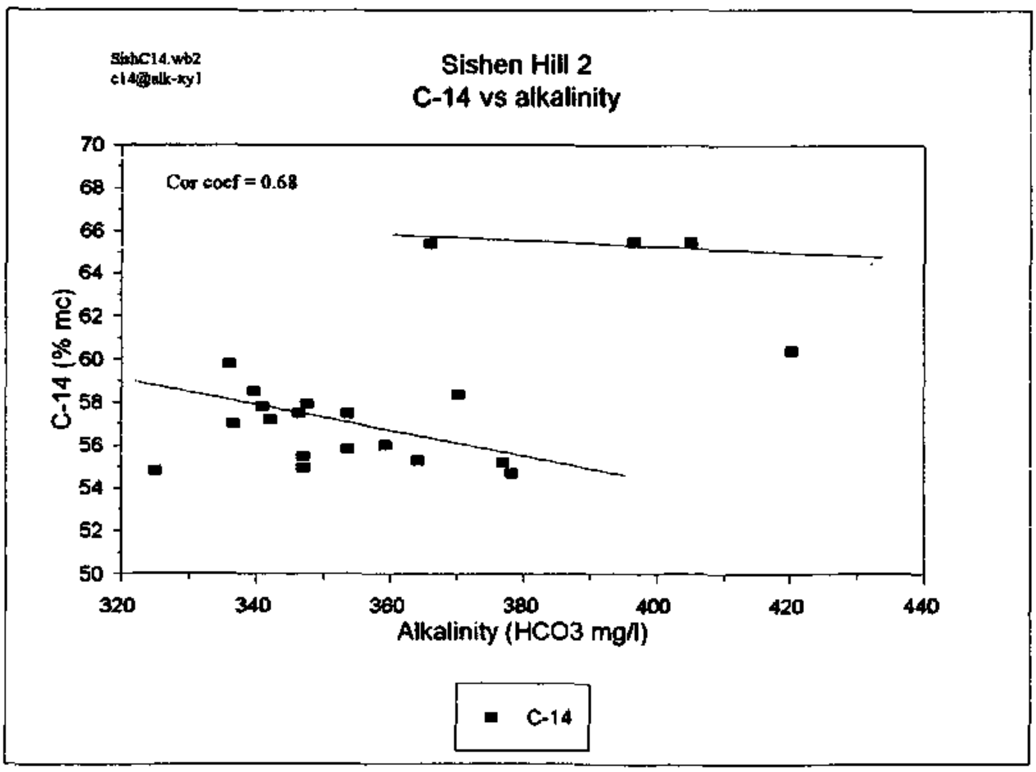


Fig. 12.6.2.9 Graph of ^{14}C concentrations against alkalinity for Sishen Hill 2 samples.

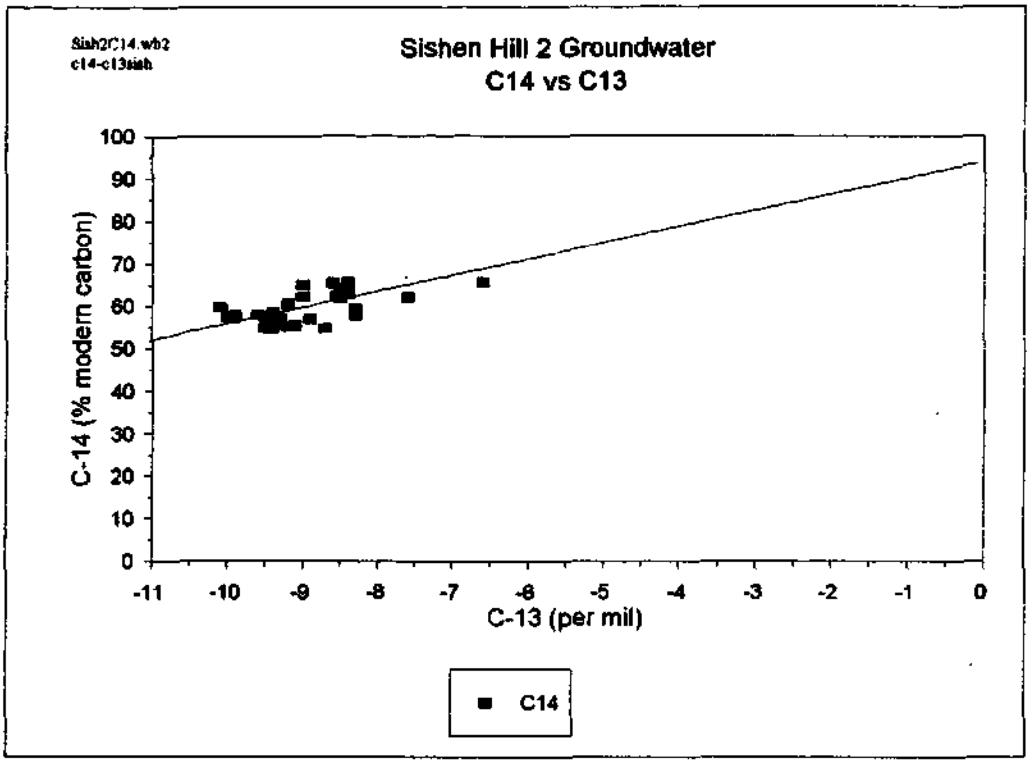


Fig. 12.6.2.10 Relationship between ^{14}C and ^{13}C concentrations.

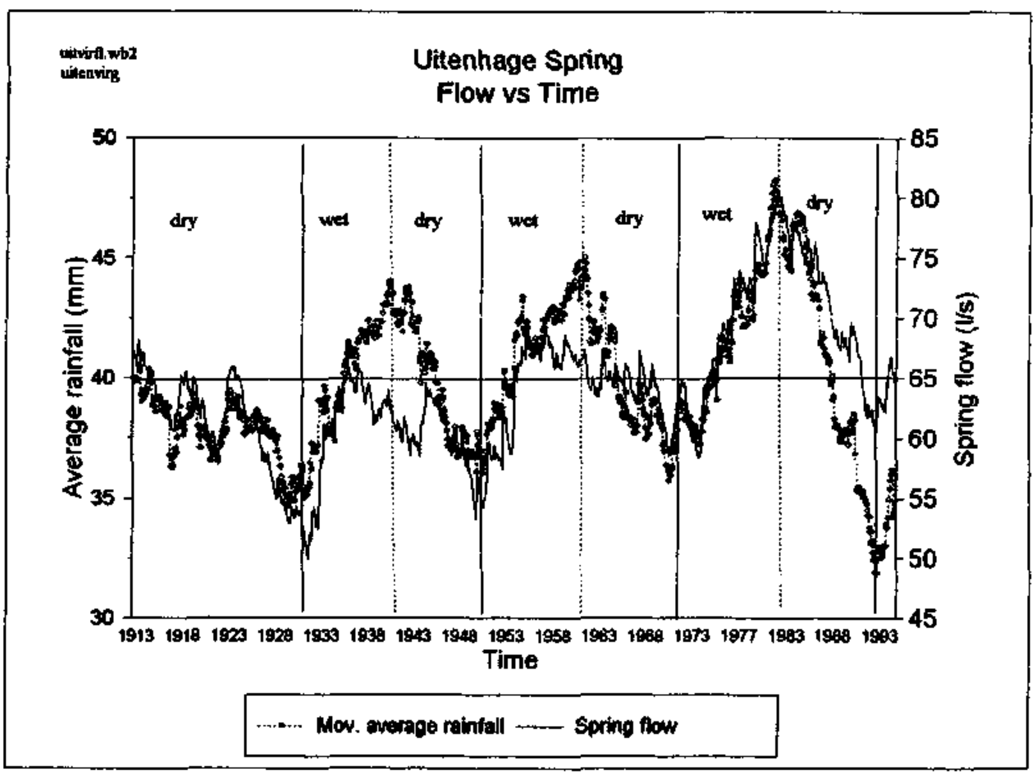


Fig. 13.1.1 Inferred natural flow of the Uitenhage springs as an indication of climate change on hydrological components.

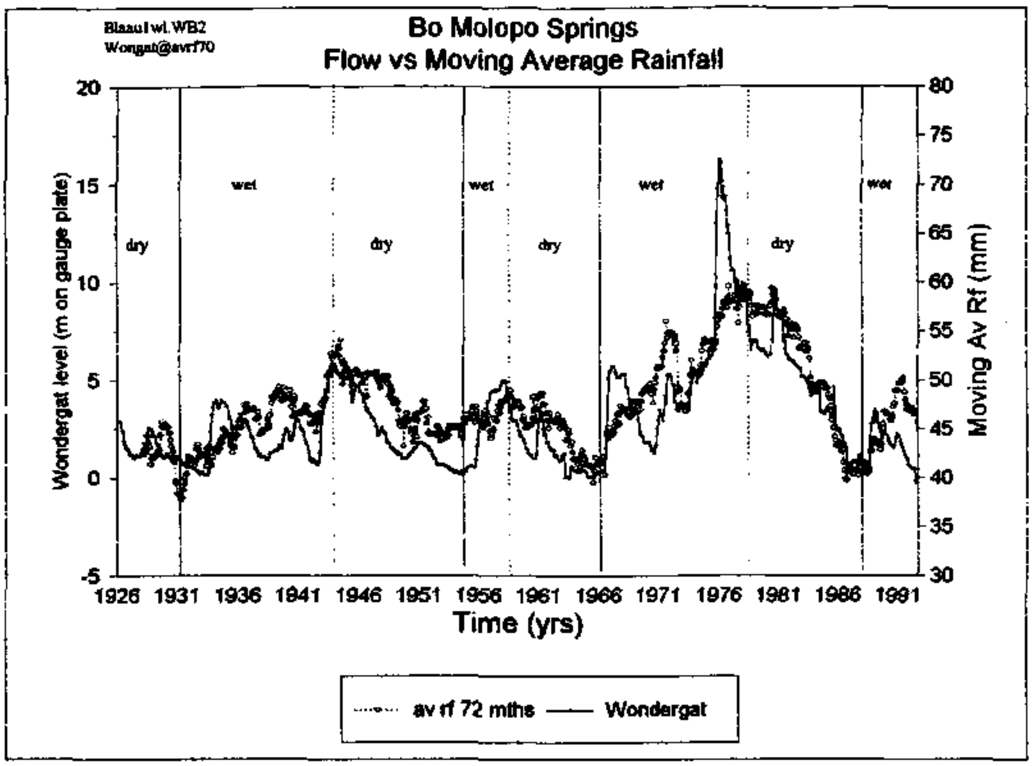


Fig. 13.1.2 Inferred natural water levels of the Wondergat sinkhole in relation to the moving average rainfall over 72 months.

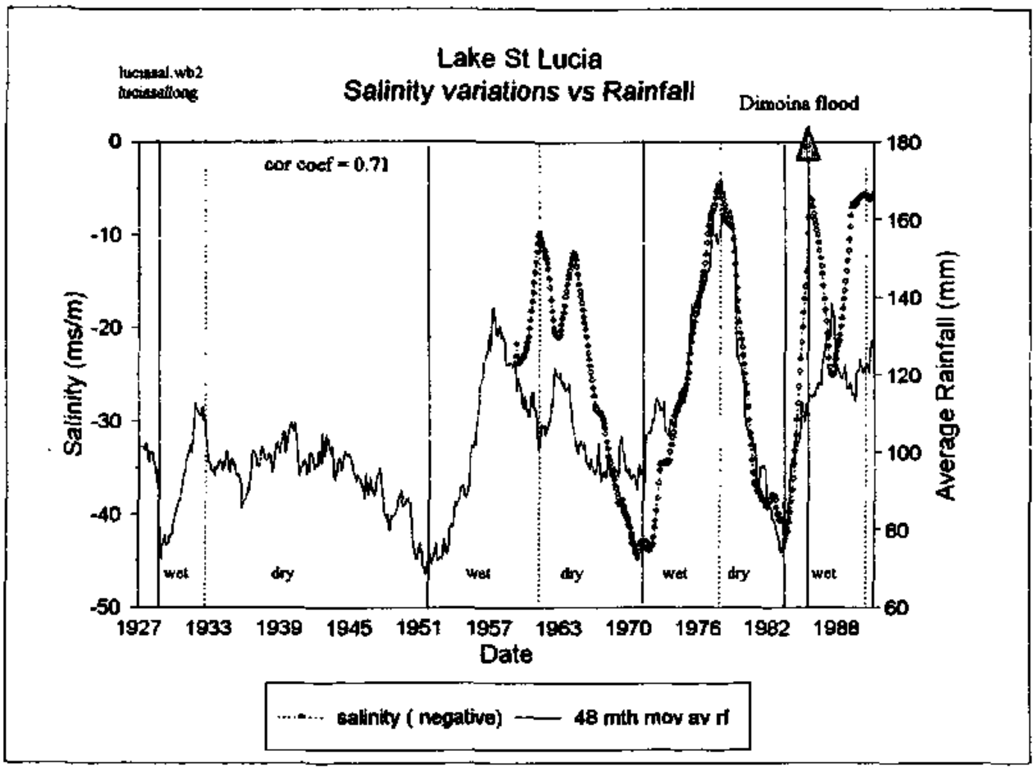


Fig. 13.1.3 Salinity variations of Lake St Lucia measured at Charter's Creek in relation to the moving average rainfall over 48 months.

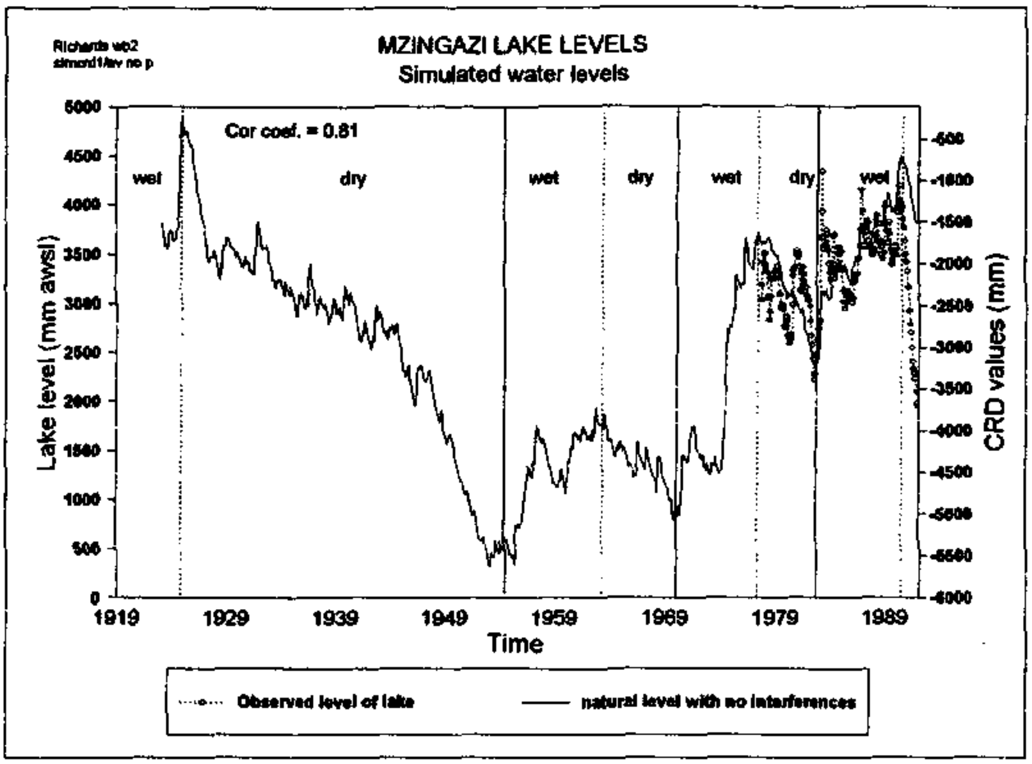


Fig. 13.1.4 Variation of the water level of Lake Mzingazi.

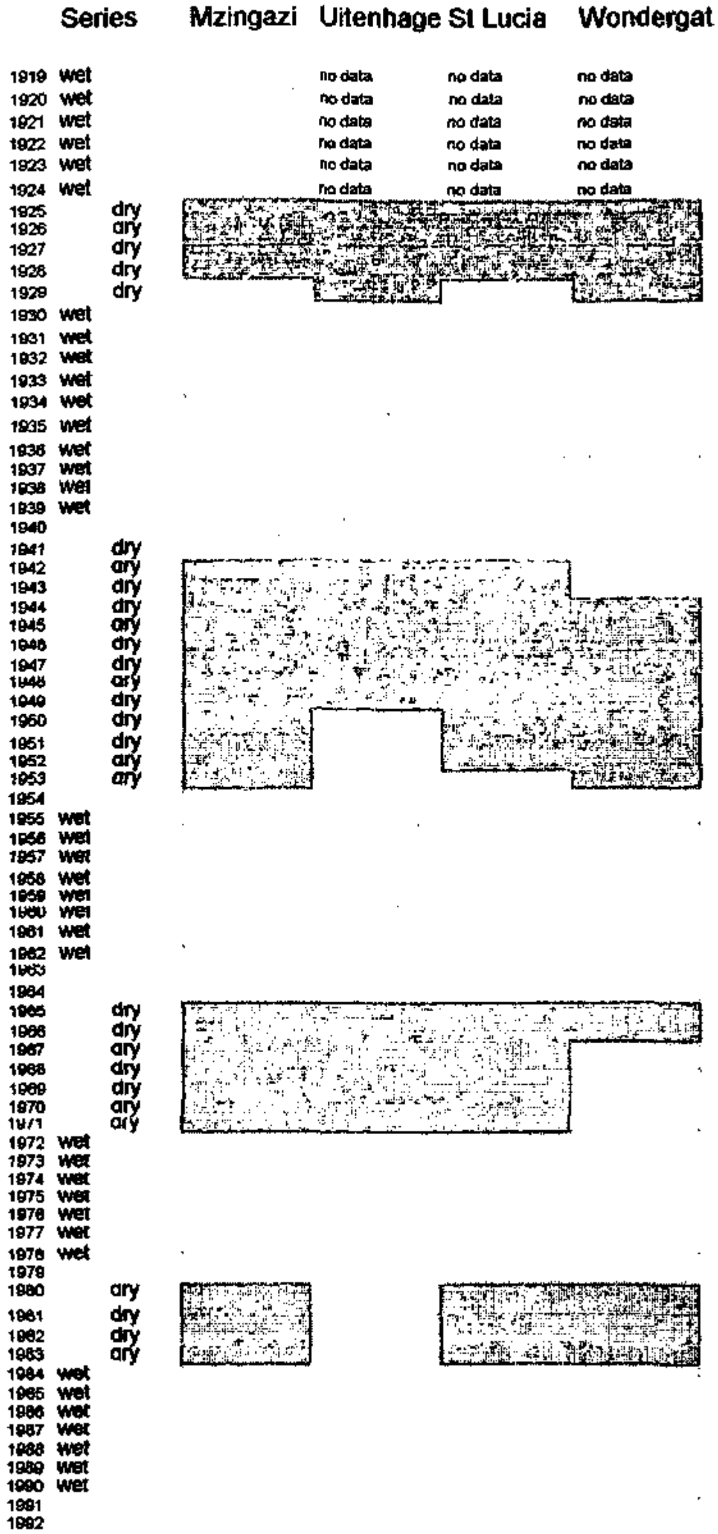
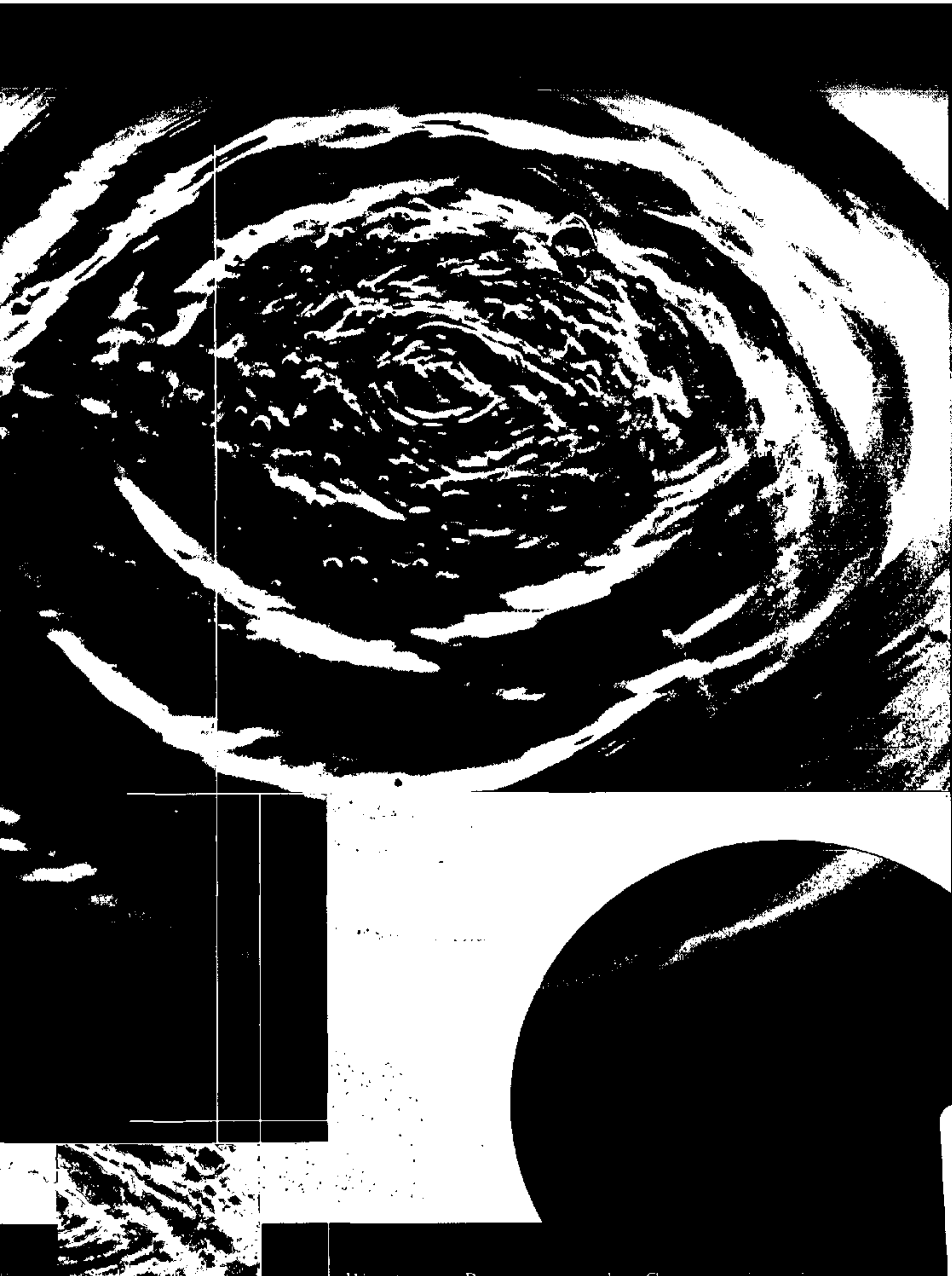


Fig. 13.2.2. **Presentation of major wet and dry cycles.**



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