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STUDIES

EXPLOITATION POTENTIAL OF KAROO AQUIFERS

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"The Exploitation Potential of Karoo Aquifers."

The Steering Committee responsible for this project consisted of the following persons:

Dr P J T Roberts	Water Research Commission (Chairman)
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Numerous other people and institutions have contributed significantly towards the completion of this project. The authors take this opportunity to thank all of them. Especially we want to thank:

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FOREWORD

The draft version of this report has been distributed amongst the members of the Steering Committee early in 1989.

Based on comments obtained, Chapter 5 was added, which gives an example how the utilization of an aquifer can be improved after the exploitation potential is determined. Checking of data used, recalculations with additional data for the alluvium at De Aar, preparation of a number of ground-water contour maps for both study areas at various times and the always necessary editorial work have deferred the publication, but have also improved the report.

It appears at this stage that *Saturated Volume Fluctuation Method* (SVF) is the major accomplishment achieved during this project. This method allows not only recharge determination, but also the assessment of the mean storativity. In secondary aquifers, a reliable value for the latter could often not be obtained from pumping tests; a problem which hampered the proper evaluation and management of aquifers in this country for a considerable period of time.

Fortunately, the delay of the publication has not affected the application of its findings. Already late in 1989, Bredenkamp, Van Rensburg, Van Tonder and Cogho presented a paper which was based on the SVF method at the International Symposium in Benidorm, Spain, on GROUNDWATER MANAGEMENT : Quantity and Quality. Last year, the Journal of Hydrology published a paper by Van Tonder and Kirchner, the authors of this report, describing the SVF method. The method also forms part of a new WRC project which is concerned with the preparation of a manual on quantitative estimation of ground-water recharge and aquifer storativity presently undertaken by Bredenkamp, Van Tonder and Van Rensburg.

It is the sincere hope of the authors that the publication of the findings of the present report will contribute to the better understanding of recharge processes, the evaluation of the exploitation potential, the utilization of aquifers and their management.

Bloemfontein, April 1991

Symbols, Units and Abbreviations

A	area
a,b,c,d	regression constants
Alat	latitude in degrees
C	concentration in ground water
C.V.	coefficient of correlation
Cl	percentage clay
D	deposition of chloride
D	number of days in month (i)
DT	dry bulk density
E_a	actual evapotranspiration
ELEV	elevation above sea level [m]
FC	field capacity [kPa]
G	gama-gamma reading
GR	recharge [m^3/d] [mm]
I	lateral inflow [m^3/d]
i	gradient
$K(\theta)$	unsaturated hydraulic conductivity
$K_s(\Theta)$	saturated hydraulic conductivity
L	length
N	measured neutron count rate
N^*	N/N_s
N_s	neutron count rate in water
O	lateral outflow [m^3/d]
P	precipitation [mm]
Q	ground-water flow [m^3/d]
q	Darcian water flux
R	mean daily temperature range [°C]
R_{an}	mean temperature difference between hottest and coldest month of the year [°C]
R_o	run-off
S	specific yield
Si	percentage silt
T	transmissivity [m^2/d]
t	time
T_a	mean monthly air temperature [°C]
W	windspeed [m/s]
WP	wilting point [kPa]
ε	porosity
ϕ	total water potential = $z + \psi(\theta)$
Θ	reduced water content
θ	volumetric water content
θ_r	residual water content
θ_s	saturated water content
ρ_B	dry bulk density
ψ	matric potential
Δh	change in ground-water level
ΔS	change in storage [m^3]
Δt	time increment
GLF	Ground-water level fluctuation
SVF	Saturated volume fluctuation
SWB	Soil water balance
ZFP	Zero flux plane

CONTENTS

LIST OF FIGURES

LIST OF TABLES

EXECUTIVE SUMMARY

With the exception of the southern and eastern coastal areas, South Africa has a semi-arid climate where potential evaporation exceeds precipitation. The water supply in the densely populated and more industrialized parts of South Africa is, to a very large degree, derived from surface water resources. Increasingly expensive schemes are built to convey water from far away to these centres.

Smaller communities away from these centres cannot and need not afford such schemes; they often rely on ground-water resources. In fact, the water of most of the towns and villages in the Karoo is supplied by ground-water schemes. As the water demand in these communities rises these schemes have to be expanded or new ones must be built. The available ground water is, of course, limited and the question arises how large are the resources. If a method can be found which allows one to determine the exploitation potential of aquifers, these aquifers can be optimally used and future planning will become easier.

Because nearly half of the country away from the major perennial rivers is underlain by Karoo sediments, a research project into the *Exploitation potential of Karoo aquifers* was funded by the Water Research Commission. In this context, the exploitation potential was defined as follows:

> *The exploitation potential is the long-term yield of an aquifer which depends on recharge due to rainfall, underground in- and outflow and ground-water storage.*

The stated objective of the project was:

- **to develop an effective elementary method for the determination of the exploitation potential of Karoo aquifers, based on available data and a minimum of additional information still to be generated.**

The three-year project, preceded by a feasibility study, was undertaken by the Institute for Ground-water Studies at the University of the Orange Free State and actively supported by the Directorate of Geohydrology of the Department of Water Affairs. The project commenced in July, 1985. Because of the exceptional rains which fell in the summer months of 1988, the duration of the project was extended by six months to allow the rising ground-water levels and the associated recharge to be studied and to include the effects that such an event has on the ground water into this report.

In order to achieve the stated objective, the following research requirements were identified:

1) The quantitative determination of ground-water recharge as a function of rainfall, type and thickness of soil cover.

2) The determination of hydraulic parameter values and their variability with depth in order to establish to what degree production boreholes can be used in excess of the recharge during times of continuous droughts.

In developing a research program, it was realized that the following topics needed consideration:

i) Geology of the aquifer systems.

ii) Hydraulic parameters of the aquifer systems.

iii) Determination of the ground-water balance.

iv) Modelling of the behaviour of the two aquifers.

v) Development of a method for the calculation of the exploitation potential.

DATA ACQUISITION

With these topics in mind, the field activities were subdivided into (i) aquifer related investigations, (ii) investigations regarding the unsaturated zone and (iii) meteorological and related measurements.

AQUIFER INVESTIGATION

First, a geohydrological investigation of the two areas was carried out. This included a census of existing boreholes and various geophysical surveys.

Because the covering soil in the Karoo is in general rather clayey, it was assumed that recharge would preferably occur in areas where this cover was thin or missing. With this hypothesis in mind, a number of borehole sites were selected: (a) along profiles from the foot of the surrounding hills towards the centre of the areas to study the amount the water level rises after rainfall and its delay, (b) to provide for observation holes at sites where pumping and/or packer tests were envisaged and (c) to fill gaps in the observation network in order to have sufficient points for later modelling.

In a second step, the drilling of 40 holes selected in Dewetsdorp and 56 holes in De Aar, was carried out by the Subdirectorate Drilling Services of the Department of Water Affairs.

Thereafter multirate and constant rate pumping tests were carried out in both areas to determine the aquifer parameters. These tests were augmented by packer tests. Ten water-level recorders were installed in Dewetsdorp and 13 in De Aar. Apart from the water-level recorders there were 76 observation holes in Dewetsdorp and 80 in De Aar. Weekly abstraction data were available for Dewetsdorp. The De Aar Municipality has supplied the monthly totals for two borehole groups.

UNSATURATED ZONE

A soil survey was carried out in both areas to identify the different soils present and to determine their structure and texture. With the soil maps as a basis, twenty and twenty-two neutron probe access tubes were installed in Dewetsdorp and De Aar respectively in an attempt to quantify soil water movement by means of neutron probe measurements. During the course of the project neutron probe measurements were taken at approximately 3-month intervals and more frequently after major rainfall events.

METEOROLOGICAL DATA

One weather station in each area registered rainfall, radiation, wind velocity, humidity and temperature and allowed the determination of the potential evaporation. Furthermore two rainfall recorders and one rain gauge were installed in Dewetsdorp; and three rainfall recorders and two rain gauges in De Aar which registered rainfall. A rainfall sampler was also installed at each weather station. Rainfall samples from these samplers taken during the first good rains in 1985 and during the exceptional rains of 1988 were analysed for stable isotopes.

RESEARCH APPROACH

In any aquifer system, the following equation applies for any specified period:

Ground-water recharge = Ground-water loss + Additional Storage

where the loss can be defined as:

loss = (subsurface outflow - subsurface inflow) + abstraction + evapotranspiration

Except for the abstraction, none of these terms can be measured directly. The Saturated Volume Fluctuation (SVF) Method and the Ground-water Level Fluctuation (GLF) Method use this approach to calculate a nett recharge, i.e. a recharge after evapotranspiration.

In another approach, the changes in the unsaturated zone are considered. Here, the following equation is used:

Ground-water recharge = Precipitation - Evapotranspiration - Run-off + Additional storage

which introduces the terms precipitation and run-off. The Soil Water Balance method is based on this approach.

RESULTS

NON-ACHIEVEMENTS

It was not possible to determine either the change of the storativity with depth nor the "thickness" of a fractured aquifer. If, however, water-level and abstraction records were available for a long enough time period the storativity could have been calculated with the SVF-method. Without knowledge on the "thickness" of an aquifer and the storativity , it cannot be determined to what degree the aquifer can be used in excess of recharge during droughts.

OBSERVATIONS

From rainfall observations, it can be deduced that:

- daily rainfall figures vary considerably between the different rainfall stations. The total annual precipitation, however, shows much less variation.

From the neutron gauge measurements in the unsaturated zone, it can be followed that in the areas investigated:

- noticeable recharge through matrix flow does occur rarely during normal rainy seasons. Only during exceptional rains and at a limited number of places does some soil water move through the unsaturated zone. This occurred more often in De Aar where the soil cover is thinner than in Dewetsdorp. Even with the exceptional rainfall in 1988, neutron gauge readings in the majority of the access tubes in both areas indicated no matrix flow. For areas underlain by Karoo mudstones and shales, matrix flow through the soil is not considered to contribute substantially to recharge during normal rainfall events.

From water-level observations it can be concluded that:

- almost immediate recharge occurs at the foot of the hill slopes where the soil is absent or has a coarse texture. There are indications that recharge occurs along preferred pathways: some boreholes with very low yields respond fast to rainfall and with high rising water levels, immediately followed by a recession to water levels similar to the level before the rainfall event. This observation can be explained by double porosity where the initial recharge takes place locally along larger joints, etc., followed by a subsequent redistribution of the water over a larger area at a slower rate through less permeable joints or the matrix pore space. It has also been observed that water levels respond with a considerable time lag in the more central parts of the aquifer and that the rise in water level in these areas is up to an order of magnitude less than in the higher lying parts.

- major recharge occurs only after exceptional rains which replenish the aquifers again. Water-level rises of more than 30 m have been observed, following the 1988 rains. It seems, however, that a second season with above average rainfall is needed after sufficient time for redistribution of the water has elapsed, i.e. in the following year or the year thereafter, to recharge all aquifers completely. This can possibly be explained by the time lag between rainfall and water-level rise in certain parts of the aquifer. These parts are recharged with water that is redistributed from other parts of the aquifer which react faster to rainfall. Additional recharge during a following season is therefore required to refill the latter.

From the pumping and packer tests carried out it is calculated that:

- the transmissivity of the Dewetsdorp aquifers varies between approximately 7 and 40 m^2/d, with a representative transmissivity value of about 10 to 13 m^2/d. The values for De Aar aquifers outside the alluvium vary between 20 and 200 m^2/d with a representative transmissivity value of about 46 m^2/d. The calculated storativities are generally in the order of 10^{-4} while a value of $3{,}5 \cdot 10^{-4}$ is regarded a representative storage value based on the pumping test results.

From the abstraction and water-level data, however, it must be concluded that:

- conventional pumping tests do not allow even an approximate determination of the storativity. This is attributed to the fact that fractured, inhomogenous and isotropic aquifers may not and cannot be analysed with the conventional methods.

From the chemical analyses it can be concluded that:

- ground water in Dewetsdorp is of suitable quality for human consumption, except for occasional higher fluoride concentrations. It undergoes changes due to ion exchange.
- ground water in De Aar is of suitable quality, although a few boreholes have elevated fluoride values. Ground water south of the Brak River differs slightly from the water found north of the river. Indications are that the water in one borehole (G 36499) which has a TDS concentration three times lower than the other samples, may originate outside the area of investigation.

From the stable isotope relation $\delta\ ^{18}O$ and $\delta\ ^{2}H$, it can be seen that:

- there is little difference between rain water from Dewetsdorp and De Aar. Ground water in De Aar contains slightly more deuterium than that from Dewetsdorp.
- To use natural isotope variations for quantitative rainfall/aquifer recharge investigations, requires a much higher sampling frequency of rainfall and ground water from boreholes that respond well and without too much of a time lag to rainfall.

MODELLING

From the results of the soil water budgeting models, it is concluded that:

- calculation of the recharge based on soil matrix flow does not give satisfactory answers. The probable reason is that in the Karoo environment investigated, the clay content is generally too high to allow such flow in noticeable quantities.

Ground-water balance calculations included the Saturated Volume Fluctuation- (SVF) and the Ground-water Level Fluctuation- (GLF) methods. From SVF calculations:

- the nett recharge can be determined, i.e. the recharge that remains after the ground water taken up for evapotranspiration has been used. This approach obviates the difficult determination of the actual evapotranspiration.

  The calculated recharge depends on the type of cover, i.e. shallow soil, deep soil or alluvium. It is also dependent on the annual precipitation.

  For the calculations, a mesh must be generated for each type. A program for this purpose (TFEM), which requires as input the coordinates of all nodes, is provided.

  A second program (SVF) which calculates the change in saturated volume requires as input the monthly rest water levels and abstractions at the respective observation points, as well as transmissivities at points where subsurface in- or outflow occurs and the applicable annual rainfall. According to the acceptable probability of failure of a scheme, the annual rainfall for the respective recurrence period must be chosen. Optionally storativities may be entered.

A minimum value of S can be calculated for periods of no recharge and preferably zero outflow. Because S / T is constant under certain conditions (Equation (3.4.10)), they can only vary within limits. From the results of the ground-water balance calculations for both research areas, it is concluded that the determination of the transmissivity through pumping tests yields acceptable values, while the determined storativity is approximately an order of magnitude too small.

Based on time intervals of 1, 3, 6 and 12 months at the beginning and the end of which the saturated volumes were equal, the recharge for Dewetsdorp can be approximated by the following equation:

Recharge = 0,039 (precipitation - 17) [mm]

For De Aar, there are more marked differences between a northern and southern Karoo- and an alluvial aquifer with recharge percentages of 1,4; 2,8 and 14,4 respectively for an average rainfall year with 287 mm/a. The calculated recharge for the whole aquifer is

Recharge = 0,048 (precipitation - 27) [mm]

or 4,3 per cent during the average rainfall year.

From the sensitivity analysis carried out for Dewetsdorp, it can be concluded that:

- the number of observation points do affect the results. The recharge calculations were repeated with 15 and 6 boreholes instead of the original 72 holes. This resulted in ± 33 per cent smaller recharge values.

From an example of the GLF method, it is deduced that:

- with known storativities, results similar to the SVF method are obtained. Recharge for Dewetsdorp was calculated as:

  Recharge = 0,035 (precipitation -181) [mm/a]

  The corresponding equation for De Aar is:

  Recharge = 0,048 (precipitation -100) [mm/a]

  with a mean recharge for the study period calculated as 1,8 per cent. For the average annual rainfall year, the recharge is 3,1 per cent which is less than the result of the SVF method.

  The GLF method is regarded less exact than the SVF method.

From inverse modelling (AQUAMOD), it can be deduced that:

- storage values determined by pumping tests are too low. Good agreement between observed and calculated water levels were obtained with a S value of 0,004, which resulted in a T of 11 m^2/d for Dewetsdorp. This coincides with the optimal parameter values for the SVF method. Varying recharge rates over the area, which were also applied in a SVF model, yielded best results.

Modelling of the De Aar aquifers for the flood period required varying storativities in the order of between 0,004 for the Karoo- and 0,03 for the alluvial aquifers. The best approximation was obtained with different storativities and different recharge values of 17 % for the alluvium and less than 5 % for the northern aquifer.

DETERMINATION OF THE EXPLOITATION POTENTIAL

According to the definition in the contract the exploitation potential depends on

- recharge due to rainfall
- underground in- and outflow and
- ground-water storage.

Recharge due to surface-water inflow and ground-water loss due to surface-water outflow e.g. discharge from fountains were not included in the definition. In the Dewetsdorp study area neither of the two occur. The alluvial aquifer in the De Aar area is, however, recharged by surface-water inflow. Due to problems in measuring the flow of the Brak River the recharge component due to inflow has not been separately determined and is included in the rainfall recharge.

AQUIFER RECHARGE

As a first approach, the following three simple equations can be used to obtain an estimate of the recharge of an aquifer:

For thin soil cover: $GR = 0{,}06\ (P - 120)\ [mm]$

For thick soil cover: $GR = 0{,}023\ (P - 51)\ [mm]$

For alluvial cover: $GR = 0{,}12\ (P - 20)\ [mm]$

where GR = annual ground-water recharge (mm) and P = mean annual precipitation (mm). These equations are based on the experience gained in the Dewetsdorp- and De Aar aquifers.

The equations were coded into program FORMULA which requires the annual rainfall of a nearby weather station and the percentage area covered by a thin soil cover, thick soil cover and alluvium cover in the area respectively. The annual record of rainfall is ranked by the program from the largest to the smallest value, whereafter the mean recharge for the aquifer is calculated for a required return period, as specified by the user.

A more reliable estimate may only be obtained with the following method:

Select information on the geological environment, positions where lateral inflow and outflow occur, positions of existing boreholes, ground-water level information, abstraction and rainfall data. Digitize the area and create a triangular mesh generated by a program named TFEM.

Run program SVF for the solution according to the SVF-method. The data requirements of program SVF are as follows:

- the triangular mesh as generated by TFEM
- monthly ground-water levels at a number of boreholes which are well-distributed over the area for a period of at least 18 months
- monthly abstraction from the aquifer
- monthly rainfall values (these values are not used in the recharge estimation procedure, but can be used later to obtain a relationship between recharge and rainfall)

- in- and outflow boundaries with transmissivity values

The output of SVF is as follows:

- the saturated volume of the aquifer is calculated on a monthly basis
- the ground-water table gradient at the in- and outflow points
- an estimate of the most probable mean S-value of the aquifer
- the recharge (in mm/time period) for a selected time period (an option in SVF allows the user to enter a different S-value at each node; in this case, the volume of ground water is calculated and not the saturated aquifer volume)
- an estimate of the lateral in- and outflow
- a file which contains information on recharge against rainfall

A much simpler program, GLF, can be used, if the user is satisfied with the assumption that the arithmetic mean of the ground-water level above a datum plane gives a reliable estimate of the saturated volume of the aquifer.

The ground-water recharge estimates by the SVF-method, can be greatly improved by the application of a two-dimensional numerical ground-water flow model AQUAMOD which enables the user to analyse the sensitivity of the aquifer in terms of aquifer response to abstraction and errors in recharge estimates. This model can also be used to estimate historical ground-water recharge.

UNDERGROUND IN- AND OUTFLOW

The ground-water flow component depends on the demarcation of the area, i.e. the length of in- and outflow boundaries, the aquifer transmissivity and the ground-water gradient. It has to be calculated for each area to be investigated.

AQUIFER STORAGE

The aquifer storage is described as

aquifer storage = aquifer thickness x storativity x area of influence.

Both the aquifer thickness and the storativity of the fractured Karoo aquifers could not be determined accurately. However, experience has shown that the thickness of a fractured aquifer is generally in the order of 20 to 25 m. As a first approximation the storativity S can be assumed to be in the order of 10^{-3} to 10^{-4} unless a better figure is available to the local geohydrologist. When a more detailed study is carried out the SVF-method allows the determination of a mean storativity value if the transmissivity T is determined by means of e.g. a pumping test.

GROOTFONTEIN AQUIFER

Independent of this project a short case study on the recharge of the dolomitic Grootfontein Compartment in the Western Transvaal was carried out in which the SVF-method was used. Recharge and storativity values obtained were in good agreement with the results obtained earlier by e.g. Bredenkamp, 1978 and Janse van Rensburg (1988).

This shows that the approaches used in this study can also successfully be applied to areas outside the Karoo Basin.

BENEFITS

The development of methods to determine the recharge due to rainfall and to approximate the ground-water storage of Karoo aquifers yields information which can be used by planners to optimally develop such aquifers. Statistically processed rainfall data allow the prediction of the ground-water recharge during drought conditions of various duration using the probability of a specified minimum rainfall. Ground-water schemes can therefore be laid out according to an acceptable risk, e.g. a failure rate of once in 20 or 25 years.

The computer programs developed as part of this research can also be used to calculate to what degree an aquifer is utilized, to determine areas for further production holes (provided suitable sites can be found) and the percentage by which the exploitation of that aquifer can be increased.

Light has also been shed on the recharge mechanism, which occurred predominantly by flow along preferred pathways and during the major rainfall events. It has also been found that the amount of ground water stored in an aquifer tends to decrease during a succession of subnormal, normal and even slightly above normal rainfall years. The aquifer is completely recharged during a succession of exceptional and very good rainy seasons.

The results gained during this project may also help other researchers working outside the Karoo and even outside South Africa on topics such as ground-water recharge, storage and exploitation.

RECOMMENDATIONS

Questions that need, in the opinion of the research team, be addressed or further investigated are:

- Methods to determine the storativity S and its variation with depth in (fine-grained) fractured aquifers.
- Methods to determine the thickness of fractured aquifers.
- Further investigation to what degree the methods developed during this project can successfully be applied under other climatic conditions, i.e. less or higher rainfall (with its influence on vegetation and soil cover) within the Karoo and to other aquifer formations outside the Karoo and the semi-arid climate zone.

1 PREFACE

The *'Klein Akwifeerprojek' (Small Aquifer Project)* as it was called in short amongst the members of the project staff, was born out of the need to supply communities in the central part of South Africa with water. In this area, rainfall is relatively low and the only perennial stream is the Orange River, which originates in the Drakensberg of Lesotho. Yet, a considerable number of villages and small towns have developed in the Karoo over the past hundred and more years, which draw their water supplies from local aquifers.

As communities grow, so rises the water demand. The former Ground-water Division of the Geological Survey, now the Directorate of Geohydrology of the Department of Water Affairs, has investigated the aquifers around many of these small towns and created new, or extended existing water-supply schemes. As aquifers are more and more exploited and additional ones have to be developed, the question of their exploitation potential arises.

Since the ground-water bearing properties of the subsoil depend mainly on the geology and because approximately fifty per cent of the surface of South Africa is underlain by rocks of the Karoo Supergroup, a project aimed at the investigation of the ground-water potential of Karoo aquifers was included in the Master Plan of the Water Research Commission (Vegter *et al.,* 1981).

Hans Willemink, an agricultural engineer, who came in 1980 from The Netherlands to South Africa to work for the Directorate of Geohydrology, joined the Institute for Ground-water Studies in 1983. He conducted a feasibility study on the exploitation potential of Karoo aquifers in 1984. Subsequently, the present project was formulated. Hans Willemink became seriously ill in 1986, and bemoaned by his colleagues and friends, died in March 1987 at the age of 36, leaving his wife Margy and his three young sons.

The authors of this report who took the project over from Hans Willemink have based their investigations on his approach. Not all of his ideas could, however, be followed up and some new avenues of approach were implemented in the later part of the project. This was necessary, because of the somewhat different professional background and the fact that not all of Hans's thoughts and findings had been laid down in a manner fully comprehensible to his successors. New findings, especially after the heavy rains of the 1988 summer, required further adaptions.

2 INTRODUCTION

With the exception of the southern and eastern coastal areas, South Africa has a semi-arid climate where potential evaporation exceeds precipitation. The water supply in the densely populated and more industrialized parts of South Africa is, to a very large degree, derived from surface water resources. Increasingly expensive schemes are built to convey water over great distances to these areas.

Smaller communities away from these centres cannot and need not afford such schemes; they rely on ground-water resources. In fact, the majority of the towns and villages in the Karoo have ground-water schemes. As the water demand in these communities rises, schemes have to be expanded or new ones must be built. The available ground water is, of course, limited and the question arises how large are the ground-water resources. If a method can be found that allows to determine the exploitation potential of aquifers, these aquifers can be optimally used and future planning will become easier.

As approximately half of the country with numerous smaller settlements, away from the major perennial rivers, is underlain by Karoo sediments, a research project into the *Exploitation Potential of Karoo Aquifers* was funded by the Water Research Commission.

2.1 OBJECTIVES

The stated objectives were:

- *to develop effective methodology which can be applied in areas where sparse data exist to determine the exploitation potential of Karoo aquifers with a minimum of additional information still to be generated, and*
- *to quantify parameters which contribute to the exploitation potential at different locations and verify methodologies developed.*

2.2 DEFINITION

For the purposes of this investigation, the exploitation potential of an aquifer is defined as follows:

> *the long-term yield of an aquifer which depends on recharge due to rainfall, underground in- and outflow and ground-water storage.*

2.3 THE PROJECT

The three-year project, undertaken by the Institute for Ground-water Studies at the University of the Orange Free State, was actively supported by the Directorate of Geohydrology of the Department of Water Affairs. The project commenced in July 1985. Because of the exceptional rains which fell in the summer months of 1988, the duration of the project was extended by six months to allow the rising ground-water level and the associated recharge caused by these rains to be studied and the results to be included into the project.

2.3.1 FEASIBILITY STUDY

A feasibility study had preceded the project. The aim of that study was to find the most suitable study areas. Criteria for the selection were:

(i) well-defined aquifer boundaries to restrict uncertainties in the water-balance calculations,

(ii) aquifers that were already exploited to a reasonable degree and for which good records were available to check the results of the study against actual abstractions,

(iii) aquifers situated in areas with different climate and preferably different geology to ensure validity of the results for a large area, and

(iv) the willingness of the local authorities to cooperate, because they had to supply the operational data and could assist by changing pumping patterns according to the needs of the research.

As a result of this feasibility study, two areas were chosen:

- Dewetsdorp with a 20 km^2 study area in the upper reaches of the Modder River, where the mean annual rainfall is 564 mm/a. The area is underlain by Beaufort sediments.
- De Aar: an area including the greater part of the *South-western Abstraction Scheme* which is approximately 120 km^2 in size, with a mean annual precipitation of 287 mm. The area is underlain by Ecca sediments.

2.4 RESEARCH REQUIREMENTS

In order to achieve the stated objective, the following research requirements were identified:

1) Quantitative determination of ground-water recharge as a function of rainfall, type and thickness of soil cover.

2) The determination of hydraulic parameter values and their variability with depth, in order to establish to what degree boreholes can be used in excess of the recharge during prolonged droughts.

In developing a research programme, it was realized that the following topics needed consideration:

a) Investigation into the physics of the aquifer systems, i.e. facies, thickness, jointing.etc.

b) Determination of the hydraulic parameters of the aquifer systems.

c) Determination of the ground-water balance under the prevailing conditions.

d) Modelling the behaviour of the two aquifers.

e) Development of a model for a "typical" Karoo aquifer and of a method for the calculation of the exploitation potential.

2.5 INVESTIGATIONS

With these topics in mind, the field activities can be subdivided into

(i) aquifer related investigations,

(ii) investigations regarding the unsaturated zone and

(iii) meteorological and other related measurements.

The aquifer related investigation comprised a geological survey of both areas assisted by aerial photo interpretation, magnetic surveys to detect dolerite dykes and resistivity work to determine the depth of the alluvium and dolerite sills.

The sedimentary facies, its diagenesis and the joint pattern determines the water-bearing properties of an aquifers. Based on the geological findings and the existing boreholes, additional sites were selected in order to get (i) a network with representative distribution of observation points, (ii) sites to carry out pumping tests, i.e. a test hole surrounded by observation holes and (iii) boreholes along sections in ground water flow direction for observation of amplitude and time lag of water-level peaks caused by recharge.

Recharge from rainfall occurs through the unsaturated zone. The investigations regarding the unsaturated zone comprised a soil survey and the creation of a representative network of neutron probe access tubes. Pits were dug for collection of soil samples for laboratory determination of soil structure, dry and wet bulk density. Gamma-gamma readings were taken at the same depths the samples were taken.

In each area a weather station was erected which measured rainfall, wind velocity, humidity, radiation and temperature. Additional rainfall recorders and rain metres were installed to obtain data on the rainfall variations within the areas. Long-term records were available from nearby rainfall stations of the Weather Bureau.

Ground-water was sampled during drilling for chemical analysis. For selected samples the concentration of stable isotope was determined. Stable isotope were also analysed for rainfall samples from a number of rainfall events.

In the following chapter, methodologies will be discussed in general.

Chapter 4 is subdivided according to the two study areas, Dewetsdorp and De Aar. After an introduction for each of the areas the various measuring networks and the data are described and evaluated according to the topics atmospheric data, data from the unsaturated zone and from the saturated zone. This is followed by modelling and the recharge estimation.

In Chapter 5 (i) the general findings made during the project are given and (ii) the results obtained from data processing and modelling of the two areas are discussed.

Chapter 6 contains three general equations proposed for first recharge evaluations, one each for Karoo aquifers under thin soil cover, under thick soil cover and for alluvial aquifers in the Karoo Basin. Procedures how to apply the computer programs developed for the evaluation of the recharge potential of Karoo aquifers, based on water-level and abstraction data from a specific area are also found in this chapter.

In Chapter 7 the results from a study of the Grootfontein Aquifer are presented in which the findings of the project under discussion have been applied. Although the study of the Grootfontein Aquifer did not form part of this project, it is included, because it shows that the proposed determination of the exploitation potential of Karoo aquifers can be sucessfully applied to other areas, even outside the Karoo.

Chapter 8 contains the references. Data not discussed in detail, are contained in the Appendix, together with the computer code of the developed programs.

With very few exceptions, all data generated and collected during the program are stored in G-BASE, a hydrological data base developed by the Institute for Ground-water studies on contract with the Water Research Commisssion. The data are available in DBASE III compatible files.

3 DISCUSSION OF GENERAL METHODOLOGIES

An investigation of this kind comprises many diverse disciplines. In general, these can be grouped into one of the following categories:

- methodologies for data collection, evaluation and storage,
- methodologies for investigation of the unsaturated zone,
- methodologies for investigation of the saturated zone, and
- methodologies for ground-water recharge estimation.

3.1 DATA COLLECTION, EVALUATION AND STORAGE

Routine data collection had, to a large degree, been automated. Weather station data were recorded on data loggers. Together with two rainfall recorders and about twenty water-level recorders in each area they formed the backbone of the measuring network. Rainfall sampling was also automatic. Depending on the season and the occurrence of rainfall events the remaining observation boreholes and the neutron probe access tubes were measured at variable frequencies.

In the beginning of the project the De Aar measuring network was attended by IGS staff transferred to De Aar. Later use was made of locally employed personnell. Abstraction data, as given by the Municipality, were supplied by the Department of Water Affairs.

Dewetsdorp data were collected by Bloemfontein based staff, assisted by the Dewetsdorp city engineer. Abstraction and some water-level data were also supplied by the city engineer.

Procedures and methodologies for data storage and geohydrological evaluation have been well-established in South Africa, in the research by Cogho *et al.* (1989) for the Water Research Commission. During the present project, extensive use was made of the National Ground-water Data Base for abstraction of historical data and storage of the basic information of newly drilled holes. At present, all data captured are stored on the micro computer data base G-BASE. They are available as DBASE III compatible and as ASCII files. Once batch programs for data loading onto the National Groundwater Data Base have been developed the data will also become available on that data base.

Apart from data storage G-BASE was used for processing of the chemical and and most of the other geohydrological data.

Potential evapotranspiration data were calculated from temperature and wind velocity data. The saturated hydraulic conductivity and retention curves have been determined using laboratory data from the soil analyses. Four models have been run to determine the flow through the unsaturated zone and various approaches were followed in the water-balance calculations.

3.2 UNSATURATED ZONE

A soil survey was carried out with a reconnaissance survey, inspection holes were dug, described and, where necessary, soil samples taken. The soil classification was done according to the binomial classification system of South Africa (MacVicar *et al.*, 1977).

The most important methods for soil water determinations are the gravimetric, tensiometric, hydrometric, electromagnetic and radiometric techniques. The indirect soil water measurement by means of the radiometric technique (neutron), was widely used during this investigation, but is still regarded with caution, because of uncertainties regarding the calibration. After Botha (1989) Neutron readings N were calibrated using gravimetric determination of moisture content θ, dry bulk density ρ_B and wet density ρ_W and gamma-gamma readings G using the following equations:

$$N = (\alpha \rho_B + \beta)\theta + \gamma \rho_B + \delta$$

$$G = a \rho_W + b$$

where α, β, γ and δ are constants.

Dry and wet densities have been determined for between four and six different depths in 31 sample pits, dug 5 m away from of the respective neutron probe access tubes. Gamma-gamma readings (density) were recorded at the same depths where the samples were taken.

Texture analyses were done on 116 samples, 44 of which came from De Aar and 72 from Dewetsdorp.

3.3 SATURATED ZONE

Drilling of the test- and observation holes was carried out by the Department of Water Affairs. Short air-lift tests were performed to establish the approximate yield of the holes. Pumping and packer testing on the holes were done to determine the transmissivity and storage properties of the aquifers. Water samples were taken of all waters struck and analysed for the usual macro elements by the Department's Hydrological Research Institute. The positions of most of the boreholes drilled during this investigation, were trigonometrically determined. The positions of other existing holes, monitored during the project, were inferred from the 1 : 50 000 topographical maps. Water levels, and their response with respect to rainfall, as well as yields from boreholes, were monitored throughout the project.

3.4 RAINFALL AND RECHARGE ESTIMATION

Estimating the rate of aquifer replenishment is but one of the components of this investigation, but probably the most difficult of all measures in the evaluation of ground-water resources. Estimates of this kind are normally and almost inevitably subject to large errors. In the preface of a workshop held by the NATO Advanced Research Group, Simmers (1987) reported that no single comprehensive estimation technique can yet be identified from the spectrum of available methods, which does not give suspect results. This book, edited by Simmers, can be regarded as possibly one of the best up-to-date references on natural recharge of ground water.

Recharge estimation can be based on a wide variety of models which are designed to represent the actual physical processes. Methods which are currently in use, include the following:

i) The soil water balance method (soil water budget).

ii) The zero flux plane method.

iii) The one-dimensional soil water flow model.

iv) The ground-water level fluctuation method.

v) The saturated volume fluctuation method.

vi) Inverse modelling for estimation of recharge (two-dimensional ground-water flow model).

vii) Isotope techniques and solute profile techniques.

The ground-water level fluctuation method, the saturated volume fluctuation method and the 2-D flow model are regarded as indirect methods, because ground-water levels are used to determine the recharge. Although the present state of knowledge in the field of ground-water recharge does not allow any particular preferred concept for research, it is, however, interesting to look at results which have been achieved elsewhere.

The Ground Water Estimation Committee of India recommended that ground-water recharge estimations should be based on the ground-water level fluctuation method (Sinha, 1987). Contradictory to this statement, Rushton (1987) and Johansson (1985, 1987) reported that this method often provides unreliable estimates. However, Bredenkamp (1987) used it with great success in the dolomitic regions of Transvaal.

In more recent studies, the unsaturated soil water flow model for recharge estimation was favoured by Johansson (1987), Rushton (1987) and Thiery (1987).

Sharma and Hughes (1985) studied the ground-water recharge in Western Australia. They concluded that more than 50% of recharge occurs through so-called preferred pathways by-passing the soil profile. If this was the case for a sandy coastal region, one can imagine that this value will be higher for the hard-rock formations of South Africa and the use of the unsaturated soil water model may be viewed with great circumspection when applied to the Karoo formations of Southern Africa. Another example of preferred flow paths in Australia, is given by Johnston *et al.* (1983).

Stephenson and Zuzel (1981) showed that in a semi-arid rangeland environment, south-west of Idaho, recharge occurs via three separate mechanisms, namely by infiltration through low-relief rubbly basalt outcrops, infiltration through shallow soils and by transmission through bed-rock channels during run-off and channel flow. To initiate the recharge, rainfall in excess of 20-30 mm over 24 hours is required. In general, the literature on ground-water recharge in arid and semi-arid regions indicates that it occurs infrequently and only in particular areas. Stephenson and Zuzel (1981) stated further that the more commonly recognized process of uniform recharge from rainfall with infiltration through the zone of aeration and percolation to the water table does not usually apply in semi-arid regions. Vanhaveren and Shiffler (1976) reported that recharge to the ground water in the semi-arid portion of the Northern Great Plains is limited to a few optimum sites.

The major use of infiltration theory has been the derivation of expressions for the amount of water flowing into the soil, given the rate of application. The first and most popular infiltration function with physical significance was developed by Green and Ampt (1911) and was adapted by Philip (1957) for instantaneous surface ponding. Philip (1957) wrote the cumulative infiltration function as a series:

$$i(t)= St^{1/2} + At + ...$$

where t is time, S is the sorptivity (defined as the amount of moisture absorbed in the absence of gravitational effects) and A is related to the saturated hydraulic conductivity K (A/K = a value between 0,3 and 0,6). The Philip infiltration function is usually truncated and only the first two terms are used.

The aim of the following sections is to give a brief theoretical review of the methods which are currently in use for the estimation of natural ground-water recharge.

3.4.1 SOIL WATER BALANCE (SWB) METHOD

Water balance models were developed in the 1940's by Thornthwaite (1948) and revised by Thornthwaite and Mather (1955), and are essentially bookkeeping procedures which estimate the balance between the gain and loss of water.

In a standard soil water balance calculation, the amount of water allocated to the various processes is usually expressed as an equivalent depth of water. The soil water balance can be represented by:

$$GR = P - E_a - \Delta S - R_o \qquad (3.4.1)$$

where

GR = recharge (i.e. water lost by percolation to below the root zone)
P = precipitation
E_a = actual evapotranspiration
ΔS = change in soil water storage
R_o = run-off

One condition that is enforced, is that if the soil water deficit is greater than a critical value (called the root constant), evapotranspiration will occur at a rate less than the potential rate. The magnitude of the root constant depends on the vegetation, the stage of plant growth and the nature of the soil (Rushton and Ward, 1979). A range of techniques for estimating E_a, usually based on Penman-type equations, can be used.

The data requirement of the SWB-method is large. When applying this method to estimate the recharge for a catchment area, the calculation should be repeated for areas with different precipitation, evapotranspiration, crop type and soil type. Khan (1980) found that recharge estimates may differ significantly between years that have similar annual rainfalls.

Gieske and Selaolo (1987) stated that the soil water balance method is of limited practical value, because E_a is not directly measurable. Moreover, storage of water in the unsaturated zone and the rates of infiltration along the various possible routes to the aquifer form important and uncertain factors. Another aspect that merits attention, is the depth of the root zone which may vary in semi-arid regions between 1 and 30 m.

Results from this model are of very limited value without calibration and validation, because of the substantial uncertainty in input data (precipitation and potential evapotranspiration). The model parameters do not have a direct physical representation which can be measured in the field (Johansson, 1987). He found that for a sandy till aquifer in Sweden, recharge had to be allowed, even when a water deficit existed, in order to correctly reproduce the dynamics, revealed as ground-water level fluctuations. Steenhuis *et al.* (1985, 1986) found that the SWB-method gave good estimates of recharge on Long Island. Alley (1984) applied the method successfully in New Jersey, while Eagleson (1978) outlined a dynamic formulation of the water budget according to the climate, soil and vegetation.

Soil water balance models, available in South-Africa, are the well-known ACRU model (Schulze, 1985), MOUSE (Steenhuis, 1985), PUTU (De Jager *et al.*, 1987) and ACRUWAT (SRK, 1988).

3.4.2 ZERO FLUX PLANE (ZFP) METHOD

The ZFP-method relies on the location of a plane of zero hydraulic gradient in the soil profile. Recharge over a time interval is obtained by summation of the changes in water contents below this plane. The position of the ZFP is usually located by installation of tensiometers. Unfortunately, the method breaks down in periods of high infiltration, when the hydraulic gradient becomes positive downwards throughout the profile (Allison, 1987).

The flux, q, defined as the volume of water per unit time passing through unit area at any depth, is given by Darcy's Law:

$$q = -K(\theta) \cdot \frac{d\phi}{dz}$$

where

$K(\theta)$ = unsaturated hydraulic conductivity

ϕ = total water potential = $z + \psi(\theta)$

Thus, knowing the unsaturated hydraulic conductivity and the potential gradient, the flux may be determined. Moisture potentials may be measured, using tensiometers or the neutron scattering technique. The hydraulic conductivity estimation presents more of a problem. Firstly, K may vary by a factor of 10^3 or so over the normal water content range of a typical soil (Hillel 1971, p.105) and, secondly, there are large variations of K from place to place, even in apparently homogeneous soils and over distances of a few metres (Nielsen *et al.*, 1973) at the same depth.

There is, however, an alternative to this which avoids the need to know values of K. From the one-dimensional vertical form of the water-balance equation:

$$\frac{\partial\theta}{\partial t} = -\frac{\partial q}{\partial z}$$

by assuming negligible lateral soil moisture flow, one obtains by integration from depth z to depth z+dz:

$$q_z = q_{z+dz} + \int_{z}^{z+dz} \frac{\partial\theta}{\partial t}\, dz$$

where q_z is the vertical component of the Darcian water flux (positive in the upward direction). Therefore, a knowledge of either q_{z+dz} or q_z and the rate of change of moisture content between the two depths allow the flux at the other depth to be calculated. At the ZFP-point, say z_0, the potential gradient is zero and the flux is also zero. If z_0 does not change with time, the accumulated flux, F(Z), between times t_1 and t_2 is

$$F(Z) = \int_{t_1}^{t_2} q(z)\, dt = \int_{z_0}^{Z} |\, \theta(t_1) - \theta(t_2) \,|\, dz$$

where $Z = z + dz$ and $z_0 = z$.

Wellings (1984) got good estimates of recharge in the UK for periods of the year when the ZFP exists. Sophocleous and Perry (1985) used the unsaturated zone hydraulic conductivity, K(θ), and the hydraulic potential data, $\psi(\theta)$, to solve Darcy's equation in the unsaturated zone below the ZFP.

Application of the Sophocleous and Perry version of the ZFP depends on the success of accurately determining K(θ), which is one of the most difficult parameters to determine directly in the field. A technique to predict K(θ) is given in the following section.

The work of Stephens and Knowlton (1986) highlights the problem of using the flux technique for estimation of ground-water recharge, especially where the fluxes are low. They found that the annual recharge flux was either 7 or 37 mm/a, depending on how they calculated their mean hydraulic conductivity.

3.4.3 SOIL WATER FLOW MODEL

Water has to move through the unsaturated zone until it reaches the water table. Flow conditions within this zone are far more complex than the flow mechanisms in a saturated aquifer. The amount of water retained in a soil has upper and lower bounds, determined respectively by inherent properties of the soil and by plant extraction. It is this retained soil water which may be redistributed under saturated conditions by drainage or under unsaturated conditions by movement up or down, depending on the relative wetness of the different soil horizons and by plant root extraction.

Definitions of three soil water constants are necessary. Porosity, ε, is the percentage soil volume occupied by voids, i.e. the maximum soil water storage or saturation. The matric potential at saturation is 0 kPa (1 kPa $\approx$ 0,1 m). When a certain maximum amount of water is stored in the soil, the remainder will drain away, leading to the concept of field capacity, the FC. The amount of water stored in the soil at field capacity is that amount of water which a soil will hold against gravitational forces. A theoretical definition in terms of matric potential is difficult, owing to the fact that the FC may vary with texture, but it is traditionally taken to fall somewhere between -5 and -33 kPa (Schulze, 1985). The wilting point, WP, is taken as the dry limit of water available to plants and the matric potential at this point is usually accepted to be -1500 kPa (150 m).

The matric potential is closely related to the volume of water contained in the geological formation. This relation, known as the water retention curve, can be conveniently expressed mathematically in terms of the volumetric water content θ, defined as:

$$\psi \equiv \psi(\theta)$$

The equation of a water retention curve is a non-linear relation of the water content. In more physical terms, it is said to show a hysteresis effect (Rushton, 1987 and Sophocleous and Perry, 1985).

This means that the curve is at least double legged, with one leg valid during the drying cycle and the other one during the wetting cycle. Care must, therefore, be taken when using the retention curve in practical applications.

Since the water retention curve can only be determined experimentally, its true behaviour in practice is only known at a finite number of points. Two methods, to obtain values at non-experimental points, can be used. The first and most obvious method is to use interpolation, but this method can only be successful in those cases where the experimental points are closely spaced. The second approach is to fit an empirical equation to the experimental points. The equations mostly used today are the Brooks and Corey function (Brooks and Corey, 1964) and the Van Genuchten function (Van Genuchten, 1980). The Van Genuchten equation deserves special attention. In this equation, the water retention curve is expressed as

$$\Theta = (1 + (\lambda\psi)^n)^{-m}$$

where λ, n and m are characteristic constants, which have to be determined for every soil type. Van Genuchten suggested that one should use the value m=1-1/n. The Van Genuchten equation expresses the water retention curve not in terms of the water content, but rather in terms of the reduced water content, defined by the equation:

$$\Theta = (\theta - \theta_r) / (\theta_s - \theta_r)$$

where

θ_s = the saturated water content

θ_r = the residual water content

The residual water content is the soil dependent amount of water remaining in the pores which cannot be drained by suction. The three parameters, namely (a) the water content, (b) the matric potential (fluid pressure) and (c) the hydraulic conductivity, are interrelated. These relationships are very sensitive, for example a change in the water content of a few per cent, often corresponds to a change in the hydraulic conductivity by two or more orders of magnitude.

The one-dimensional equation for vertical flow in the unsaturated zone can be expressed as (Richards, 1931)

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left(K(\psi)\cdot\frac{\partial\psi}{\partial z}\right) + \frac{\partial}{\partial z}\left(K(\psi)\right) \qquad (3.4.2)$$

where

θ = volumetric water content

K = hydraulic conductivity [= $K(\theta)$ or $K(\psi)$]

ψ = matric potential

Both θ and K are functions of the unknown potential ψ. Studies by Jayawardena and Kaluarachchi (1984) indicate that the solutions of (3.4.2) are more sensitive to ψ(θ) variations than K(θ) variations. No evidence in the literature exists that the K(θ) relationship exhibits significant hysteresis (Steenhuis *et al.*, 1979), so that it is safe to assume that K is a unique function of θ.

Unsaturated flow can only occur if the fluid molecules move in pores that remain filled at the given suctions or in fluid films that surround partially drained pores (Botha *et al.*, 1988). The Darcy velocity for unsaturated flow should, therefore, be less than that of saturated flow and steadily decrease as the water content of the porous medium decreases. This implies that the unsaturated hydraulic conductivity, for a given porous material, should be smaller than its saturated hydraulic conductivity and, moreover, that it should decrease as the water content decreases. Following Richards (1931), Darcy's law for unsaturated flow can be expressed as:

$$q=-K(\theta)\cdot\frac{d\phi}{dz}$$

where $\phi = z + \psi(\theta)$ and $K(\theta)$ is related to the relative permeability given by Van Genuchten (1980):

$$k_r(\Theta) = \Theta^{1/2}\,(1-(1-\Theta^{1/m})^m)^2$$

where

$K(\Theta) = K_s \cdot k_r(\Theta)$

K_s = saturated hydraulic conductivity

Equation (3.4.2) can be solved by either a finite difference or a finite element model. Computer code (SUFF), developed by Botha (1988), can be used to solve Equation (3.4.2) via the finite element method. Wagenet & Hutson 's (1987) finite difference approach (LEACHM) can also be used. This method has the same disadvantages as the technique used by Sophocleous en Perry (1985), which was described in Section 3.4.2.

Freeze (1969) stated that of all the parameters controlling ground-water recharge, it is likely that the antecedent soil water regime is the most important. He also found that:

(i) Estimates of recharge of a soil based on a knowledge of the saturated hydraulic conductivity and its textural classification can be misleading.

(ii) If all other controlling parameters are equal, the possibility of a water-table rise is greater for:

(a) rainfalls of long duration rather than those of short duration,

(b) low-intensity rainfalls of long duration rather than high-intensity rainfalls of short duration,

(c) high ground-water recharge rates rather than low,

(d) wet antecedent soil water conditions rather than dry,

(e) soils with high permeability, and/or low specific water capacity, and/or high water content over a considerable range of tension head values, and

(iii) The depth of ponding has a considerable effect on the surface infiltration rate, but very little effect on the nature of the infiltrating wetting front.

3.4.4 GROUND-WATER LEVEL FLUCTUATION (GLF) METHOD

An indirect method of deducing the ground-water recharge is from the fluctuation of the ground-water level. It is recognized that factors such as pumping or irrigation do have an influence on ground-water recharge. The fluctuation of the water table is a function of the recharge, the water yielding properties (T,S) and the geometry of the aquifer.

The ground-water recharge (GR) during the time period Δt, can be written as:

$$GR = \Delta h \cdot S \text{ (for a point value)} \qquad (3.4.3)$$

where

Δh = change in ground-water level

S = specific yield

For a specified area the following equation applies:

$$GR = \Delta h \cdot S \cdot A + Q_T \, \Delta t$$

where

A = area where the change Δh takes place

Q_T = ground-water inflow - outflow - abstraction

Although the method is attractive to use, due to its dependence on available ground-water level observations, it can lead to unreliable results (Rushton, 1987). One of the fundamental inadequacies of the method is the implicit assumption that every inflow or outflow is uniformly distributed over the area. Estimation of recharge, using Equation (3.4.3), is greatly dependent on the value of S.

The significance of S, determined from a pumping test, is equally open to question. Nwankwor *et al.* (1984) estimated that the S of a clean unconsolidated sand can vary between 0,07 to 0,3, depending on the method used for determining S. In another study, Sophocleous (1985) showed that calculation of S by any inverse procedure, based on water-level changes and on estimated recharge values, may be erroneous due to the influence of the capillary fringe, especially in shallow, fine-grained water-table aquifers. Air entrapment may be an additional difficulty of this method, particularly at high percolation rates.

Based on ground-water level fluctuations and rainfall amounts, Chaturvedi (1943) (as quoted by Sinha, 1987) derived the following relationship:

$$GR = 2\ (P - 15)^{0,4}\ [\text{inches}].$$

Studies in India by the Central Ground Water Board lead to the following equation:

$$GR = 0{,}26\ (P - 23)\ [\text{mm}].$$

Bredenkamp (1987) favours the GLF-method and derived the following relationship for the dolomitic regions of Transvaal:

$$GR = 0{,}3\ (P - 313)\ [\text{mm}].$$

A method, proposed by Hill (as reported by Domenico, 1972), used the GLF-method for the determination of safe yield for an aquifer. When the ground-water level fluctuations are graphed against the abstraction rates, the value of the abstraction rate which equals the mean GLF, is regarded as the safe yield of the aquifer.

Figure 3.1 shows the reaction of the water table for a recharge of one millimetre for different specific yield values of the aquifer. For example, a recharge of one millimetre to an aquifer with a S which equals 1×10^{-3}, leads to a rise in the water table of 1 000 mm. S-values of 1×10^{-5}, as sometimes quoted for Karoo formations, will lead to a water-table rise of 2 500 metres, for an annual recharge of 5% and rainfall of 500 mm. To compare favourably with ground-water levels as measured in the field, either the recharge must be much lower or the S-value much higher than 1×10^{-5}.

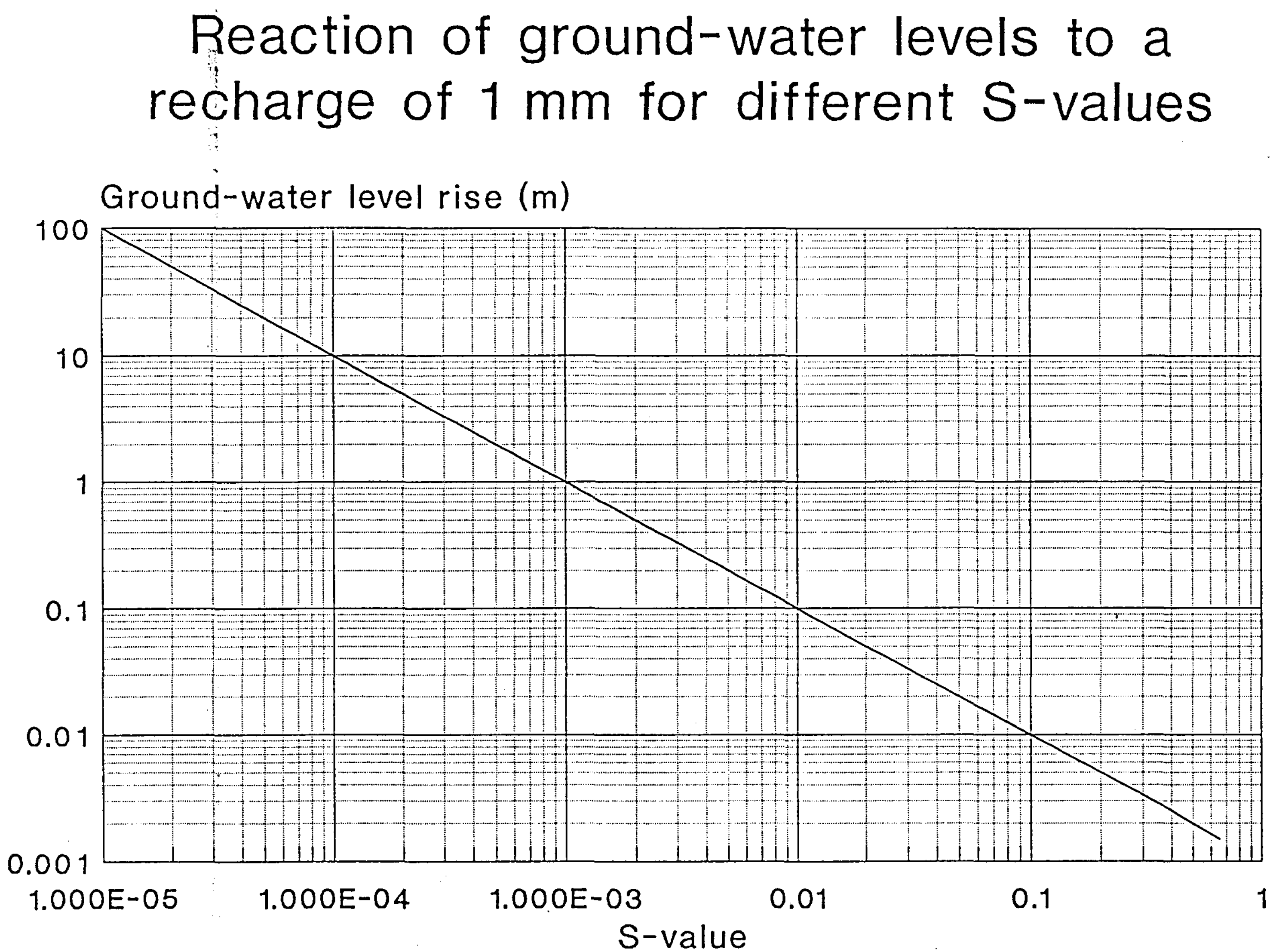


Figure 3.1. Rise of ground-water levels for a recharge of 1 mm for different specific yield values.

A computer program GLF has been written for Equation (3.4.3) and is given in the Appendix. In a following section, the GLF-method will be adapted to a proposed saturated volume fluctuation method. PC-MINE (SRK, 1988) is also a program that can be used for the GLF-method for estimating of recharge to an aquifer.

3.4.5 INVERSE MODELLING TECHNIQUE

The inverse modelling technique to be discussed, is a two-dimensional finite element (or finite difference) ground-water model of the saturated zone. Current methods of calibrating ground-water flow models are either indirect or direct. The indirect approach is essentially a trial and error procedure that seeks to improve an existing estimate approach of the parameters in an iterative manner, until the model response is "sufficiently" close to that of the real system. The direct approach is different in that it treats the model parameters as dependent variables in a formal inverse boundary value problem.

One of the main difficulties in dealing with the inverse problem stems from the inherent non-uniqueness of its solution (Simpson *et al.*, 1971). Many of the data entered into the inverse problem, represent imprecise measurements and processed information that gives a distorted picture of the system's true state (Neuman, 1973).

The Boussinesq equation, which permits the estimation of recharge of steady ground-water flow by taking derivatives, leads to numerical instability (Allison, 1987).

The calculation of recharge to an aquifer by the inverse modelling technique must be regarded with great suspicion, if the true S-values of the aquifer are not known. If, however, the calibrated S-values can be regarded as very close to the real values, this technique can be of much use in describing the behaviour of the aquifer to the recharge phenomena in general.

3.4.6 SATURATED VOLUME FLUCTUATION (SVF) METHOD

3.4.6.1 THEORETICAL DEVELOPMENT

Inputs and outputs for conventional hydrological models are generally water volumes per unit time, such as recharge and discharge and surface in- and outflows. The fundamental idea common to a variety of situations, is that the hydrological balance equation or some other equation, empirically derived, is usually employed, e.g:

$$I - O = \frac{\Delta W}{\Delta t}$$

In the same manner, the geohydrological balance equation for a ground-water reservoir is given as:

$$I - O + GR - Q = \frac{\Delta W}{\Delta t} \qquad (3.4.4)$$

where

- I $= \frac{I_1 + I_2}{2} =$ mean lateral inflow (m^3/d) during time $t_2 - t_1 = \Delta t$
- O $= \frac{O_1 + O_2}{2} =$ mean lateral outflow (m^3/d) during Δt
- GR = ground-water recharge into the reservoir in m^3/d (also named percolation or deep infiltration)
- Q = discharge out of (or into) the reservoir (boreholes, rivers, etc.) in m^3/d during Δt
- ΔW = change in ground-water volume (m^3) = $S \cdot \Delta V$, where
- S = specific yield (or effective porosity)
- ΔV = change in saturated volume aquifer material (= $V_2 - V_1$)
- Δt = $t_2 - t_1$ = time increment

In Equation (3.4.4), it is assumed that there is no vertical movement through the base of the water-table aquifer. In the case where the root of plants extract water from the aquifer, the evapotranspiration must be added to the Q term in Equation (3.4.4). The accuracy of estimating this evapotranspiration term is not very reliable, and, for this reason, it is recommended that the GR-term already includes the evapotranspiration, i.e. effective GR = actual GR - evapotranspiration. In the application of the SVF-method in this report, the estimated ground-water recharge must be viewed as the effective ground-water recharge.

The term I in (3.4.4) can be expanded with the aid of Darcy's law :

$$I = T_I \cdot L_I \frac{i^1{}_I + i^2{}_I}{2} = A_1 \cdot T_I \qquad (3.4.5)$$

where

T_I = mean transmissivity at the inflow boundary (i.e. T_I =KD for a water-table aquifer)

L_I = width of the inflow boundary

$i^1{}_I$ = ground-water gradient at inflow at time t_1

$i^2{}_I$ = ground-water gradient at inflow at time t_2

The same reasoning can be followed at the outflow boundary to yield:

$$O = T_O \cdot L_O \frac{i^1{}_O + i^2{}_O}{2} = A_2 \cdot T_O \qquad (3.4.6)$$

Substitution into (3.4.4) yields:

$$GR + A_1 \cdot T_I - A_2 \cdot T_O - A_3 \cdot S = Q \qquad (3.4.7)$$

where $A_3 = \frac{\Delta V}{\Delta t}$.

Equation (3.4.7) is the general ground-water balance equation for an unconfined aquifer. The boundaries of an area studied usually do not represent streamlines, i.e. they are not perpendicular to the equipotential lines. Hence the lateral in- and outflow of ground water crossing the area's boundaries must be accounted for in the balance equation. One of the factors influencing the change in water table is the effective porosity, S, of the zone in which the water-table fluctuations occur. It has been recognized that S changes as the depth to the water table changes (Kessler and De Ridder, 1974), especially for shallow water tables, less than 3 m. Furthermore, it should be noted that if the water table drops, part of the water is retained by the soil particles; if it rises, due to infiltration, air can be trapped in the interstices that are filling with water. Hence S for rising water is, in general, less than that for a falling water table (Kessler and De Ridder, 1974).

To apply Equation (3.4.7) effectively, it is essential that both the area and the period for which the balance are assessed, be carefully chosen. By a comparison of ground-water levels of boreholes, with similar water-table fluctuation patterns, holes with the same pattern can be grouped together. It is also conceivable that the whole area be divided into sub-areas by the Thiessen method. A different approach shall be described at the end of the chapter.

Equation (3.4.7) can be applied for a number of specific conditions (assumptions):

(a) Where the inflow terms are balanced by the outflow terms, the change in ground-water storage is zero (i.e. $\Delta V = O$). This provides the necessary conditions to derive at safe yield estimates and to predict recharge from precipitation:

$$GR = A_2 \cdot T_O - A_1 \cdot T_I + Q \tag{3.4.8}$$

When outflow occurs in the absence of inflow, a general recession model may be formulated. This permits an evaluation of the outflow quantities, the effects on ground-water storage, or the inflow that takes place following the recession.

(b) By incorporating the "no recharge" recession (i.e. $GR = O$) for ΔV = maximum decrease during Δt, Equation (3.4.7) reduces to :

$$A_1 . T_I - A_2 . T_O - A_3 \cdot S = Q$$

from which S can be calculated as:

$$S = \frac{A_1 . T_I - A_2 \cdot T_O - Q}{A_3} \tag{3.4.9}$$

(c) By choosing a sub-area from the aquifer such that $T_I = T_O = T$ and $Q = 0$, Equation (3.4.7) can be written as:

$$GR + (A_1 - A_2) \cdot T - A_3 \cdot S = 0$$

In the case where $GR = 0$ (ΔV = maximum decrease) we have:

$$\frac{S}{T} = \frac{A_1 - A_2}{A_3} \tag{3.4.10}$$

(d) If the aquifer is bounded by impervious dykes or by ground-water divides, A_1 and A_2 in (3.4.7) are zero. For this case

$$GR - A_3 \cdot S = Q \tag{3.4.11}$$

from which the ground-water recharge can be calculated, if S is known. If $GR = 0$ is assumed for ΔV = maximum decrease, a mean S-value for the aquifer can be calculated.

The following procedures for the application of Equation (3.4.7) are recommended:

(i) Ground-water levels (mamsl) in observation boreholes, which are well-distributed over the whole of the aquifer within certain well-defined ground-water boundaries (such as ground-water divides or no-flow boundaries), are required on a regular, preferably a monthly basis.

(ii) The region must be discretized in a number of small triangles (constructing a mesh).

(iii) Monthly water levels should be interpolated to every node of the mesh after which the saturated volume, V_i at time t_i, is calculated. An arbitrary base level can be assumed, because only the difference ΔV over a Δt is needed.

(iv) Repeat (iii) for all months.

(v) Construct a graph of V_i against time (months). For times where $V_i = V_j$ (i.e. $\Delta V = 0$), Equation (3.4.8) can be used to estimate the ground-water recharge GR, during the time $t_j - t_i$.

(vi) An estimate of the S-value of the aquifer (if T_I and T_O are known) can be obtained from Equation (3.4.9). For each month, the ground-water recharge, GR, is calculated from Equation (4.4). For certain special conditions, one of the other equations above can be used.

It is very important to realize that Equation (3.4.7) is subject to a number of possible errors. Furthermore, the equation is a finite difference approach, with a solution accuracy which is dependent on the size of Δt. Interpolation errors always occur, but can be minimized if the boreholes are well-distributed over the domain of interest. The same boreholes must be used when interpolating ground-water levels of different periods. It is equally important to always use the same interpolation technique (e.g. kriging) during the above calculations. If the aquifer is bounded by a flow or constant head boundary, the solution of Equation (3.4.7) is dependable on the accuracy of the transmissivity values at these boundaries.

A micro computer program which deals with the calculation of saturated volumes and ground-water recharges was developed to deal with the above-mentioned equations. It is named SVF and the code is given in the Appendix. A program, called TFEM (developed by IGS), can be used for the construction of the triangular mesh. TFEM requires the nodal coordinates of points selected by the user in the aquifer, whereafter automatic constructing and numbering of triangles are performed. Because of the length of this program, the code is not given here, but is available from IGS.

3.4.7 ISOTOPE AND SOLUTE PROFILE TECHNIQUES

^3H, ^2H, ^{18}O and ^{14}C are commonly used in recharge studies, of which the first three most accurately simulate the movement of water, because they form a part of the water molecule. Many studies on recharge estimation using natural tritium, are listed in the literature (Bredenkamp, *et al.*, 1974 and Foster & Smith-Carrington, 1980). Although a proven tool for qualitative recharge estimation, environmental tritium has several disadvantages (Edmunds, 1987) such as:

(i) Tritium is not conservative and is lost from the system by evapotranspiration.

(ii) Contamination during sampling and processing is a factor which is enhanced in remote areas and at low total moisture levels.

(iii) Analysis is highly specialized and costly.

(iv) Quantitative studies are difficult to achieve, since it is difficult to determine a tritium mass balance.

Adar (1984) used natural isotopes, together with chemical analyses. He was able to differentiate between different aquifers, recharge areas and flow paths, but could not quantify recharge with the aid of these parameters.

An environmental tracer most suitable for following the movement of water, must be highly soluble, conservative and not substantially taken up by vegetation (Sharma, 1987). The chloride ion satisfies most of these criteria and is therefore considered a suitable tracer, particularly in coastal areas where large quantities of aeolian chloride are deposited.

If the assumption of chloride as a conservative ion is accepted, the ground-water recharge is given by

$$GR = \frac{D}{C} [mm/a]$$

where

D = wet and dry chloride deposition ($mg/m^2/a$)

C = concentration in ground water

The method is convenient, fast and cheap. The drawback of the technique is the uncertainty in the determination of the wet and dry deposition. The principal source of chloride in ground water, if there are no evaporite sources, is from the atmosphere. In this case, the recharge can be expressed as (Houston, 1987):

GR = rainfall x Cl of rainfall / Cl of ground water.

The chloride method must be treated with caution, as accumulation of chloride near the soil surface, because of evapotranspiration, may violate the assumption of a steady state chloride flux density throughout the unsaturated zone (Allison *et al.*, 1984). Furthermore, recharge under conditions of extremely high rainfall with a long recurrence period, is likely to influence the chloride concentration of ground water to a high degree, resulting in an overestimate of the mean annual recharge. Houston (1987) used this method for recharge calculations in Victoria Province, Zimbabwe, and arrived at values between 2 - 5% of the annual rainfall. The chloride content of rainfall in Zimbabwe is in the order of 0,5 mg/l .

4 RESEARCH AREAS

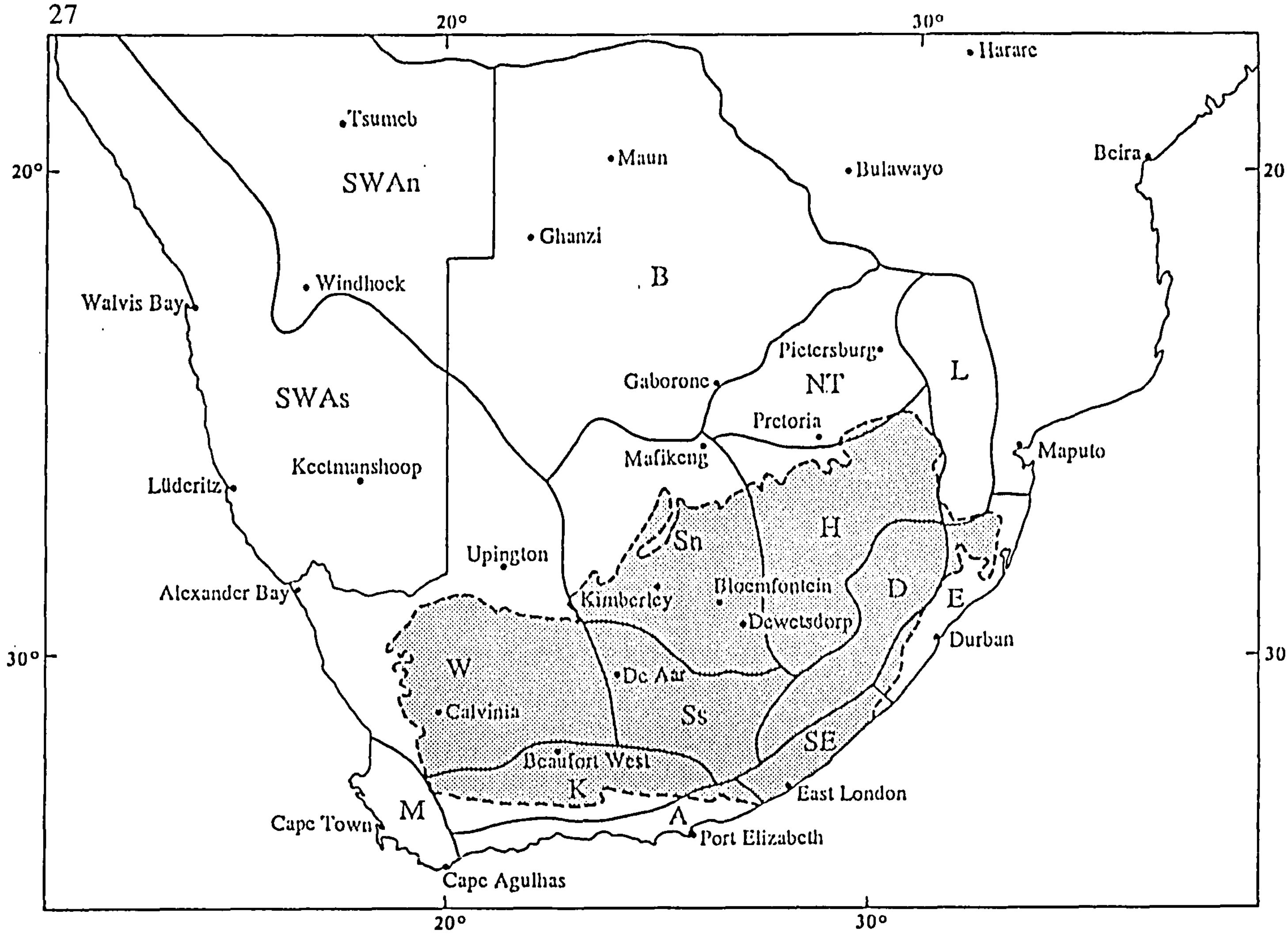
Following a feasibility study, the research areas for this study were selected out of twenty potential areas in the Cape and the Orange Free State, according to the following criteria (Willemink and Kirchner, 1984):

(i) The surface area must be relatively small.
(ii) It must consist of a system with a well-defined underground watershed.
(iii) The exploitation potential must, within limits, be known.
(iv) Alternative water resources must exist, in order to allow the change of abstraction rates and study the aquifer response.
(v) Other aspects such as available data and cooperation with local authorities.

Based on these criteria and because of differences in rainfall, topography and vegetation, two sites, namely Dewetsdorp and De Aar, were selected. Figure 4.1 shows the climatic regions of Southern Africa (after B.R. Schulze, 1986). The shaded area comprises the Karoo Basin. The situation of Dewetsdorp and De Aar is indicated on this map.

Dewetsdorp lies in the eastern part of the Karoo Basin. Annual precipitation is about 600 mm. Geologically, it lies on the boundary of the Permian Adelaide- and the Triassic Tarkastad Formations which form the Beaufort Group.

De Aar is located in the centre of the Great Karoo and receives approximately half of the amount of rainfall of Dewetsdorp. The De Aar research area lies near the boundary of the Adelaide Formation and the Tierberg Formation, which belongs to the middle Permian Ecca Group.



REGION	AREA and CLIMATE
A	Southern Cape Coastal Belt : Rain during all seasons
B	Botswana : Warm, Steppe, summer rain
D	Drakensberg region : Temperate to warm with summer rain
E	East Coast : Warm to hot and humid subtropical climate
H	Highveld : Temperate to warm with summer rain
K	Little and Great Karoo : Mainly semi-arid climate
L	Lowveld : Warm, Steppe
M	Southwestern Cape : Mediterranean climate
NT	Northern Transvaal : Warm, Steppe, summer rain
SE	Southeastern coastal region : Temperate to warm, mainly summer rain
Sn	Northern Steppe : Semi-arid climate
Ss	Southern Steppe : Semi-arid climate
SWAn	Northern South West Africa : Warm, Steppe, summer rain
SWAs	Southern South West Africa : Desert and poor Steppe
W	Namaqualand and Northwestern Cape : Desert climate

Figure 4.1 Locality map showing Dewetsdorp and De Aar in the (shaded) Karoo Basin. The climatic regions of Southern Africa are also indicated.

4.1 DEWETSDORP

4.1.1 PHYSIOGRAPHY

LOCATION - TOPOGRAPHY - DRAINAGE

The Dewetsdorp research area lies approximately 65 km south-east of Bloemfontein in the upper reaches of the Modder River. The largest portion of the research area lies on the farm Kareefontein 66, overlapping onto parts of the farms Frankfürt 602 and Jouberts Rust 456 (borehole JPN1 north of Frankfürt on Figure 4.20).

The research area lies in the uppermost part of the drainage region 352 (Modder River) along the drainage regions 351 (Kaffir River) and 420 (Caledon River). It is drained by the Kareefontein Spruit. Botanically, the area is classified as transitional *Cymbopogon - Themeda veld* which resorts under the pure grass velds (Acocks, undated). The terrain-morphology is described as *lowlands with hills* (Kruger, 1979).

The research area stretches approximately 8,5 km in a SSW - NNE direction, is about 4 km wide and has a surface area of 21,1 km^2. It consists of a well-defined ground-watershed in the west, south and south-east, which follows a chain of rather steep sloping hills, partly capped by dolerite sills. The eastern boundary of the area is formed by a shallow topographical ridge between the Kareefontein Spruit and the Modder River Valley. In the north, where the Kareefontein Spruit joins the Modder River, there is no boundary and surface- as well as ground-water discharge occurs here.

The elevation at the watershed reaches a maximum of 1620 mamsl, while the confluence of Kareefontein Spruit and Modder River lies at a height of approximately 1475 mamsl. In the uppermost parts of the area, the gradient is ± 1: 30. It steepens to less than 1 : 10, just south of the village. In the lower part of the study area, the gradient is about 1 : 40.

4.1.2 Climate

According to the Köppen classification (Schulze, B.R., 1986), Dewetsdorp lies in the boundary area between the climatic region Sn (Northern Steppe) and H (Highveld, see Figure 4.1). It has an annual precipitation of between 500 and 650 mm. Rainfall is largely due to showers and thunderstorms falling between October and March, with the peak in the months January to March. Continued summer rains usually occur if an equatorial low-pressure trough is established over South Africa. The wave in the west wind area lies further south than usual and is followed by a succession of anticyclonic cells moving around the coast, while the low-pressure trough remains stationary. This situation occurs not frequently, but is important because of the extensive and continued rainy weather associated with it (Schulze, 1986). Such a situation set in at the end of November 1985, when a rainy period started in Dewetsdorp and De Aar: a deep low pressure system over the west of the subcontinent fed in moist air, while cool moist air moved in from the south-east of the country which caused general rain and showers over most of the country. A synoptic weather map (after the Daily Weather Bulletin of the Weather Bureau for November 28, 1985) reflects this situation (see Figure 4.2).

During the usually dry winter, occasional rains do occur, on average $1^1/_2$ days per month. Hail is sometimes associated with thunderstorms, predominantly in the early summer. Based on the data from Schulze, 1986 the annual rainfall distribution is shown on Figure 4.3.

Temperatures show large diurnal and seasonal variations. Temperature statistics for Dewetsdorp are not on record. The temperature statistics of the JBM Airport near Bloemfontein (Weather Bureau 1986), are regarded as representative for Dewetsdorp:

	January [° C]	July [° C]
Maximum	39.3	24.0
Mean daily maximum	30.5	17.4
Mean daily temperature	22.9	7.6
Mean daily minimum	14.8	-2.1
Minimum	0.9	-9.6

North-westerly to northerly winds prevail which attain their maximum velocities in the afternoon. Recorded velocities for Bloemfontein show for 50 per cent of the time a

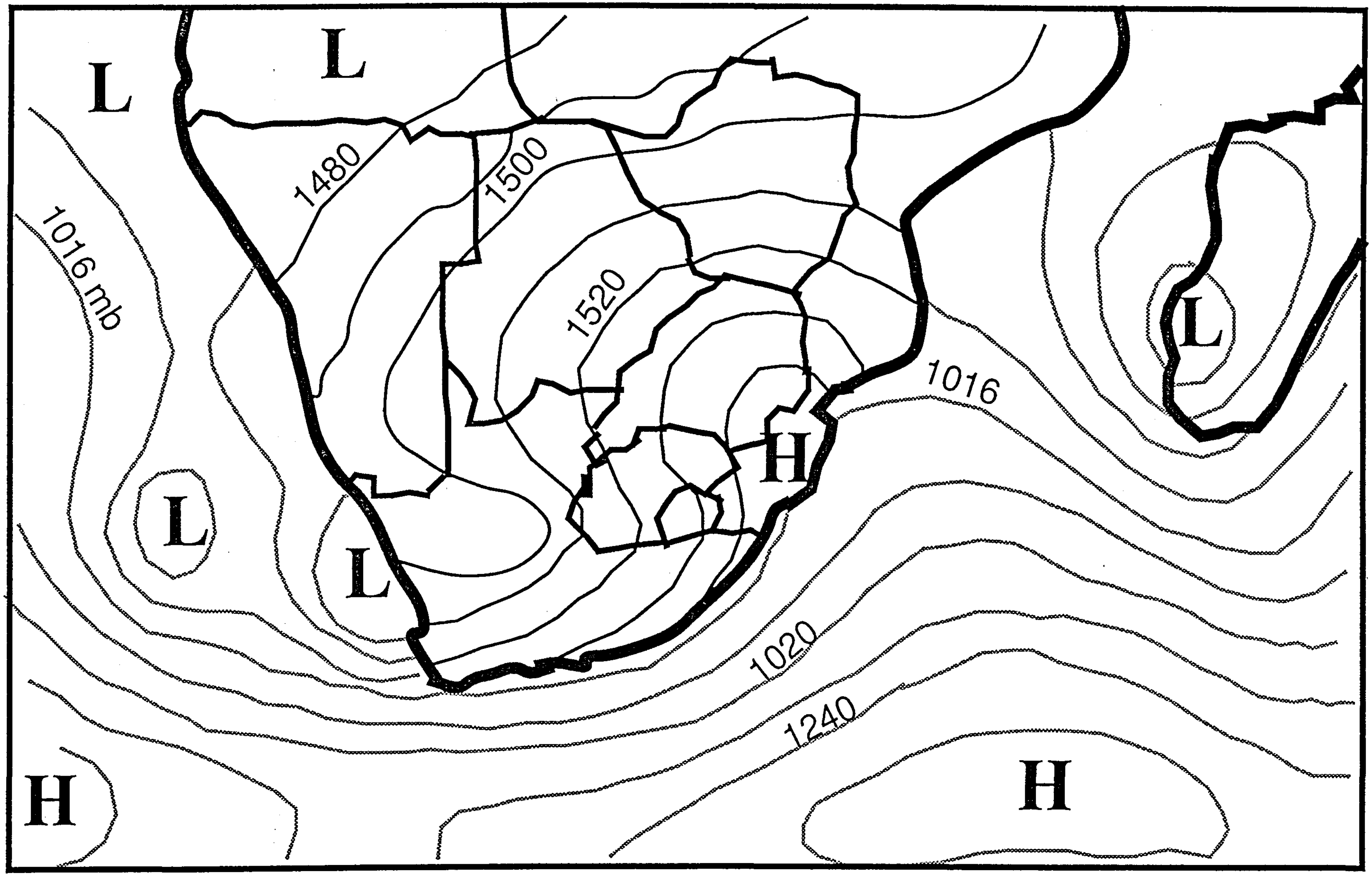


Figure 4.2 Synoptic chart for 28 November 1985 (after Daily Weather Bulletin, Weather Bureau).

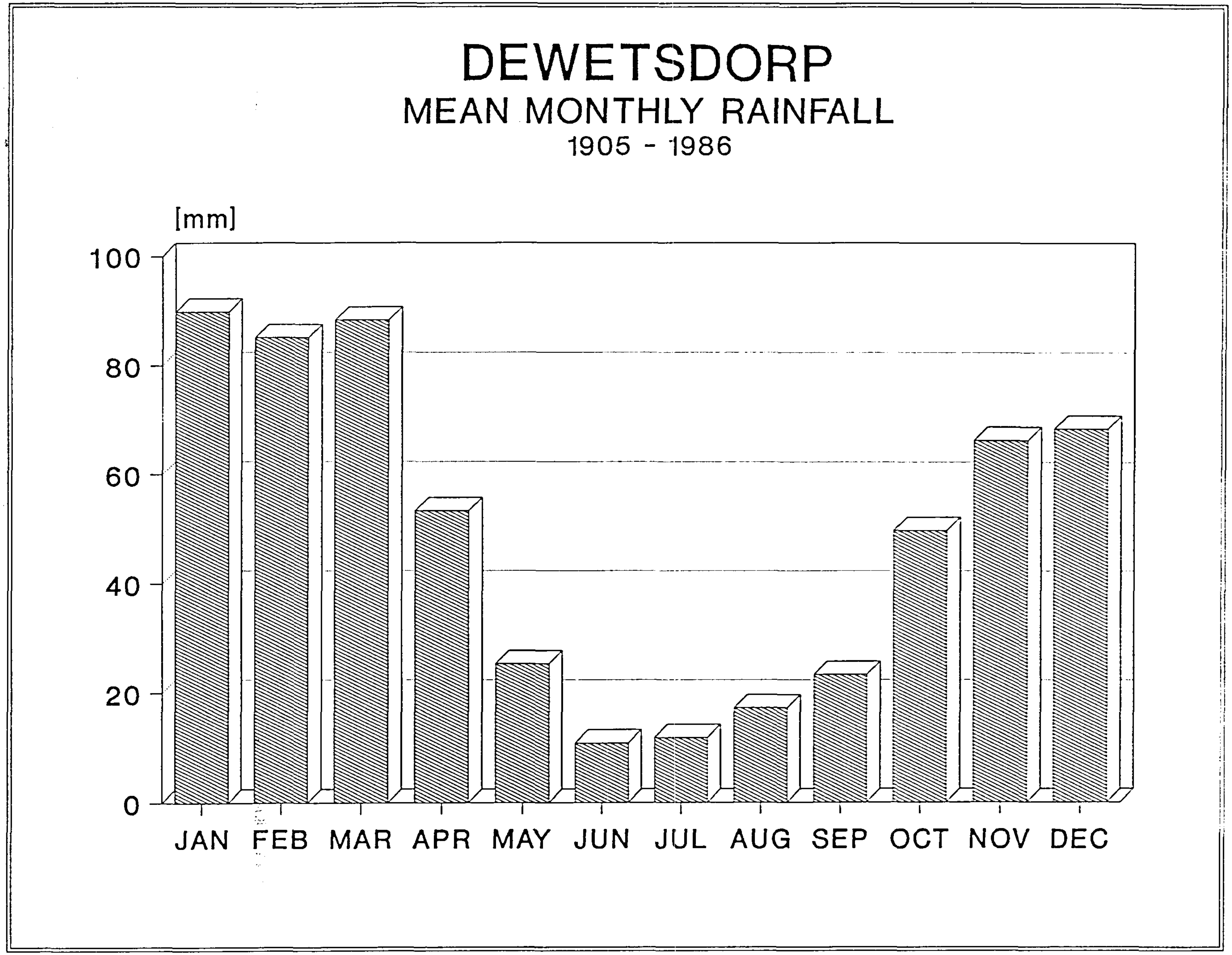


Figure 4.3 Mean monthly rainfall for Dewetsdorp.

wind-speed of less than approximately 20 km/h (Schulze, B.R., 1986). Slightly higher speeds which rarely exceed 40 km/h are experienced between August and January.

Values for the long term average monthly evaporation [mm/d] and the total annual evaporation [mm/a] for a Class A Pan, as extracted from Weather Bureau data for the two stations nearest to Dewetsdorp, are given in the table below. They vary between ±2,5 mm/d in June and a maximum of approximately 9 to 9,5 mm/d in November or December.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	*YEAR*
Bloemftn.	9,4	8,3	6,2	4,6	3,1	2,6	3,0	4,8	7,3	8,6	9,8	9,7	*2350*
Wepener	8,5	8,1	5,7	4,3	3,0	2,3	2,7	4,0	6,1	7,7	8,8	8,8	*2120*

Much effort has been put into obtaining meteorological data during this project. A network consisting of a weather station, two automatic rainfall recorders and a rain meter were set up. The weather station was erected at the municipal sewage works. It recorded rainfall at five minute intervals and hourly mean values for the following time dependent variables: wind velocity, radiation, humidity and temperature. The station also collected separate rainfall samples of each 5 mm that fell (see Figure 4.4).

The one rainfall recorder was installed above the village along the Reddersburg Road and the other in the lower part of the research area along the road to the homestead of the farm, Frankfürt. Available on a daily basis, were the records of the Dewetsdorp Police Station, Weather Bureau Station No. 322275, since 1905. Outside the research area, west of the watershed, rainfall has been measured since September 1986 at the homestead of the farm, Glen Heights.

In the light of later experience, it seems doubtful whether all the additional meteorological data generated, helped to obtain a better estimate of the aquifer recharge. The main reason for data losses was that operators were not stationed in the research areas. Recorders, especially the tipping bucket with weekly charts, could not always be serviced in time. They were also occasionally blocked. Solar panels should

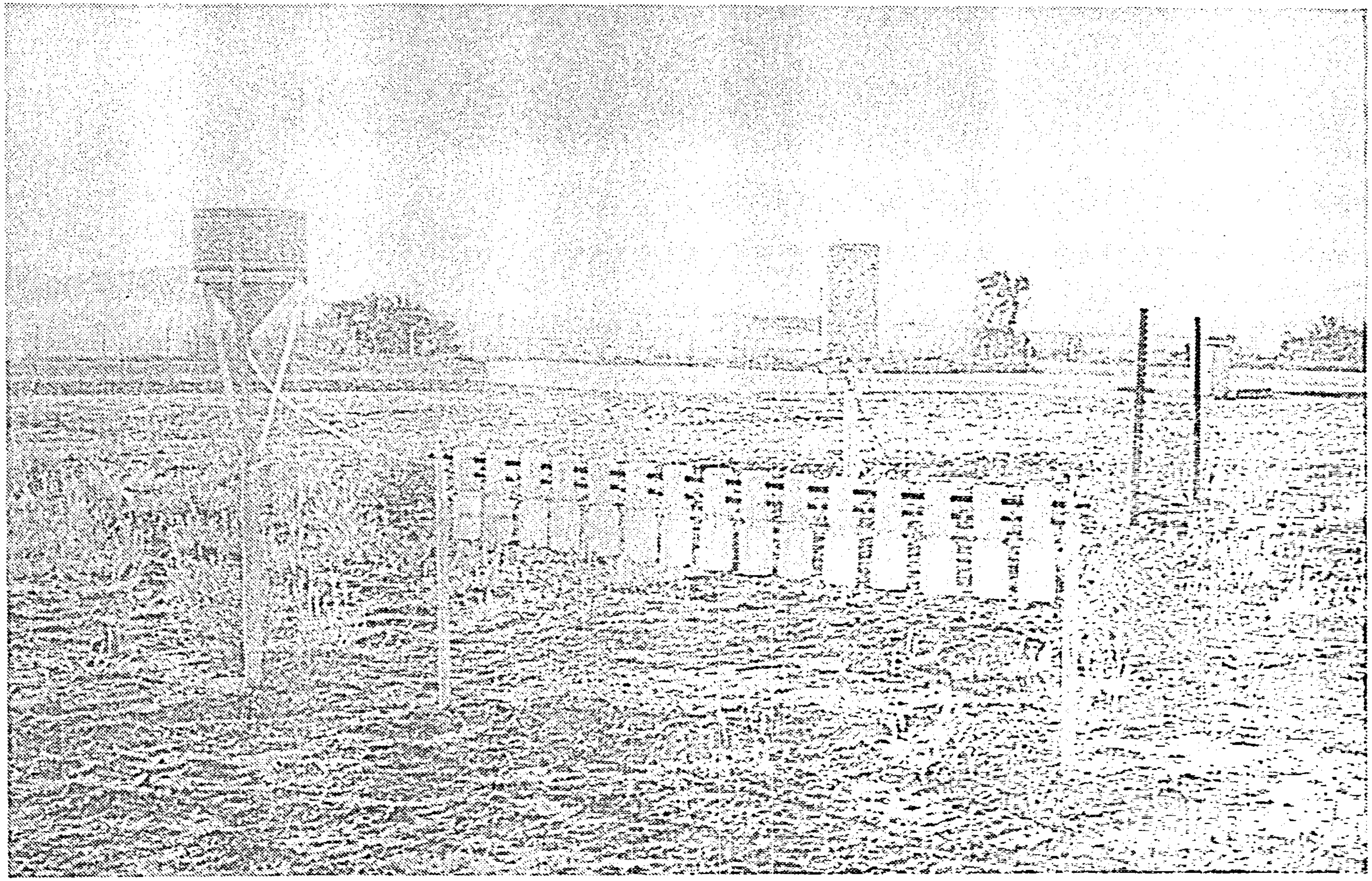


Figure 4.4 Automatic rainfall sampler.

be used for the power supply of data loggers to avoid loss due to battery failure. Especially during 1986 and 1987, rainfall data for the weather station were lost.

It is known that the spatial rainfall distribution for individual occurrences is rather variable, while the total precipitation during a rainy spell varies much less. The mean annual precipitation shows even less variations. The spatial variability is illustrated in Figure 4.5 for the November and December rains of 1985 that were recorded in Dewetsdorp. For comparison the De Aar data are shown in Figure 4.6. It is impossible to link the water-level rise in a certain borehole to a specific rainfall event which was measured in a certain recorder; instead the total rainfall is considered which fell during a certain wet period, which shows less local variation. The most important recharge, as will be shown later, occurs during exceptional high rainfall events caused by general rain with a more equal distribution rather than by thunderstorms. The total rainfall can be equally well or better assessed from a few uninterrupted, than from numerous occasionally patched records.

4.1.2.1 RAINFALL

In ground-water potential calculations, it is common practice to establish a relationship between the annual rainfall and the annual ground-water recharge. The yearly exploitable ground-water volume is then defined as the average annual recharge in terms of a certain percentage of the average annual rainfall. As far as arid and semi-arid regions are concerned, this approach could be erroneous in two aspects. Firstly, the annual rainfall is highly variable and, therefore, the ground-water supply can be depleted seriously during a low rainfall period of several consecutive years, especially if the storage is limited. Secondly, the annual recharge is not merely dependent upon the annual rainfall depth, but upon the annual distribution, magnitude and intensity of storm events as well.

Observing 82 storm events in the Negev desert, Asher (1986, personal communication) showed that rainfall occurs in bursts of pulses with different wave forms and some random variations. The rainfall rate increases steadily, until it reaches a certain maximum after several minutes and then decreases according to a recession curve. Asher defined the annual rainfall in terms of three independent stochastic

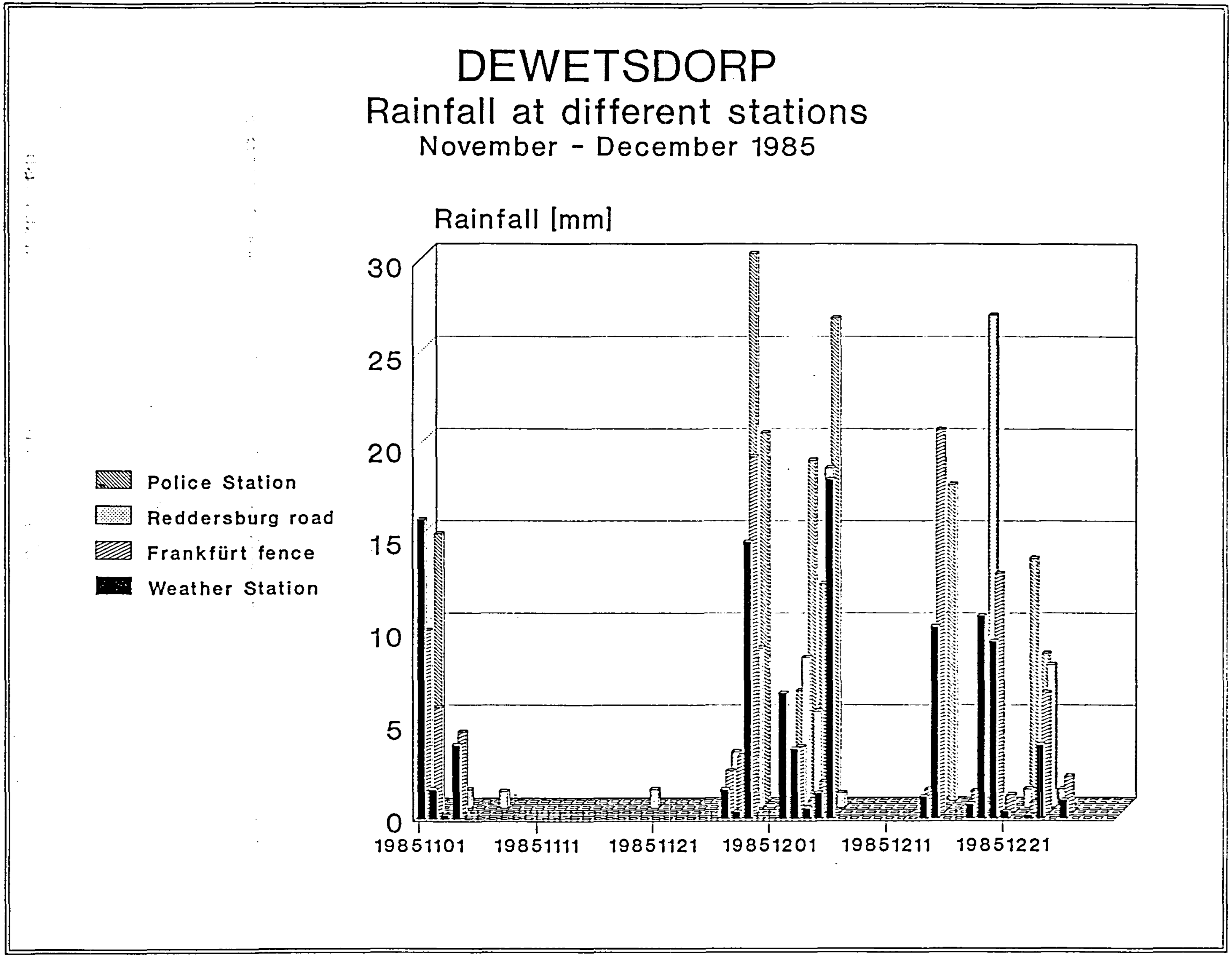


Figure 4.5 Rainfall for November and December 1985 measured at different stations in the Dewetsdorp area.

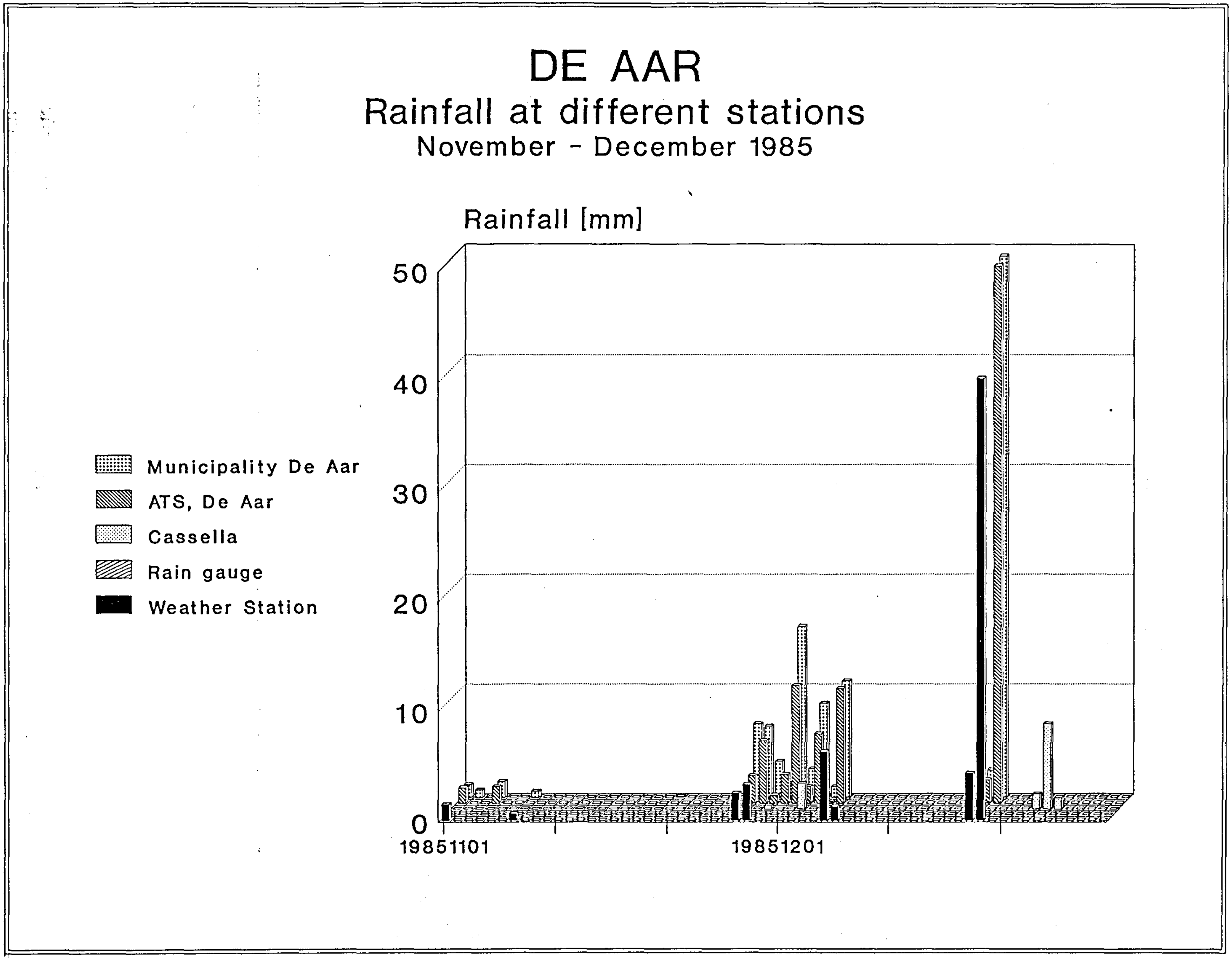


Figure 4.6 Rainfall for November and December 1985 measured at different stations in the De Aar area.

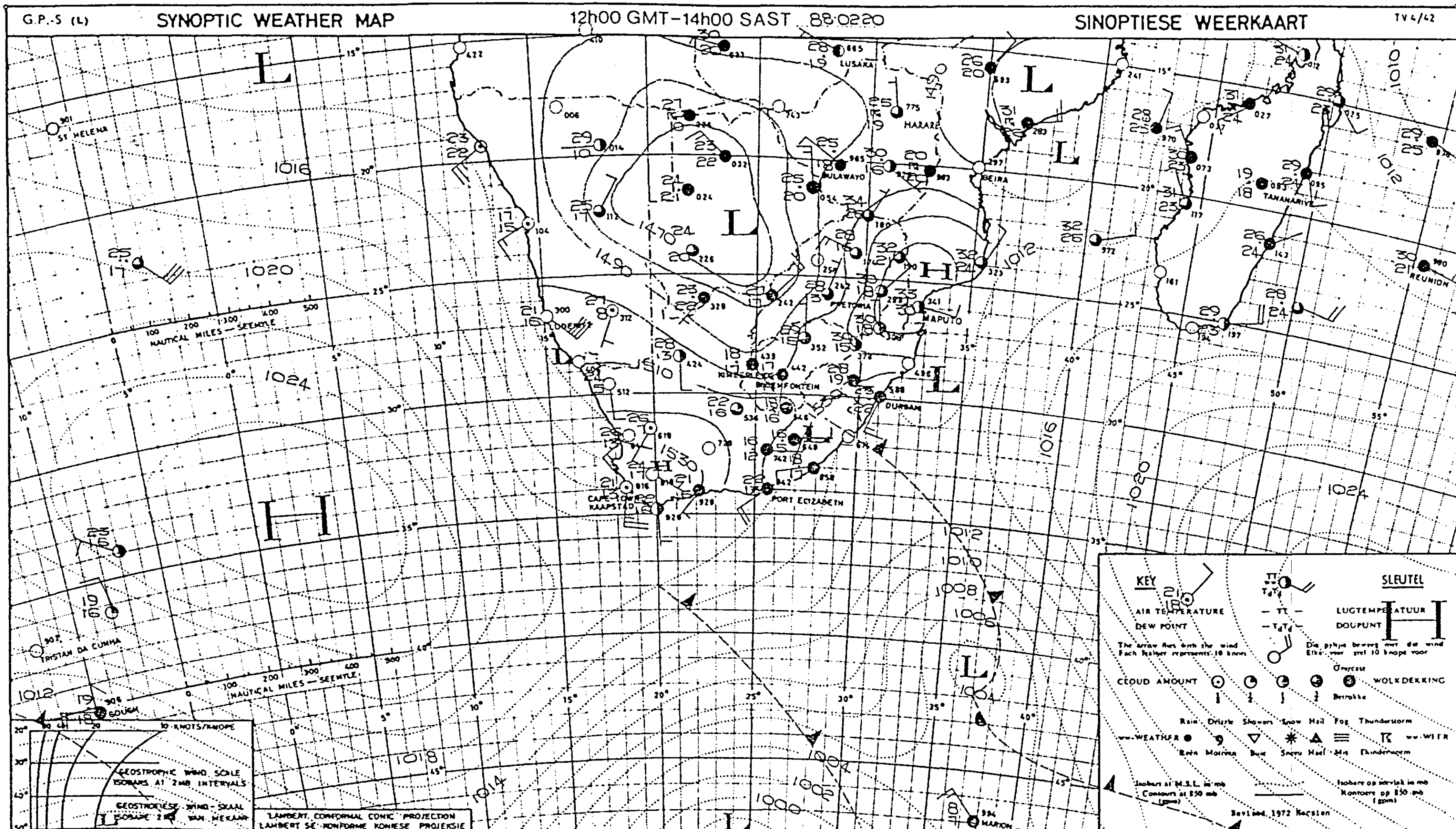
parameters, e.g.: (a) the number of storms per year, (b) the duration of each storm and (c) an intensity parameter, which can be derived from an intensity duration probability analysis. One of the main problems encountered was to determine the intensity duration relationship for storms with different recurrence intervals, because the rainfall intensity records, covering a sufficient number of years, were not available for the study area.

The temporal distribution of rainfall in the Karoo (i.e. short convectional showers characterized by a highly variable intensity during an individual storm and long duration rainfall events with a more or less uniform intensity) will probably be different from the stochastic nature of rainfall intensity and depth in the Negev.

The inherent variability of annual rainfall in the Karoo is of importance in many applications from agriculture to water resources planning and is usually mapped as the coefficient of variation (C.V.) of the annual rainfalls (Zucchini and Adamson, 1984). The mean annual rainfall for Dewetsdorp, as measured from 1905 to 1987 at the Police Station, amounts to 587 mm (median = 564 and standard deviation = 152) with a C.V. of 26%. The minimum annual rainfall during this period was 341 mm in 1919 and the maximum 957 mm in 1925. The rainfall between October 1985 and September 1986 equaled 564 mm and for the consecutive year it was 612 mm. Except for the flood event, the rainfall could be considered as quite normal during the study period. Precipitation in the 1987/88 season amounted to 1075 mm and during the period 1st November 1987 to 31st October 1988, it was even higher at nearly 1200 mm.

The reason for this high record was, of course, the February, 1988, rain. After an extended dry spell with soaring temperatures (JBM Hertzog Airport at Bloemfontein, measured 38,9 °C on 4th February) soaking rains started on 17th February. A low over northern Botswana fed in moist air, which was undercut by cool air from a high south of the country (see Figure 4.7). Within 48 hours, rain exceeding 250 mm fell within a ±40 km wide band, stretching from south of Vryburg to Smithfield (see Figure 4.8).

A probability analysis was performed on the maximum 30-day rainfall amount per year. From Figure 4.9, it can be seen that the 417 mm of the February 1988 flood, has a return period of 500 years. The non-exceedence probability for a specific annual rainfall is given in Figure 4.10, from which it can be seen that the probability of a



INFERENCE The low over Botswana and the frontal trough over the south-eastern parts is causing scattered showers and thundershowers over the northern, central and south-eastern parts of the country. These conditions will spread to the eastern parts tomorrow. Heavy falls are expected at places. Gradual clearing will take place along the south coast.

AFLEIDING Die laag oor Botswana en frontale trog oor die suidoostelike dele veroorsaak verspreide buie en donderbuie oor die noordelike, sentrale en suidoostelike dele van die land. Die toestande sal na die oostelike dele môre uitbrei. Swaar neerslae word op plekke verwag. Geleidelike opklaring sal môre langs die suidkus plaasvind.

WEERBURO—DEPARTEMENT VAN OMGEWINGSAKE

WEATHER BUREAU—DEPARTMENT OF ENVIRONMENT AFFAIRS

Figure 4.7 Synoptic weather map for 20 February 1988.

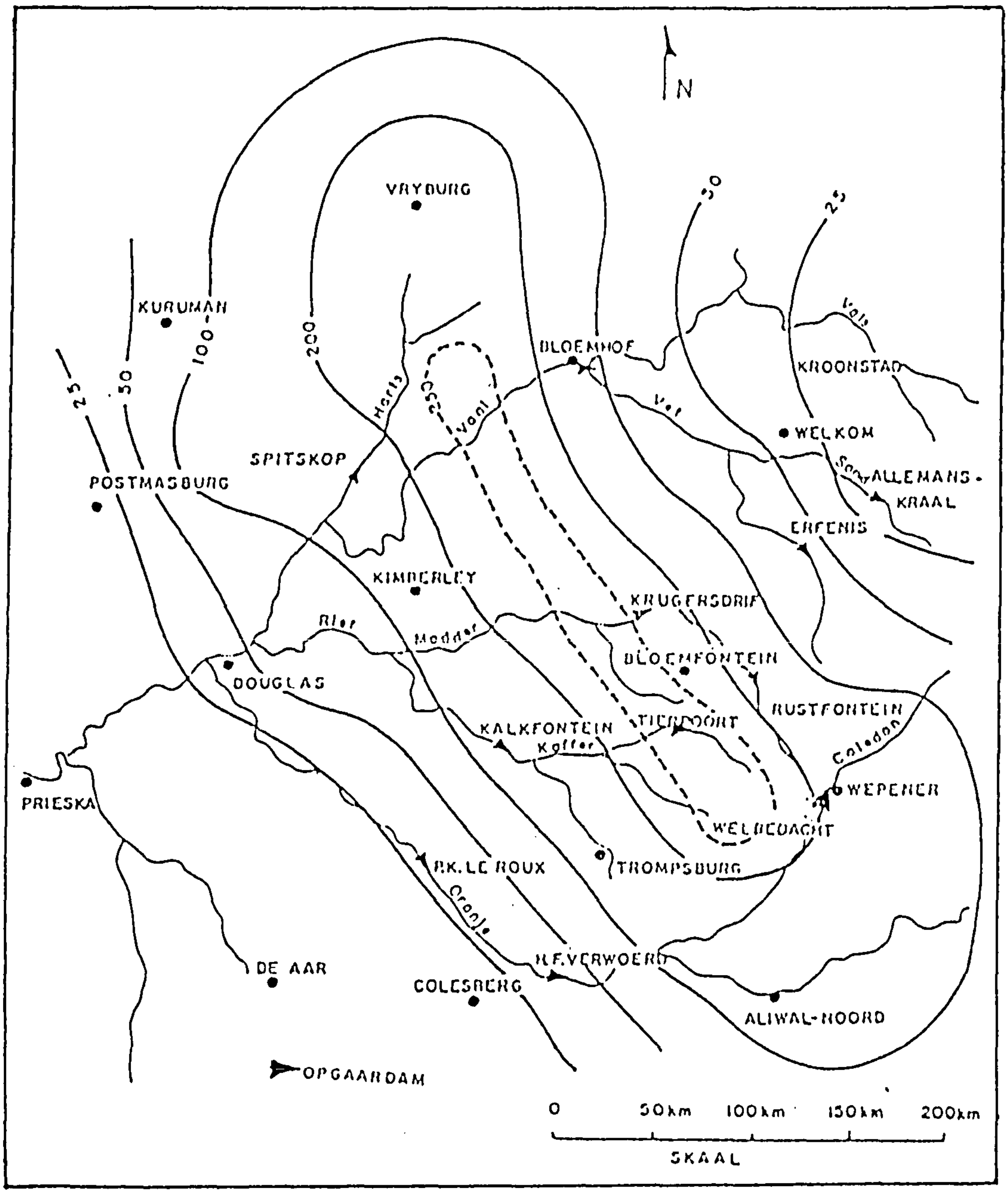


Figure 4.8 Rainfall in millimetres for the 48-hour period ending 8:00 on 22 February 1988 (after Du Preez, 1988).

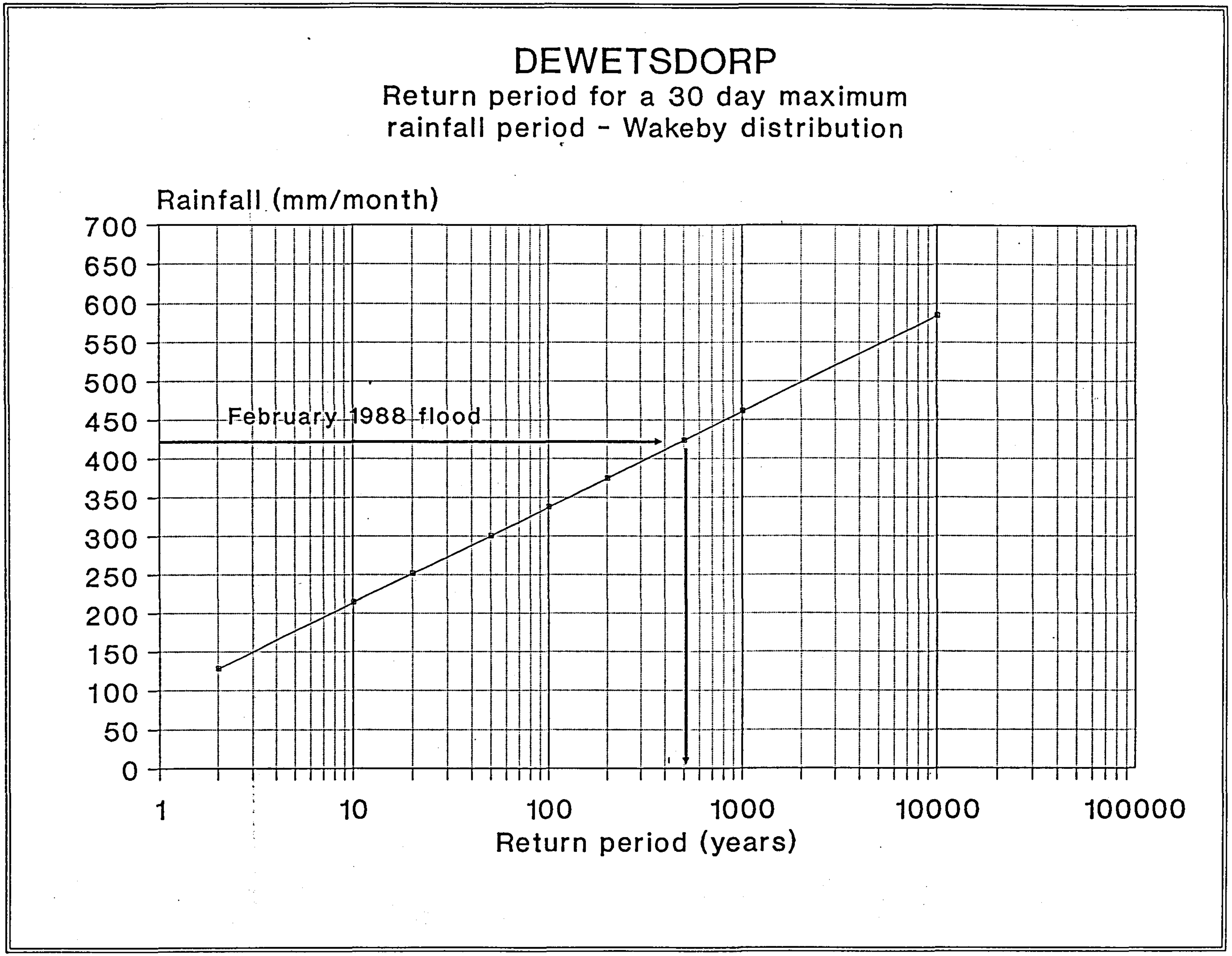


Figure 4.9 Return period for above average rainfall for a 30-day period for Dewetsdorp. It can be seen that the February 1988 flood has a return period of 500 years.

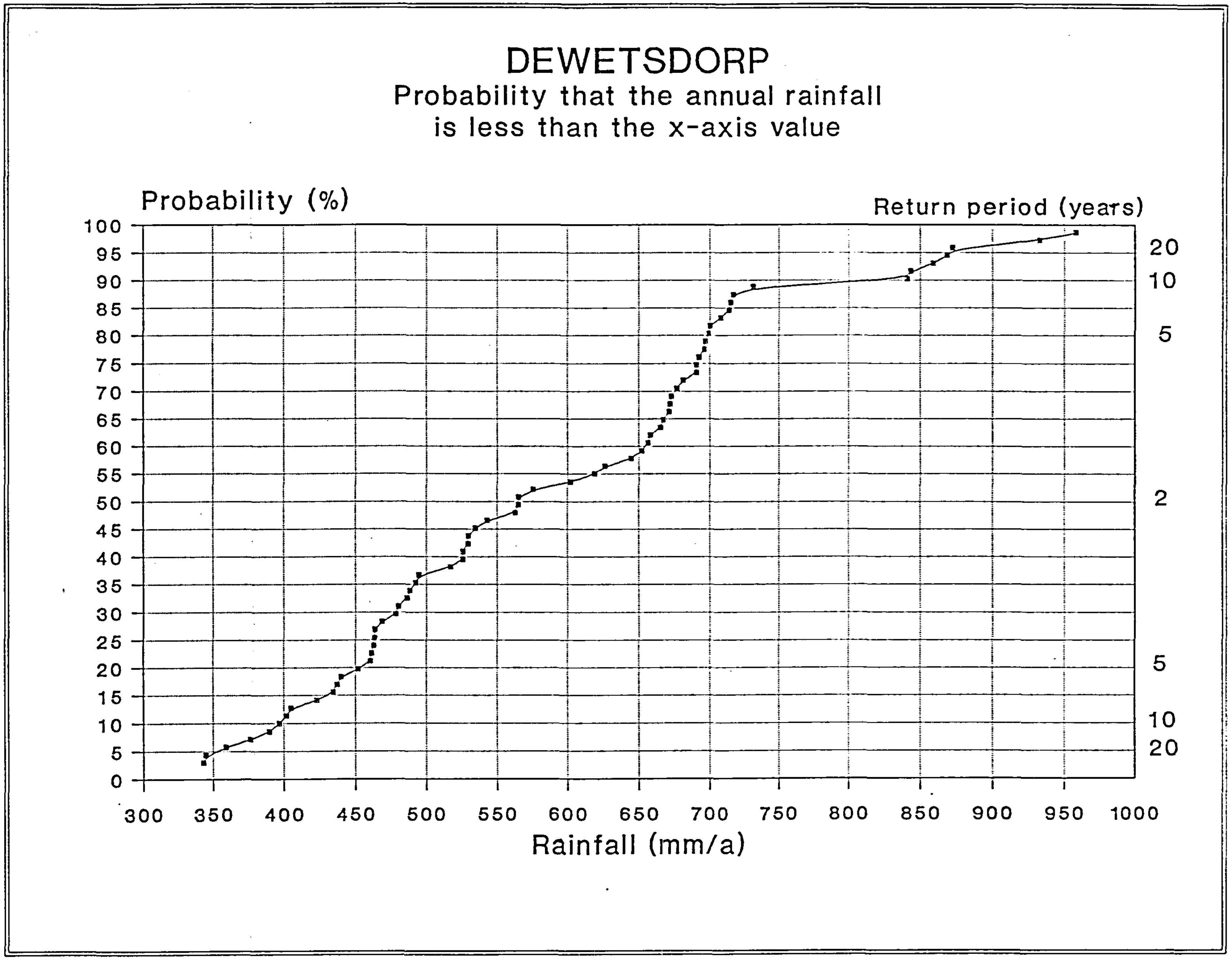


Figure 4.10 Probability that the annual rainfall in Dewetsdorp is less than the value on the x-axis.

rainfall, of say higher than 700 mm/a to occur in any year, is 20%. For the determination of the minimum recharge, geohydrologists are, however, more interested in the probability of occurrence of a specific minimum annual rainfall. In the case of Dewetsdorp, Figure 4.11 can be used for this purpose. From this figure, it can be seen that Dewetsdorp may expect an annual rainfall of less than 420 mm, once every ten years on average. Return periods for 2- and 3-year droughts are also shown. To conclude with the probability analyses, the rainfall events of every year were divided into three four-months periods, namely the January-April, May-August and September-December period, and a probability analysis were performed on these data. The results in Figure 4.12 show that the period January to April has a higher recurrence rainfall value than the other periods.

In Figure 4.13, a binominal filter, as used by Dyer, is applied to the annual rainfall data of Dewetsdorp. Although not very obvious the figure shows that there exists a quasi-18-year cycle between wet and dry periods. Wet and dry years can be distinguished by the frequency of rainy spells of four days and longer (Tyson, 1986).

Taking into account that some rainfall data were lost, Figure 4.14 shows that differences in the monthly totals of four rainfall stations in the Dewetsdorp study area were between ±10 and ±60 mm with an average of approximately 30 to 40 mm during the wet months. If one looks at the accumulated totals in Figure 4.15, which cover records of different lengths, it is evident that the general slope of the various records during the rainy spells is nearly the same. The deviation of the annual precipitation of the highest lying station at Glen Heights from that of the Police Station was +72 mm or +12% in 1987 and -20,5 mm or - 2% between January and October 1988. Because the differences are comparatively small, it is generally not possible to link water-level rises in a borehole to a specific rainfall event measured at a specific station. For future calculations of the exploitation potential elsewhere, use will have to be made of the records of the next lying rainfall station. For the calculations in this study the data from the Police Station were used.

4.1.2.2 EVAPOTRANSPIRATION

There are various processes through which water can be added or removed from the unsaturated zone of the soil: mainly infiltration percolation, capillary rise, transpiration

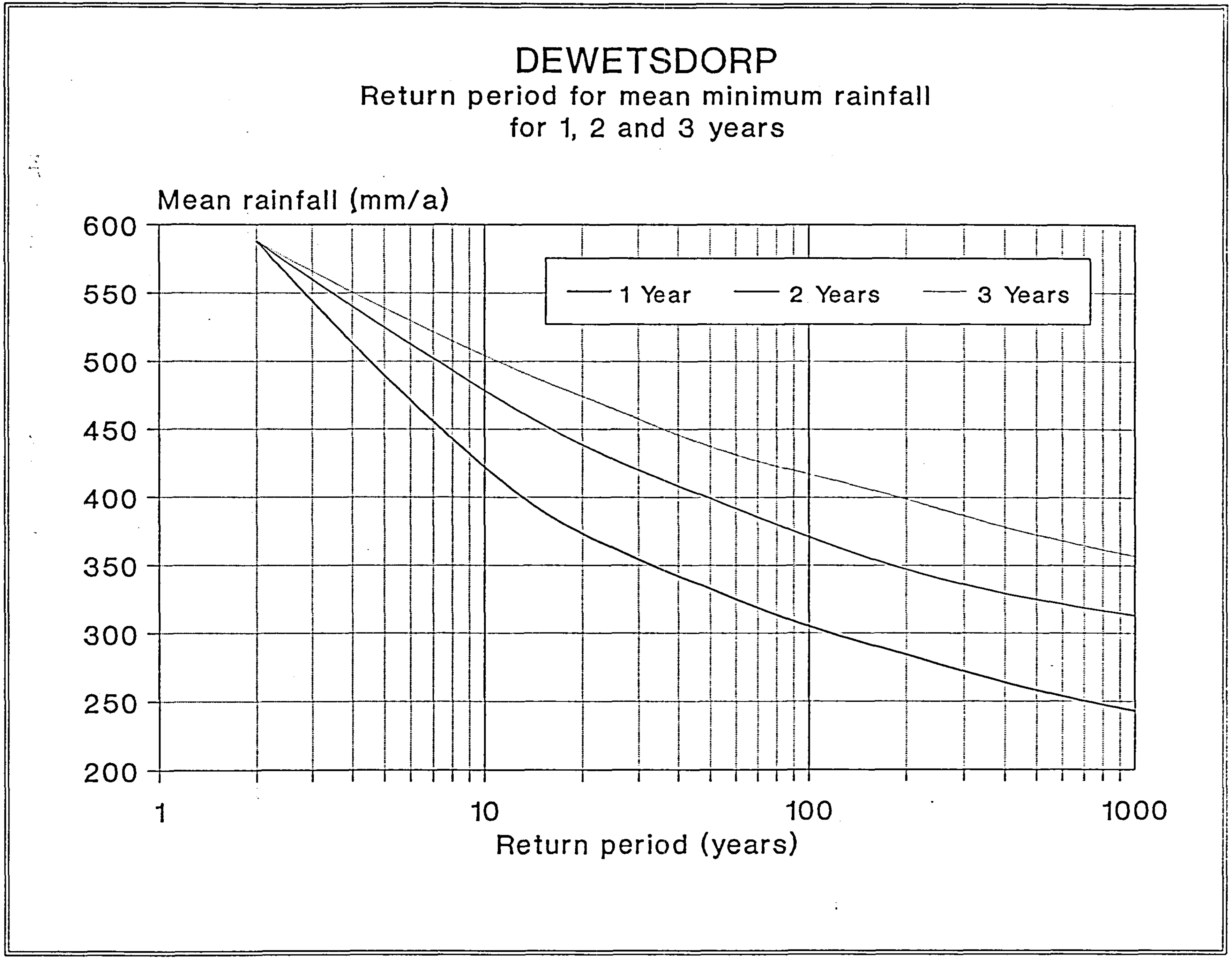


Figure 4.11 Return periods for minimum mean annual rainfall for 1-, 2- and 3-year periods.

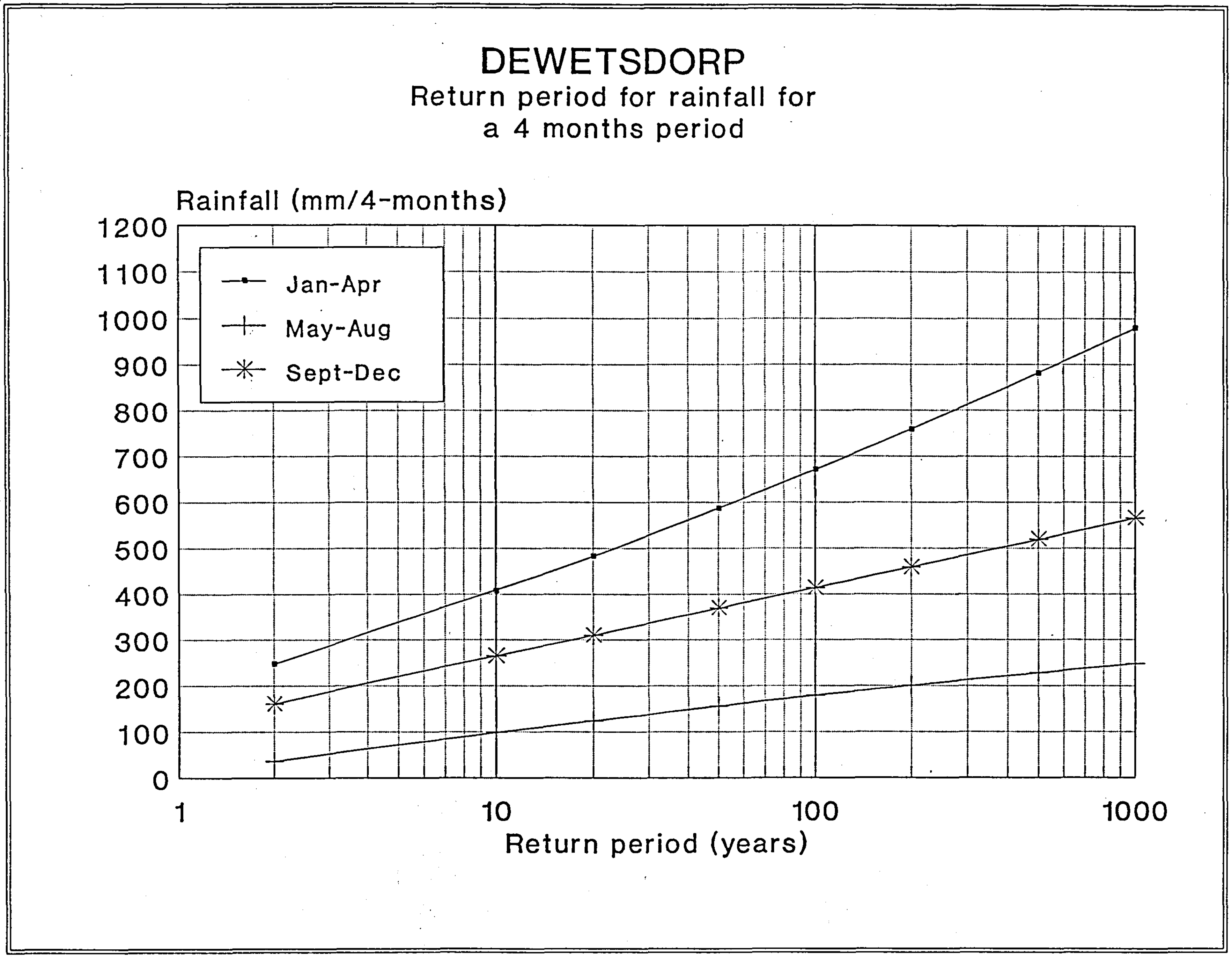


Figure 4.12 Return period for maximum rainfall for the months January to April, May to August and September to December.

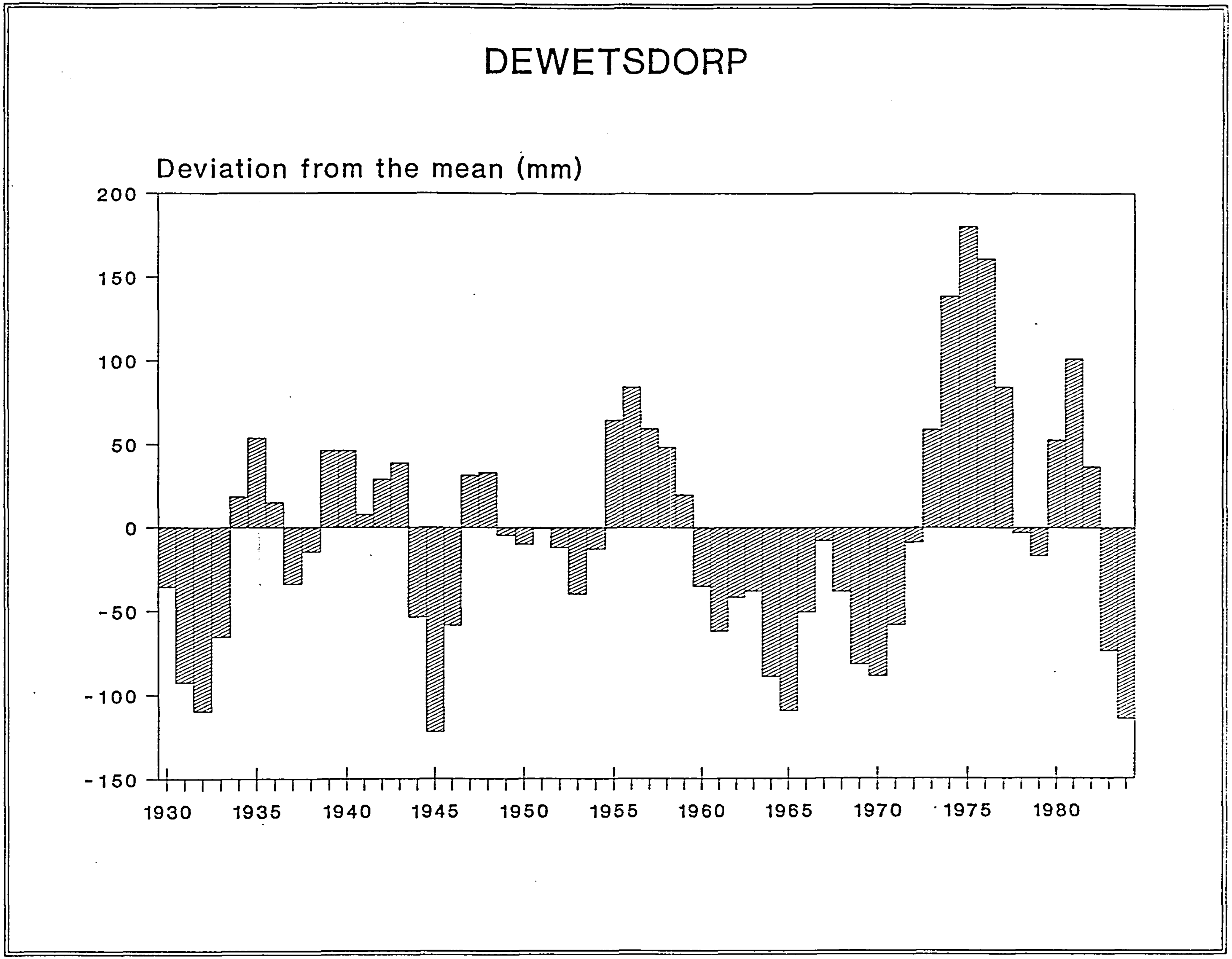


Figure 4.13 Dyer formula applied to the annual rainfall of Dewetsdorp.

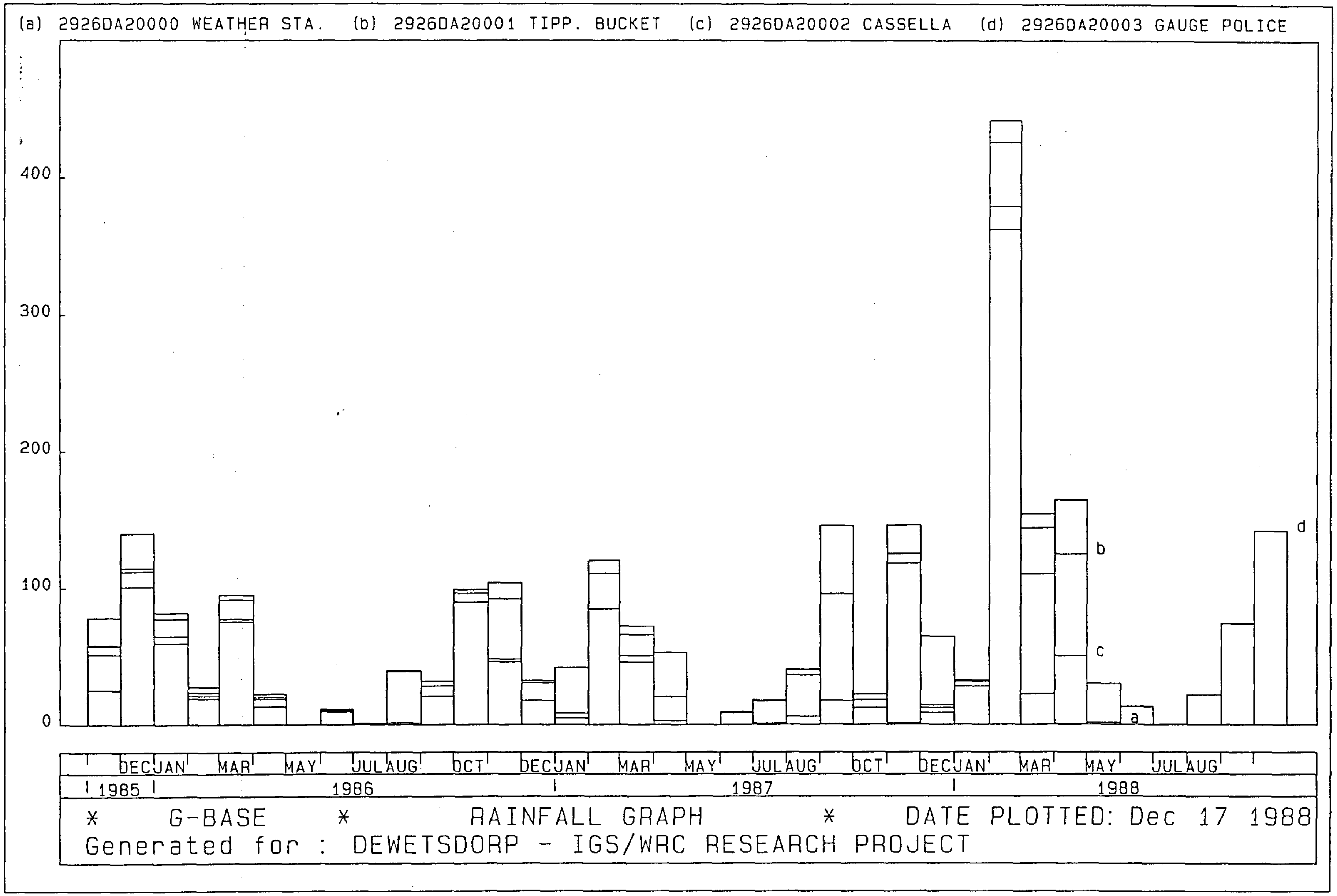


Figure 4.14 Total monthly rainfall between November 1985 and October 1988, measured at four stations.

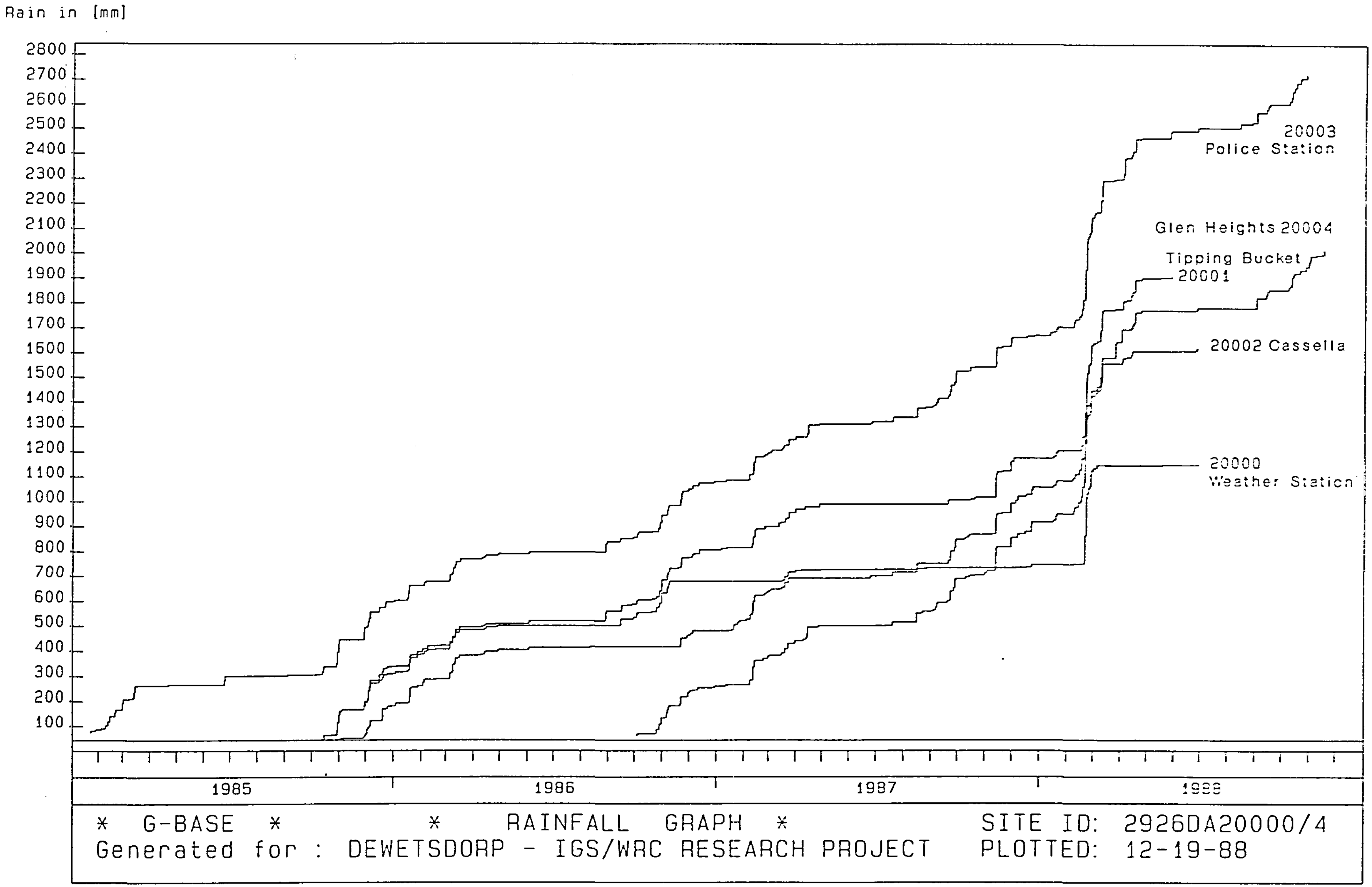


Figure 4.15 Cumulative rainfall measured at five stations in the Dewetsdorp area.

and evapotranspiration. In applying the water balance, it is necessary to obtain estimates of these parameters.

The potential evapotranspiration is the maximum amount of vapour which could be transferred from an area to the atmosphere under the existing meteorological conditions, whereas the actual evapotranspiration is the actual amount of vapour transferred to the atmosphere, which depends not only on existing meteorological conditions, but also on the availability of water to meet the atmospheric demand and, in the case of vegetation, its ability to extract water from the soil (Kijne, 1974).

Estimations of potential evapotranspiration are usually based on Penman-type equations. These equations need a wide spectrum of energy data, which are not always readily available. Linacre (1977) attempted an approximation of the Penman equation for cropped surfaces by involving temperature only:

$$PE\ (\text{mm/month}) = \frac{[650 \cdot T/(85\text{-Alat})\text{-}56 + (5+4W) \cdot C] \cdot D}{(80\text{-}T_a)}$$

where

W	= wind-speed in [m/s]
T	$= T_a + 0{,}006$ ELEV
T_a	= mean monthly air temperature [°C]
ELEV	= elevation above sea level [m]
D	= number of days in month (i)
Alat	= latitude in degrees
C	$= 0{,}0023\text{ELEV} + 0{,}37T_a + 0{,}53R + 0{,}35R_{an} - 10{,}9$ [°C]
R_{an}	= difference between the mean temperature of the hottest and coldest month of the year [°C]
R	= mean daily range of temperature [°C]

This equation has been tested by Schulze (1983) and was reported to give markedly more reliable simulations of A-pan values, when compared with other temperature-based equations commonly in use.

Figure 4.16 shows the potential evapotranspiration for Dewetsdorp, as calculated with the Linacre formula. During the study period, it varied between ±70 and 240 mm/month.

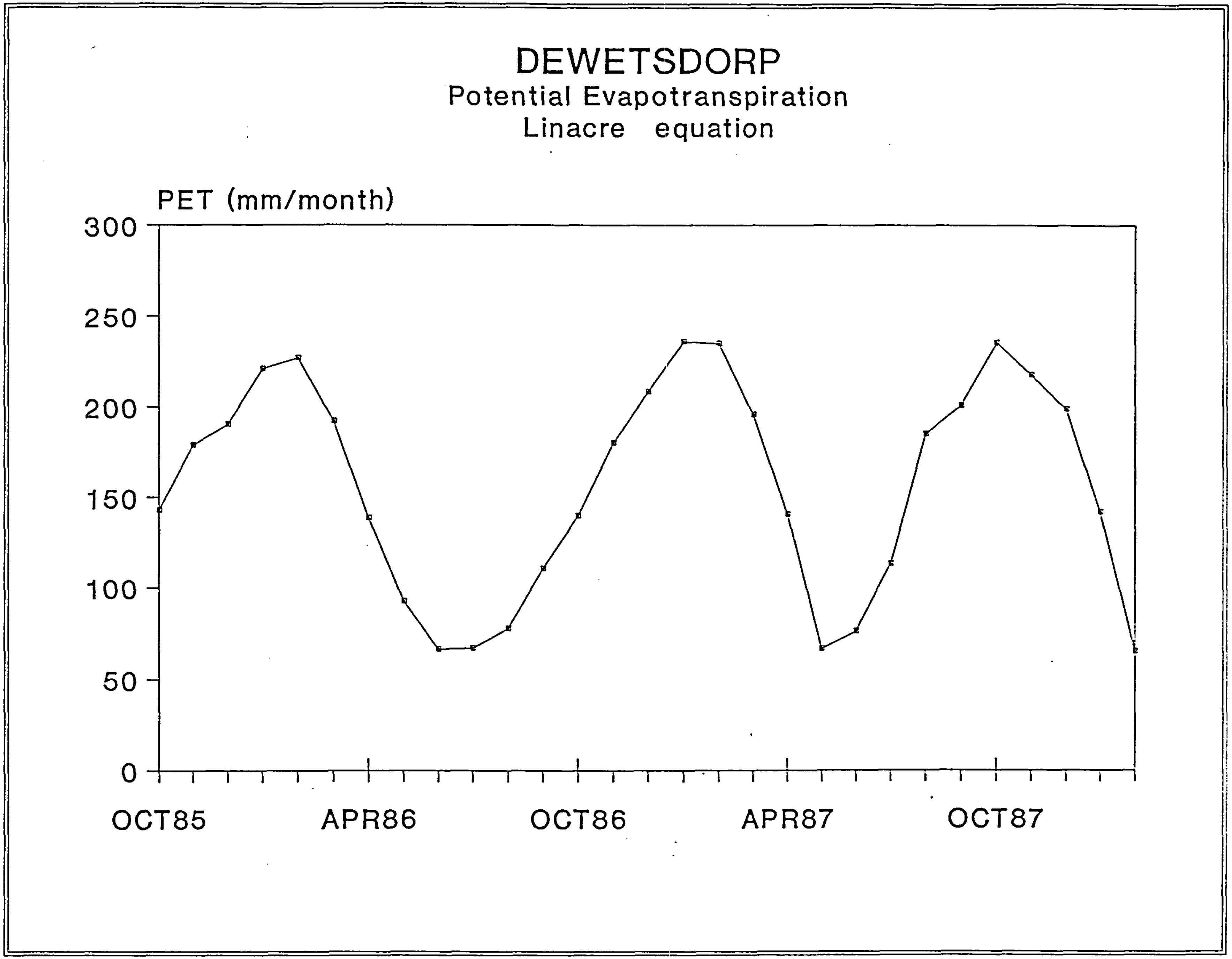


Figure 4.16 The potential evapotranspiration at Dewetsdorp for the period October 1985 to February 1988.

4.1.3 GEOLOGY

SOIL

Soil types encountered in the Dewetsdorp area are:

MAPPING UNIT	SOIL FORM	SOIL SERIES
M	Mispah	Mispah
O	Oakleaf	Limpopo
Sw	Swartland	Nyoka
S1 and S2	Sterkspruit	Sterkspruit
E	Estcourt	Rosemead
V	Valsrivier	Lindley
B	Bonheim	Bonheim

The S1 phase has a higher clay content in the A-horizon than the S2 phase.

The results of the soil survey are shown on Figure 4.17. The mapped area is in the land type Dc, i.e. *Prismacutanic* and *pedocutanic* B-horizons are dominant with vertic, melanic or red-structured diagnostic horizons, while *neocutanic* horizons are scarce. The soil definitely follows a topographical pattern.

The high-lying dolerite hillocks are covered with shallow soil of the MISPAH form. Depending on texture and gradient, it absorbs water readily. It forms an interrupted thin layer and retains little water.

With decreasing gradient, the SWARTLAND form appears, followed by the STERKSPRUIT and the ESTCOURT form. The latter covers the lowest lying areas. These soils are thicker and are probably more important as water reservoirs.

The secondary watercourses are bordered by BONHEIM soils. Along the major watercourses, BONHEIM and VALSRIVIER soils alternate.

Buried soil occurs where road building material has been spread. This is marked with an 'X' on the map.

The OAKLEAF form covers a small patch where colluvial transport has caused accumulation of material.

SWARTLAND, STERKSPRUIT and VALSRIVIER forms are soils with a low infiltration capacity. They have a reasonable thickness (± 100 cm) and rest on thick layers of SAPROLITE (± 50 cm) and alluvial material (± 100 cm).

Lateral flow occurs in ESTCOURT soil and is also possible in SAPROLITE. It is unlikely in the other forms.

Profile descriptions and analytical data are contained in the Appendix.

HARD ROCK

The area has not been mapped in detail previously. The only geological map available was the 1 : 250 000 sheet, Bloemfontein, with its outdated lithostratigraphical classification. It was therefore necessary to map the study area in collaboration with the Department of Geology at the UOFS.

The area is underlain by sediments of the Beaufort Group. This group can be subdivided into the lower Adelaide- and the upper Tarkastad Subgroup (SACS 1980). The Balfour Formation is part of the former and is found in the north-east of the study area, while the Katberg Sandstone Formation forms the base of the Tarkastad Subgroup and is found in the south-west. The Balfour Formation consists of fine-grained cross-bedded sandstone and coarse arkose, alternating with greenish-, bluish grey and greyish red mudstone, which reflect a cyclical succession of channel- and flood plain deposits typical for the Beaufort Group. The sharp contrast in elevation within the area is caused by the geology: the higher lying parts are underlain by the Katberg Sandstone Formation which is predominantly arenaceous and harder than the argillaceous Balfour Formation. It offers more resistance to weathering. It is assumed that the boundary between the Katberg and the Balfour Formations coincides with the base of a prominent sandstone layer noticeable in the field, because of the change in the topographical gradient. This morphological feature can be distinguished on the stereo photographs and follows more or less the 1500 m contour line.

Dolerite sills occur on the hills west and north-west of the village. A number of prominent dolerite dykes are present in the area. They crop out in the high-lying areas and in the watercourses where they can be easily detected on the aerial photos. In the lower parts, they are frequently masked by thick layers of soil where they had to be detected and traced by magnetometer. They strike in various directions. These dykes give rise to springs in the upper parts of the Townlands, i.e. on Kareefontein 66 and to a number of good yielding boreholes which have been sited along them (see Figure 4.18).

4.1.4 GEOHYDROLOGY

Information on the regional geohydrological features has not been collected. The geohydrology of the research area is delt with under Section 4.1.5.3.

SOIL MAP : DEWETSDORP

LEGEND

- MISPAH
- SWARTLAND
- STERKSPRUIT 1
- STERKSPRUIT 2
- ESTCOURT
- VALSRIVIER
- BONHEIM
- OAKLEAF
- FOREIGN MATERIAL

SCALE

0 250 500 1 000 1 500 m

Figure 4.17

GEOLOGICAL MAP : DEWETSDORP

LEGEND

BEAUFORT SEDIMENTS covered by soil

DOLERITE DYKE

DOLERITE

KATBERG SANDSTONE FORMATION

Expected boundary between Balfour- and Katberg Sandstone Formations

Watershed

Test borehole

Observation borehole

SCALE

0 250 500 1 000 1 500 m

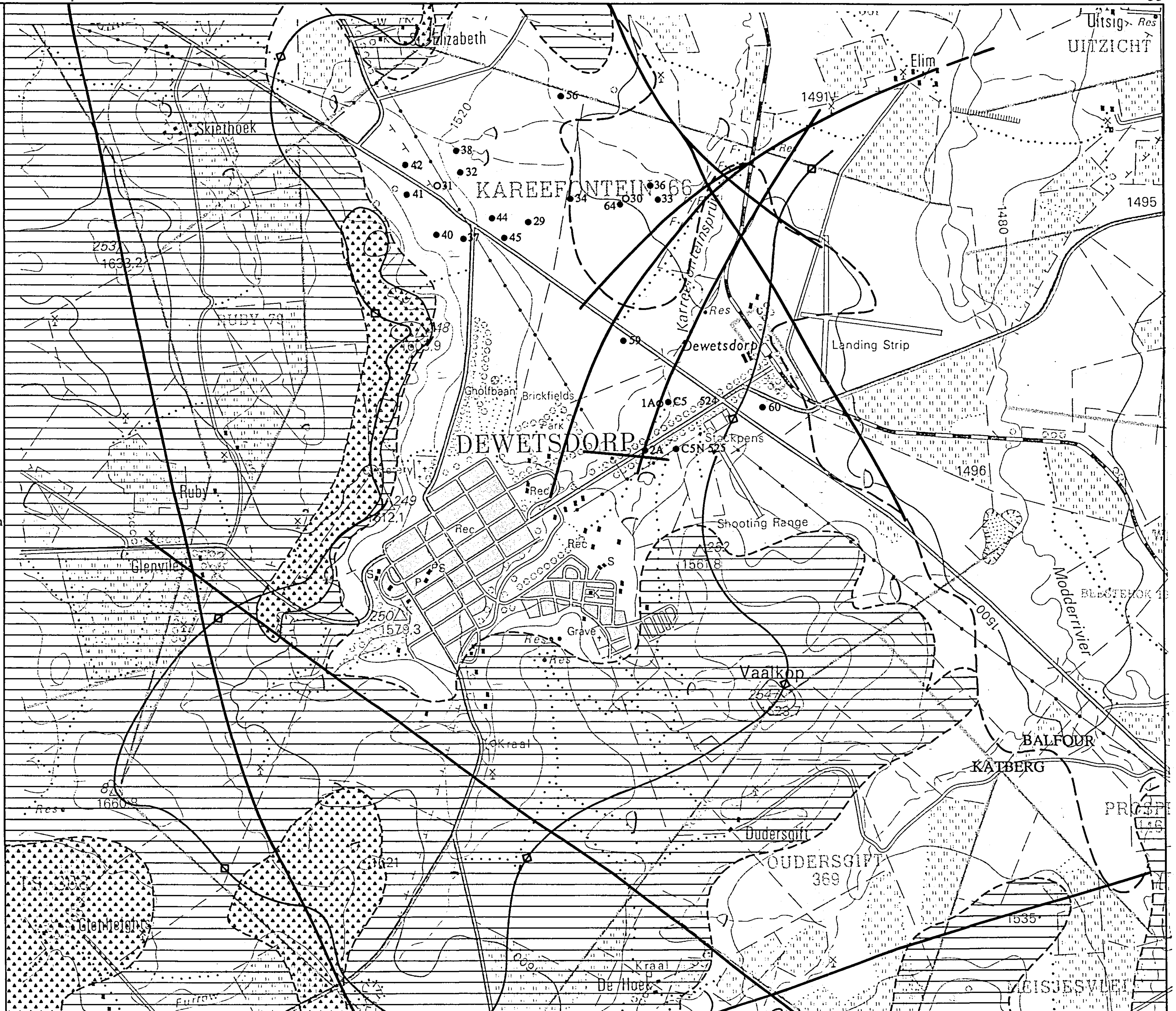


Figure 4.18

4.1.5 MEASURING NETWORK

4.1.5.1 UNSATURATED ZONE

For the determination of the recharge through the soil cover, twenty 2,0 m deep neutron probe access tubes were installed in the lower flat part of the research area, allowing measurement to a depth of up to 1,8 m. Thirteen sites were selected along two profiles, one in the assumed approximate direction of the ground-water flow from G 36439 to the municipal production hole 3 A and a second ± parallel to the ground-water contours along the Main Road, Bloemfontein - Wepener. The other seven tubes were installed over the remainder of the area to obtain a representative coverage of the area (see Figure 4.19).

Mispah and Oakleaf soils were not monitored, because of either insufficient thickness of the soil in the case of Mispah or insignificant representation (Oakleaf). Swartland (Sw) and Sterkspruit (S1 and S2) are the predominantly occurring soils. This is reflected in the following table which shows the soils in which the neutron probe access tubes were installed.

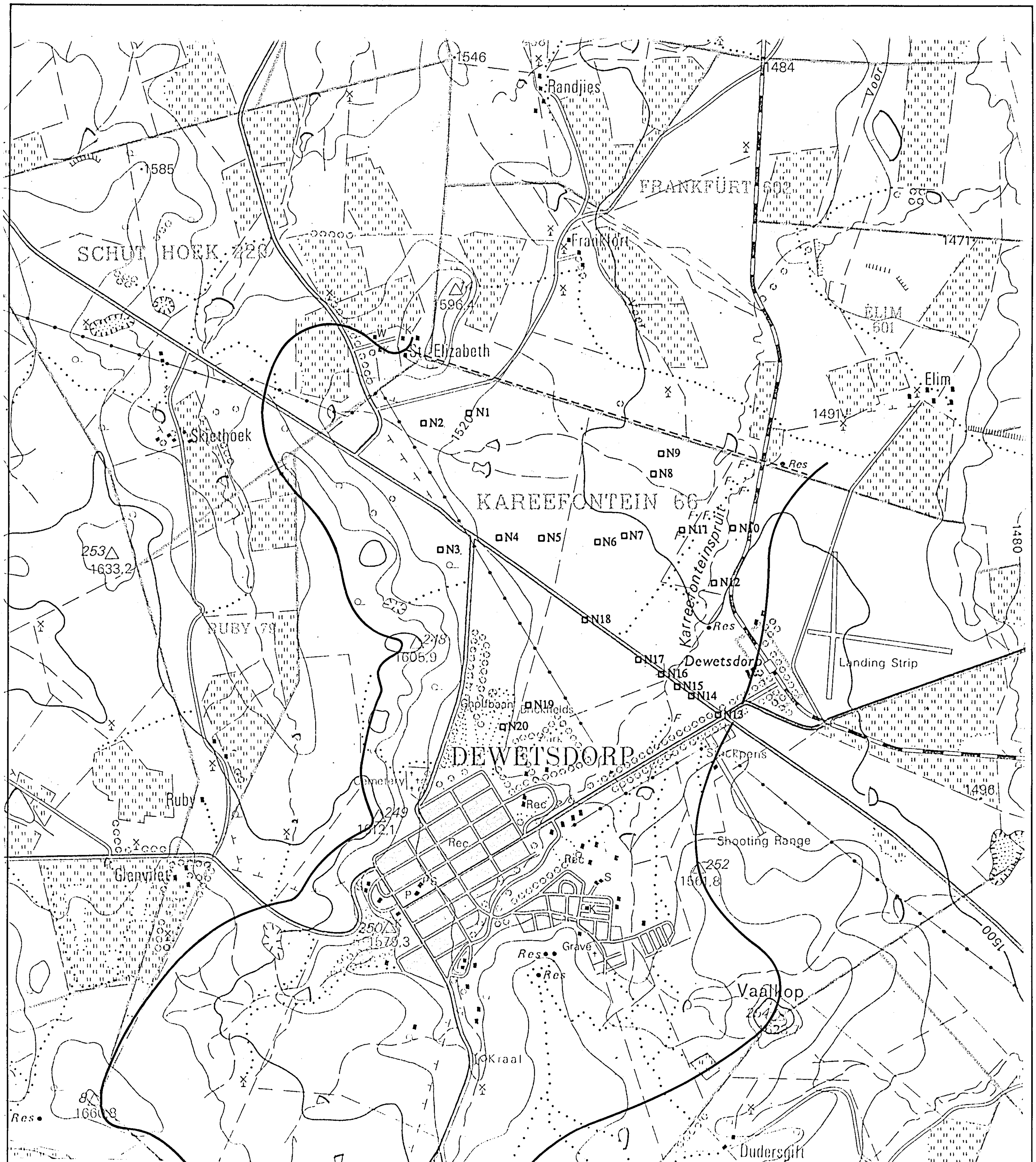
Soil	Neutron probe access tube no.
S1	6, 7, 11, 18
S2	1, 2, 13
V	10, 12, 15, 17
Sw	3, 4, 5, 8, 9, 16, 19
B	20
E	14

Neutron gauge measurement

In 1985, measurements were taken eight times after completion of the access tubes on 17th October, seven times during 1986, twelve times during 1987 and five times during 1988.

The soil water distribution observed will be discussed, according to the different soils the neutron probe access tubes are situated in. Water contents were calculated from the neutron counts, using the method described in Chapter 3. As described previously, the uppermost (two) values are generally too low, because the measurement is taken near the surface.

In Figure 4.20, the maximum and minimum volumetric water content observed in each tube are shown. Where the difference between maximum and minimum was less than



NEUTRON PROBE ACCESS TUBES : DEWETSDORP

SCALE

0 250 500 1 000 1 500 m

Figure 4.19

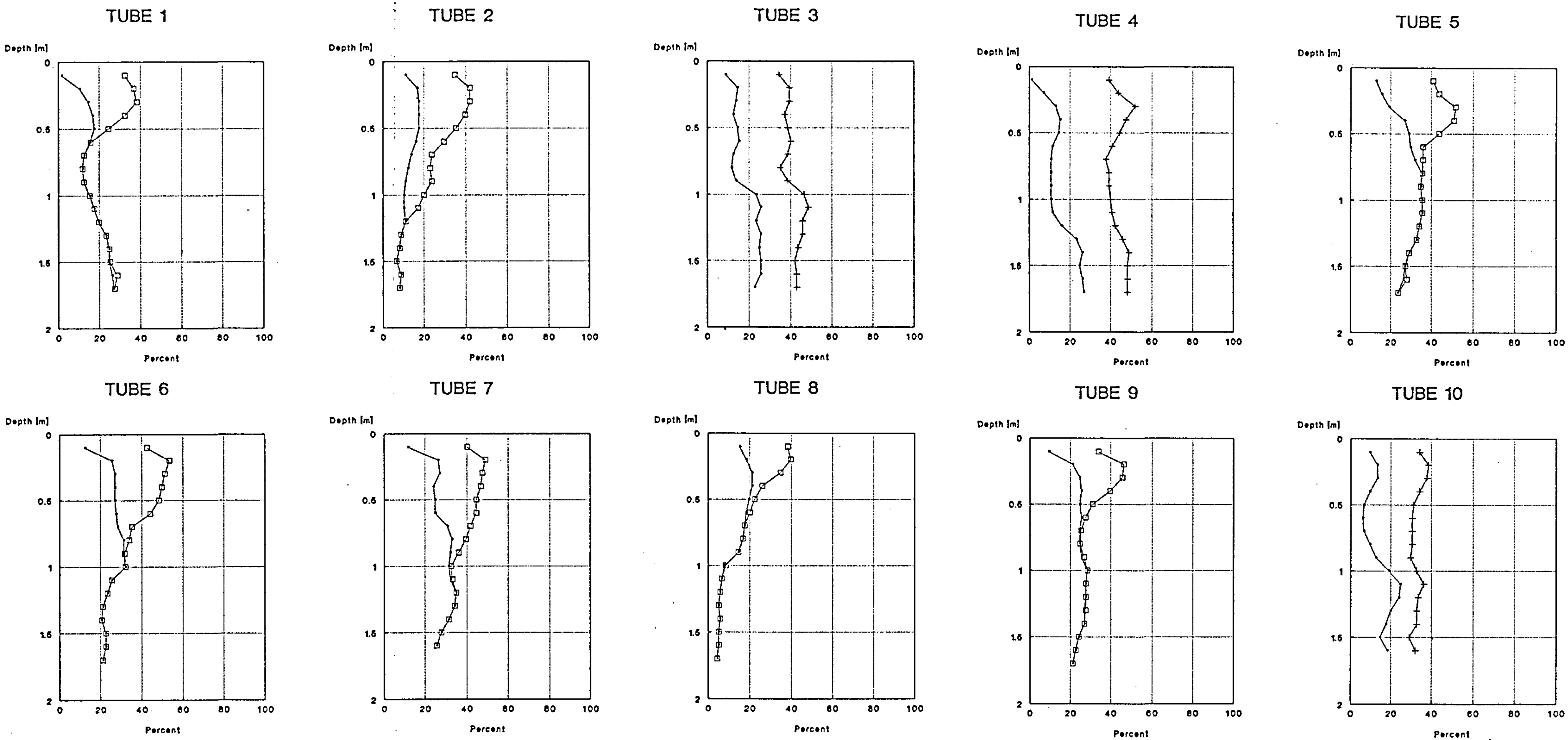


Figure 4.20 Minimum and maximum volumetric water content determined in the Dewetsdorp neutron probe access tubes.

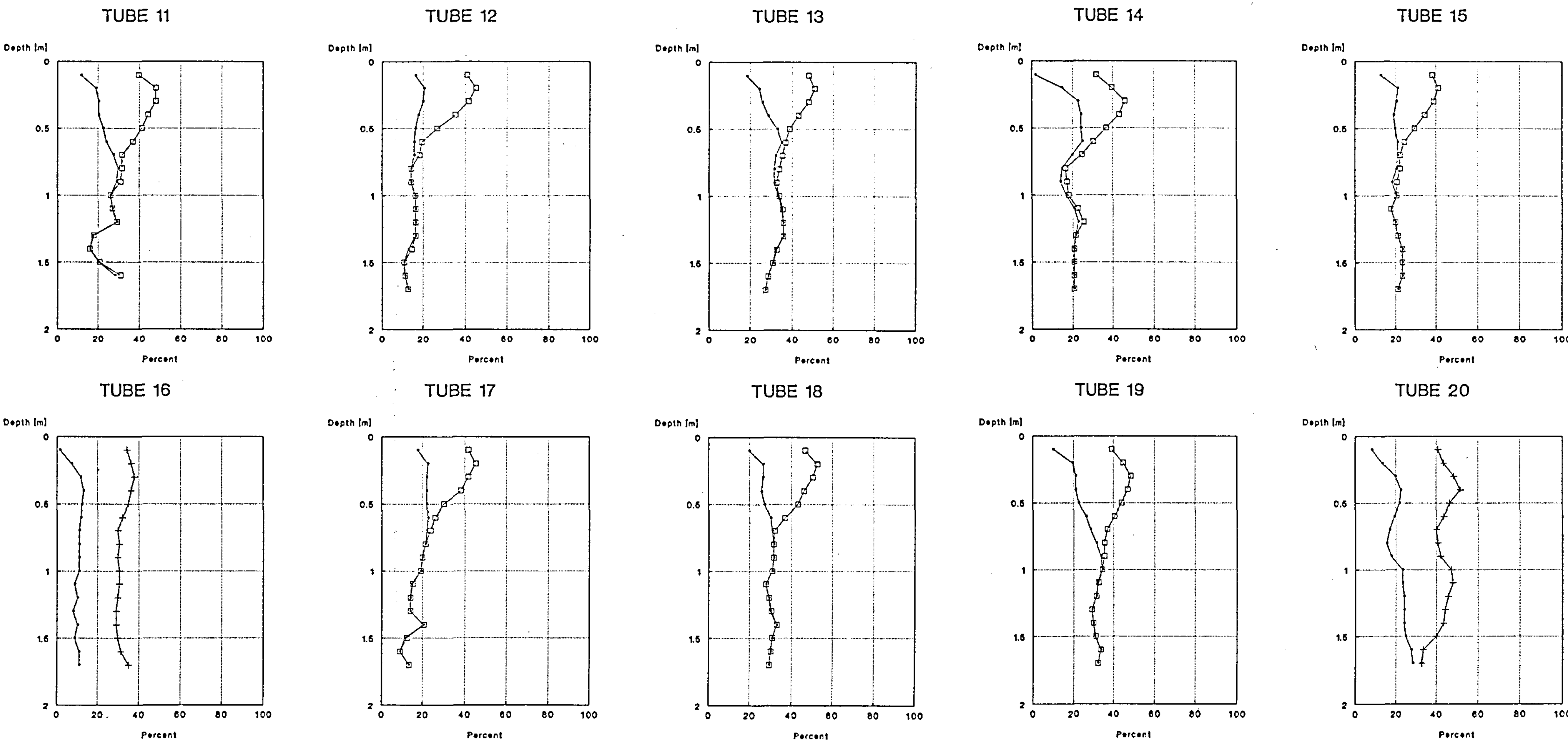


Figure 4.20 (cont.) Minimum and maximum volumetric water content determined in the Dewetsdorp neutron probe access tubes.

5 per cent, i.e. the difference was below the accuracy of the instrument, the two values were plotted on top of one another (plots of the original data are contained in the Appendix).

Sterkspruit (S1): Minimum water content is in the range of approximately 20 to 35 per cent. Variations occur down to a depth of ± 0,6 to 0,9 m. Maxima of up to 50% were observed above 0,5 m.

Sterkspruit (S2): Except for tube 13, the minimum water was generally below 20 per cent. Maxima of 40 (to 50)% were found at very shallow levels. Soil water variations were irregular, but not exceeding 1,2 m.

Valsrivier (V): Soil water distribution in tubes 12, 15 and 17 is quite similar, with a minimum of about 20 per cent, a shallow maximum at just over 40 per cent and practically no changes below 0,6 m. Tube 10 in the municipal well field displayed slightly lower minima and maxima. Only after the rains in February/March, water penetrated downwards, indicating recharge through the soil matrix and levels of nearly 40 per cent were measured.

Swartland (Sw): Tube 3 is situated at the foot of the hills, north of the town near the quarry and tube 4 lies 200 to 300 m further downstream, just NE of the main road near a little spruit. Tube 16 lies further SE, also near a little spruit on the south-western side of the main road. Their minimum soil water levels are between 15 and 25 per cent. In all three tubes, the water content rose by ± 20 per cent to maximum values of between 30 and 50 per cent. Only at tube 3, water transport had been observed at previous occasions, indicating possible ground-water recharge during normal rainfall events. Soil water distribution in the remaining tubes 5, 9 and 19 resembles that of 6 and 7 (S1), while in tube 8, it was similar to that in S2-tubes.

Bonheim (B): The tube is situated on the golf-course between borehole G36446, which started flowing in July, 1988, and borehole G36447, where water levels started to rise at an increased rate in May, 1988. Only the last measurement, after the great rains, showed increased soil water levels down to 1,5 m. The distribution is similar to that of tube 3. At late stages, soil water may therefore have penetrated to deeper levels. The water may also have risen from deeper lying aquifers.

Estcourt (E): Similar to tubes 12, 15 and 17 in the Valsrivier Form nearby. No soil water changes were noticed below 1,2 m.

There seems to be no obvious correlation between the form and position of these curves and the soil classification.

From the soil water observations, it is concluded that in the flat area below the village no significant recharge takes place by means of matrix flow in the unsaturated zone, because no significant change in the soil water was observed in the deeper soil layers. Recharge through the soil does, however, occur in the higher lying areas near the foot of the hills and possibly along the major spruits, especially during extended high rainfall periods. Field observations, like rising water levels in G 36457 soon after the rain with no corresponding increase of soil water in tube 17 twenty metres away, indicate, however, that some local recharge does take place in the lower lying parts of the research area. This can probably be ascribed to flow along preferential pathways like root channels, burrows and other openings like cracks or joints. At this hole G 36457 a rise due to aquifer recharge in the higher lying parts is regarded less likely because of the fastness of the water-level response and because water levels in holes upstream did not show a response to the rainfall at this stage.

4.1.5.2 SATURATED ZONE

After a borehole survey had been carried out, 40 boreholes, with a total depth of 1970 m, numbered G 35429 to G 36468, were drilled by the Department of Water Affairs for water-level observations and for various tests (see Figure 4.21). These holes varied in depth between 40 and 100 m. As it was the intention to generate representative information for e.g. the geology and yields of boreholes and to create a network which would provide good coverage of the area (away from the existing holes), sites for the holes were not selected at the most promising places, but rather with the objective of conducting certain tests and to obtain representative data. Observation holes were drilled around G 36430 and G 36431 to serve as piezometers during pumping tests and others along certain assumed flow paths to study water-level response to rainfall.

The logs for these boreholes and for some of the municipal holes are shown in the Appendix.

In addition, 68 boreholes and one well situated on Kareefontein and the adjoining farms Schut Hoek, Randjies, Frankfürt, Jouberts Pan, Elim, Oudersgift, De Hoek, Glen Heights, Glenvilet and Ruby were located during the duration of the project. These include four former production holes (10 A, 12 A, 12 B, KRN 10) above the village and four other holes (SHK 1, ODT 3, 4 and 5) which had all collapsed. Data from these holes and historical data about 117 holes in Dewetsdorp (DEW 18 to 135), as well as G 2366, G 2377 and the destroyed G 9992 collected by Dziembowski (1977), have also been entered into the data base.

ANNEXURE E

Analysis method for Total Fe, SG and Free Acid

ANNEXURE E

ANALYSIS METHOD FOR FERRIC CHLORIDE/SULPHATE SOLUTION

1. PROCEDURE FOR TOTAL Fe CONTENT

A. REAGENTS

(i) 0,10N POTASSIUMDICHROMATE SOLUTION

Dissolve 4,903 grammes of A.R. potassium dichromate in 200ml of distilled water and dilute to one litre with distilled water in a volumetric flask.

(ii) CONCENTRATED HYDROCHLORIC ACID C.P. (300 g/l HCI min.)

(iii) MIXED ACIDS

To 700 ml of cold distilled water add 100ml of concentrated sulphuric acid with caution. When ambient temperature add 100ml of concentrated phosphoric acid and dilute the mixture to one litre.

(iv) STANNOUS CHLORIDE SOLUTION 6% m/v

Dissolve 60g of stannous chloride dihydrate C.P. in 300ml of warm concentrated hydrochloric acid. Cool and dilute to one litre with distilled water.

(v) SATURATED MERCURIC CHLORIDE SOLUTION C.P. 7%

Dissolve 70g of mercuric chloride in 1 000ml of hot distilled water.

(vi) SODIUM DIPHENYLAMINE - 4 - SULPHONATE 2g/100ml, (SAARCHEM CAT. NO. 126200)

B. PROCEDURE

Weigh out 25ml of ferric solution accurately and transfer to a 500ml volumetric flask. Dilute to the mark with water. Invert sufficiently to ensure thorough mixing.

Pipette a 25ml aliquot into a 500ml wide mouth Erlenmeyer flask. Add 25ml of concentrated hydrochloric acid and heat to boil. Add 6% stannous chloride solution dropwise until the yellow colour is discharged. Add a few drops in excess. Cool rapidly and add 10ml saturated mercuric chloride solution and let stand for 10 minutes. Add 30ml of mixed acids and dilute the contents of the flask to 200-300ml with distilled water. Add 2,0ml of Sodium diphenylamine sulphonate indicator solution and titrate to a permanent purple end point with 0,10N Po tassium dichromate solution.

Collapsing holes

The depth of the recorder boreholes and of a few others were measured again during 1988. Within four years after completion, three of the fourteen holes measured had completely collapsed, seven to a lesser degree:

Hole Number	Date measured	Depth completed	Latest depth
G 36430	14 Nov. 1988	100.00	95.73
G 36439	14 Nov. 1988	40.00	39.41
G 36443	14 Nov. 1988	100.00	96.15
G 36447	14 Nov. 1988	40.00	38.98
G 36451	30 May 1988	30.00	6.92
G 36353	10 Mar. 1988	41.00	1.00
G 36354	14 Nov. 1988	41.00	39.64
G 36457	14 Nov. 1988	48.00	38.99
G 36462	27 Apr. 1988	30.00	1.20
G 36463	14 Nov. 1988	30.00	27.64

Yields

Borehole yields were generally very weak (< 1 l/s), if water was struck in the weathered zone only. Higher yields were associated with dolerite intrusions or prominent jointed zones. Of the 40 exploration holes drilled in Dewetsdorp in the beginning of the project, only G 36430, G 36450 and G 36467 had yields in excess of 1 l/s.

Original yields of the municipal production holes vary between ± 3.5 l/s and 6.5 l/s. The only higher recorded yield of 8,8 l/s is for Elm 4. All higher yielding holes lie in a ±200 m wide stretch along the Kareefonteinspruit.

Water levels

The water-level observation network consisted of 95 boreholes (excluding the private holes within the village, G 2366 and G 2377 owned by SATS, and six holes which could not be measured).

Recorders

In addition to the two existing recorder boreholes of the Department of Water Affairs from which the data were available, ten of the newly drilled boreholes were equipped with recorders. They were (from north to south and west to east): G 36431, G 36451, G 36449, G 36441, G 36430, G 36429, G 36437, G 36439, G 36443 and G 36458. The recorder on hole G 36437 was very near to G 36439. It was later shifted to G 36463, because of the lack of continuous records in the southern part of the aquifer.

After borehole G 36451 had collapsed, this recorder was installed on G 36454. The recorder on G 36431 showed little variations and was also not far from other recorders. It was moved to a private borehole, 'Ruby 8', just west of the research area beyond the watershed.

Oscillating water levels (e.g. in G 36429), chemical composition of the water and analysis of pumping test data indicated, at least for some of the boreholes, (semi-) confined conditions. In order to investigate whether such conditions existed in other holes too, recorders were temporarily installed on most of the open boreholes in the area to monitor their water-level behaviour. No oscillations were observed. Although this does not prove that (semi-) confined conditions do not exist in these holes their existence was not confirmed either. In the light of the behaviour of the water levels in boreholes G36429 and G36430 (see below), it is questionable, however, whether water-level observations, at times when levels are high in Dewetsdorp, can provide such information.

Other holes

Water levels in the exploration holes, not equipped with recorders, were measured on a monthly basis. Water levels in the remaining accessible boreholes, which were traced during the borehole survey in and around the research area, were measured on a quarterly basis, but more frequently during and after the heavy rains of February to April, 1988, and thereafter.

Water levels in the Dewetsdorp area vary between flowing (G36446) and 40,70 m (RBY6 in January 1988). Figures 4.22 and 4.23 show the highest and lowest water levels observed in boreholes during the three-year period October, 1985, to October, 1988. Water levels are generally deeper near the watershed. Water-level contours for October 1985 (see Figure 4.24) show that the flow is in a north-easterly direction with steeper gradients in the central part of the area (additional water-level contours appear in the Appendix). The municipal well field is indicated by the closed 1 500 m contour.

Water levels in the various boreholes responded differently to rainfall, abstraction, in- and outflow:

(i) A first group responded only to major recharge events and displayed during the remainder of the time virtually unchanged water levels. Examples are G 36464, G 36461, G 36441 and G 36457 (Figures 4.25 - 4.28). The difference between them lies in:

a) the magnitude of the rainfall event which caused a noticeable rise in the water level or the extent to which the water level rose, and

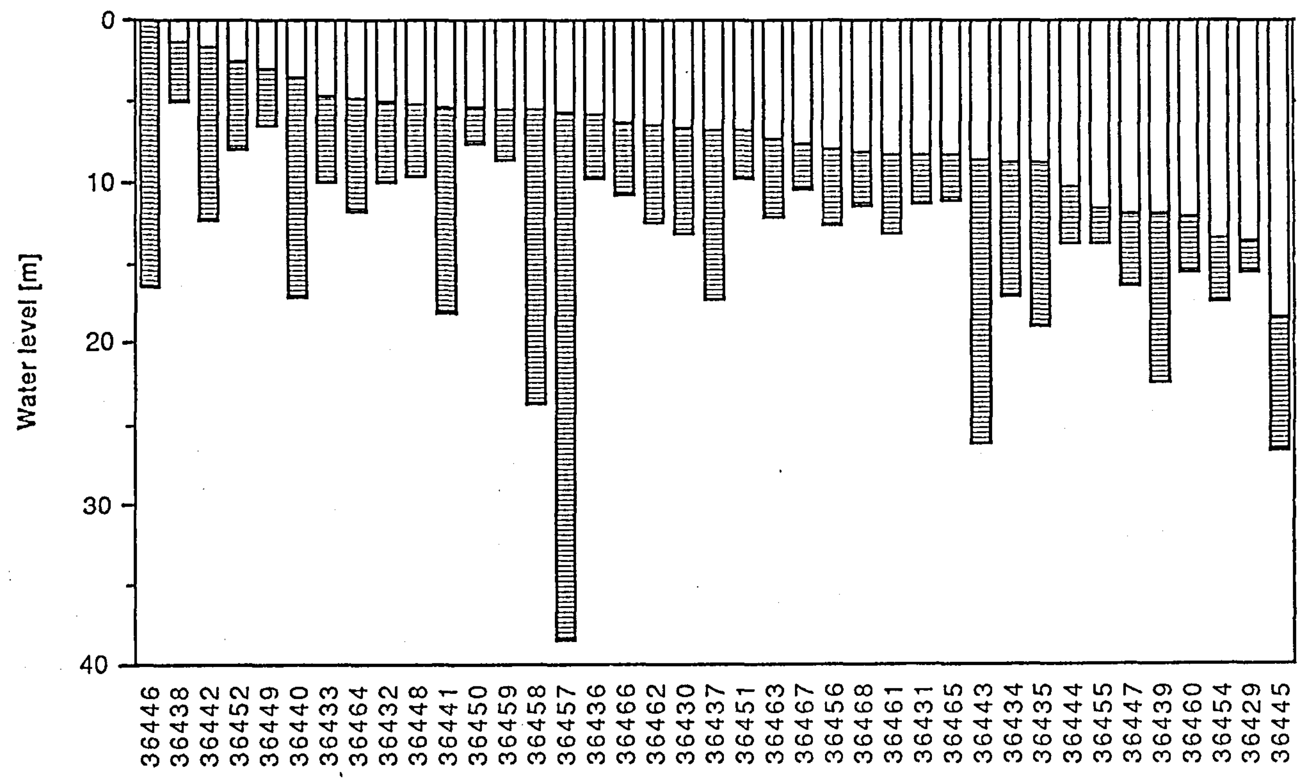


Figure 4.22 Highest and lowest water levels of G-number holes.

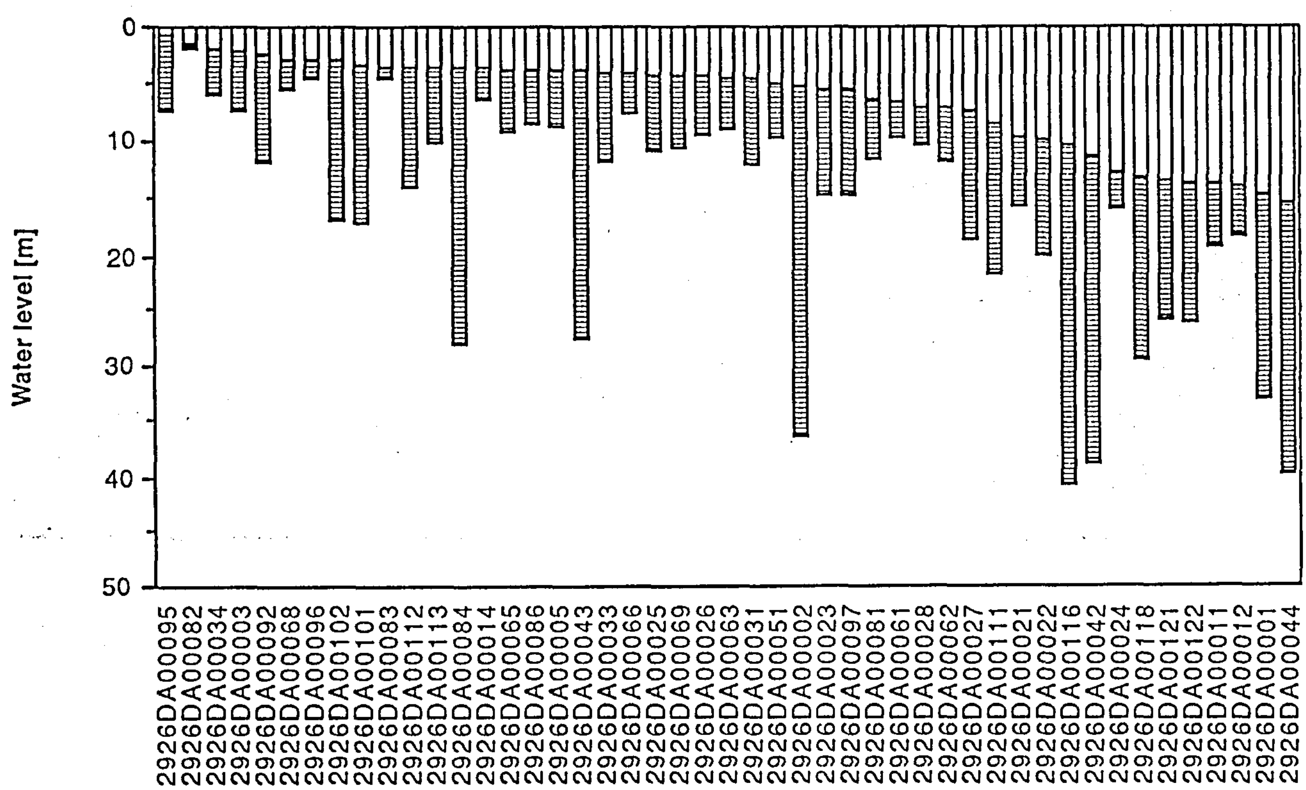


Figure 4.23 Highest and lowest water levels of ID-code holes.

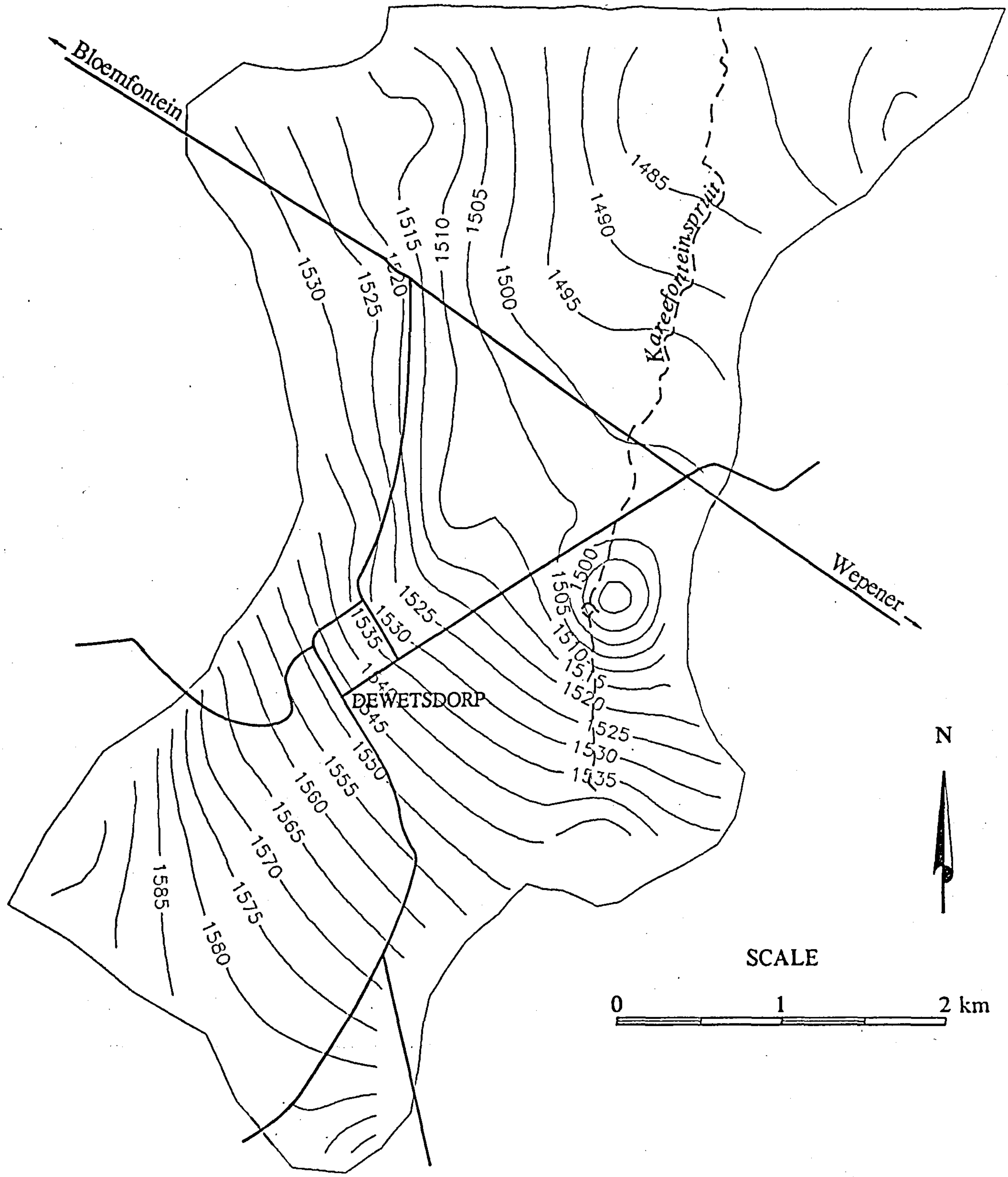


Figure 4.24 Boundaries of the Dewetsdorp aquifer showing the ground-water contours for October 1985. The western, southern and eastern boundaries are ground-water divides.

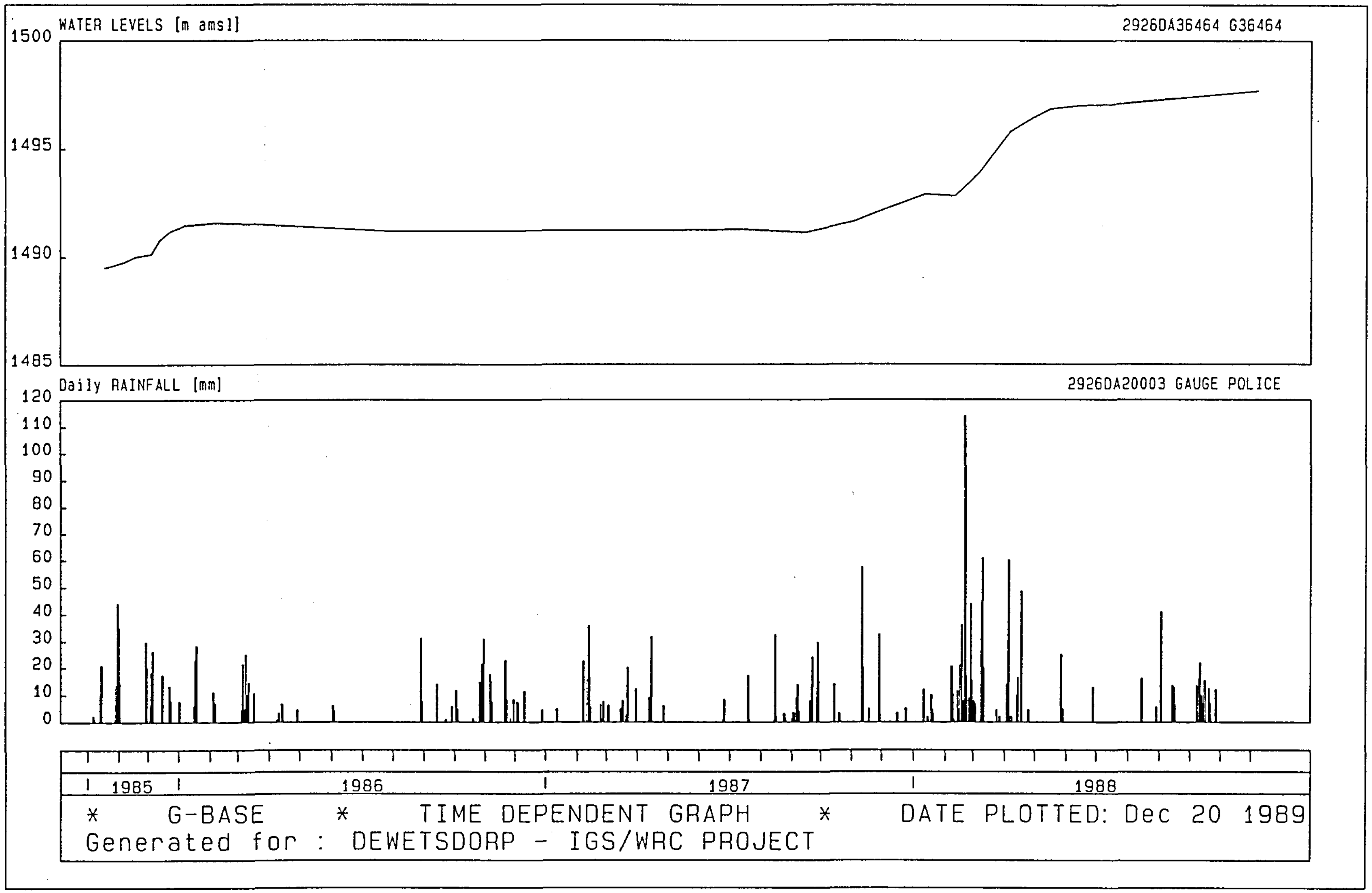


Figure 4.25 Water-level graph for borehole G 36464 and rainfall measured at the Dewetsdorp Police Station.

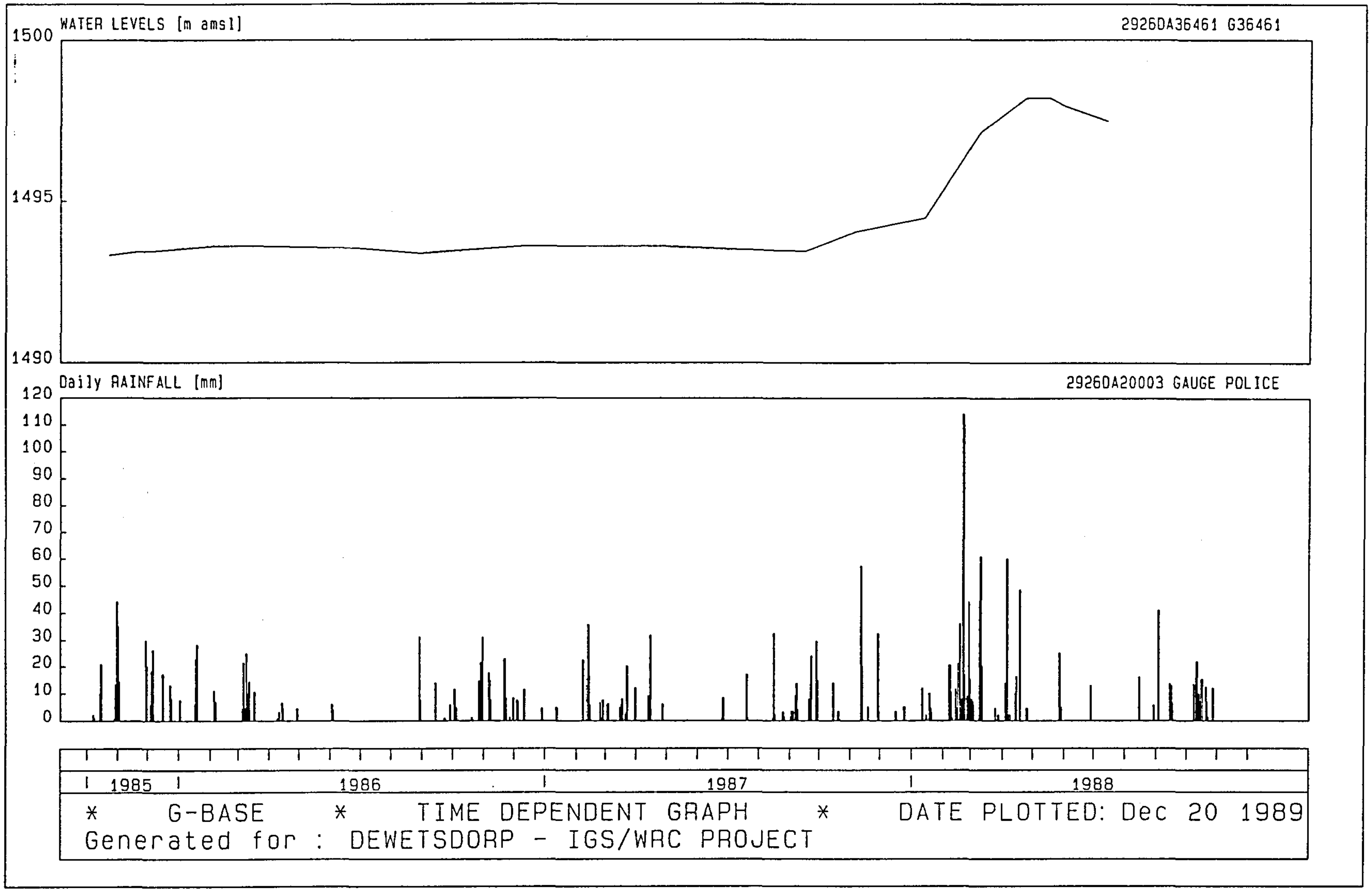


Figure 4.26 Water-level graph for borehole G 36461 and rainfall measured at the Dewetsdorp Police Station.

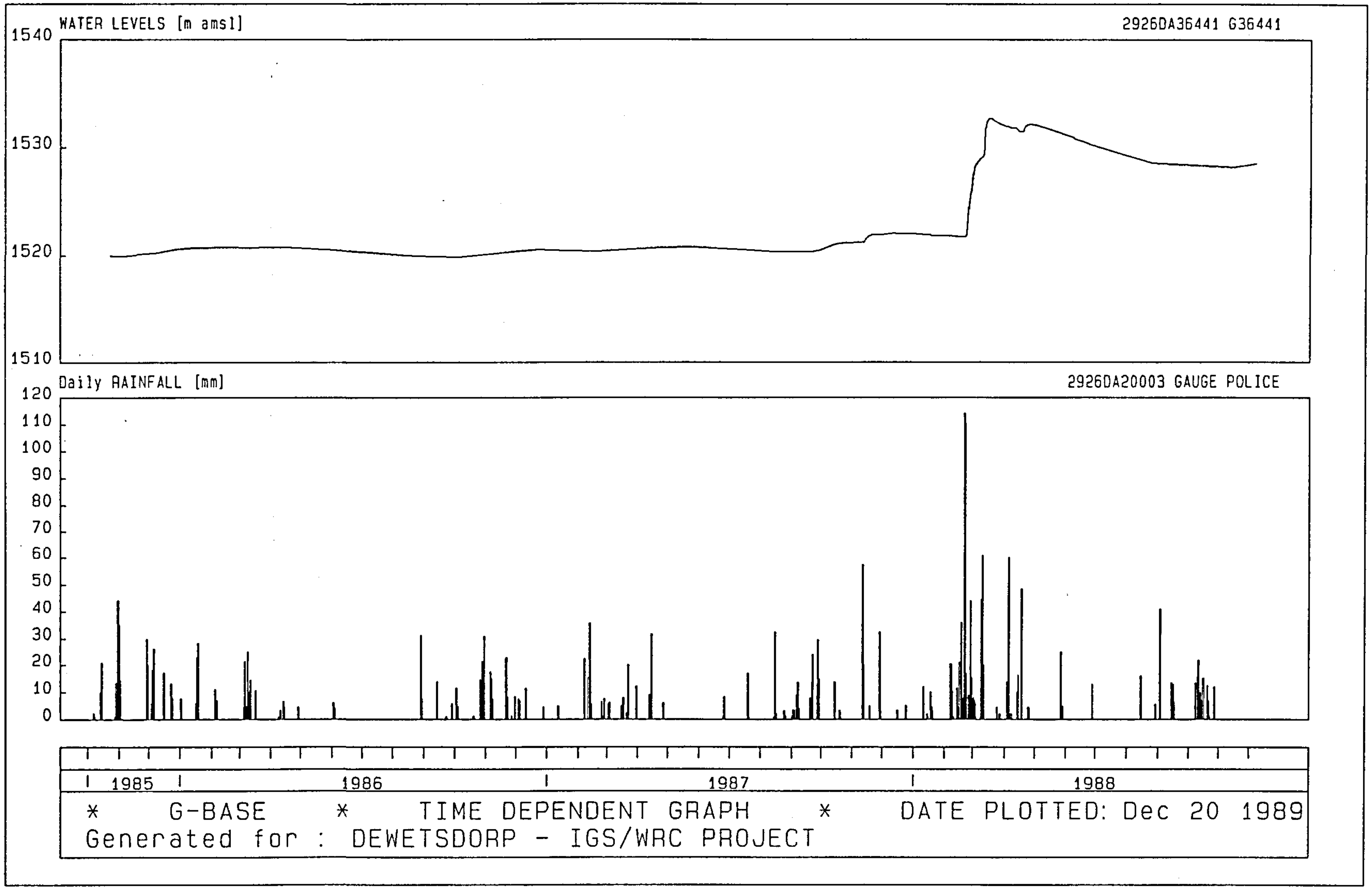


Figure 4.27 Water-level graph for borehole G 36441 and rainfall measured at the Dewetsdorp Police Station.

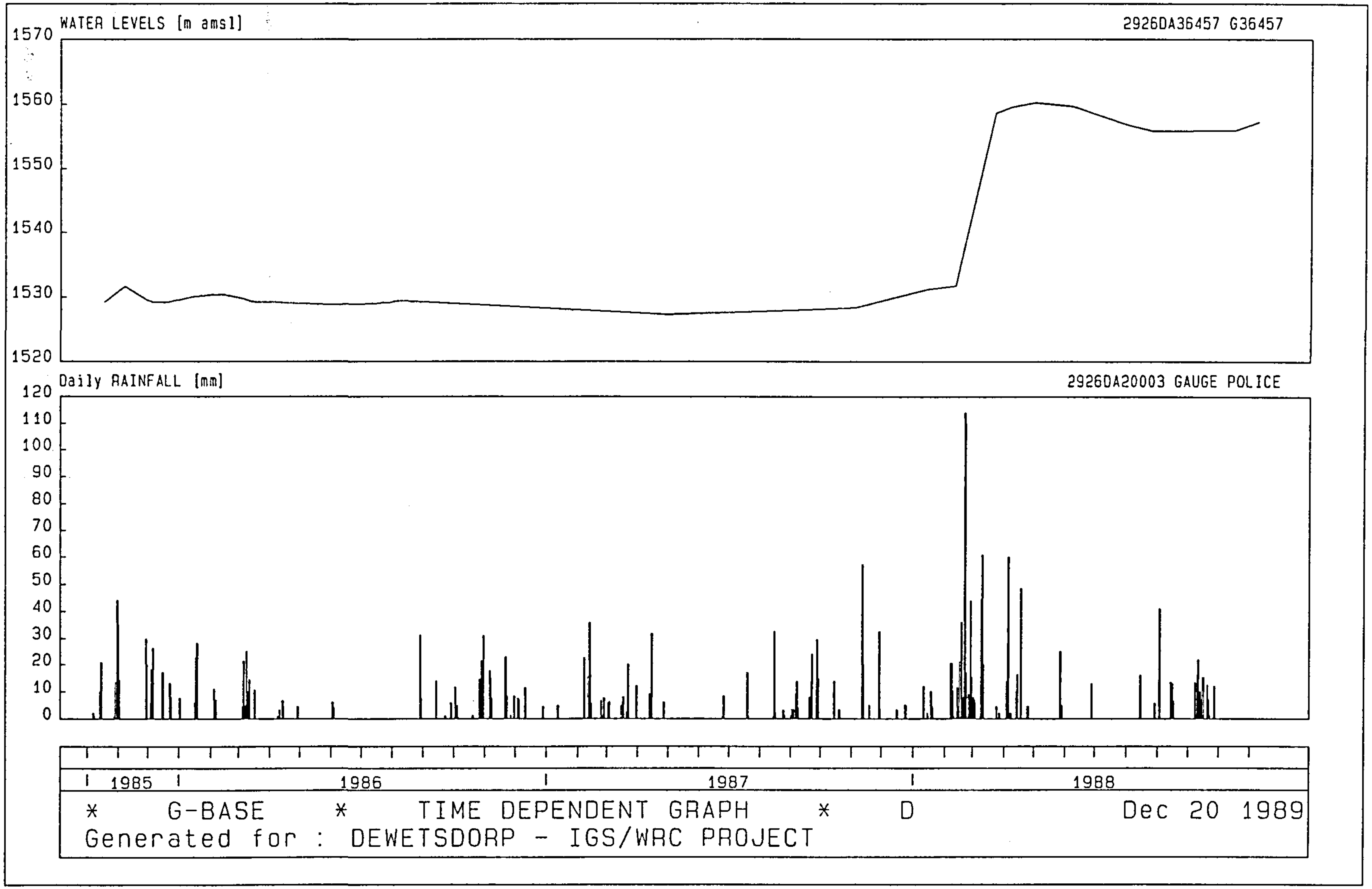


Figure 4.28 Water-level graph for borehole G 36457 and rainfall measured at the Dewetsdorp Police Station.

b) the time lag between rainfall and water-level rise.

(ii) The second group also shows a water-level rise after major rainfall events. Recharge due to normal rainfall is hidden in a reduced recession rate. The water level in FRT 7 in Figure 4.29 is an example. Extent of water-level rise and time lag may vary.

(iii) Water levels in boreholes of this group respond to any rainfall. Depending on the time lag, the peaks and lows of the graphs give it the appearance of a saw. Figures 4.30 and 4.31 show graphs for the two recorders, G36439 and G36463, which vary in the time lag and the relation between event and response.

(iv) Special cases:

a) Boreholes affected by pumping in the well field (e.g. G36449, see Figure 4.32).

b) G36429: Superimposed on water-level fluctuations, similar to group (i), are oscillations with two amplitudes (tides). NB: These tide effects are only displayed when the water level is deep (see Figure 4.33).

c) G36430: Similar to G36429, but oscillations are caused by pumping. They also peter out with rising water level (see Figures 4.34 and 4.35).

Boreholes under b) and c) are 100 m deep and not far from each other. Apparently, a composite aquifer is encountered in these boreholes: the lower aquifer is semi-confined, while the upper one is unconfined. When the latter is full, it suppresses the oscillations.

d) Irregular response: Recorder borehole G36458 displays abnormal and unexplained high or low water levels for relatively short periods, but only when the general water level is low (see Figure 4.36).

Recorder borehole G36443 showed a relatively sudden recession at the end of 1986, followed by a rise at a ± constant rate after the subnormal rainfall in the summer 1987 (see Figure 4.37).

All holes can be allocated to one of these groups.

Within a month after the heavy rains fell in February 1988, water levels rose between zero and more than 28 m. The water-level rise is schematically depicted in Figure 4.38. It can be seen that the highest rise occurred at the foot of the hills, NNW of Dewetsdorp (28,37 m just north of the cemetery), further north at Randjies (12 - 14 m) and also in certain parts of the higher lying areas, SW of the village on Ruby and Glenvilet (14 - 23 m).

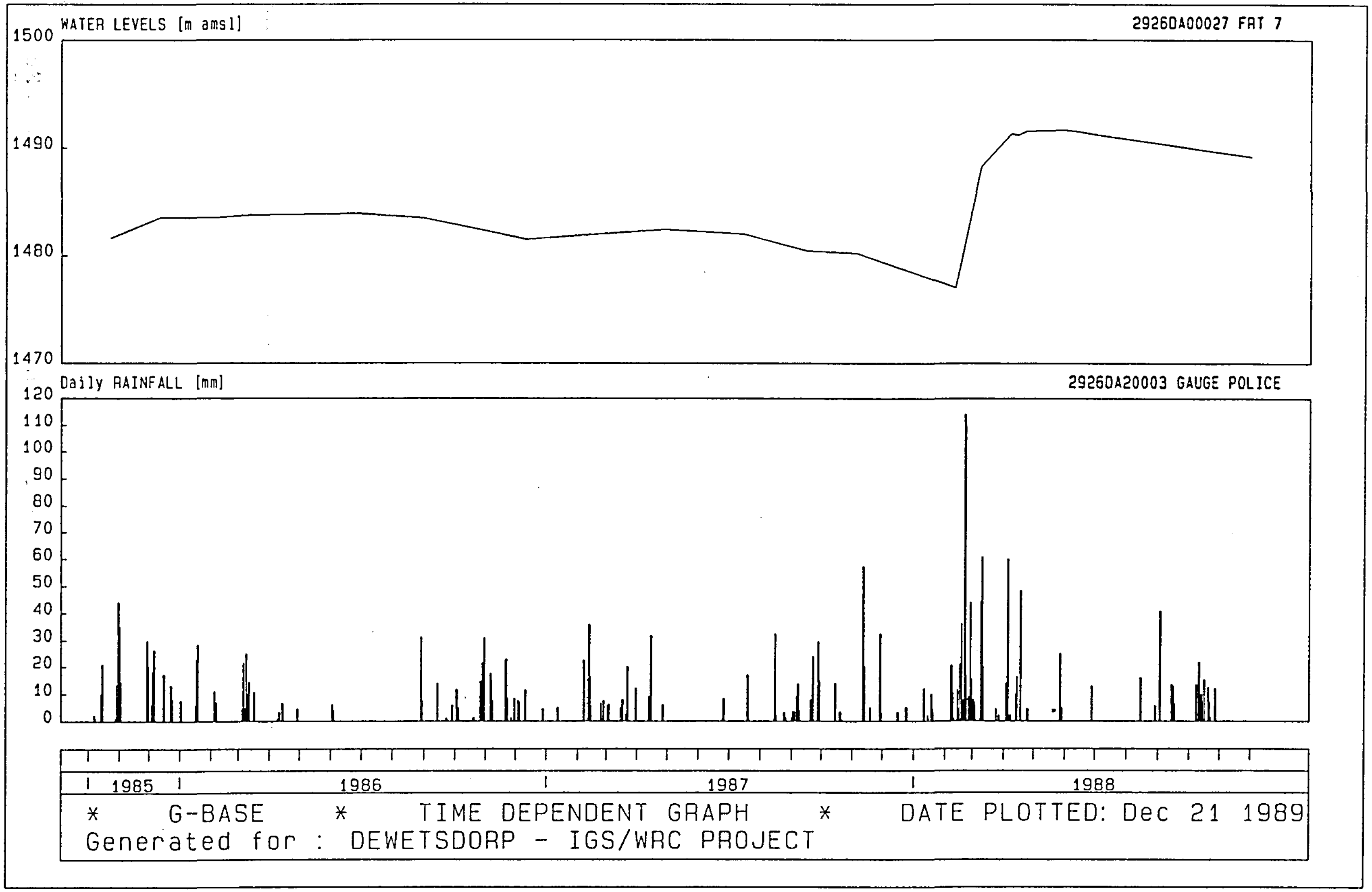


Figure 4.29 Water-level graph for borehole FRT 7 and rainfall measured at the Dewetsdorp Police Station.

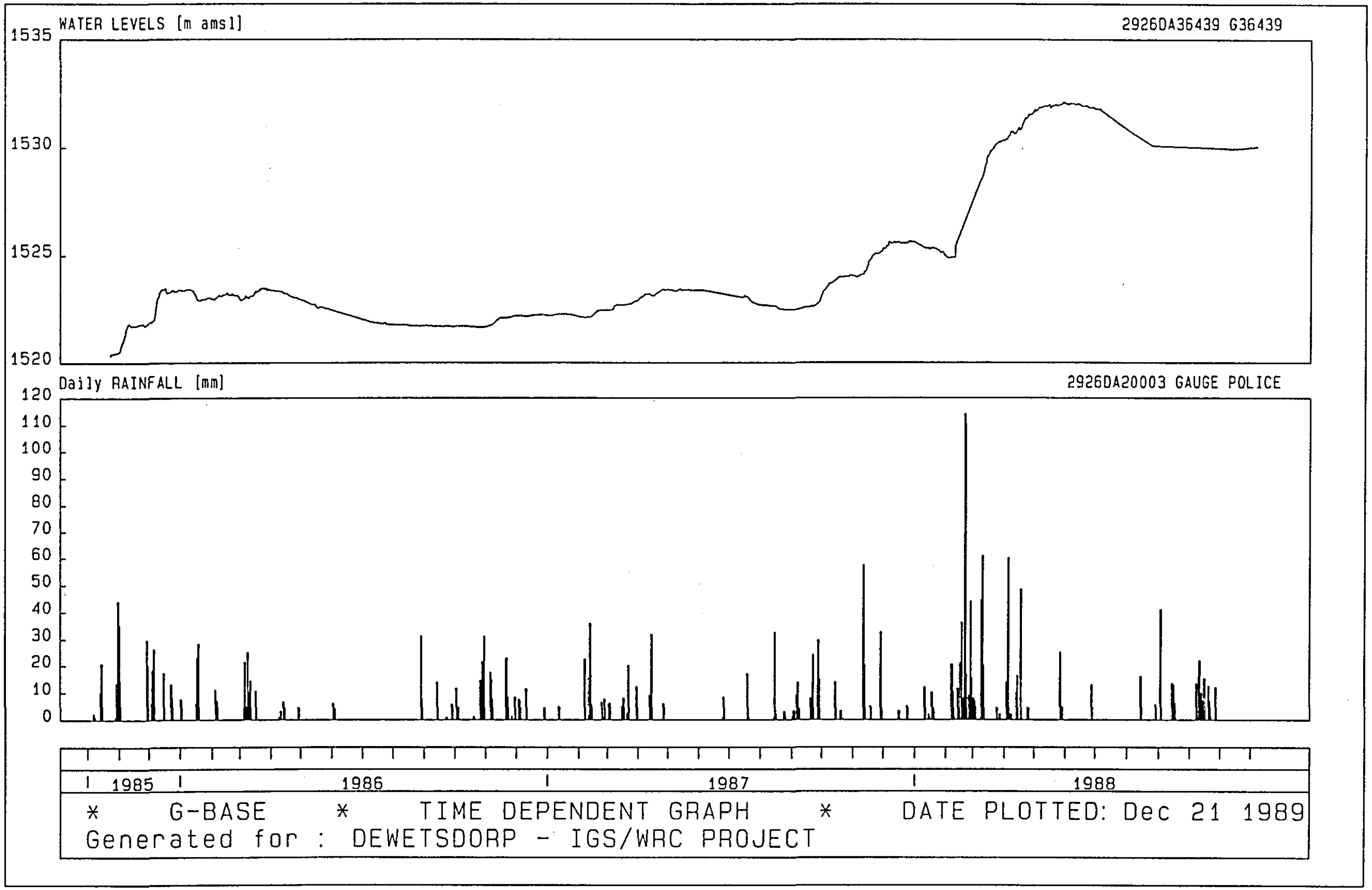


Figure 4.30 Water-level graph for borehole G 36439 and rainfall measured at the Dewetsdorp Police Station.

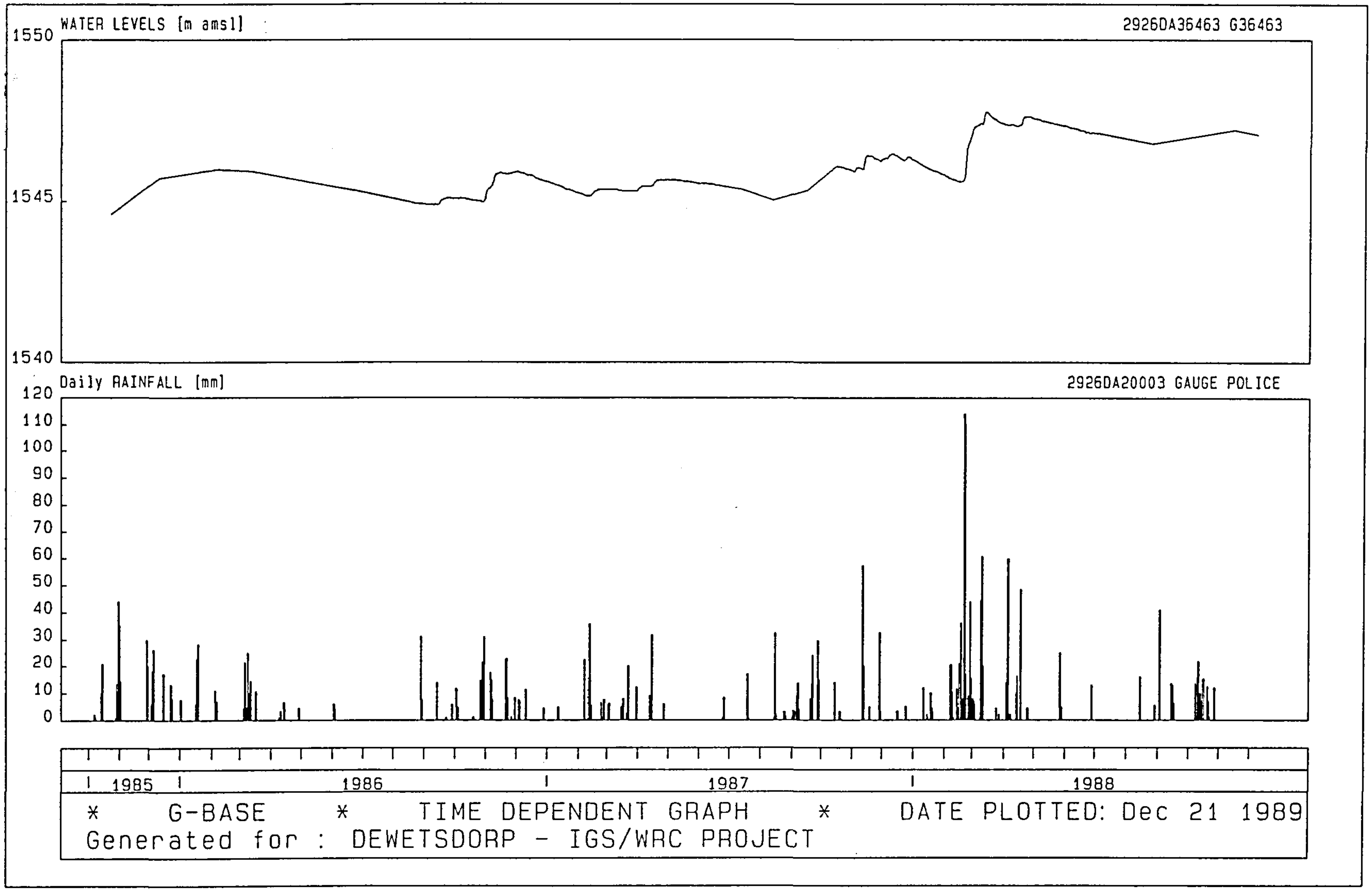


Figure 4.31 Water-level graph for borehole G 36463 and rainfall measured at the Dewetsdorp Police Station.

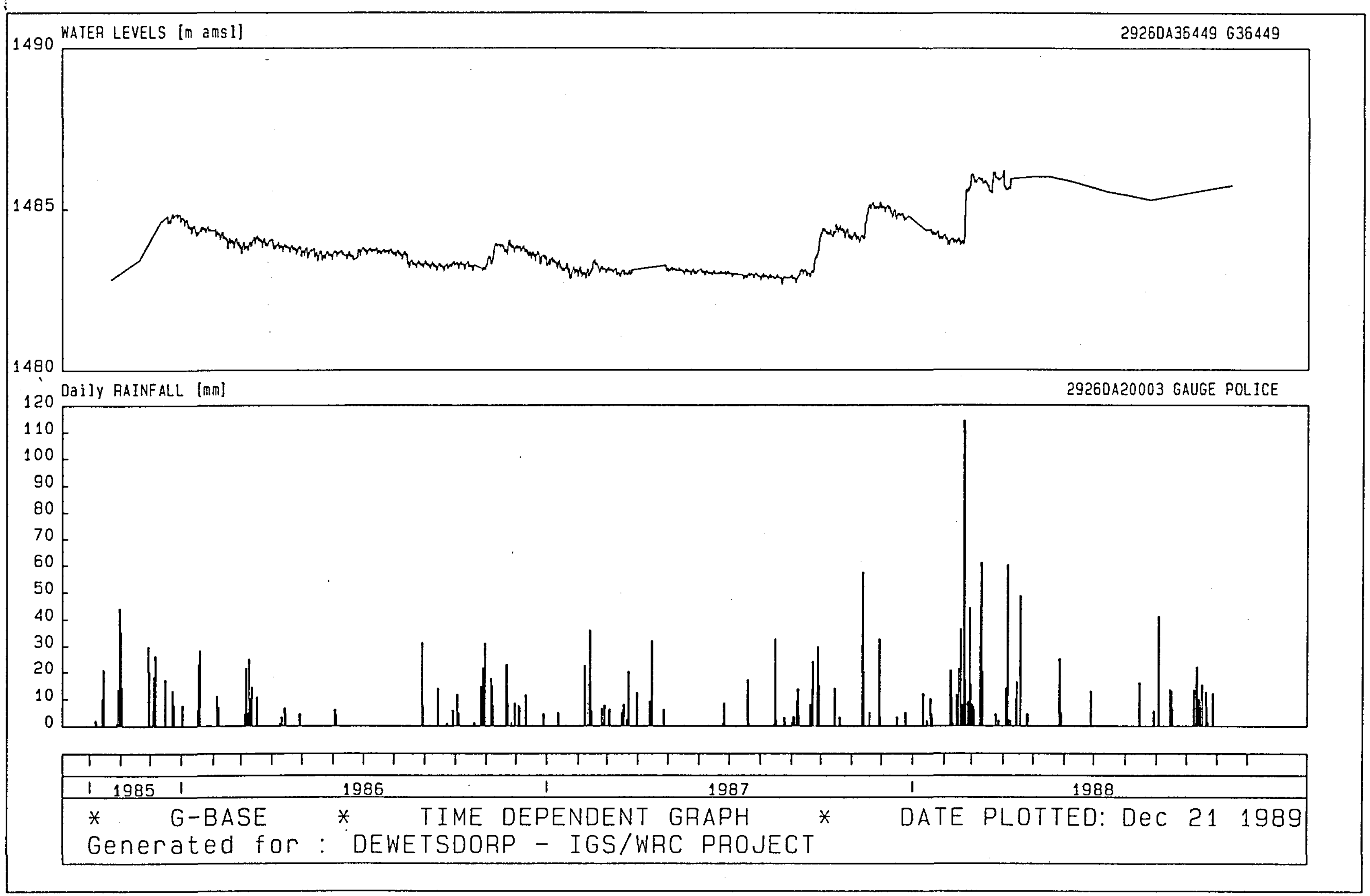


Figure 4.32 Water-level graph for borehole G 36449 and rainfall measured at the Dewetsdorp Police Station.

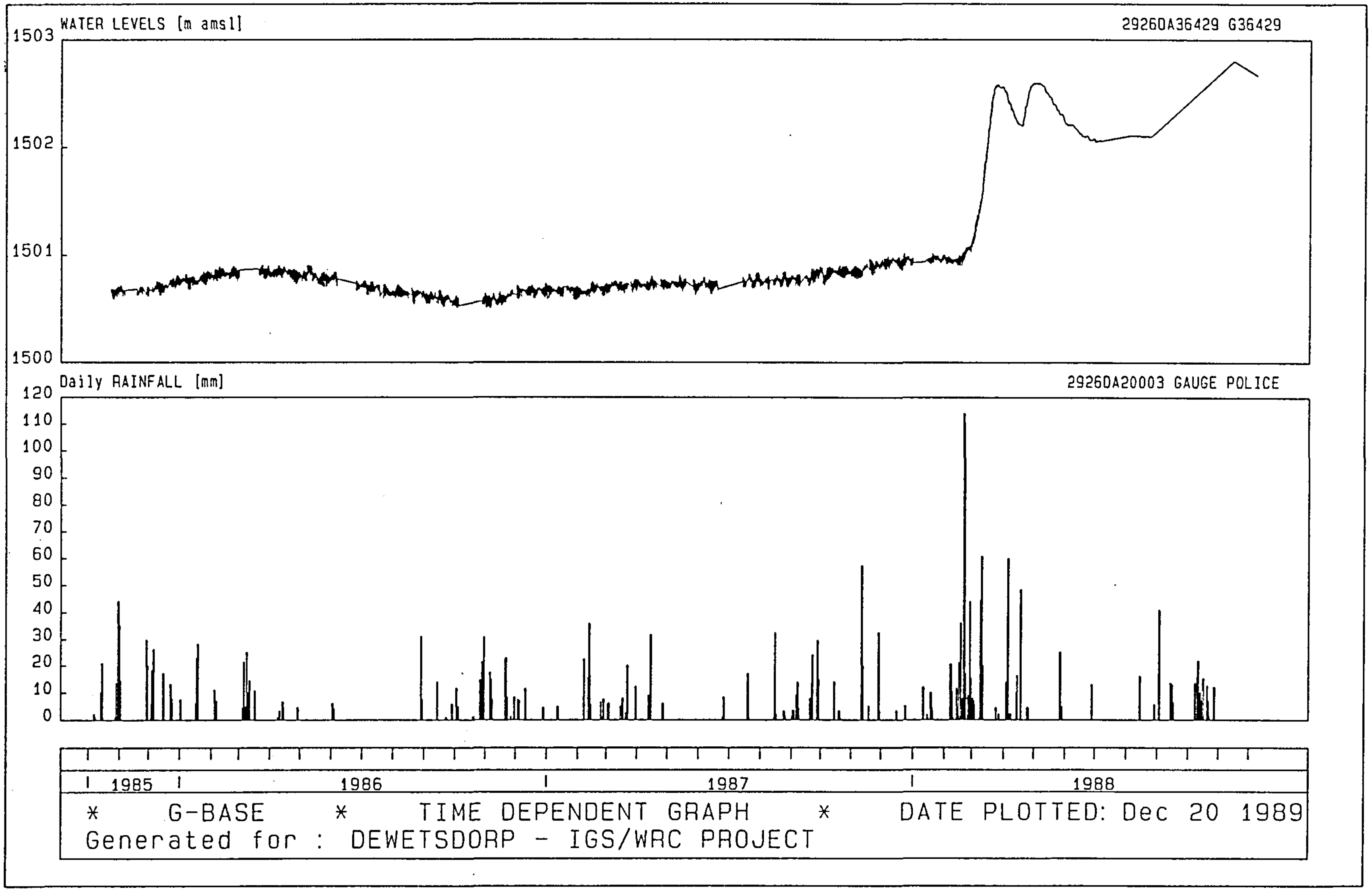


Figure 4.33 Water-level graph for borehole G 36429 and rainfall measured at the Dewetsdorp Police Station.

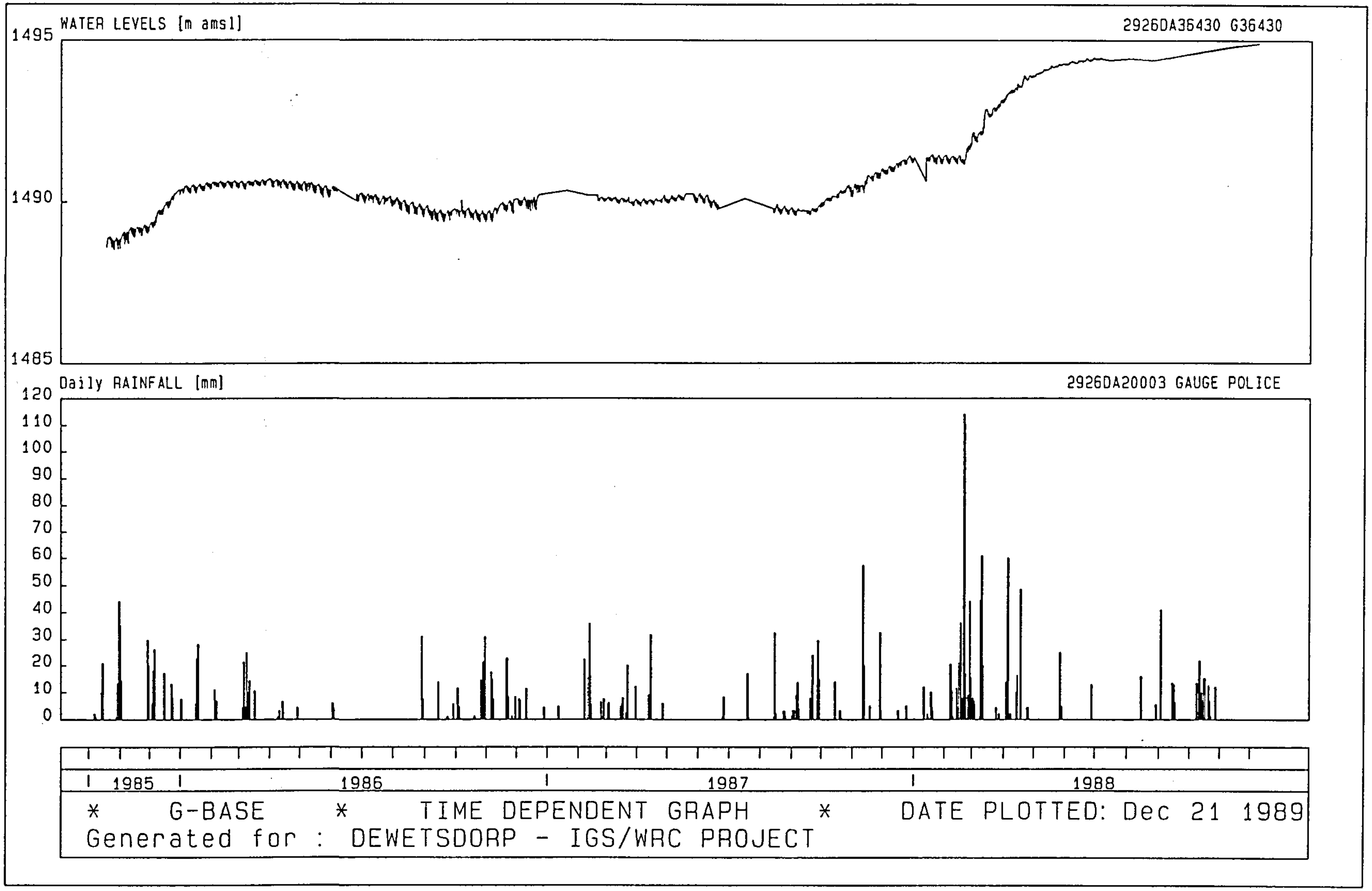


Figure 4.34 Water-level graph for borehole G 36430 and rainfall measured at the Dewetsdorp Police Station.

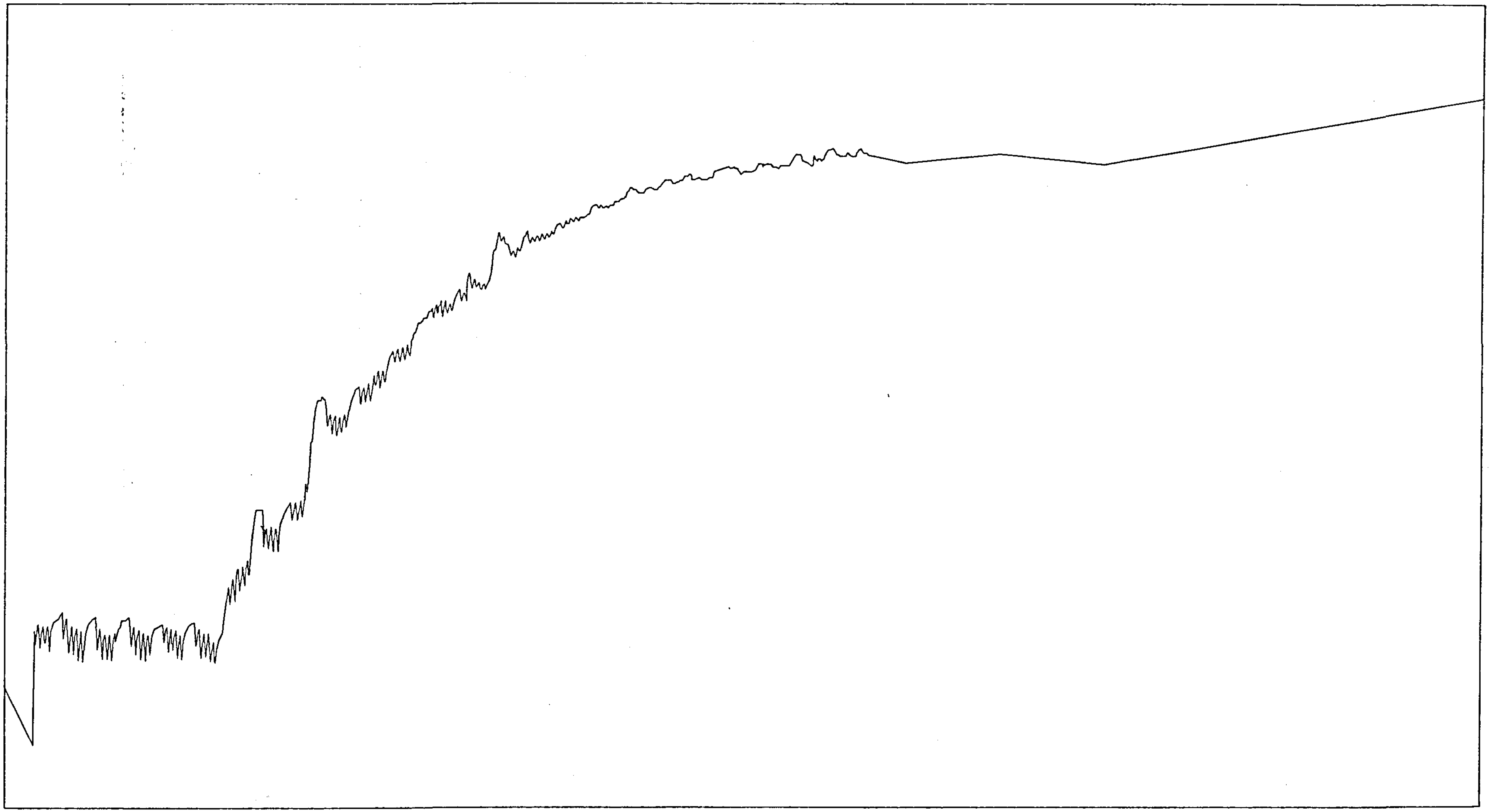


Figure 4.35 Water-level graph for borehole G 36430 (detail).

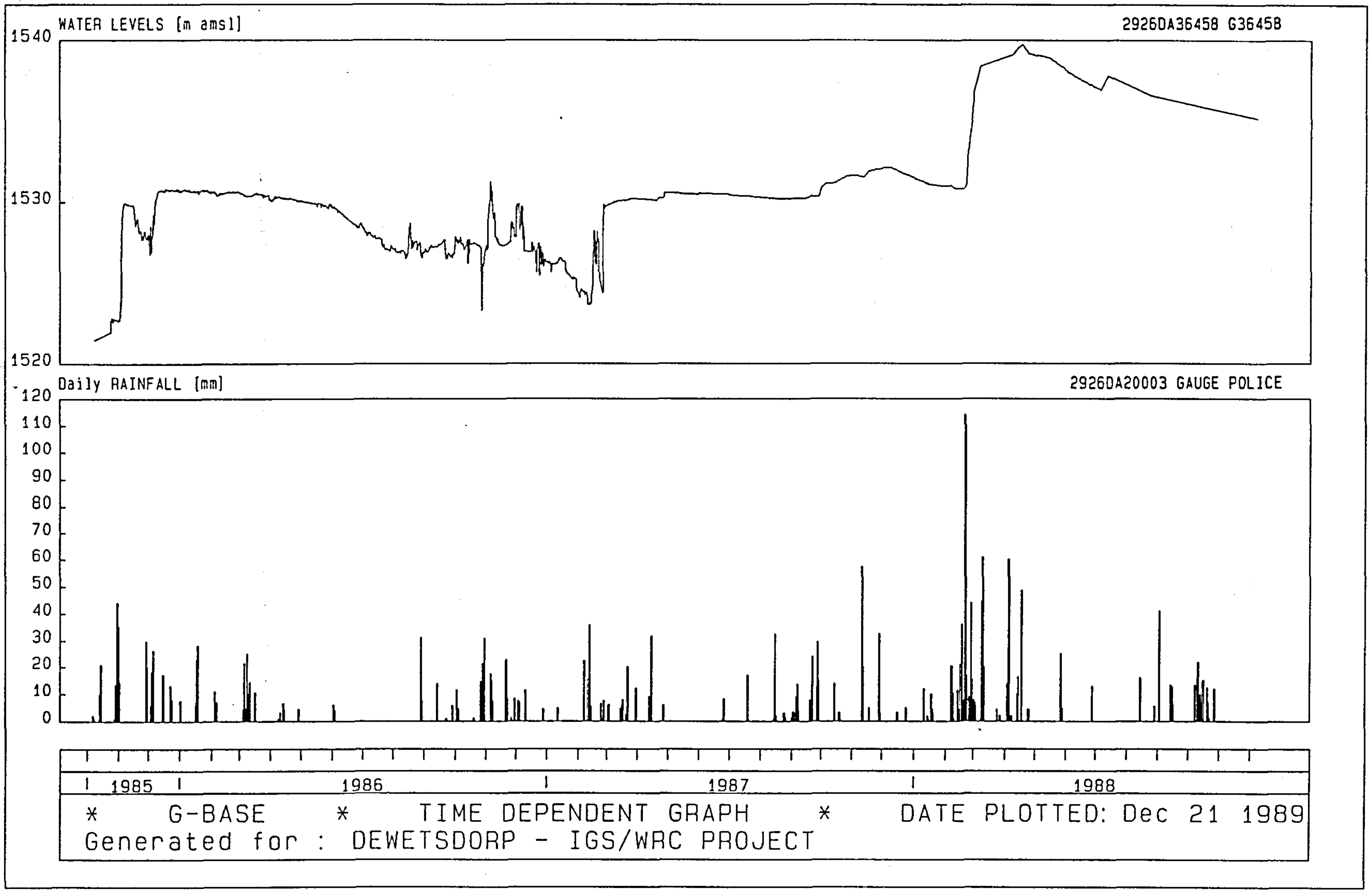


Figure 4.36 Water-level graph for borehole G 36458 and rainfall measured at the Dewetsdorp Police Station.

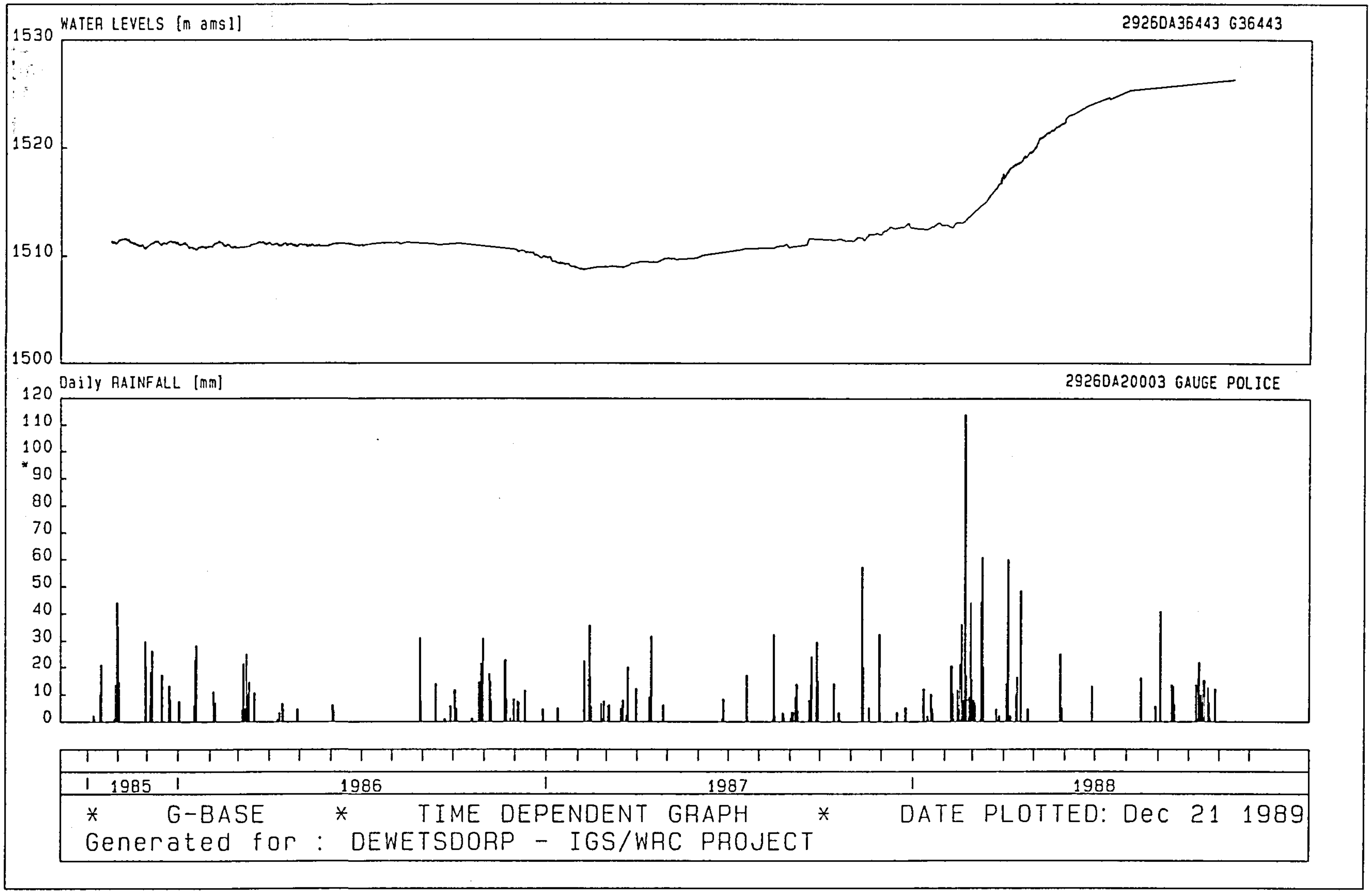


Figure 4.37 Water-level graph for borehole G 36443 and rainfall measured at the Dewetsdorp Police Station.

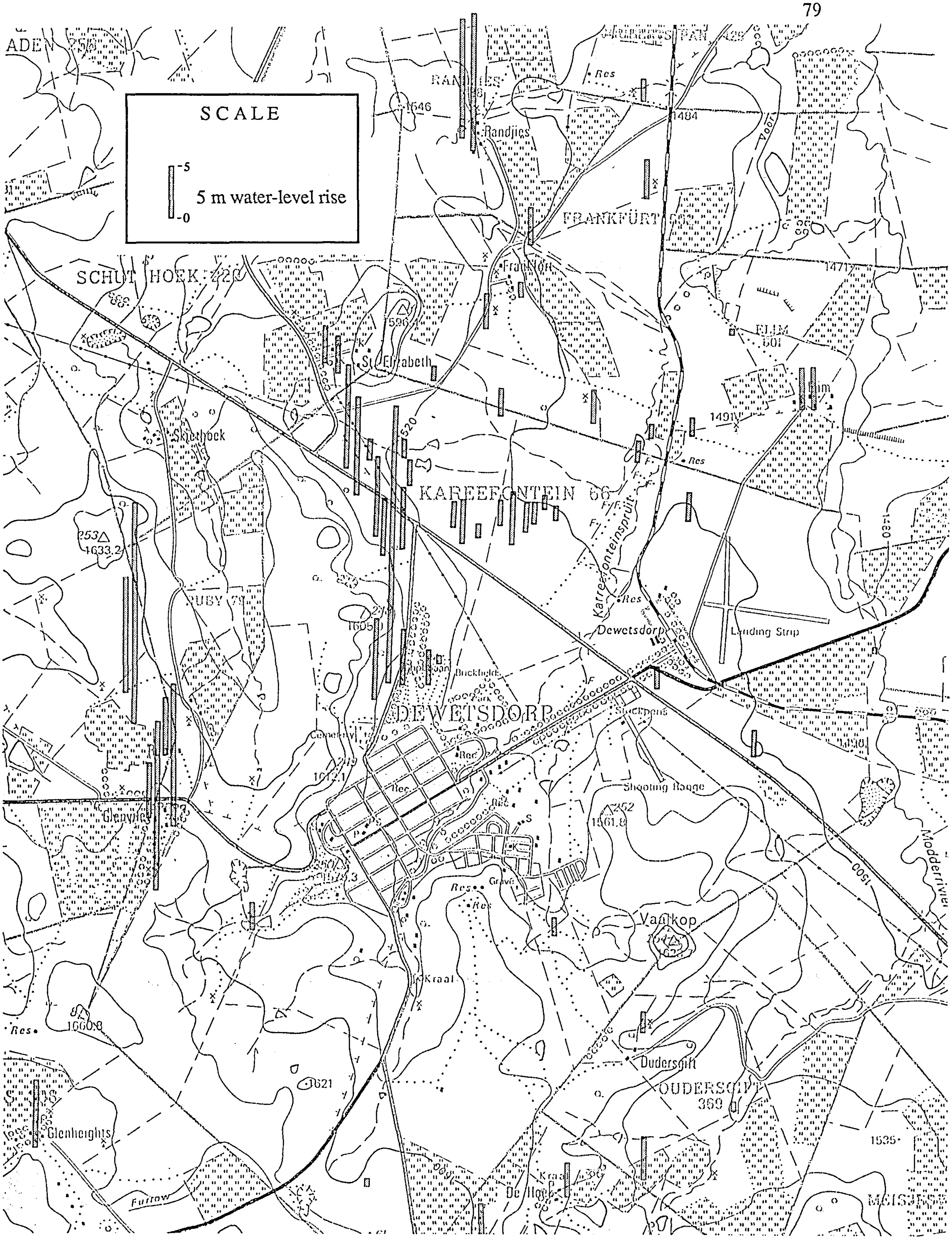


Figure 4.38 Relative water-level rise within 14 days after the February 1988 flood.

Looking at the time lag between rainfall and water-level response, Table 4.1 lists the abbreviated ID-codes of the various boreholes, according to the dates the minimum water levels were measured. Also stated, is the maximum level and the date it was recorded. In the last column, previous minimum levels are given. For better orientation, the Site ID codes and the corresponding Map numbers are listed in Table 4.2.

In analysing the information, it is noticed that minimum levels below the surface were reached *within a month* after the February rains:

(i) along the Kareefonteinspruit and tributaries, and
(ii) in selected areas covered by Mispah soils.

In these areas, there was a fast response to the rain. Recharge rates were high, but the ground-water flow rate was also high. This caused the ground water to spread into adjoining parts of the aquifer.

Two to three months after the heavy rains, areas adjoining those above, i.e. slightly further away from the spruits and the slopes, reached maximum water levels (Schut Hoek, Frankfürt, Randjies, Elim, the eastern part of Kareefontein, Oudersgift and Glen Heights).

In June/July, maxima followed in areas which were even further away from the Kareefonteinspruit, as well as on the higher lying areas (De Hoek, Glenvilet, Ruby).

A large area, stretching from the golf-course to the northern boundary in the flat part of Kareefontein where water levels are relatively deep, responded very late. This delayed response is solely ascribed to the time lag between rainfall and water-level rise. Late water-level maxima in the higher lying areas are thought to be caused by the continuing rainfall in the second half of the year.

This response of the aquifers is also seen in Figure 4.39, which shows the delay in water-level rise after the flood. The water levels in the northern part of the study area and in the deepest part of the aquifer started to rise only after 90 days or later.

Discussion

Conceptual model: The first rain percolates at a maximum rate which is governed by rainfall intensity, local hydraulic conductivity and head. As the vertical hydraulic conductivity (due to blocked pores, swelling clay) and head decrease, the infiltration rate is reduced. Selected aquifers or parts thereof become 100% filled. From these parts, ground water is draining at a slower rate to the 'remoter' parts of the aquifer which recharge and dewater at a slower pace. If good rains continue within the same year or not too late thereafter, all the aquifers will be filled.

Number	Start	Minimum	Date	Maximum	Date	
36445	1985	18.50	19880307	26.75	19851018	
00011	1976	13.91	19880308	19.44	19851021	
00012	1986	14.07	19880308	18.44	19860313	
36463	1985	7.22	19880312	12.11	19880414	
00069	1985	4.40	19880314	10.85	19851128	
00065	1954	3.78	19880314	9.35	19861013	5.50
36441	1985	5.31	19880317	18.17	19861006	
00062	1954	7.20	19880321	11.98	19860728	10.97
00043	1976	3.86	19880322	27.40	19860205	
00082	1985	1.38	19880322	1.82	19860628	
00064	1954	3.49	19880331			3.96
36449	1985	3.04	19880331	6.53	19870831	
00102	1985	2.87	19880407	16.90	19870714	
00101	1985	3.40	19880407	17.34	19870915	
36458	1985	5.46	19880417	23.75	19851009	
00068	1985	2.76	19880418	5.60	19870202	
00092	1976	<2.33	19880422	(12.00)		3.92
00025	1985	4.28	19880422	10.98	19851022	
36461	1985	8.24	19880422	13.12	19851022	
00096	1976	2.77	19880422	4.44	19870715	2.06
00083	1976	3.47	19880422	4.48	19870715	7.92
00097	1985	5.57	19880422	14.84	19870915	
00023	1985	5.42	19880422	14.86	19870915	
36452	1985	2.45	19880422	7.89	19871103	
00024	1985	12.83	19880422	15.97	19880712	12.10
00026	1988	4.42	19880422	9.56	19881130	
36450	1985	5.37	19880425	7.58	19870916	
36457	1985	5.62	19880430	38.48	19870427	
36437	1985	6.76	19880430	17.18	19870715	
00111	1985	8.51	19880430	21.76	19870915	
00032	1988	1.43	19880502			
36462	1985	6.42	19880502	12.39	19851211	
00086	1985	3.81	19880502	8.64	19860829	
00003	1976	2.16	19880502	7.40	19860929	
00033	1985	3.98	19880502	11.85	19871103	
36442	1985	1.63	19880502	12.28	19880127	
00034	1988	1.80	19880502	5.93	19881201	
36451	1985	6.87	19880516	9.76	19851022	
36456	1985	7.96	19880516	12.59	19851022	
36466	1985	6.23	19880516	10.79	19851203	
36465	1985	8.31	19880516	11.14	19870714	
36467	1985	7.61	19880516	10.50	19870916	
00051	1985	5.07	19880530	9.70	19851022	
00121	1985	13.66	19880530	25.81	19851022	
00122	1976	13.84	19880530	26.18	19851022	14.35
00085	1988	6.78	19880530			
00028	1985	7.10	19880530	10.56	19860205	
00014	1985	3.70	19880530	6.40	19860829	2.78
00021	1985	9.69	19880530	15.90	19860829	
00095	1985	0.66	19880530	7.35	19870715	
36460	1985	12.13	19880530	15.64	19870915	
00027	1985	7.33	19880530	18.60	19870915	
00066	1954	4.00	19880613	7.55	19851021	5.50
00041	1988	16.20	19880613			
36448	1985	5.08	19880613	9.69	19851024	
00113	1985	3.56	19880613	10.28	19851211	
00044	1985	15.47	19880613	39.62	19860627	
00042	1976	<11.51	19880613	38.85	19861210	
00031	1985	4.61	19880613	12.15	19870714	
00112	1985	3.52	19880613	?14.00		
00022	1985	10.04	19880712	20.14	19870915	
00081	1985	6.45	19880713	11.77	19860312	
00116	1987	10.59	19880713	40.70	19880113	
00005	1988	3.85	19880713	8.79	19881130	
00002	1985	5.34	19880714	36.30	19861210	
00080	1954	7.95	19880719			
00061	1954	6.65	19880719	9.83	19870907	5.50
36439	1985	11.94	19880802	22.39	19851023	
36454	1985	13.43	19880802	17.50	19851024	
36468	1985	8.07	19880802	11.41	19870916	
00063	1954	4.60	19881107	8.99	19851028	6.40
36429	1985	13.57	19881114	15.67	19861105	
36443	1985	8.63	19881114	26.23	19870205	
36455	1985	11.64	19881130	13.85	19870916	
00118	1987	13.31	19881201	29.48	19870715	
00084	1985	3.58	19881202	28.00	19860312	
00001	1985	14.95	19881202	33.10	19871014	
36446	1985	0.00	19881205	16.40	19860117	
36447	1985	11.89	19881205	16.47	19870915	
36431	1985	8.30	19881206	11.21	19850926	
36459	1985	5.40	19881206	8.66	19851022	
36440	1985	3.48	19881206	17.10	19851205	
36430	1985	6.64	19881207	13.12	19850920	
36433	1985	4.62	19881207	10.03	19850921	
36438	1985	1.36	19881207	4.97	19851018	
36436	1985	5.81	19881207	9.80	19851018	
36434	1985	8.77	19881207	17.16	19851106	
36435	1985	8.83	19881207	18.85	19851106	
36432	1985	5.00	19881207	9.98	19851203	
36464	1985	4.85	19881207	11.76	19851212	
36444	1985	10.22	19881209	13.73	19851917	

Table 4.1 Highest and lowest water levels recorded.

G-BASE DEWETSDORP Site ID and Map numbers.

All Site ID's start with 2926DA.....

S-ID	Num.on Map	S-ID	Num.on Map	S-ID	Num.on Map
00001	DHK 1	00086	KRN 6	20002	CASSELLA
00002	DHK 2	00087	KRN 7	20003	GAUGE POLICE
00003	DHK 3	00088	KRN 8	20004	GAUGE GLEN HTS
00004	DHK 4	00090	KRN 10	36429	G36429
00011	ELM 1	00091	ODT 1	36430	G36430
00012	ELM 2	00092	ODT 2	36431	G36431
00013	ELM 3	00093	ODT 3	36432	G36432
00014	ELM 4	00094	ODT 4	36433	G36433
00021	FRT 1	00095	ODT 5	36434	G36434
00022	FRT 2	00096	ODT 6	36435	G36435
00023	FRT 3	00097	ODT 7	36436	G36436
00024	FRT 4	00101	RNS 1	36437	G36437
00025	FRT 5	00102	RNS 2	36438	G36438
00026	FRT 6	00103	RNS 3	36439	G36439
00027	FRT 7	00111	RBY 1	36440	G36440
00028	FRT 8	00112	RBY 2	36441	G36441
00031	GHS 1	00113	RBY 3	36442	G36442
00032	GHS 2	00116	RBY 6	36443	G36443
00033	GHS 3	00117	RBY 7	36444	G36444
00034	GHS 4	00118	RBY 8	36445	G36445
00036	GHS 6	00121	SHK 1	36446	G36446
00041	GLT 1	00122	SHK 2	36447	G36447
00042	GLT 2	00123	SHK 3	36448	G36448
00043	GLT 3	10001	NEUTRON	36449	G36449
00044	GLT 4	10002	NEUTRON	36450	G36450
00051	JPN 1	10003	NEUTRON	36451	G36451
00061	1 A	10004	NEUTRON	36452	G36452
00062	2 A	10005	NEUTRON	36453	G36453
00063	3 A	10006	NEUTRON	36454	G36454
00064	4 A	10007	NEUTRON	36455	G36455
00065	5 A	10008	NEUTRON	36456	G36456
00066	6 A	10009	NEUTRON	36457	G36457
00067	7 A	10010	NEUTRON	36458	G36458
00068	8 A	10011	NEUTRON	36459	G36459
00069	9 A	10012	NEUTRON	36460	G36460
00070	10 A	10013	NEUTRON	36461	G36461
00072	12 A	10014	NEUTRON	36462	G36462
00073	12 B	10015	NEUTRON	36463	G36463
00079	C5N 524	10016	NEUTRON	36464	G36464
00080	C5N 525	10017	NEUTRON	36465	G36465
00081	KRN 1	10018	NEUTRON	36466	G36466
00082	KRN 2	10019	NEUTRON	36467	G36467
00083	KRN 3	10020	NEUTRON	36468	G36468
00084	KRN 4	20000	WEATHER		
00085	KRN 5	20001	TIPPING BUCKET		

Table 4.2 G-Base Site ID- and Map numbers.

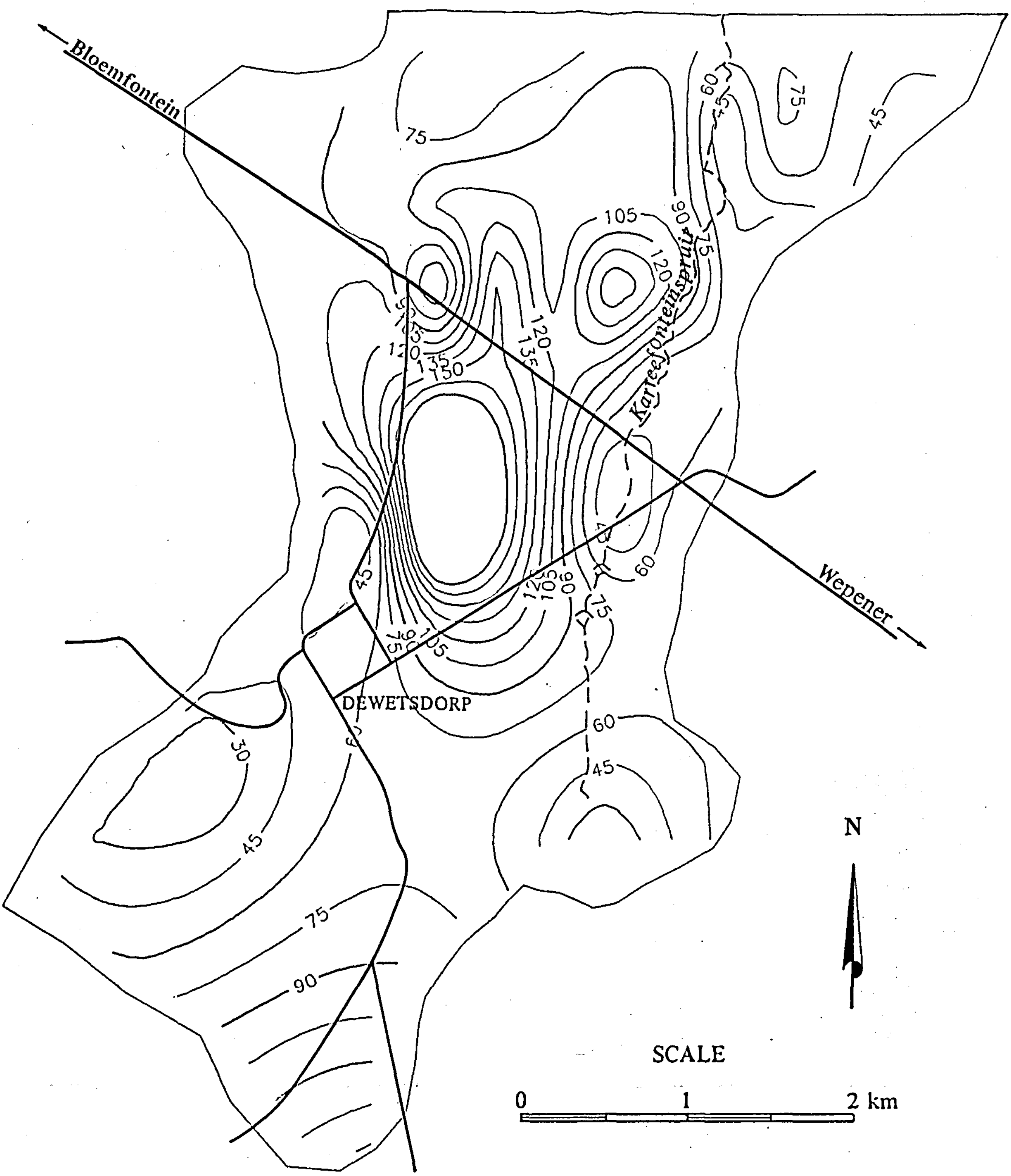


Figure 4.39 Delay in days of water-level rise after the February 1988 flood.

Such behaviour of aquifers was observed in Dewetsdorp in 1976. Overlying the annual trend of high water levels in the winter months and low levels at the end of the year, the water-level records for C5N 525 showed a high in ± 1958 and from there a declining level which was at minimum in 1965. From then on, it rose to the highest levels recorded so far in 1976, after very good rains had fallen in 1974 and again in 1976 (see Figure4.40). The same observation has been made in southern SWA where the good 1974 rains filled aquifers to a large degree. The smaller rains two years later caused a further water-level rise and some boreholes, e.g. between Mariental and Maltahöhe, began to overflow.

It then appears that water levels, in semi-arid climates where rainfall variability is high, show a general downward trend during good, normal and low rainfall years, in spite of occasional limited recharge. During the exceptional rainfall year or a succession of an exceptional and a good year, aquifers are completely recharged with springs appearing widespread and the soil soaking wet long after the rain. The water-level graph of borehole FRT 7 in Figure 4.29 possibly illustrates a part of such a cycle.

Hydraulic parameters

Pumping tests

Evaluation of pumping test data and double-packer testing is used to determine hydraulic parameter values.The positions of the test- and observation holes are indicated on the geological map Figure 4.18.

Pumping test G 36430

A step drawdown test, comprising of three steps with pumping rates of 1.18, 2.38 and 4.09 l/s followed by a 1-hour recovery period, was carried out on 19th September

Borehole	Method	Transmissivity [m2/d]	Storativity
G36430	Theis	7	0.2
G36433	Theis Cooper Jacob I Recovery	35 40 48	0.0002 0.0002 -
G36434	Theis Cooper Jacob I	7 12	0.0001 0.00007
G36435	Theis Cooper Jacob I	9 13	0.00005 0.00002
G36436	Theis Cooper Jacob I	13 17	0.00005 0.00003
G36464	Theis Cooper Jacob I Recovery	16 17 30	0.0006 0.0004 -

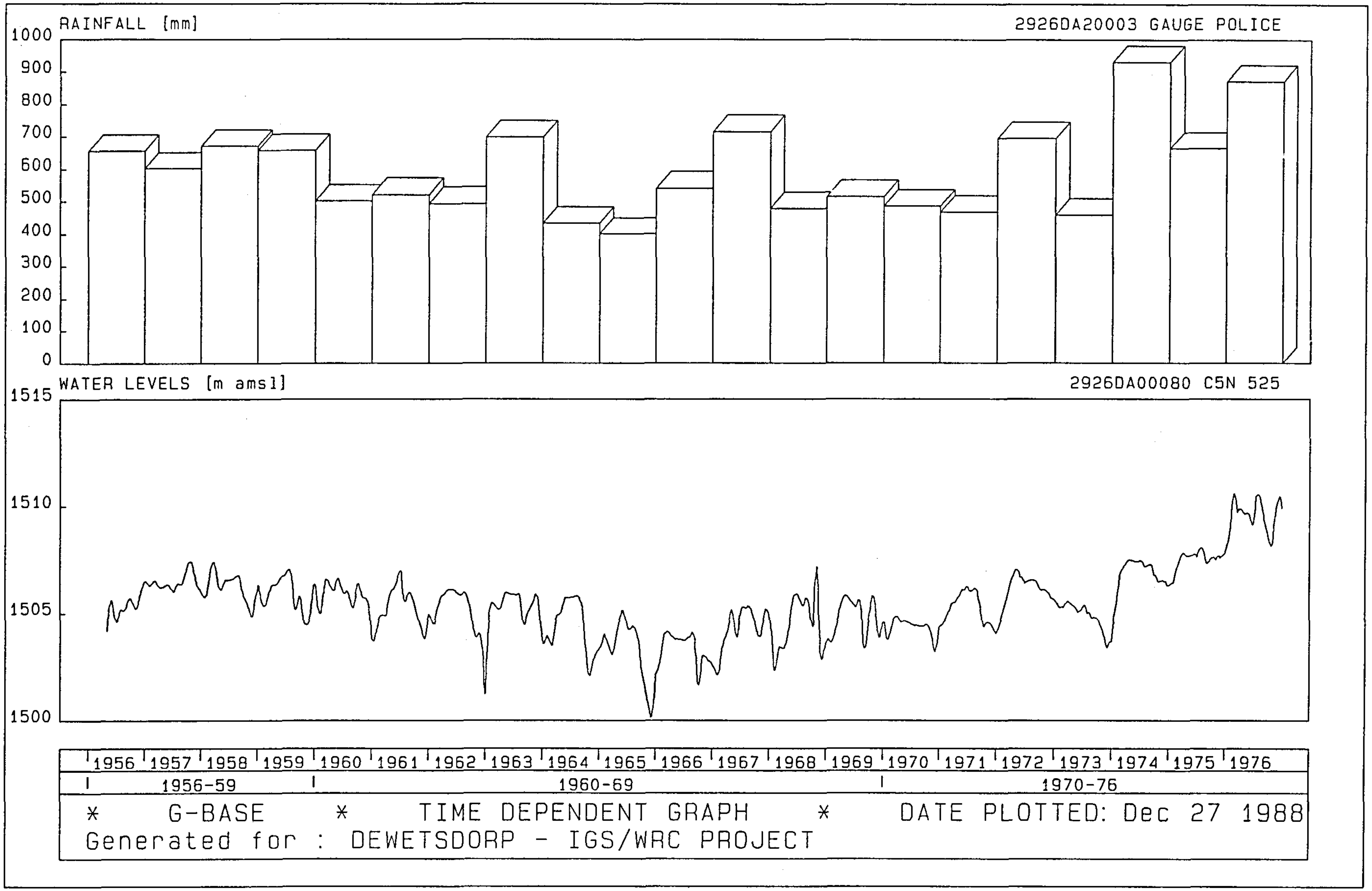


Figure 4.40 Water-level graph of C5N 525 and histogram of the annual rainfall for the years 1956 - 1976.

1985. The following day a 72-hour pumping test at a constant pumping rate of 2,45 l/s commenced. Water levels were monitored in the test borehole and the following observation holes: G 36429, G 36433, G 36434, G 36436, G 36437, G 36444, G 36445, G 36456, G 36459 and G 36464. Levels in the underlined boreholes did not respond to the extraction.
The graphs for these pumping tests evaluations are shown in the Appendix.

Pumping test G 36431

A step drawdown test, comprising of two steps with pumping rates of 0.65 and 1.8 l/s followed by a one-hour recovery period, was carried out on 26 September 1985. Four days later a pumping test at a constant pumping rate of 0,67 l/s commenced. Due to excessive drawdown, the test was stopped after six hours. Water levels were monitored in the test borehole and the following observation holes: G 36432, G 36437, G 36438, G 36440, G 36441, G 36442, G 36444, G 36465, G 36466, G 36467 and G 36468. Levels in the underlined boreholes did not respond to the extraction.

Pumping test 1A

Towards the end of three-day pumping tests carried out in 1985, indications were that delayed yield conditions are encountered in the Karoo aquifers in the Dewetsdorp catchment area. In collaboration between the Dewetsdorp Municipality and the IGS, a pumping test with a duration of 14 days, followed by a two-week recovery test, was performed on one of the municipal production holes. These tests were carried out between 21st June and 19th July 1988. The objective of the tests was to ascertain whether such phenomena can be observed and to establish whether transmissivity- and storativity values change with increased pumping time, because modelling of the aquifer and ground-water balance calculations indicated that storativity values, determined by shorter pumping tests, are an order too low.

Production boreholes, 1A and 2A, are situated ± 280 m apart from each other. They lie more than a kilometre away and upstream of the remaining five production holes of the Municipality. Boreholes 1A or 2A were chosen for the pumping test, because no interference from pumping in the main production field or of any other hole was expected. The final choice was borehole 1A. Five observation holes were available: a recorder C5N 524 at a distance of 21 m, another recorder C5N 525 ± 240 m away, borehole 2A and two further holes, G 36460 and G 36459, at a distance of approximately 600 m. The last- mentioned borehole showed no response to the pumping.

The physical parameters of the test- and observation holes are as follows:

Borehole	1A	C5N 524	C5N 525	2A	G36460
Depth [m]	23,30	26,50	48,20	46,00	30,00
Diameter [mm]	165	165	165	165	165
Rest water level [m]	7,23	7,28	8,11	8,45	12,40
Distance [m]	0,08	21,0	242,3	282,3	608,4

Usually, the Dewetsdorp production holes are pumped intermittently between 07:00 and 17:00, from Monday to Friday or Saturday, and they rest over the weekend. Boreholes 1A and 2A were switched off at 18:30 on Saturday, June 18, 1988, and recovered for approximately 61 hours before the pumping test commenced.

Main test

For the test, the submersible pump was used which was installed in the borehole. The pump inlet was at a depth of 23,2 m and the pump has a capacity of approximately 1,35 l/s.

During the day hours, the water was pumped into a pipeline, leading to the balancing reservoir at the old power station, from where it was pumped by a booster pump to the main reservoir. During the night, the water was discharged through a backwash pipe in the vicinity of the borehole.

The discharge was measured by a totaling meter. Measurements were taken frequently during the first hours of the test and thereafter twice daily when the discharge was switched over from network to waste and vice versa. This allowed the determination of the influence of the changing head on the discharge from the borehole.

On the seventh day, Monday, 27 June, a major power failure occurred in the Central and Southern Orange Free State and pumping was interrupted for approximately 9 hours. Because there was sufficient time left for the water level to reach equilibrium before the end of the pumping test and also because no recovery measurement was taken, it was decided not to abandon the test.

The average daily discharge rate declined slightly during the test from ± 1,46 l/s in the beginning to 1,33 l/s after the power failure (see Figure 4.41). This was not due to a declining water level, because maximum drawdown in the tested hole was observed after three days. Wearing of the pump seems to be the only explanation for this observation. Variations between day and night were 0,05 l/s or less, i.e. one to three per cent. The average discharge during the 14-day period was 1,38 l/s or 116 m^3/d.

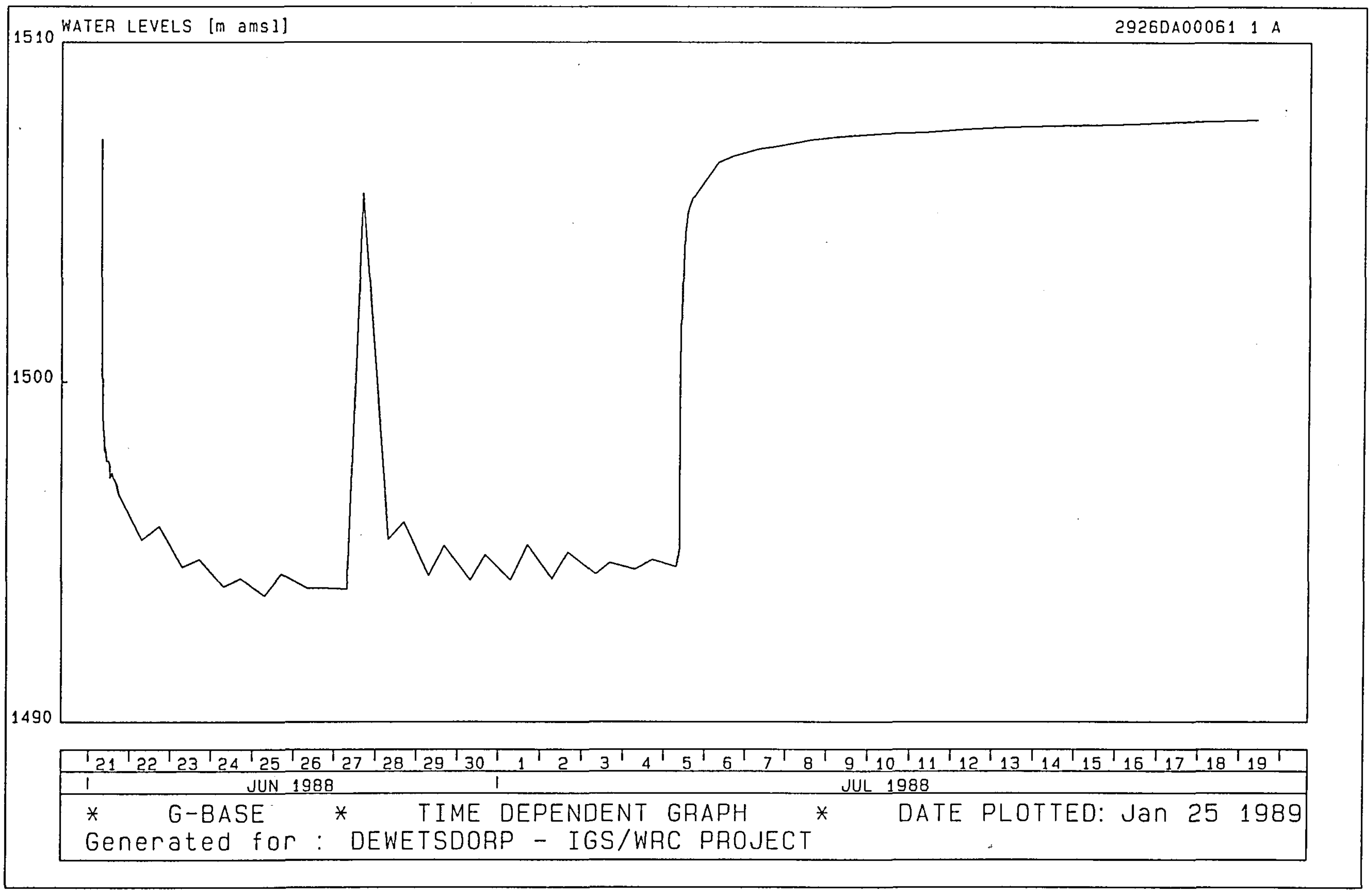


Figure 4.41 Water level in borehole 1 A during the pumping test in July 1988.

The above-mentioned observation that water levels in the tested hole were at minimum four days after the start of the test may, in addition to the declining discharge rate, have been caused by a certain amount of recirculation. Four days after the power disruption the drawdown in C5N 524 was the same as before the failure, but a slight recovery was noticeable during the last three days of the test. This may have been either due to possible recirculation or because of the slightly decreased pumping rate. Apart from a limited recovery during and shortly after the power failure, boreholes 2A and C5N 525 showed a declining water level throughout the whole test. In G 36460, water levels began to respond approximately 10 hours after commencement of the test and showed no recovery due to the power failure.

Borehole	Method	Transmissivity [m2/d]	Storativity
1A	Cooper Jacob I (first 6 days)	10	0.024
C5N 524	Theis	11	0.00065
	Cooper Jacob I (first 4 days)	10	0.00054
C5N 525	Theis	66	0.00036
	Cooper Jacob I (first 4 days)	50	0.00027
	Cooper Jacob I (last 10 days)	181	0.0000001
2A	Theis	65	0.000066
	Cooper Jacob I (first 6 days)	50	0.000082
G36460	Theis	12	0.00043
	Cooper Jacob I (last 10 days)	24	0.00032
All holes	Cooper Jacob II for 5750 min	13	0.00028
	Cooper Jacob II for 8620 min	13	0.00041

The graphs for these pumping test evaluations are shown in the Appendix.

Analysis of the drawdowns in all five boreholes, according to the distance/drawdown diagram for two times before the power disruption, yields an average transmissivity value of approximately 13 m^2/d and a storativity of ≈ 0,00035. The plotting positions in the distance/drawdown diagram (Cooper Jacob II) for the drawdown after t=5750 min and t=8620 min lie roughly on a straight line. Noticeable well loss would cause a greater drawdown in the pumped hole. Because this is not the case, the diagram also indicates that the drawdown in the pumped borehole is caused by aquifer loss only and that the well loss for a pumping rate of ± 120 m^3/d is negligible.

Recovery

Water levels of the test- and the observation holes were monitored for two weeks after pumping had been stopped. With the exception of borehole G 36460, water levels recovered quickly within less than four days in the test hole and six days in borehole C5N525. Residual drawdown in borehole G 36460 was 0,16 m after 14 days.

Borehole	Method	Transmissivity [m2/d]
1A	Theis Recovery	11
C5N 524	Theis Recovery	14
C5N 525	Theis Recovery	37
2A	Theis Recovery	43
G 36460	Theis Recovery	27

The slow recovery here is attributed to the fact that the hole lies very near to the watershed on the downstream of the cone of depression, where it will take the longest time to recover.

Analysis of the observed data yielded transmissivity values which correspond well with the main test results. Air entrapped in the system during recovery, results in the slightly higher calculated transmissivity values. The graphs for the recovery evaluations are also included in the Appendix.

Discussion

The aquifer is semi-confined. Based on the main- and recovery tests, the following parameter values are considered representative for the pumping test:

Transmissivity [m^2/d]	13
Storage coefficient [%]	0.0004

Along certain structures, e.g. along the dyke between A1 and A2, a higher than average transmissivity is observed.

Packer tests

Double-packer tests were carried out in the following seven boreholes to establish the average permeability for Karoo rocks and its variability with depth: G 36429, G 36430, G 36431, G 36437, G 36443, G 36449 and G 36450. When double-packer testing was applied, a 5 m long section of the borehole was sealed off by the packers and water was injected at varying pressures, e.g. 0, 10, 20 and 30 m water pressure. After a certain time, which depends on the nature of the formation, the flow stabilizes. In the case of Karoo Formations below the water level, a constant flow was reached within minutes. The constant flow rate was first measured at increasing pressures and a second time at decreasing pressures. This was done to identify changes in the permeability during the test, which may have been caused by deformation and opening or closing of pathways, due to moving material during the test. Such changes were, however, neither noticed in Dewetsdorp nor in De Aar.

Because the gauge is situated above the surface, the reading does not show the pressure between the two packers. The water column between the water level and the gauge causes an additional pressure. There may, on the other hand, be friction losses in the pipe from the pump to the packer. The friction losses in the pipe were measured and compared well with the calculated losses. In cases where water does freely run out of the supply pipe (when joints are open and no pressure can be built up), the flow rate is at maximum and friction losses are equal to the gauge reading. The quantity of water taken up by the joints may be higher than the maximum flow rate of the supply pipe. The consequences thereof are discussed below.

When there is laminar flow, the constant flow rate Q is proportional to the effective head H and the tangent of the gradient of the Q/H curve is proportional to the average permeability of the tested section. If the capacity of the joints is higher than the measured flow rate, the effective head will be constant and the Q/H curve will plot as a vertical which suggests an infinite permeability.

Evaluation of the Dewetsdorp data shows that the measured flow rates and the effective heads generally lie on a straight line through the origin. This suggests laminar flow through very small joints.

Because the aquifer consists of layers with different permeabilities, the transmissivity (a term which is found in the ground-water flow equation) is equal to the sum of the product of permeability k and thickness of each layer L., i.e. the average permeability times the length of the test section. The k x L graphs show a general decrease of the permeability with depth. Anomalies are associated with more permeable contact zones along dolerite intrusions. The decrease of the transmissivity with lower piezometric level can also be deduced from these graphs. On Figure 4.42, the ground surface has been chosen as a reference level to show how the transmissivity decreases exponential with dropping water level. This probably explains why, during successive years of low recharge, borehole yields decrease disproportionately when water levels become relatively deep.

In order to verify the packer test, the results have been compared with the 72-hour pumping test done on borehole G 36430. The test was evaluated, using the Hantush method which has been developed for ± steady-state flow in semi-confined aquifers (Kruseman & De Ridder, 1970, p. 88). The calculated transmissivity value of T = 6,58 m^2/d was not essentially different from the result of the Theis method which yielded T = 7 m^2/d, although a slightly better fit of the curve was obtained with the former method.

It must be mentioned that in accordance with a general decreasing permeability with depth, the weathered zone in Karoo aquifers reacts like a phreatic aquifer. Locally

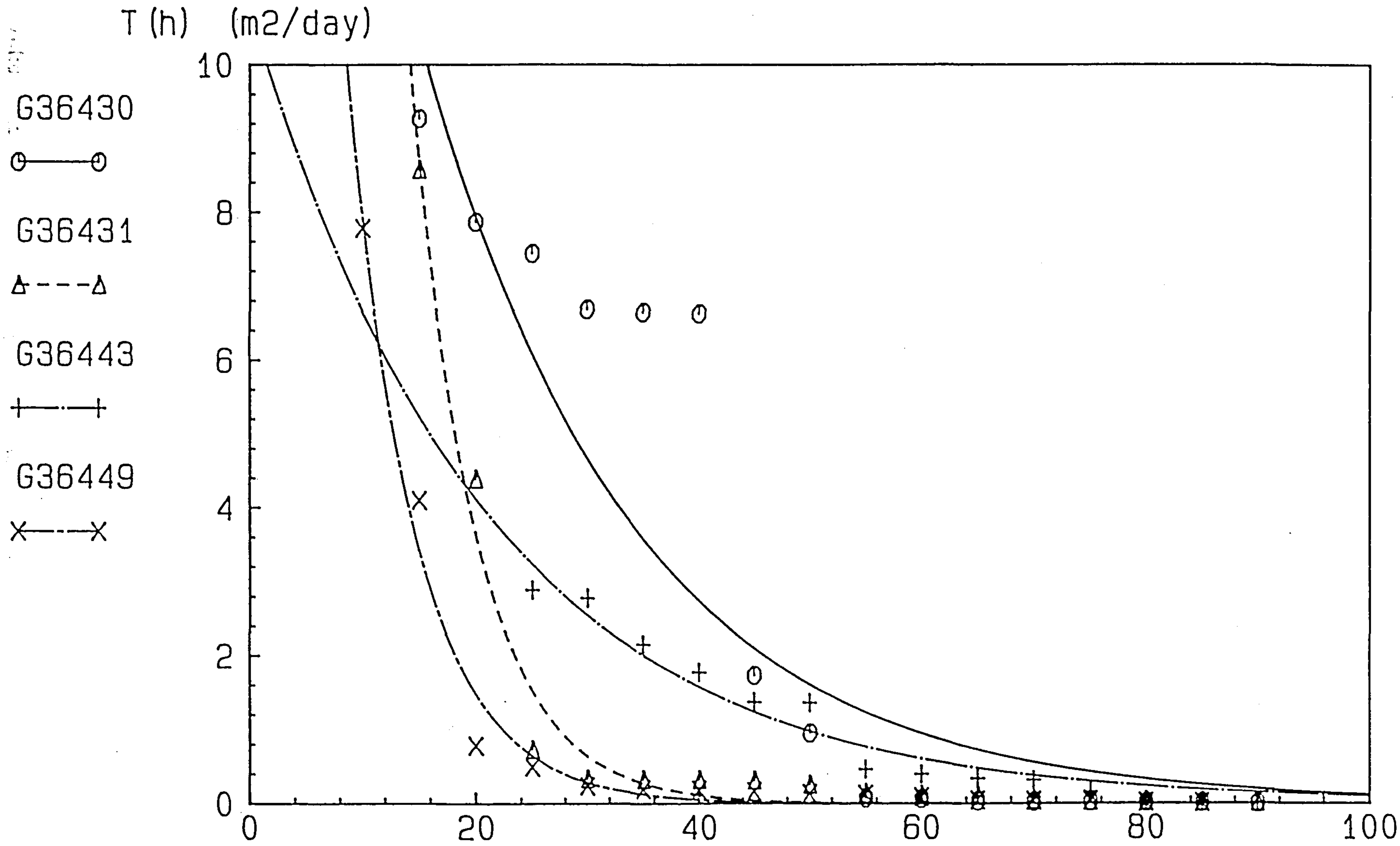


Figure 4.42 Transmissivity vs. depth of water table below surface.

semi-confined conditions may occur where relatively high permeable zones are encountered at depth as in borehole G 36430. The top of the permeable layer is approximately 40 m below the surface (see Figure 4.43). The transmissivities for the respective 5 m sections are shown in Figure 4.44. Should the water table be at this depth, the transmissivity would be about 6,6 m^2/d, as calculated according to Hantush.

Quality

The quality of the ground water in Dewetsdorp has received early attention. High fluoride concentrations gave rise to an investigation by Kent (1950). Thirty-seven analyses, most of them taken in 1985, show fluoride values between 0,2 and 9,9 mg/l, with the following distribution:

F (mg/l)	0 - 0,5	0,51 - 1,0	1,01 - 1,5	1,51 - 2,0	2,01 - 3	above 3
Number of samples	6	14	8	3	4	2
Per cent of samples	16	38	22	8	11	5

The analyses of the municipal boreholes had fluoride values below 1,0 mg/l. However, borehole 2A had a relatively high nitrate concentration of 18,3 mg/l. Higher fluoride values were found in:

G36429: 2,8 mg/l at 77 m and 2,2 mg/l at 101 m.
G36430: 2,7 mg/l at 45 m, 1,9 mg/l at 100 m (but a low fluoride value at 19 m).
G36435: 9,8 mg/l at 19 m and 7,1 mg/l at 41 m.

The analyses are found in the Appendix and will not be discussed in detail. Except for the fluoride found in some of the holes, the quality is generally within the drinking standard limits with TDS values between ± 250 and 400 mg/l. Higher values of up to 640 were measured in some of the private holes in the village. These holes also have increased sodium-, chloride- or nitrate levels which indicate local pollution.

Piper diagram:

Boreholes 12 A, G36437, 6 A, 3 A, 5 A and, to a certain degree, the uppermost sample from G36430 plot near the recharge corner of the diagram (see Figure 4.45). Also near that corner, but along a parallel line with slightly lower bicarbonate concentrations, lie 1 A, 2 A and 8 A. The first line continues with G36452, G36461, DEW 51 (in the village); the same cation relation as these samples, but less bicarbonate have G36442 and G36435. Without repeating the prefix G364, the analyses 33, 67 31, 36, 65, 39 and 41 follow the first line. DEW 47 (in the village) lies on the parallel line. Continuing along the first line towards the bottom corner of the diagram, lie 55,

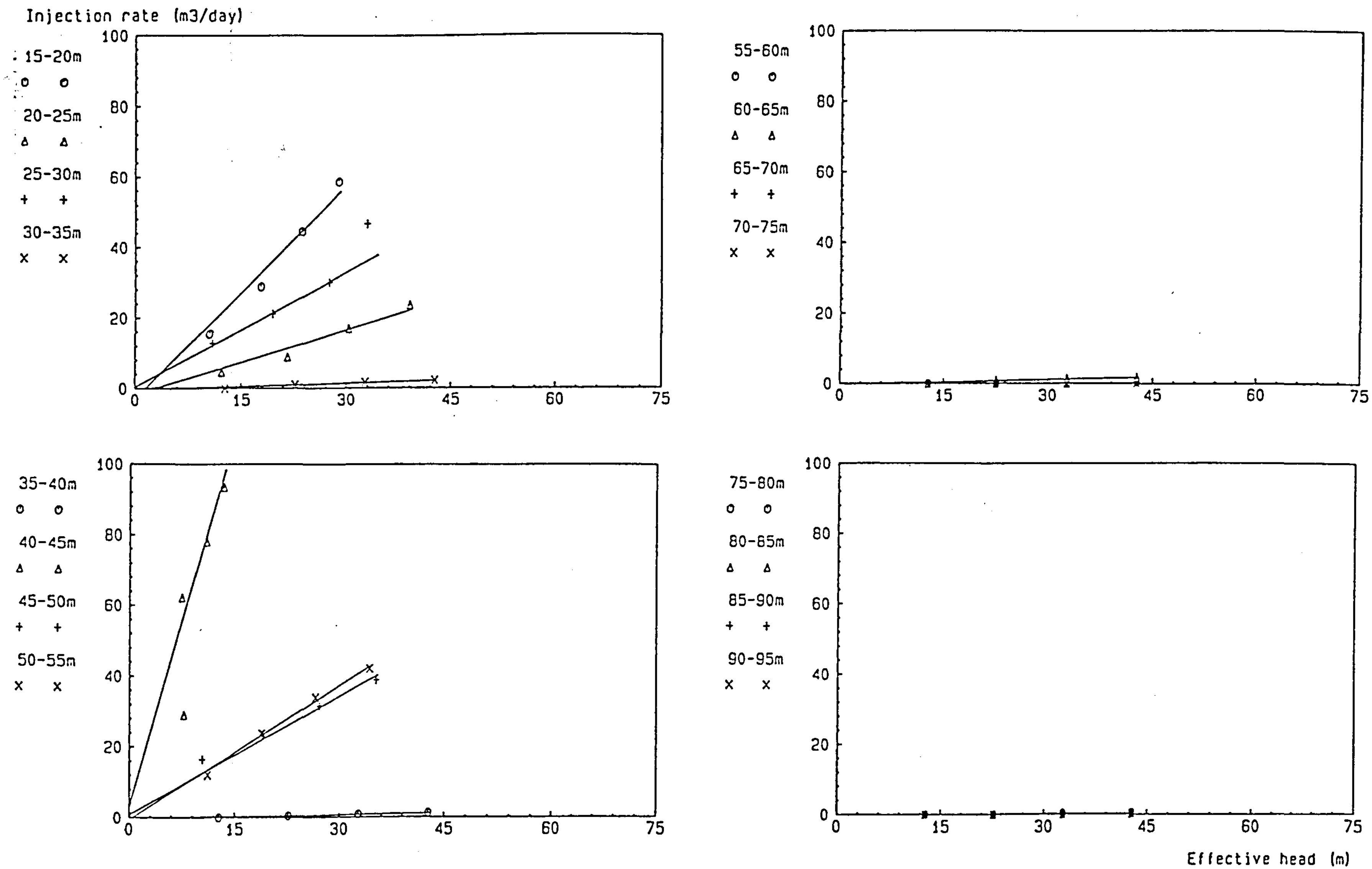


Figure 4.43 Packer test of borehole G 36430: injection rate vs. effective head.

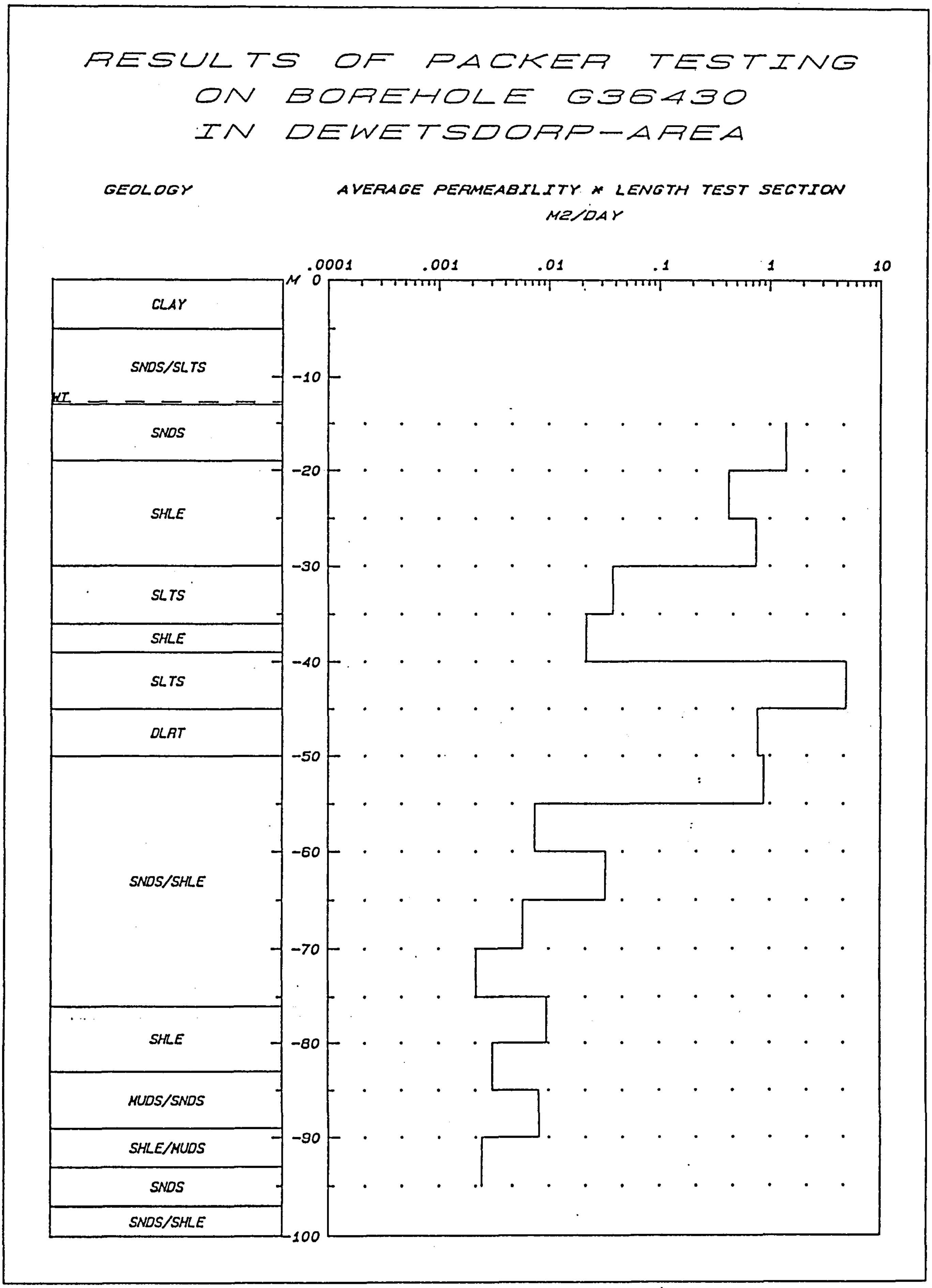


Figure 4.44 Packer test of borehole G 36430: transmissivities of test sections.

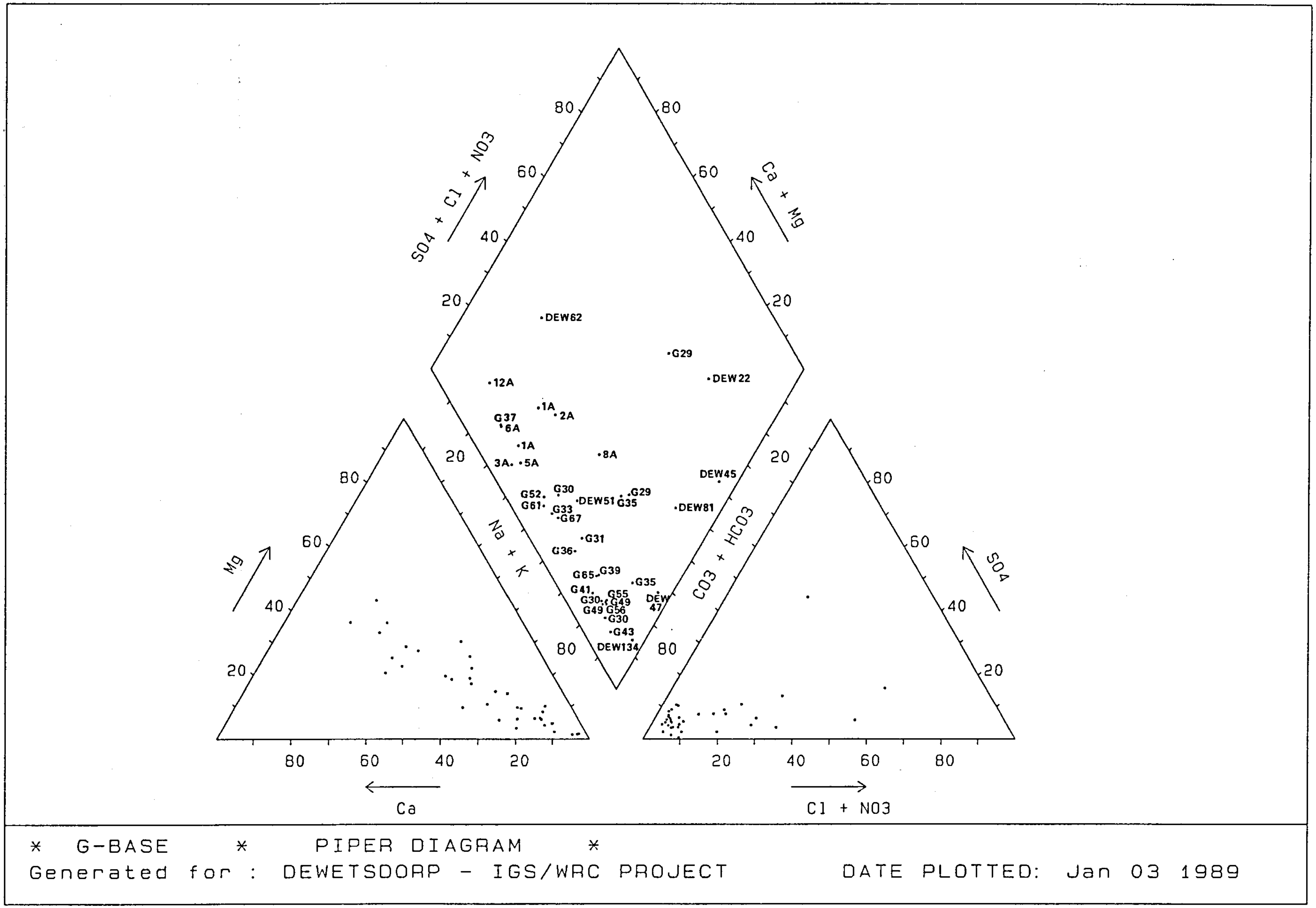


Figure 4.45 Piper Diagram : Dewetsdorp analyses.

30 (at 45 m), 49 (at 30 m), 56, 49 (at 100 m), 30 (at 100 m), 43 and DEW 134 (in the village).

Along the lower right side of the diamond towards the 'stagnant water', lie the remaining holes in the village (DEW 81, DEW 45 and DEW 22). The lowermost sample from G 36429 plots not far from the right corner of the diamond. DEW 62 lies in the upper half of the diamond.

If one looks at the plot positions for the samples from three boreholes at which samples were taken at different depths, one finds that the uppermost samples from G 36429 and G 36430 lie approximately halfway down from the recharge corner, although the former has considerably less bicarbonate. The two deeper samples of G 36430 and both samples from G 36449 lie in the lowermost corner of the diamond, while the second sample from G 36429 plots near the right corner of the diamond.

These findings are somewhat contradictory to the deductions made from the response to recharge, in that the samples from the municipal production holes all lie in or next to the recharge corner of the diagram, while recorder holes, such as G 36439 and G 36441 with 70 to 80% (Na + K), react directly to rainfall. A tentative explanation might be found in the theory that due to continued pumping and the corresponding increased flow velocity, less exchangeable sodium might be available along the flow paths leading towards the production holes. Both samples from G 36449 which lies just downstream of the well field and is in a similar position with respect to the Kareefontein Spruit, plot next to the bottom corner. Similarly surprising, though slightly less, are the plotting positions of samples from G 36455 and G 36456 along the Frankfürt fence near the bottom of the diamond and those of the holes in the village. These latter ones may, however, be affected by irrigation return flow; a hypothesis supported by the observed higher nitrate concentrations.

Another explanation for the unexpected plotting position of the municipal boreholes could be sought in the fact that the respective samples were taken in 1976, after a series of very good rainfall years and might as such not be representative. More recent analyses, especially ones from samples taken during the drought, are not available.

From the plotting positions of the samples, it can also be deduced that cation exchange takes place over very short distances of probably not more than a few hundred metres, e.g. at the recorder hole on the golf-course, G 36443.

The diagram confirms the deductions regarding different aquifers encountered in boreholes, G 36429, G 36430 and possibly G 36435, indicating (semi-)confined conditions for the lower aquifers.

SAR diagram:

The majority of the holes plots in the C_2S_1 field of the diagram (see Figure 4.46). The exception are DEW 134, DEW 45 and DEW 47, which have higher sodium values and lie in the C_2S_3 field, DEW 22 and DEW 81 in the C_3S_2 field and DEW 51 in the C_3S_1 field (on the diagram the numbers are shown without the prefix DEW).

Natural Isotopes

For determination of natural isotopes, ground-water samples have been collected from boreholes after completion and rain samples between November, 1985, and January, 1986, and again during the heavy rains in February/March, 1988.

In Figure 4.47, the oxygen-18 and deuterium concentrations of the rain samples which were taken in 1985/86 are plotted in chronological order. The sample numbers on the graph consist of an identifyer DR (Dewetsdorp Rain), the date of the rainfall event and a consecutive number. Each sample represents a precipitation of 5 mm. Samples from the different rainfall events are separated by vertical lines.

The general trend of high concentrations at the beginning of a rain event which decrease during the later stages is clearly visible. Also evident is that $\delta^{18}O$ and δD are running parallel, although deuterium displays much larger variations. In Figure 4.48, the $\delta^{18}O$ is plotted versus δD for ground water and rain water. The good correlation of the two stable isotopes of $R = 0,97$, as well as the usual slope of the regression line is evident in Figure 4.49. Ground-water samples show a certain small depletion in deuterium and lie not far from the (weighted) mean of the precipitation that has fallen during this period. All in all, the stable isotopes display as postulated in the literature, i.e. rather varying within a rainfall event and increasing depletion of the isotopes during the storm. The results of the Dewetsdorp samples support the assumption that local rainfall is responsible for recharge and no indication of ground water of really different origin was found.

Twelve rainfall samples collected by the automatic sampler at the beginning of the heavy rains between 16 February and 19 February, 1988, had a mean $\delta^{18}O$ of -6,14‰ which compares relatively well with the mean of the December, 1985, rains. A composite sample taken from the Casella rainfall recorder comprising 337,4 mm rainfall that had fallen in the period 10 February to 3 March, 1987, yielded a noticeably higher $\delta^{18}O$ value of - 8,35.

It is doubtful whether natural isotopes variations can be used for quantitative aquifer recharge assessment through rainfall. Such an approach would at least have required a much higher sampling frequency of rainfall and of ground water from boreholes that have been found to respond well and without too much lag in time to rainfall.

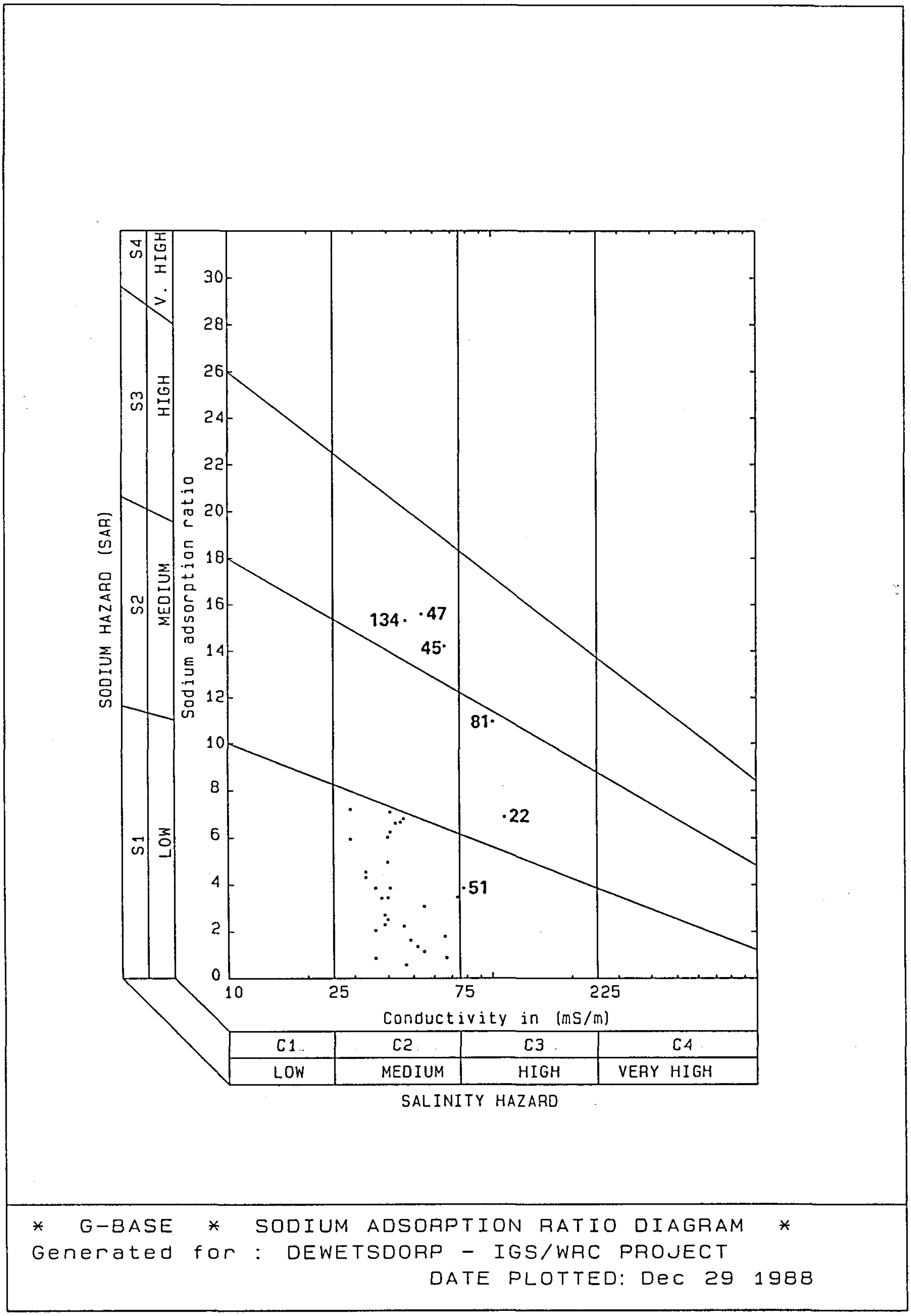


Figure 4.46. SAR Diagram : Dewetsdorp analyses.

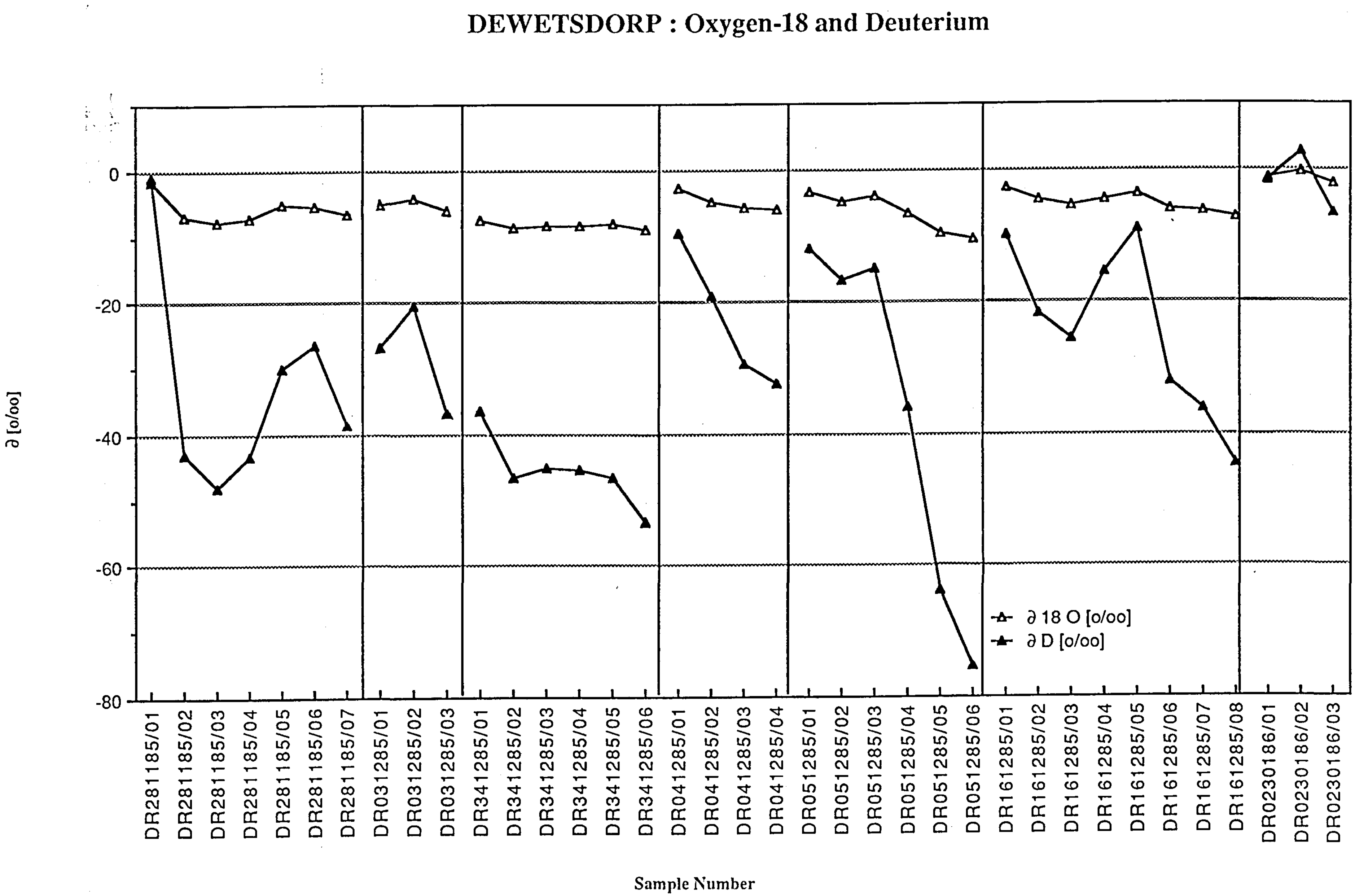


Figure 4.47 Dewetsdorp oxygen-18 and deuterium concentrations in chronological order.

DEWETSDORP: Ground-, Rain-, and Mean Rain Water

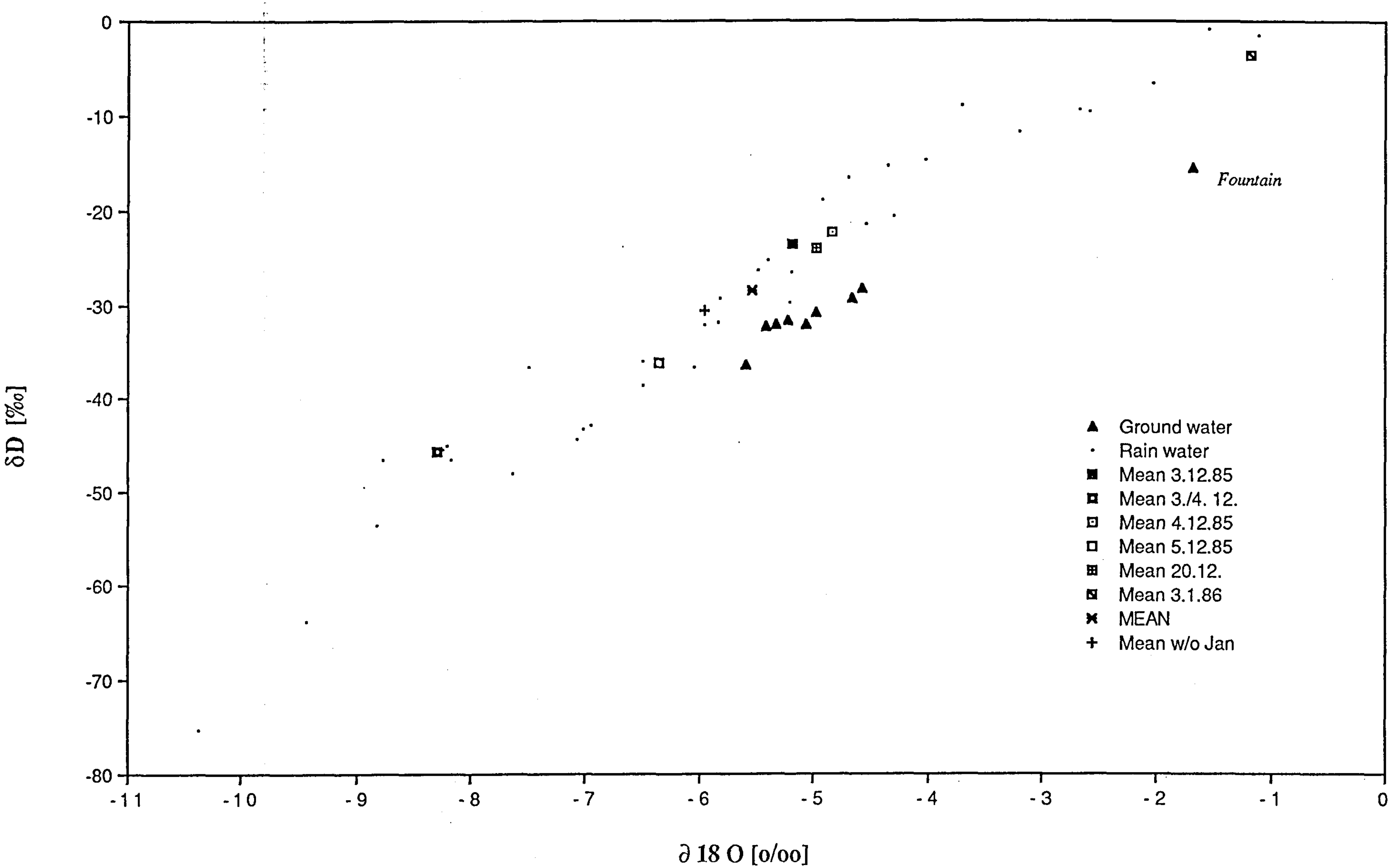
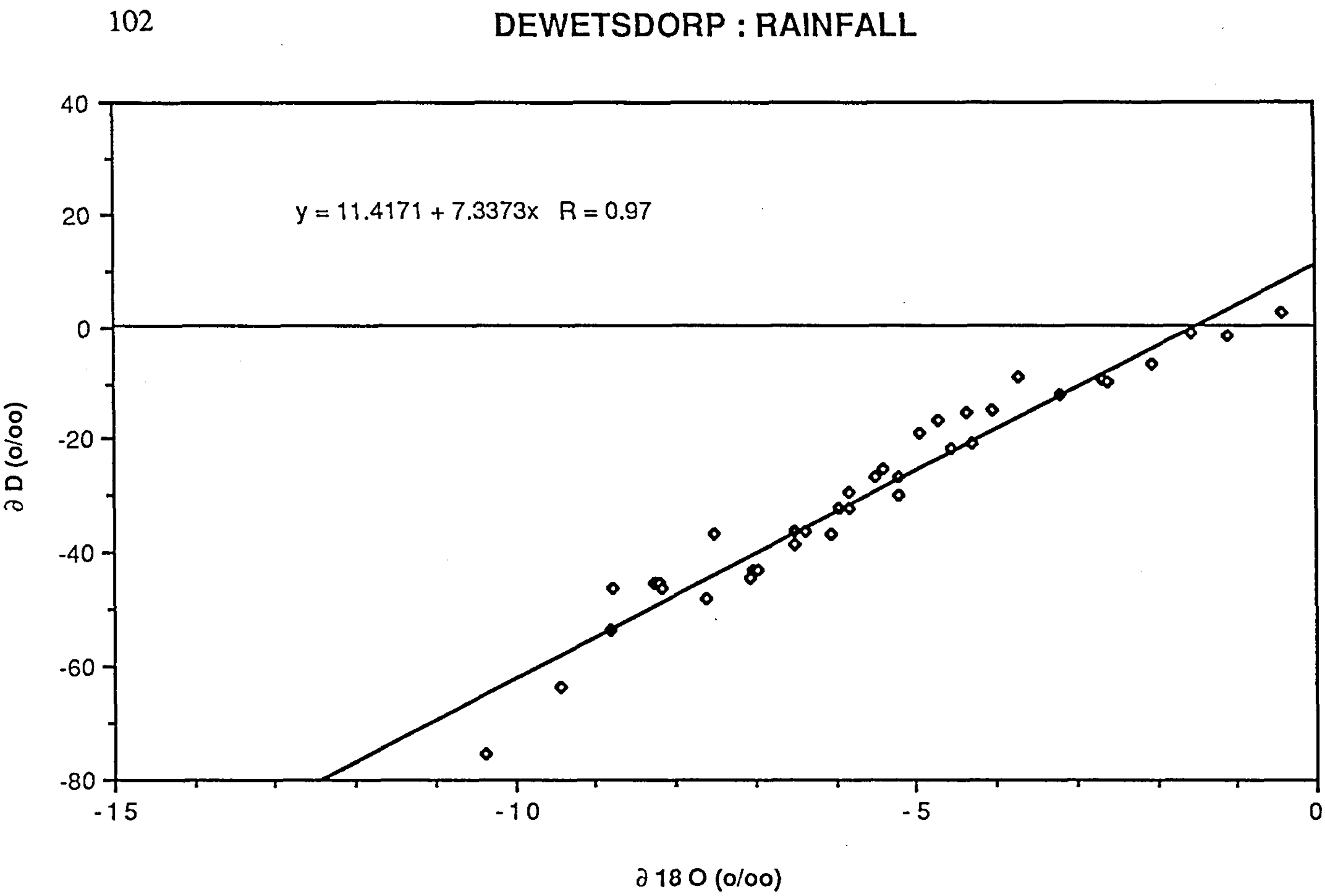


Figure 4.48 Oxygen-18 and deuterium concentrations of rainfall and ground water for various rainfall events.



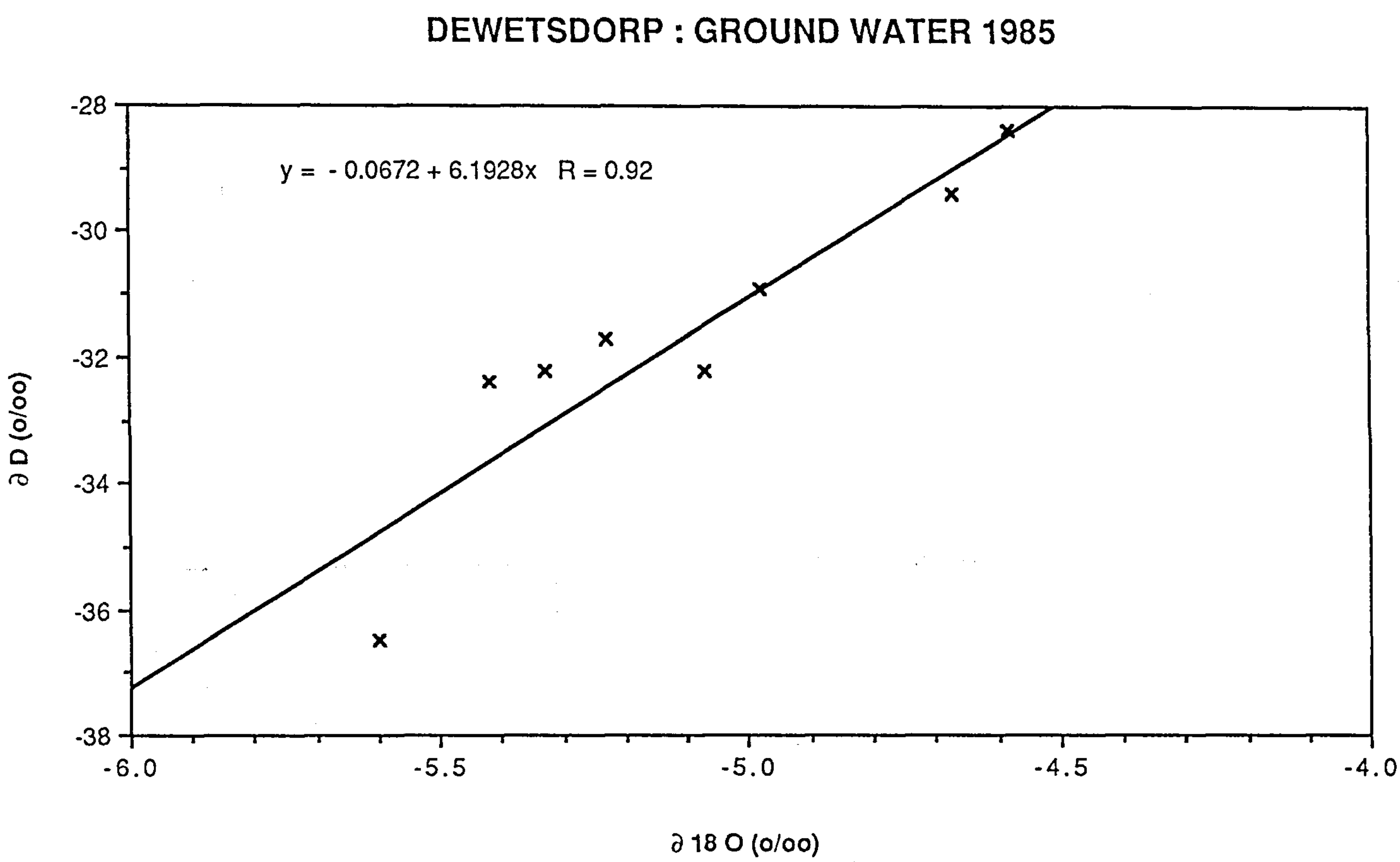


Figure 4.49 Oxygen-18 and deuterium concentrations of rainfall and ground water.

4.1.6 ESTIMATION OF RECHARGE

4.1.6.1 THE SVF-METHOD

The saturated volume fluctuation (SVF) method transforms ground-water level fluctuations above a given datum plane to equivalent amounts of water from a constructed saturated volume graph and the specific yield concept, together with the net inflow and outflow from the aquifer, as described by Equation (3.4.7).

The procedure followed for the calculation of the saturated volume for the Dewetsdorp aquifer is outlined below:

(i) Firstly, the position of the topographic ground-water divide was deduced from reports (Dziembowski, 1977). Figure 4.17 shows the position of these boundaries for the southern, western and eastern parts of the aquifer. The ground-water level contours confirm the existence of these divides. As can be seen from Figure 4.25, ground-water outflow occurs in the northern part of the aquifer.

(ii) The part of the aquifer, outlined on Figure 4.17, was discretized into 934 triangles with 513 nodes (Figure 4.50) and covered an area of 21 km^2. The program TFEM was used to construct this mesh.

(iii) Monthly ground-water levels at 72 boreholes were used by the program SVF to calculate the saturated volume for each triangle and consequently for the whole aquifer. An arbitrary water level, lower than any observed level, was used as the basis of the aquifer. The minimum saturated volume was calculated for October, 1985. This volume has been subtracted from that of all other months. Figure 4.51 shows the saturated volume for the period October, 1985, to July, 1988, (a period of 34 months) with the volume of October, 1985, as a basis.

The SVF-method will subsequently be applied to a number of special cases for Dewetsdorp.

Case 1

Equation (3.4.8) describes the case where the change in ground-water storage is zero (i.e. $\Delta V = 0$):

$$GR = A_2T_0 - A_1 T_I + Q$$

which simplifies to

$$GR = Q + A_2 T_0 \qquad (4.1)$$

for Dewetsdorp, because $A_1T_0 = 0$ for the region.

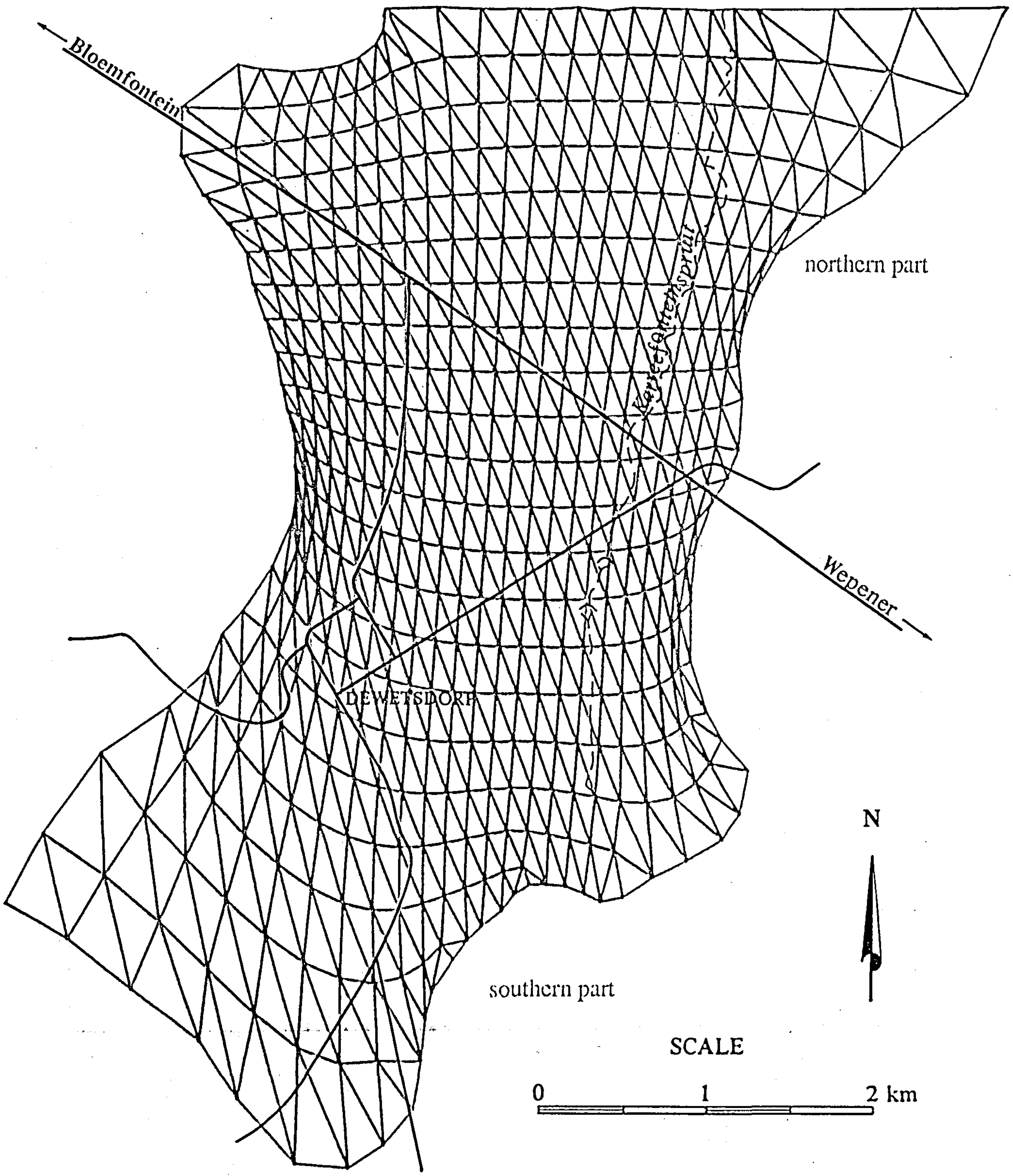


Figure 4.50 Triangular mesh constructed for the Dewetsdorp aquifer for the SVF-method.

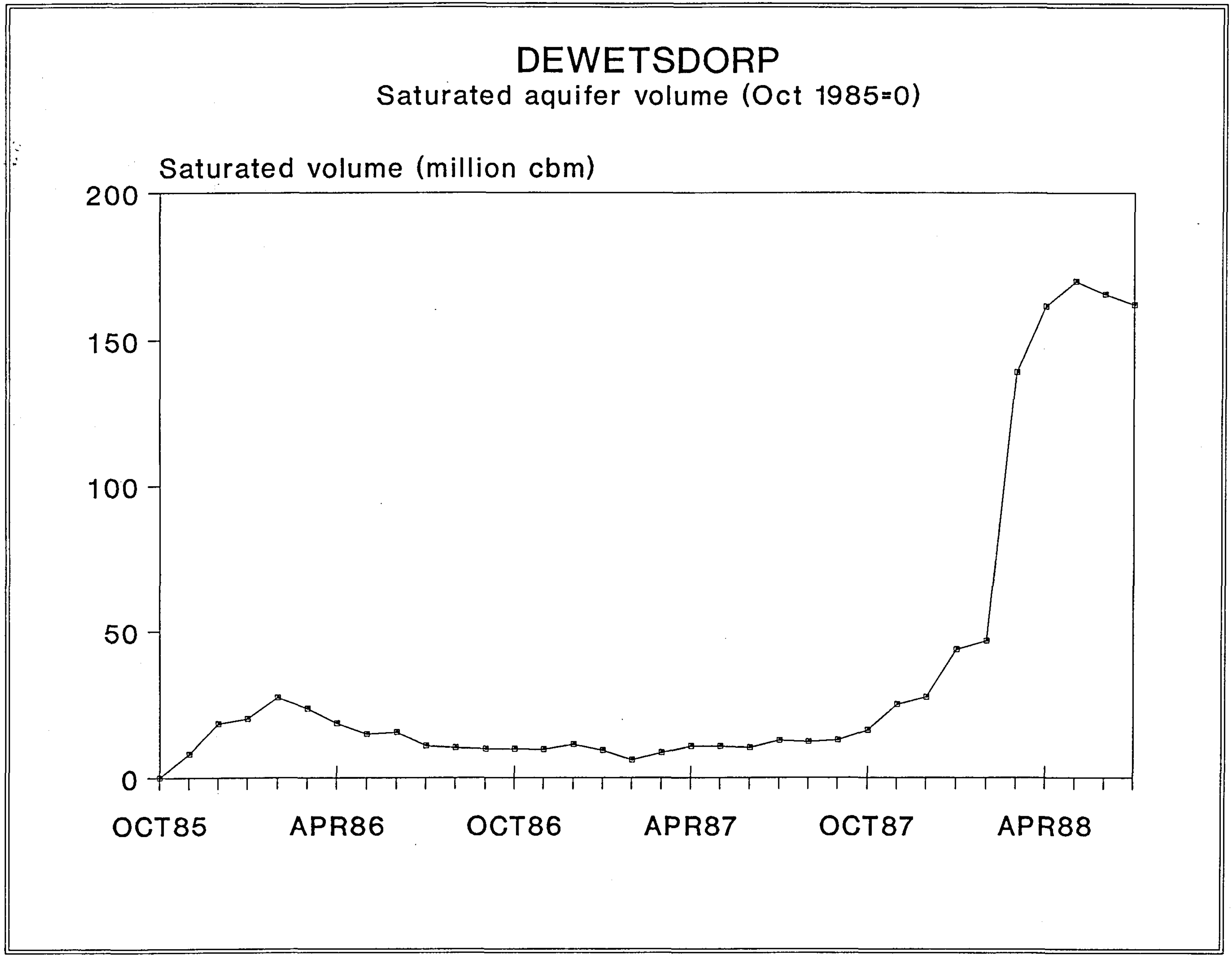


Figure 4.51 Saturated volume of the Dewetsdorp aquifer obtained with program SVF.

The Q-term in Equation (4.1) can easily be quantified, because the abstraction rate of the municipality boreholes is monitored (= about 50 000 m^3/a). Ground-water abstraction by other consumers in the township amounts to more or less 13 500 m^3/a (Dziembowski, 1977). The ground-water abstraction by farmers in the aquifer is an unknown quantity and was disregarded in this study. It is, however, believed that this quantity is less than 5% of the total abstraction.

The quantification of T_0 in the second term of Equation (4.1) is very important. This means that transmissivity of 10 m^2/d at the outflow boundary of the Dewetsdorp aquifer was deduced from (a) pumping tests and (b) packer tests performed in the aquifer. To obtain a suitable calibration of the two-dimensional ground-water flow model (which will be described in a following section), it was necessary to allow for a mean outflow at this boundary, of 470 m^3/d, which is in agreement with an outflow T of 10 m^2/d. The A_2-term in Equation (4.1) represents the mean ground-water gradient times the width of the outflow boundary and is an output of SVF.

From Figure 4.52, which is a shortened version of Figure 4.51, periods of $t_i - t_j = \Delta t$ for which $V_i = V_j$ can be calculated. For example, December, 1985, to April, 1986, was a period for which $\Delta V = 0$.

If Equation (4.1) is applied for all such periods, the graph illustrated in Figure 4.53 is obtained, which shows the relationship between rainfall and the ground-water recharge (infiltration) for different periods of time.

A regression analysis yielded the following relationship between rainfall and recharge:

$$GR = 0{,}024 \ (P - 51) \ [mm] \tag{4.2}$$

with a correlation coefficient r = 0,99

where

GR = ground-water recharge [mm]
P = rainfall [mm]

With the exception of one recharge value, Equation (4.2) was determined, using periods of at least three months. When using Equation (4.2), only periods longer than three months are recommended.

The coefficient of variation (C.V.) is defined as the standard deviation divided by the mean, and equals 26% for the mean annual rainfall in the Dewetsdorp region.

For the mean annual rainfall of 587 mm, the mean annual ground-water recharge of 12,8 mm was calculated, which is equal to a recharge of 2,2% of the annual rainfall. The C.V. for these values is 26%, which implies that the mean annual ground-water recharge of the Dewetsdorp region usually varies between 9,5 and 16,1 mm.

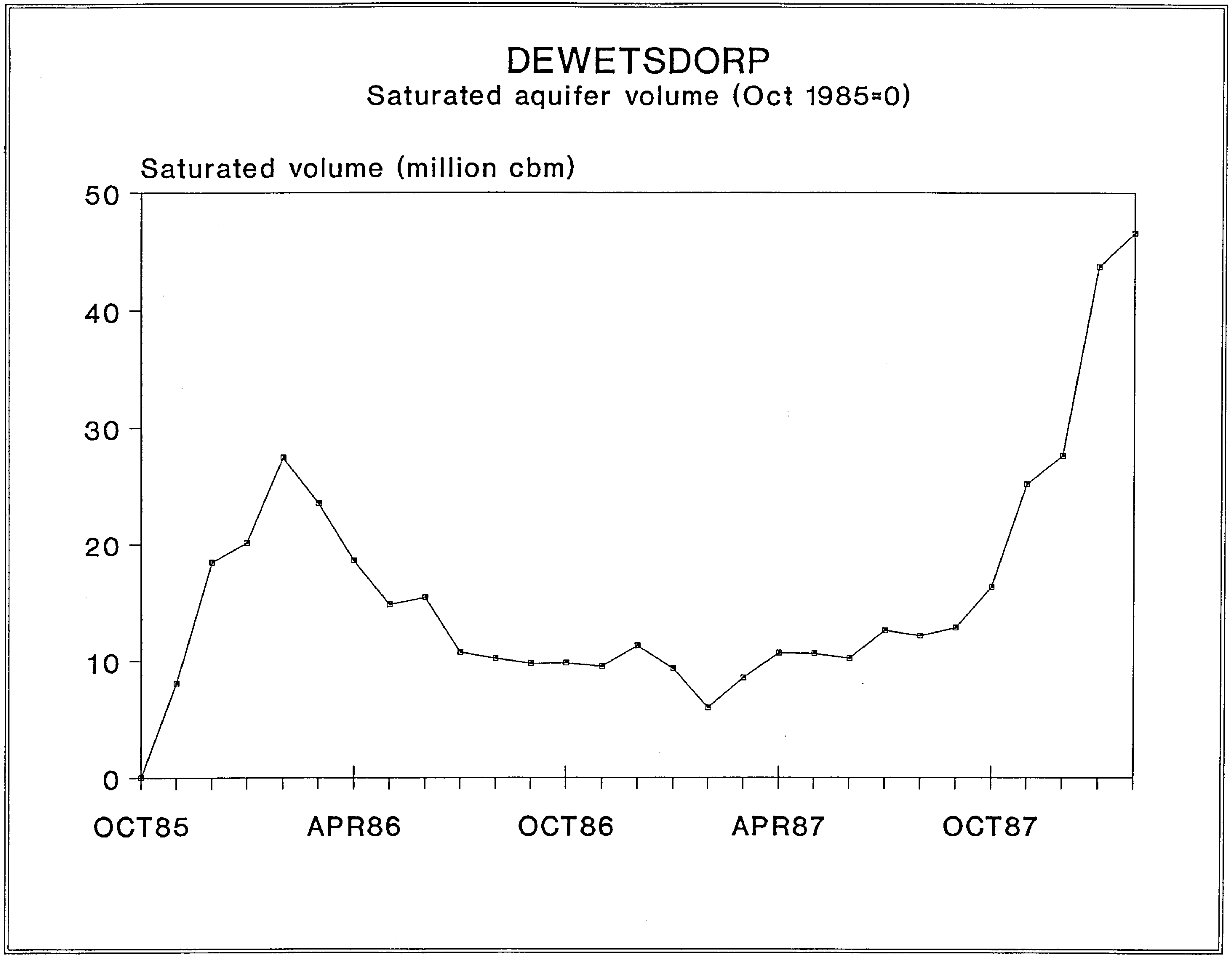


Figure 4.52 The saturated volume of the Dewetsdorp aquifer until the February 1988 flood.

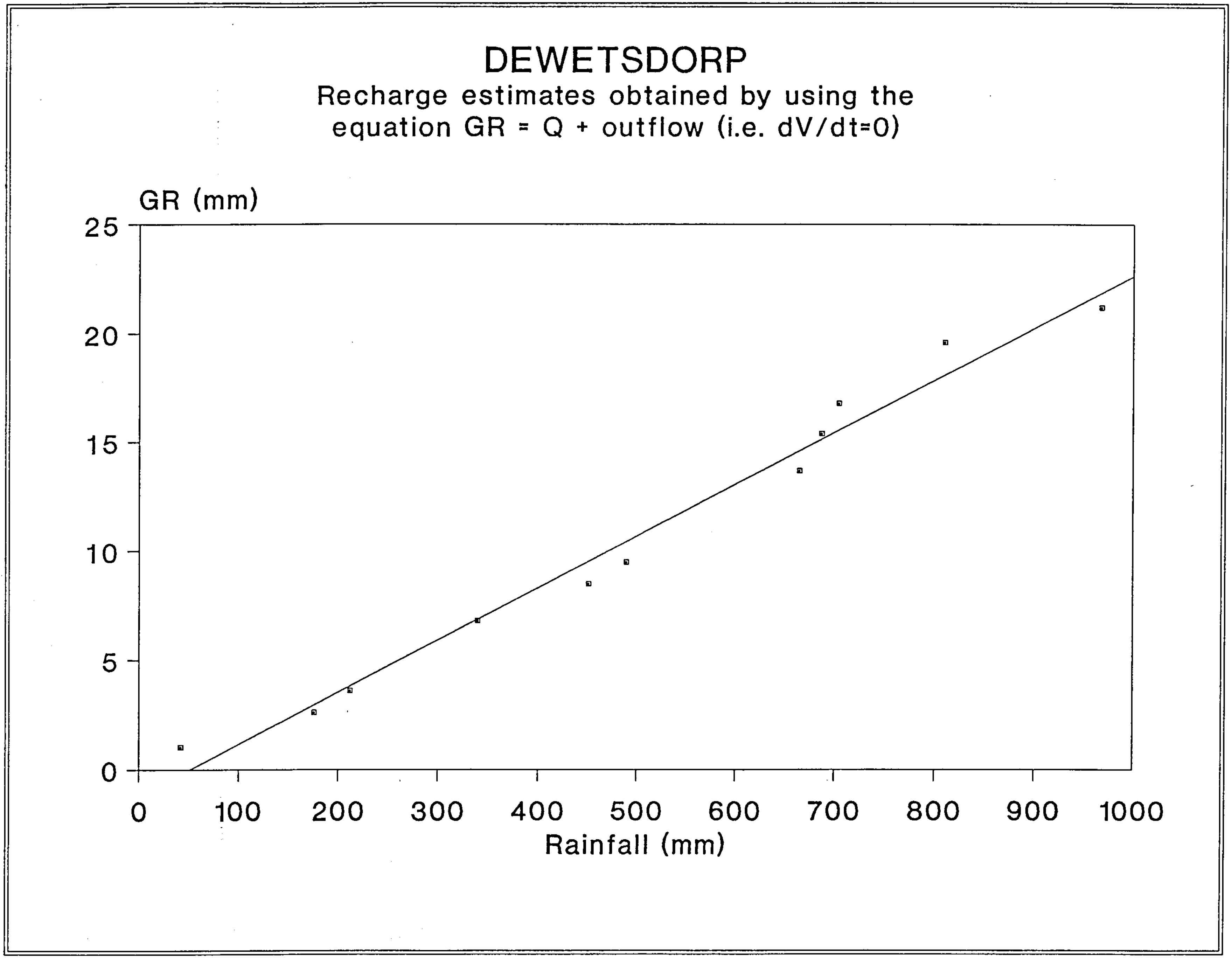


Figure 4.53 Graph showing the relationship between rainfall and ground-water recharge as obtained with the SVF-method for the case $\Delta V/\Delta t = 0$.

Case 2

Equation (3.4.7) is the general ground-water balance equation for an unconfined aquifer and for the Dewetsdorp aquifer reduces to:

$$GR = Q + A_2\,T_O + A_3\,S \tag{4.3}$$

where

$$A_3 = \frac{\Delta V}{\Delta t}$$

Whereas Equation (4.1) is independent of the specific yield (S), it is not the case with Equation (4.3). If it is assumed that a maximum decrease in saturated volume constitutes a "no recharge" or minimum recharge period and if the abstraction rate remained approximately the same, an estimate of the mean S-value for the aquifer can be obtained.

During the period 1 April, 1986, to 1 June, 1986, the maximum decrease in saturated volume occurred and equaled -4,938 $\cdot 10^6$ m^3, so that

$$S = \frac{-A_2\,T_O - Q}{A_3}$$

$$= -\frac{-14\ 100 - 6\ 020}{-4.938 \cdot 10^6}$$

$$= 4{,}1 \cdot 10^{-3}$$

An absolute minimum for S can be obtained by taking $T_O = 0$:

$$S = \frac{-6\ 020}{-4.938 \cdot 10^6} = 1{,}2 \cdot 10^{-3}$$

The above calculation of S is only shown for clarity and is automatically calculated by the program.

We can now proceed and use Equation (4.3) to calculate the recharge during the study period for the Dewetsdorp aquifer via SVF with a time step of $\Delta t = 1$ month. A graphical representation of the results is depicted in Figure 4.54, which is a cumulative plot of rainfall and recharge for the 34-month study period.

A factor to be considered before proceeding, is the time-lag which exists between a rainfall event and the infiltration period of the rain water to reach the ground-water table. (see Figure 4.55). This time-lag depends on a number of factors that cannot usually be measured for each rainfall event (e.g. soil water deficit and rainfall intensity). However, we can use the well-known cross-correlation function (c.f. Davis, 1973) to calculate the lag between the rainfall and the ground-water recharge. Table 4.3 shows that a lag of one month yields the highest cross-correlation coefficient.

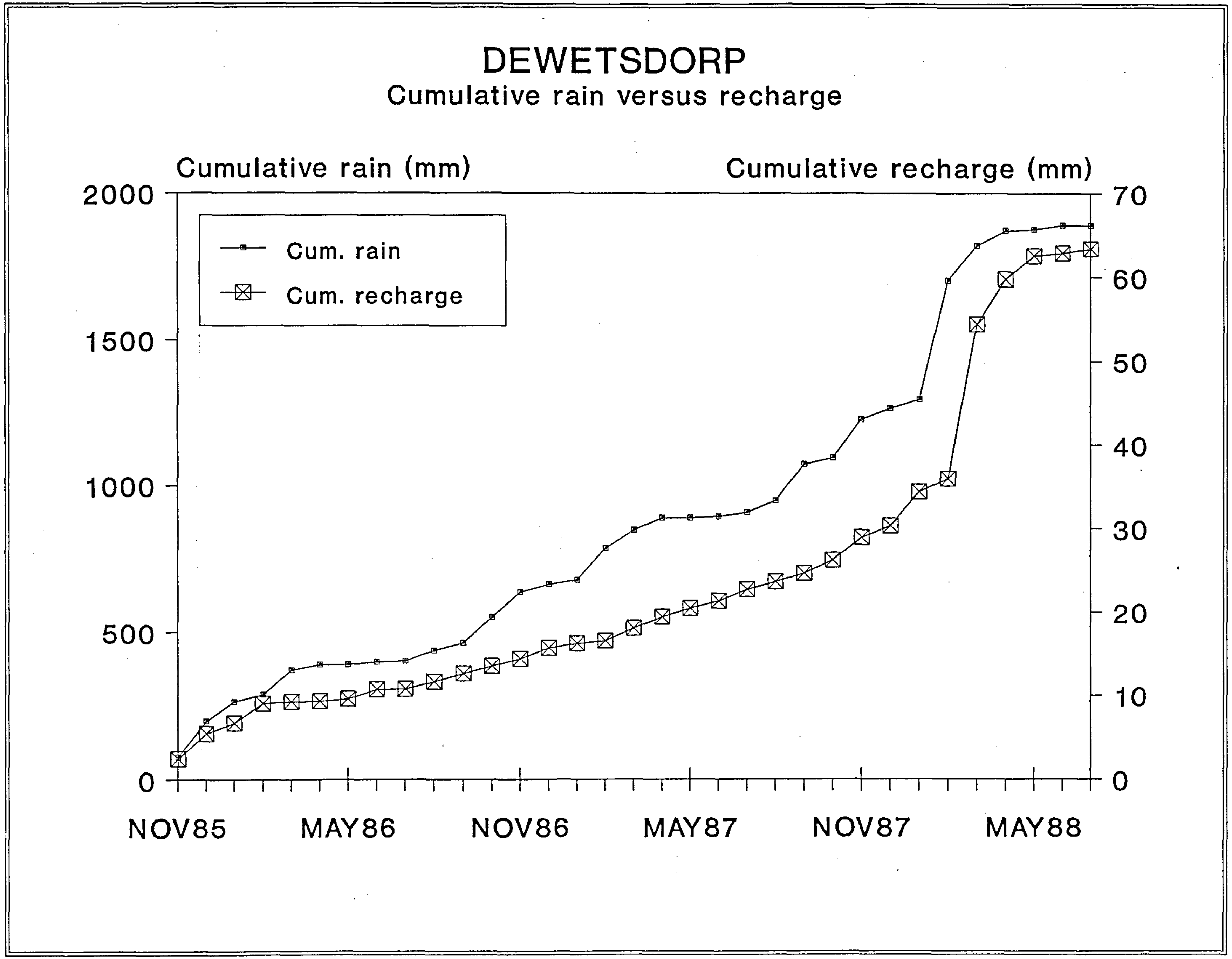


Figure 4.54 Cumulative rainfall and ground-water recharge as obtained with the SVF-method for $\Delta t = 1$ month.

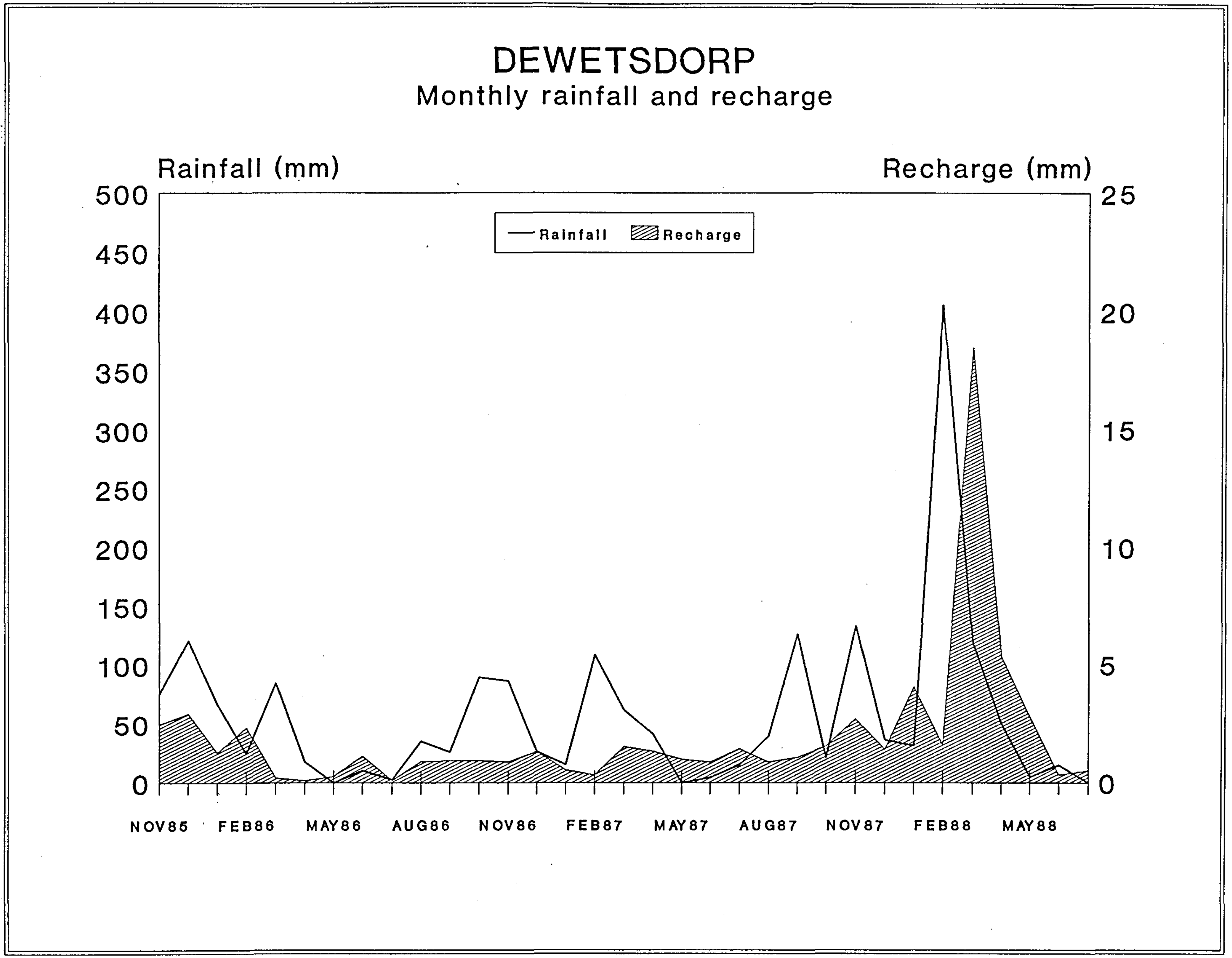


Figure 4.55 Graph showing the lag between rainfall and recharge. The recharge was calculated with the SVF-method with Δt = 1 month.

Lag (months)	Cross-correlation
0	0,12
1	0,85
2	0,20

Table 4.3. Cross-correlation of the lag between rainfall and ground-water recharge.

By lagging the rainfall with one month, the following relationship between monthly rainfall and ground-water recharge was obtained :

$$GR = 0{,}039\ (P - 17), \qquad (r = 0{,}88)$$

which implies that a monthly rainfall of lower than 17 mm would not result in any recharge. This equation is only stated to serve as a guide-line between monthly rainfall and recharge and is very dependable on the soil water deficit of the previous months.

By lagging the rainfall with one month, annual recharge values, shown in Figure 4.56, were obtained (for $\Delta t = 12$ months). The following equations were obtained via a regression analysis:

$$GR = 0{,}046\ (P - 263)\ \text{mm/a}, \qquad (r = 0{,}80) \qquad (4.4)$$

and

$$GR = 0{,}063\ (P - 335)\ \text{mm/a}, (r = 0{,}98) \quad \text{(flood included)}$$

For $P = 587$ mm/a, the $GR = 14{,}9$ mm/a = 2,5% of the annual rainfall (from 4.4).

According to the C.V. which equals 26%, the rainfall will usually be scattered between 434 and 739 mm per annum. When judging Figure 4.54, it must be kept in mind that Equation (4.4) is only valid between the range of rainfall values for which it was attained. Although the value of 14,9 is higher than the 12,8 mm obtained with Equation (4.1), the two values are regarded to be in close resemblance with each another. Two possible reasons for the higher value of 14,9 mm are (a) a too high value of S (in fact a value of $S = 2{,}0 \cdot 10^{-3}$ yields a $GR = 13{,}1$ mm/a) and (b) Equation (4.3) is similar to the finite difference approach, which always includes an error depending on the size of Δt. We ascribed the higher estimate of GR to the last factor. Furthermore, it is important to note that Equation (4.4) is highly dependent on the value of T and thus the lateral outflow rate (calculated as 10 m^2/d and 470 m^3/d respectively), at the outflow boundary, shown in Figure 4.57, whereas values of S, ranging between $1 \cdot 10^{-5}$ to $1 \cdot 10^{-1}$, still yield annual recharge values between 2% and 3,3%. The reason for this is that the total amount of ground water did not show any

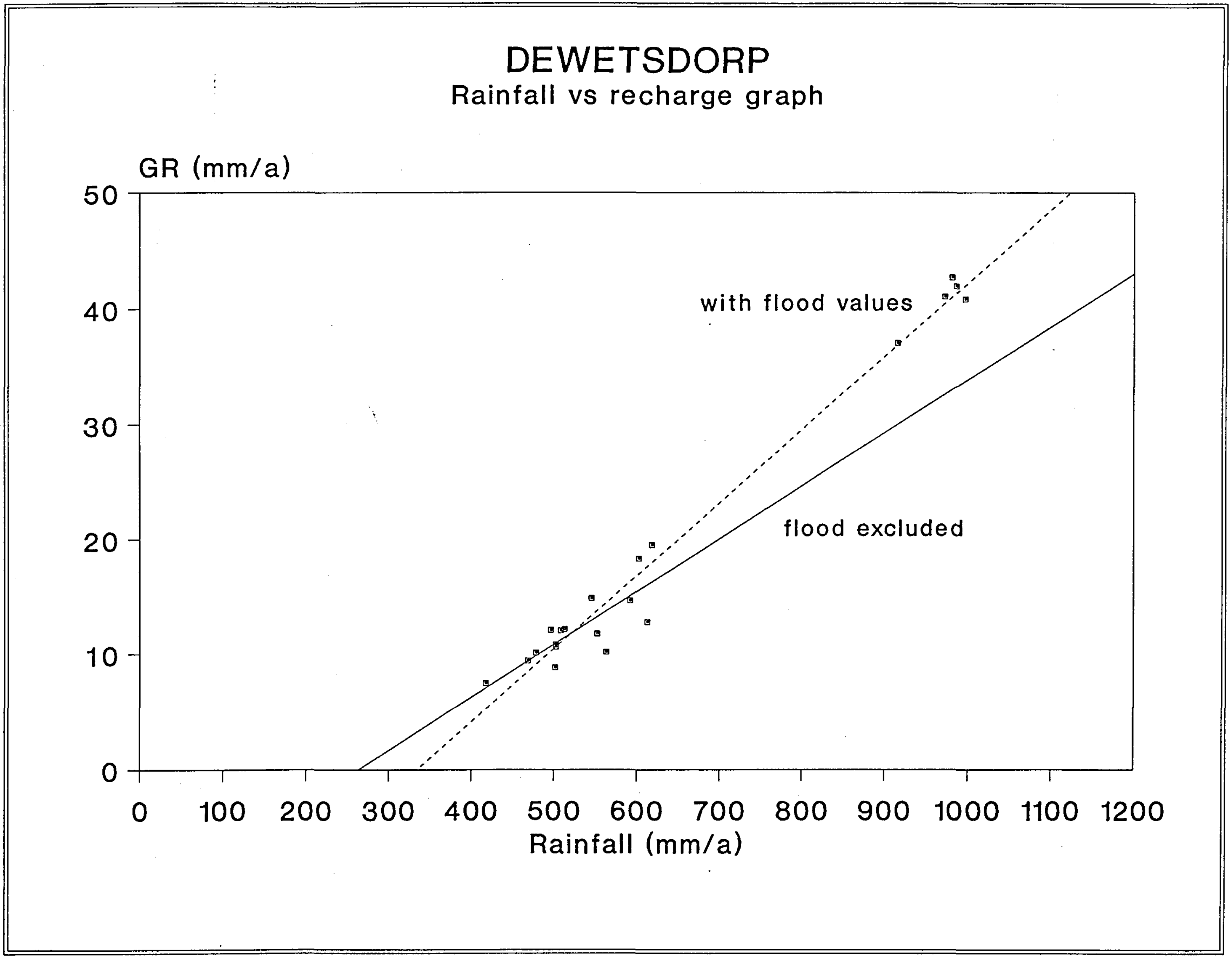


Figure 4.56 The annual ground-water recharge (SVF-method) when plotted against the annual rainfall values.

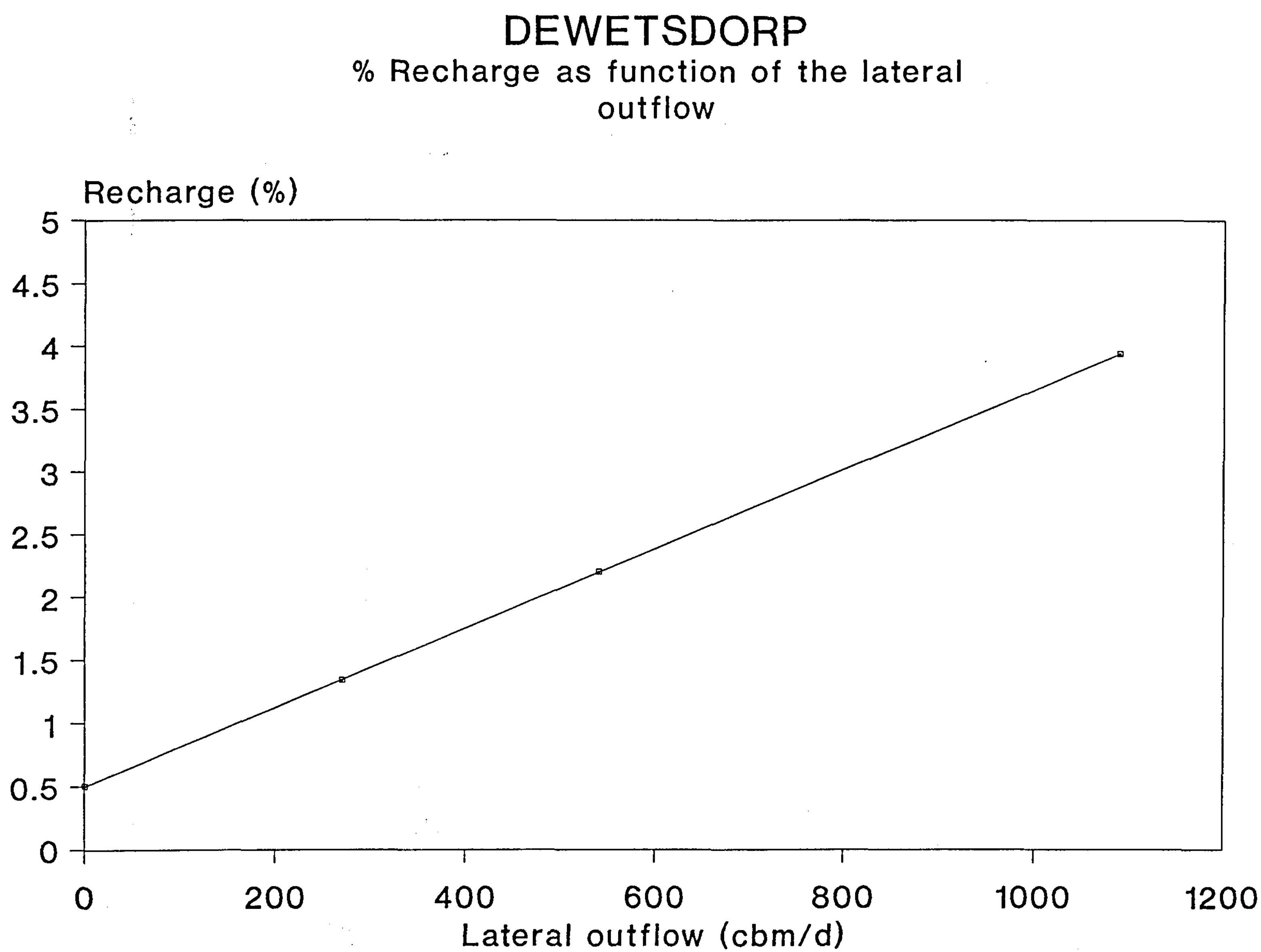


Figure 4.57 In the SFV-method's application the lateral outflow of ground water at the northern boundary was calculated with Darcy's Law. This figure shows the influence a wrongly calculated outflow has on recharge estimates.

drastic decrease or increase during this period. If this was the case, the S-value would definitely have a noticeable influence on the estimated recharge values.

Figure 4.58 shows similar calculations with SVF for different values of T and S. At this stage, we regard a $S = 0{,}004$ and $T = 10\ m^2/d$ to be quite realistic for Dewetsdorp.

The percentage annual recharge-rainfall relationship can be visualized from Figure 4.59 and has the form:

$$GR(\%) = 0{,}0043\ P, \qquad (r = 0{,}95).$$

Figure 4.60 shows the percentage annual recharge for each month, starting from October, 1986, as well as the percentage mean recharge. The percentage mean recharge for a month was obtained by dividing the total cumulative infiltration by the total rainfall up to that month. In doing so, the influence of the time delay between GR and P was reduced. It is evident from Figure 4.60 that the recharge amounts to 2,2% of the rainfall, except for the flood period when between 4% and 5% of the rainfall recharged the aquifer.

Similar calculations as above, with $\Delta t = 1$ month, yielded the results shown in Figure 4.61. The ground-water recharge varies between 0% and 7% (mean = 2,7%) of the monthly rainfall, except for two values of 32% during the mid-winter of 1986 and 1987 and the two values of approximately 12%, which occurred during the late summer of 1987. The reason for the higher recharge values during the winter is that the evapotranspiration is less during this time of the year. Figure 4.16 shows the potential evapotranspiration for the study period and from this figure, it can be seen that the rainfall which occurs during the winter months has a higher percentage recharge potential ability, because the potential evapotranspiration is much less.

From the cross-correlation calculations, it can be concluded that specific ground-water recharge at a specific time is highly dependent on the rainfall that occurs during the previous three months. Calculations using a $\Delta t = 3$ months, strengthen the belief that more rainfall, which occurs during the winter months, reaches the water table, relative to the rainfall during the rest of the year (c.f. Figure 4.62). The mean value for a three-month period was once again in the order of 2,2% of the rainfall.

To conclude Case 2, it was estimated that the mean recharge for the 34 months was 2,6%.

Case 3

Dziembowski (1977) expressed the opinion that the recharge is possibly much higher in the hilly areas, surrounding the township in the south and south-west. The aim is to

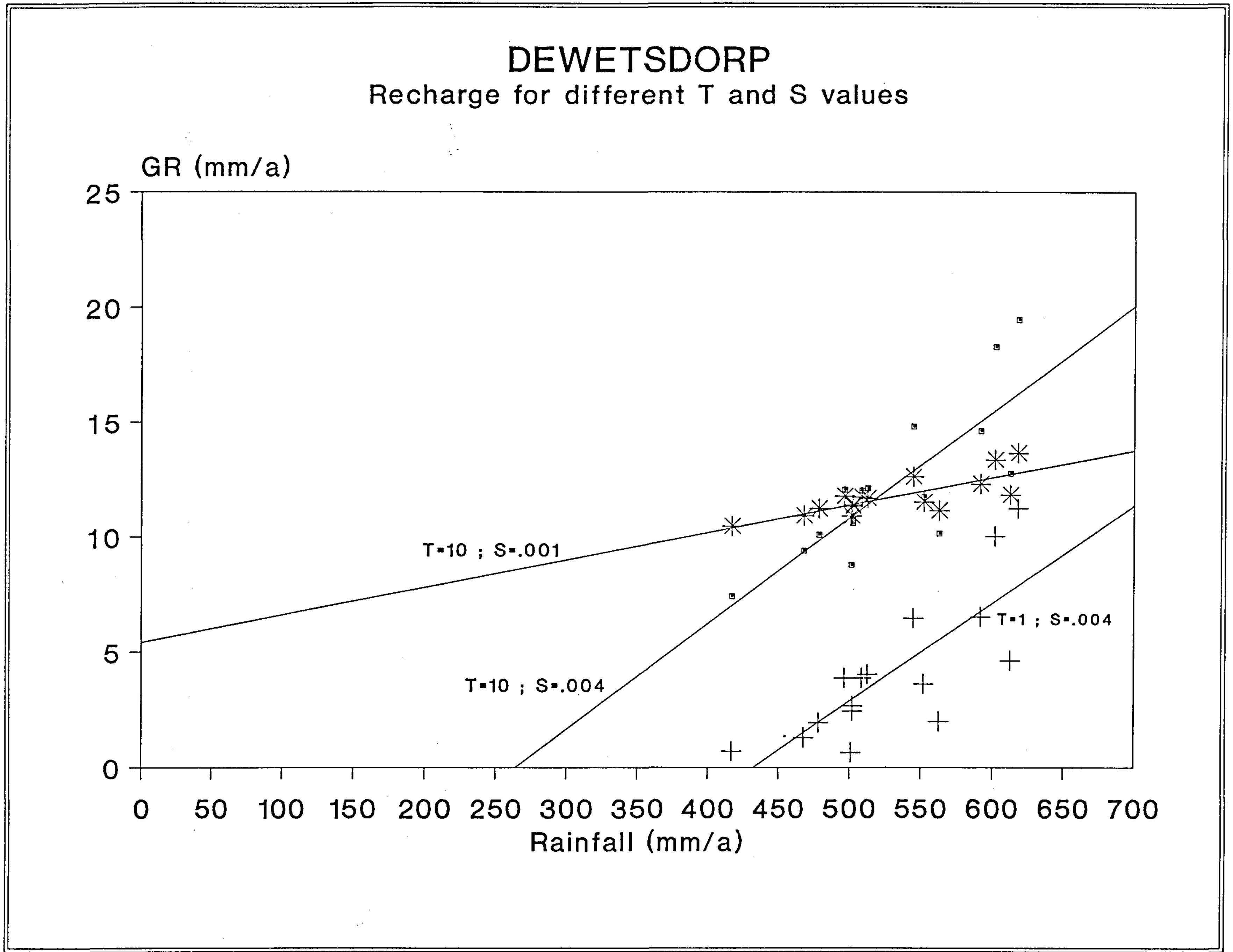


Figure 4.58 Sensitivity analysis of the SVF-method for T- and S-values in Dewetsdorp.

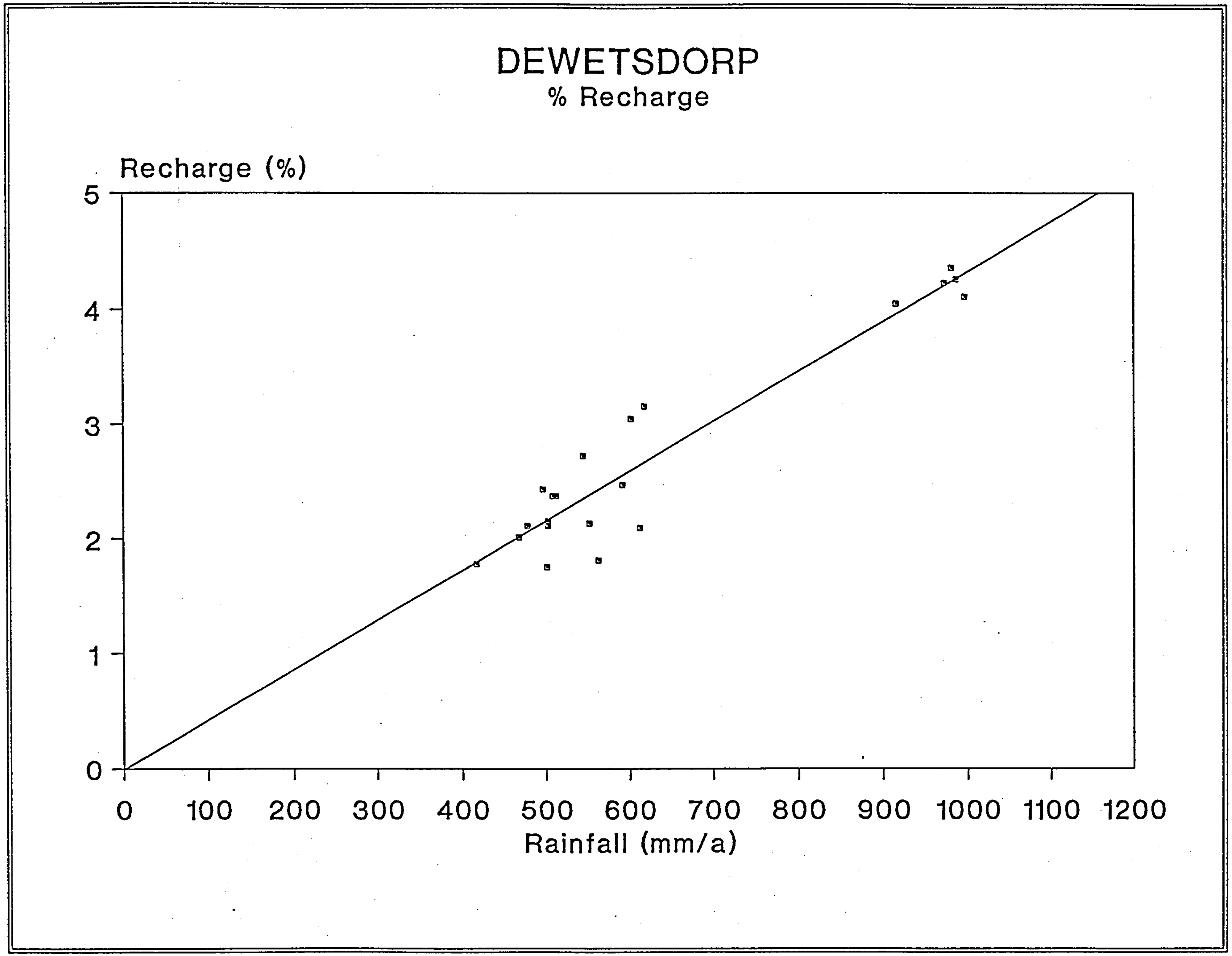


Figure 4.59 The annual percentage recharge of the Dewetsdorp aquifer as obtained with the SVF-method.

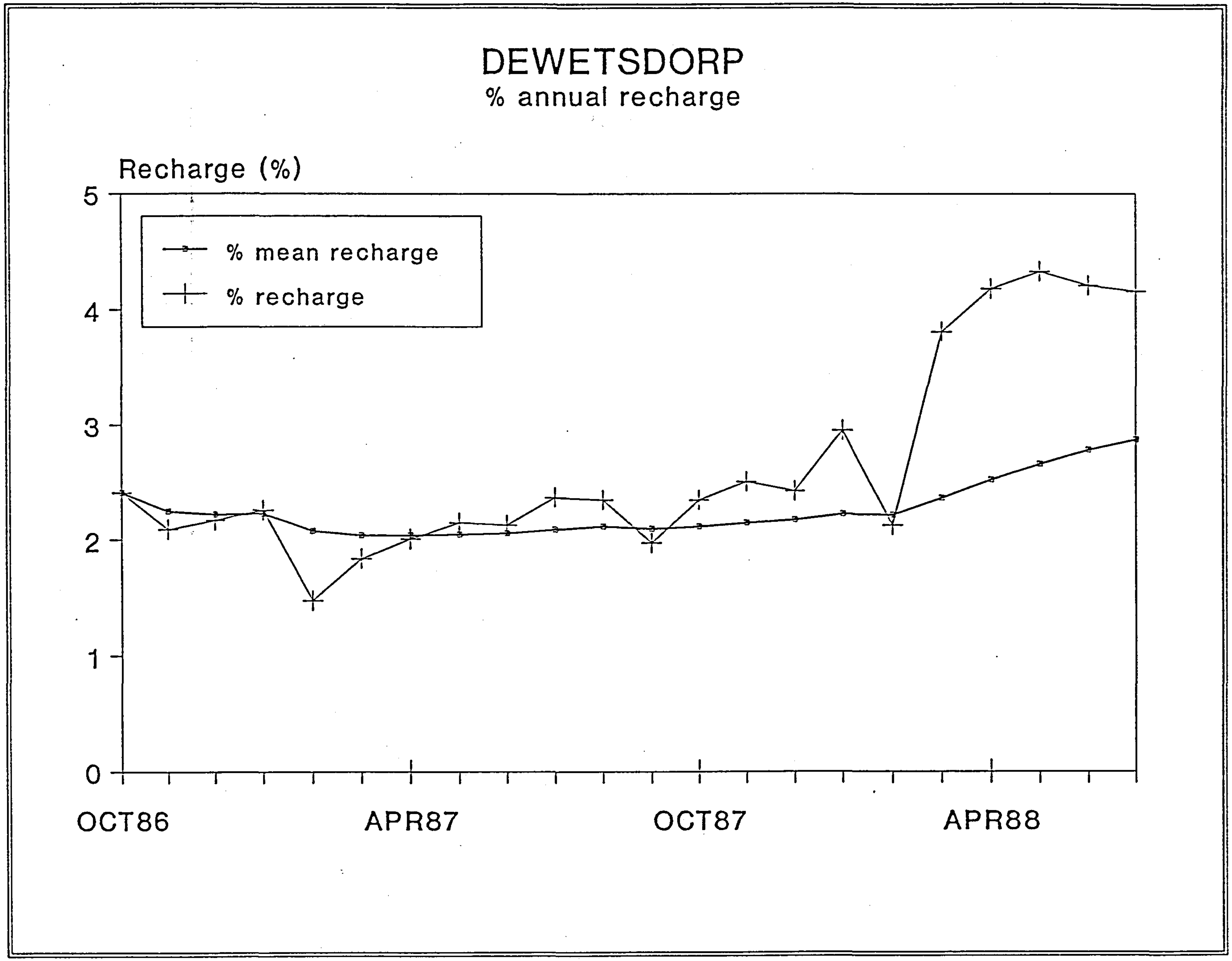


Figure 4.60. The percentage annual and mean recharge during the study period for the Dewetsdorp aquifer as estimated with the SVF-method.

Figure 4.60 Percentage of the 12-months' mean, and of the total recharge since commencement of the project.

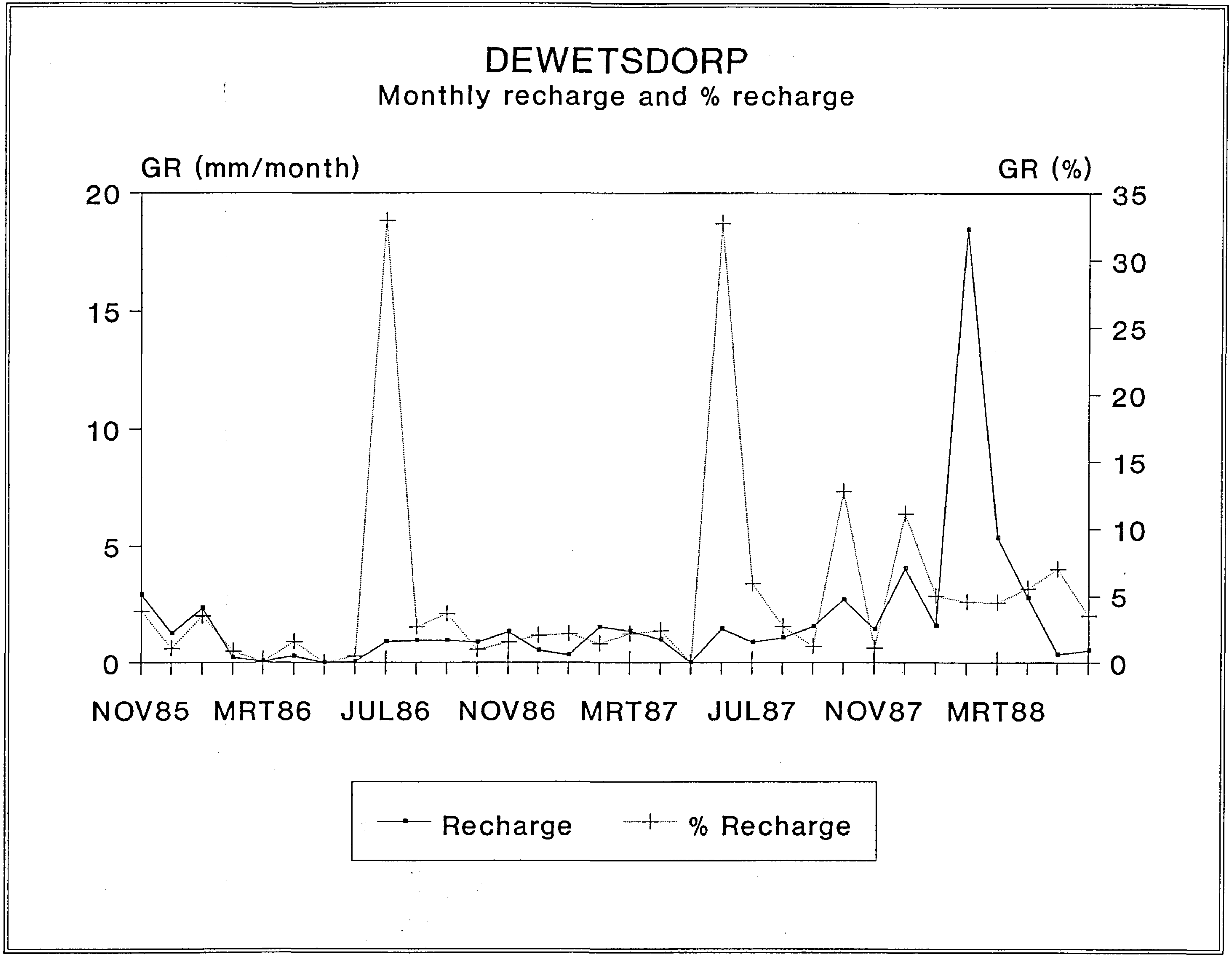


Figure 4.61 Monthly recharge values for Dewetsdorp as estimated with the SVF-method.

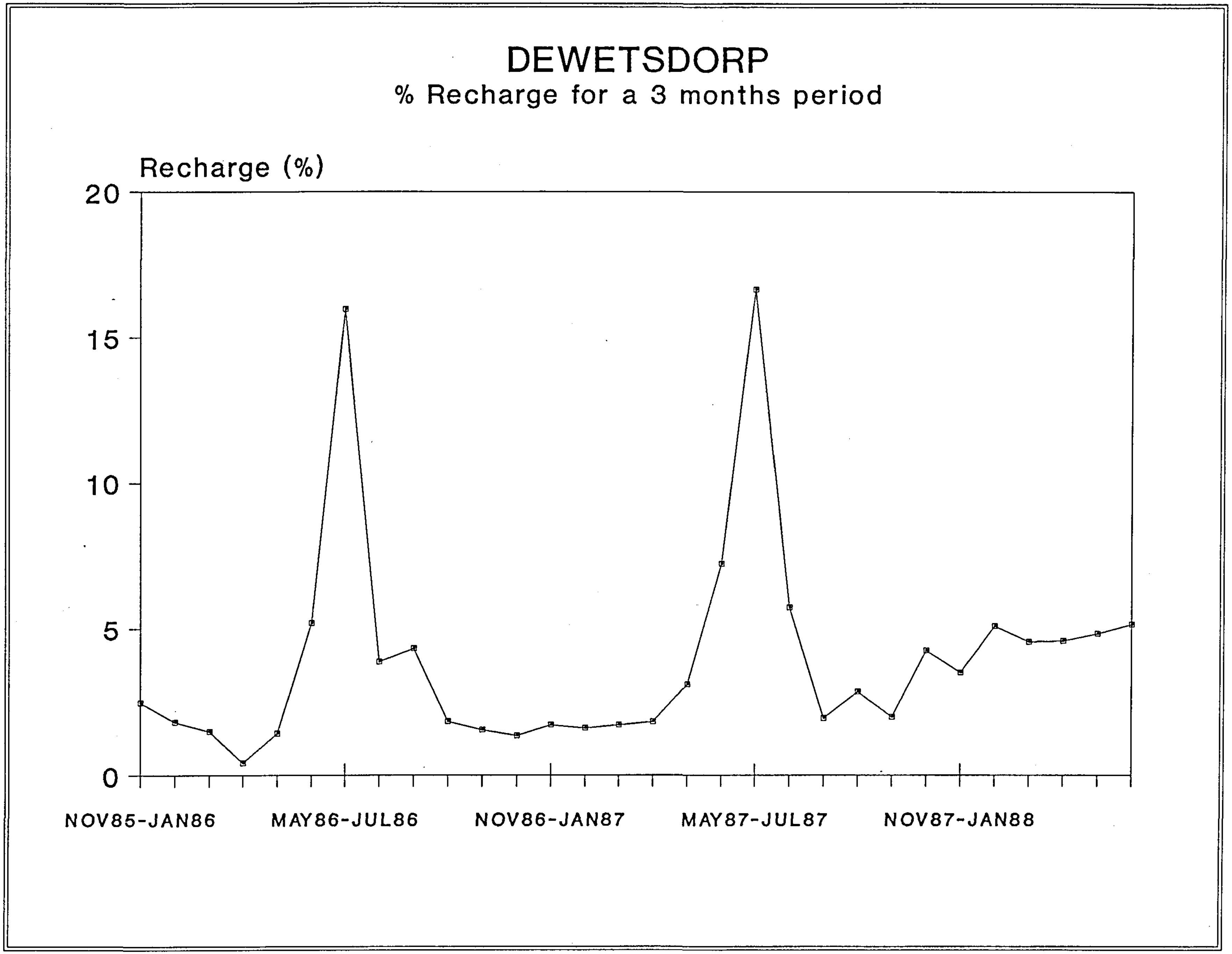


Figure 4.62 Percentage ground-water recharge (SVF-method) for a three months period. It is evident that the recharge during the winter months is higher, possibly because of the fact that evaporation is lower during this time of the year.

try to quantify and differentiate between the southern and northern parts of the aquifer in terms of recharge. It was therefore decided to divide the aquifer into a northern and southern part (Figure 4.50).

A T-value of 10 m^2/d was used for the northern outflow boundary. For the southern part, an outflow T of 3 m^2/d was used. This value was obtained from the two-dimensional finite element model, which will be discussed in the next section. This value of T accounts for the steeper water-table gradients in the vicinity of the northern boundary in the southern part. The steeper gradients, however, may be due to other reasons such as evapotranspiration. Whatever the case may be, the results will be the same. The cumulative infiltration after 34 months was 40 mm for the northern part and 102 mm for the southern part, which is equivalent to a recharge of 2,1% and 5,4% of the rainfall (see Figure 4.63) respectively. By excluding the flood values, recharge values of 22 mm (= 1,7%) and 65 mm (= 5%) were obtained.

If, however, a T-value of 5 m^2/d, instead of 3 m^2/d, was used for the southern part, recharge estimates of 7,6% (with flood) and 7,3% were obtained.

The relationship between annual rainfall and recharge is depicted in Figure 4.64. The following relationships exist:

Southern part:	$GR = 0{,}058 \ (P - 132)$,	$(r = 0{,}81)$
Northern part:	$GR = 0{,}035 \ (P - 324)$,	$(r = 0{,}74)$

where GR and P have dimensions mm/a.

Figures 4.65 and 4.66 show estimates of annual recharge percentages from October, 1986, for the northern and southern parts.

From these figures, it can be concluded that the ground-water recharge is approximately three times higher in the hilly areas than in the areas with a deeper soil.

4.1.6.1.1. Sensitivity analysis

In the previous cases, 72 boreholes were used for the SVF-method. The standard deviation (std) of the ground-water levels in the Dewetsdorp aquifer is 33 m. To determine the sensitivity of the SVF-method to the number of boreholes used in the analysis, firstly 15 (std = 38,7) and then 6 (std = 41,8) boreholes were randomly selected in the Dewetsdorp area (see Figure 4.67). Figure 4.68 shows the calculated saturated volume. The saturated volume, as calculated with the different number of boreholes, shows a resemblance as is evident from this figure. Periods where Vi = Vj

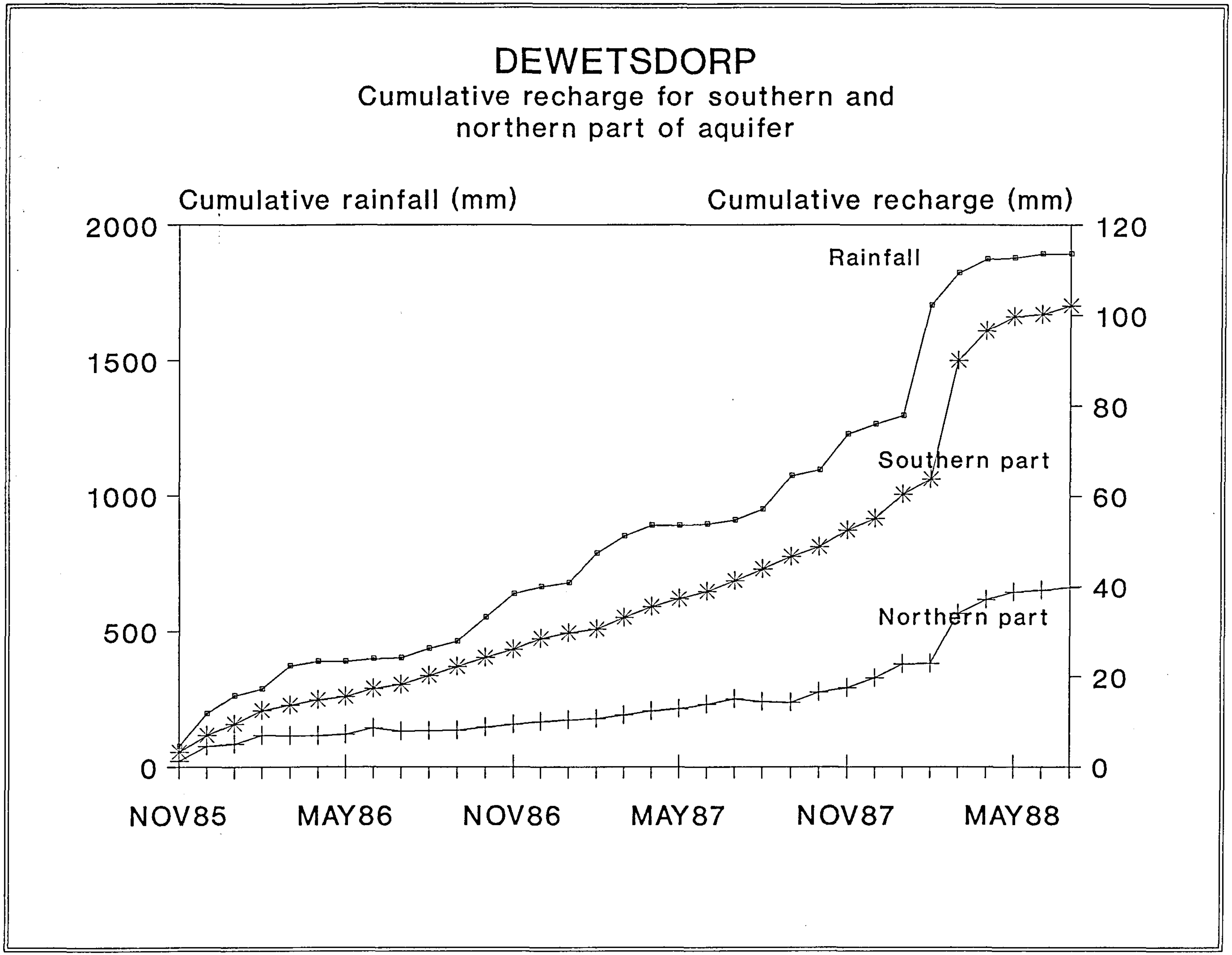


Figure 4.63 Cumulative ground-water recharge (SVF-method) for the southern- and northern parts of the Dewetsdorp aquifer.

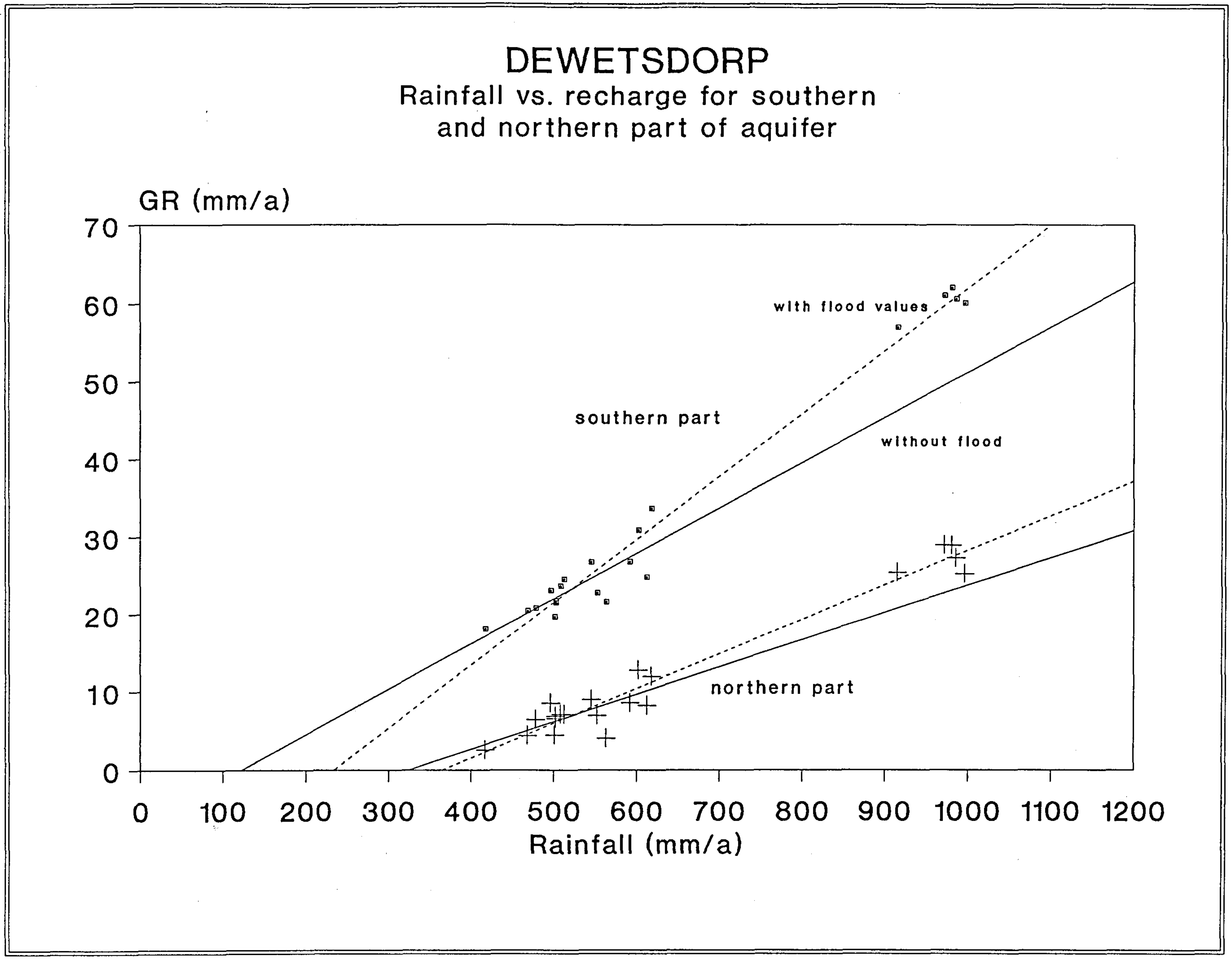


Figure 4.64 Graph showing that the annual recharge in the southern part is higher than in the northern part; the reason being the occurrence of outcrops in the southern part.

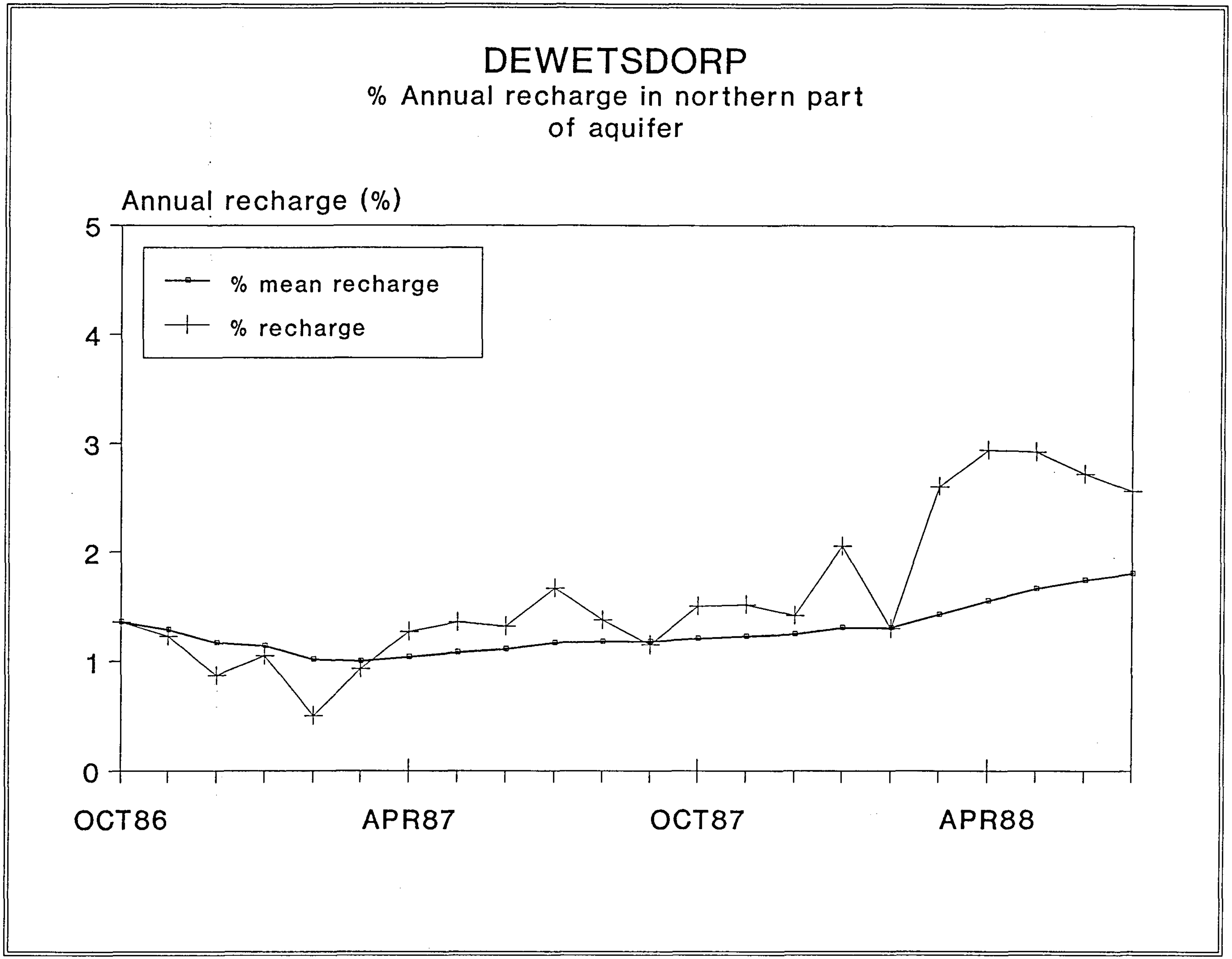


Figure 4.65 Percentage annual recharge in the southern part of the Dewetsdorp aquifer (SVF-method).

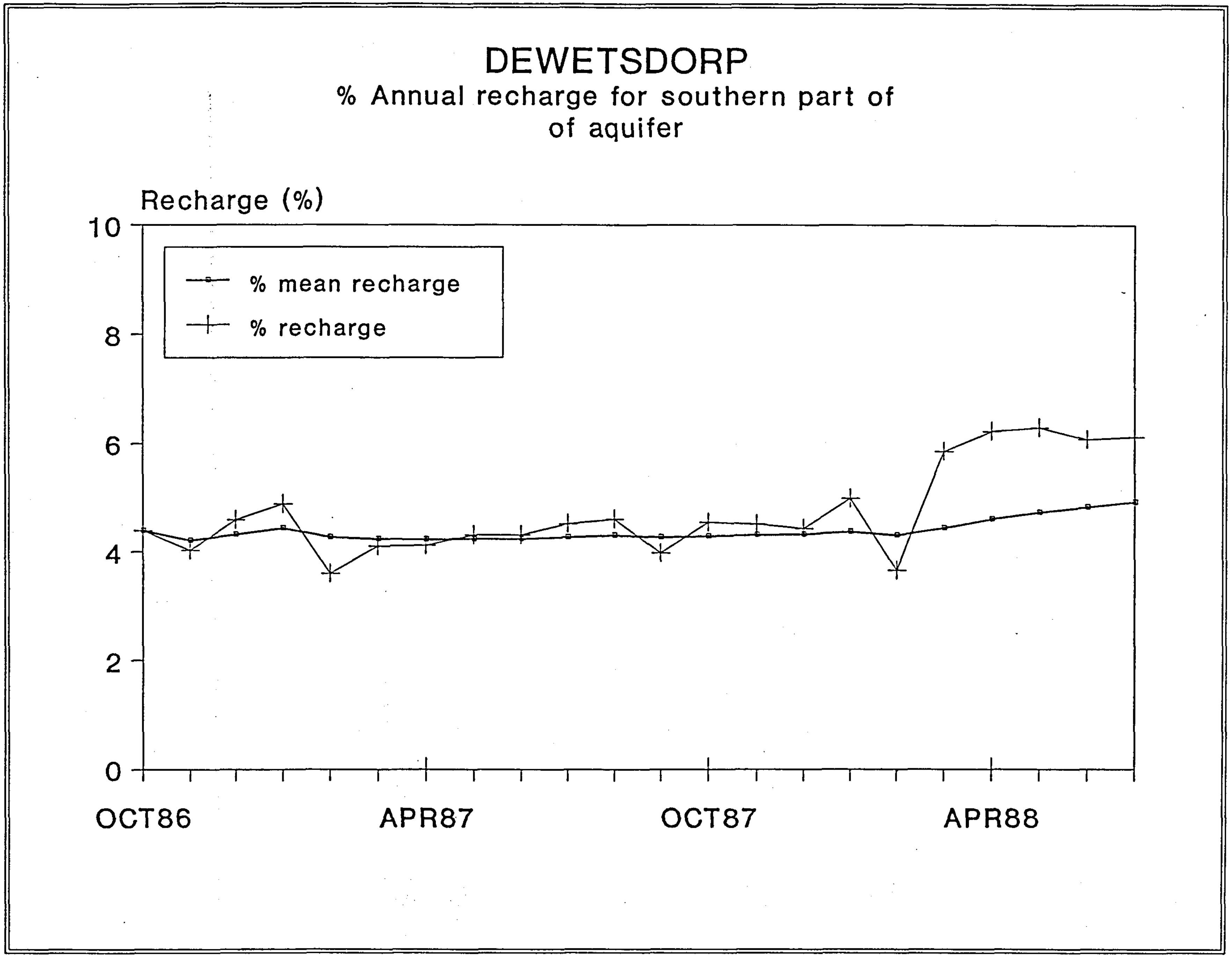


Figure 4.66 Percentage annual recharge in the northern part of the Dewetsdorp aquifer (SVF-method).

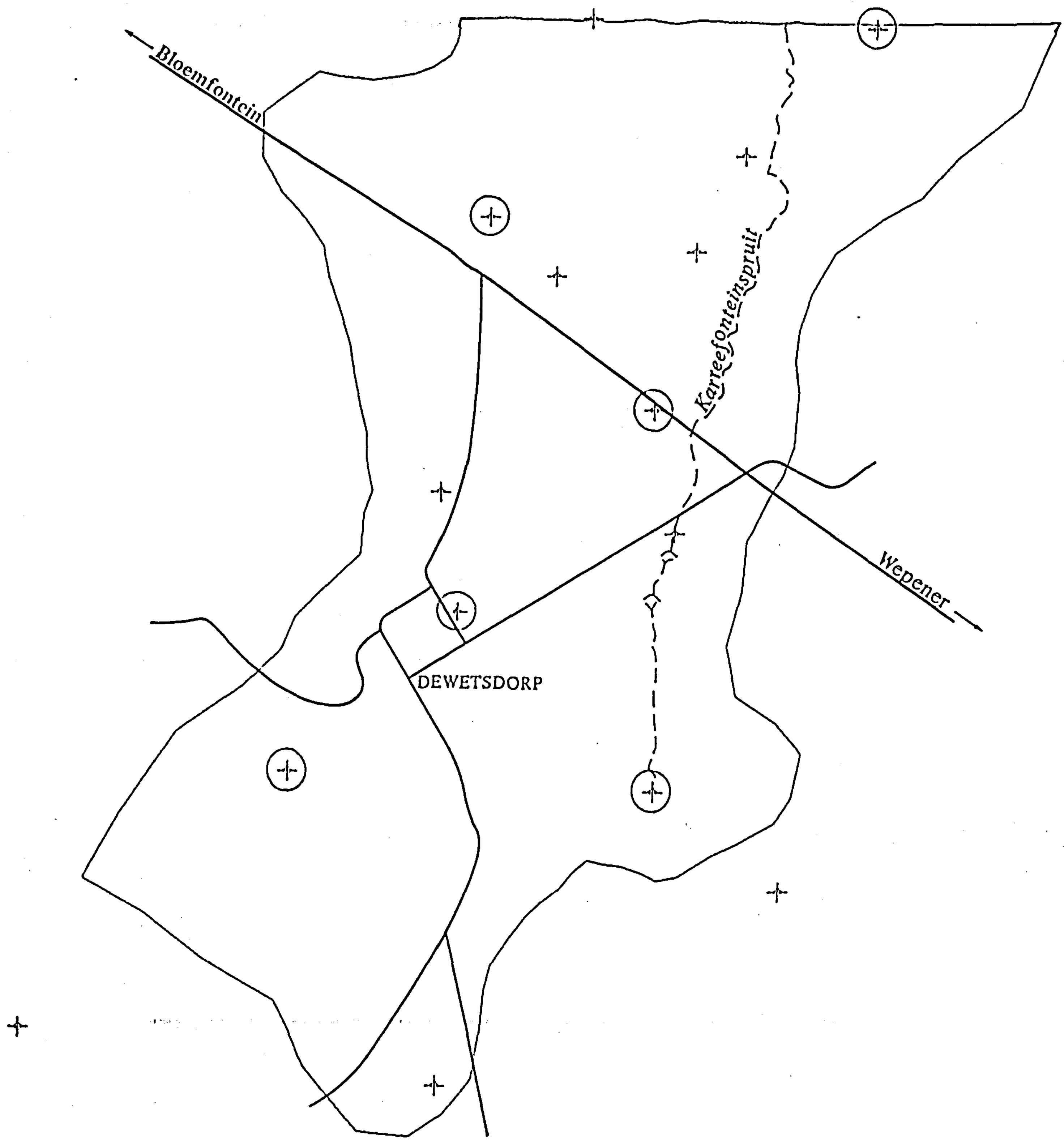


Figure 4.67 Positions of the selected 15 (+) and 6 (O) boreholes used for the sensitivity analysis on the SVF-method for the Dewetsdorp aquifer.

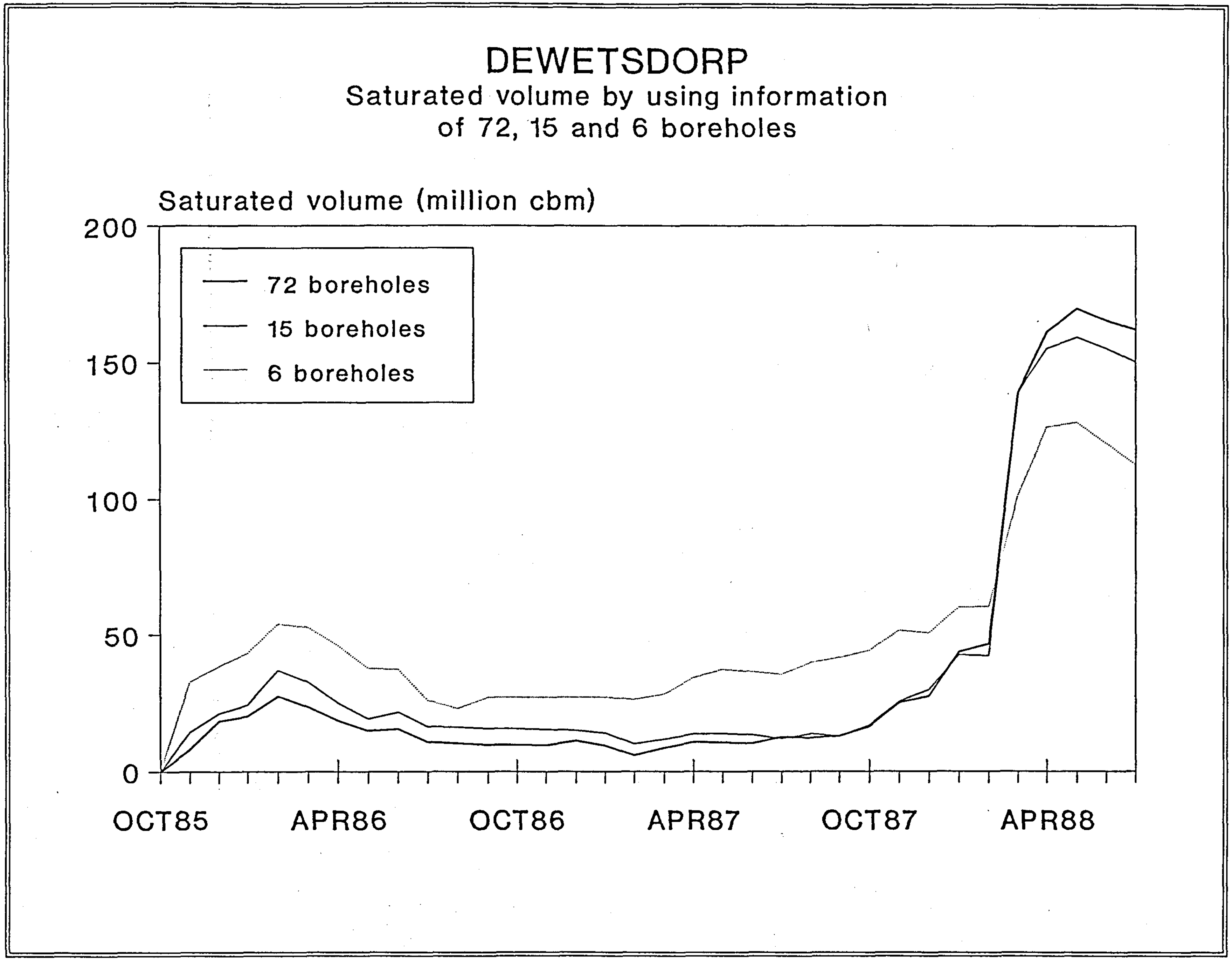


Figure 4.68 Saturated volume of the Dewetsdorp aquifer using information from 72, 15 and 6 boreholes respectively.

were selected for both the 15- and 6-borehole calculations and the ground-water recharge was calculated. Figure 4.69 shows the relationship between the recharge and rainfall for the three cases. If the results obtained with the 72 boreholes are supposed to be 100% correct, then the results of the 15 boreholes are 68% and those of the 6 boreholes 67% correct.

The annual ground-water recharge-rainfall relationship, for $\Delta t = 12$ months, is given below (flood excluded) :

72 boreholes :	GR = 0,046(P-263) mm/a,	(r = 0,80)
15 boreholes :	GR = 0,046(P-371) mm/a,	(r = 0,83)
6 boreholes :	GR = 0,049(P-380) mm/a,	(r = 0,74)

It is thus clear that the SVF-method loses accuracy with a decrease in the number of boreholes used. It is, however, felt that the method may still yield acceptable results as long as the standard deviation of the water levels is not " too high". At present, it is very difficult to quantify the term "too high". All that can be said, is that care must be exercised when using the SVF-method for an aquifer where boreholes are irregularly spaced and ground-water levels have a relatively large standard deviation. Because ground-water levels tend to follow the topography, special care must also be taken in areas where the topography varies considerably.

A major difficulty encountered with all indirect methods (i.e. methods using water-level changes for recharge calculations), is that the S-value usually changes with depth, so that the use of a constant S-value may be in question. This is an unresolved problem for the present study (because of a lack of data on ground-water levels in the upper part of the aquifer) which definitely needs attention in the near future, as well as the calculation of S-values from pumping tests. To obviate this difficulty, the recession of the ground-water levels after a flood can be used to calculate a S for the upper part of the aquifer.

To conclude the SVF-method, Figure 4.70 was obtained with $\Delta t = 1, 3, 6$ and 12 months, yielding the following regression equation :

$$GR = 0{,}039\,(P - 17), \qquad (r = 0{,}88) \qquad (4.5)$$

Equation (4.5) may be used as a first approach for recharge estimates in the Dewetsdorp aquifer. Furthermore, the relation is independent of the rainfall period, i.e. rainfall may be measured in weeks, months or years. For management purposes, the geohydrologist is interested in periods during which the aquifer is subject to a depletion of the ground-water resource. Once every ten years, the Dewetsdorp aquifer

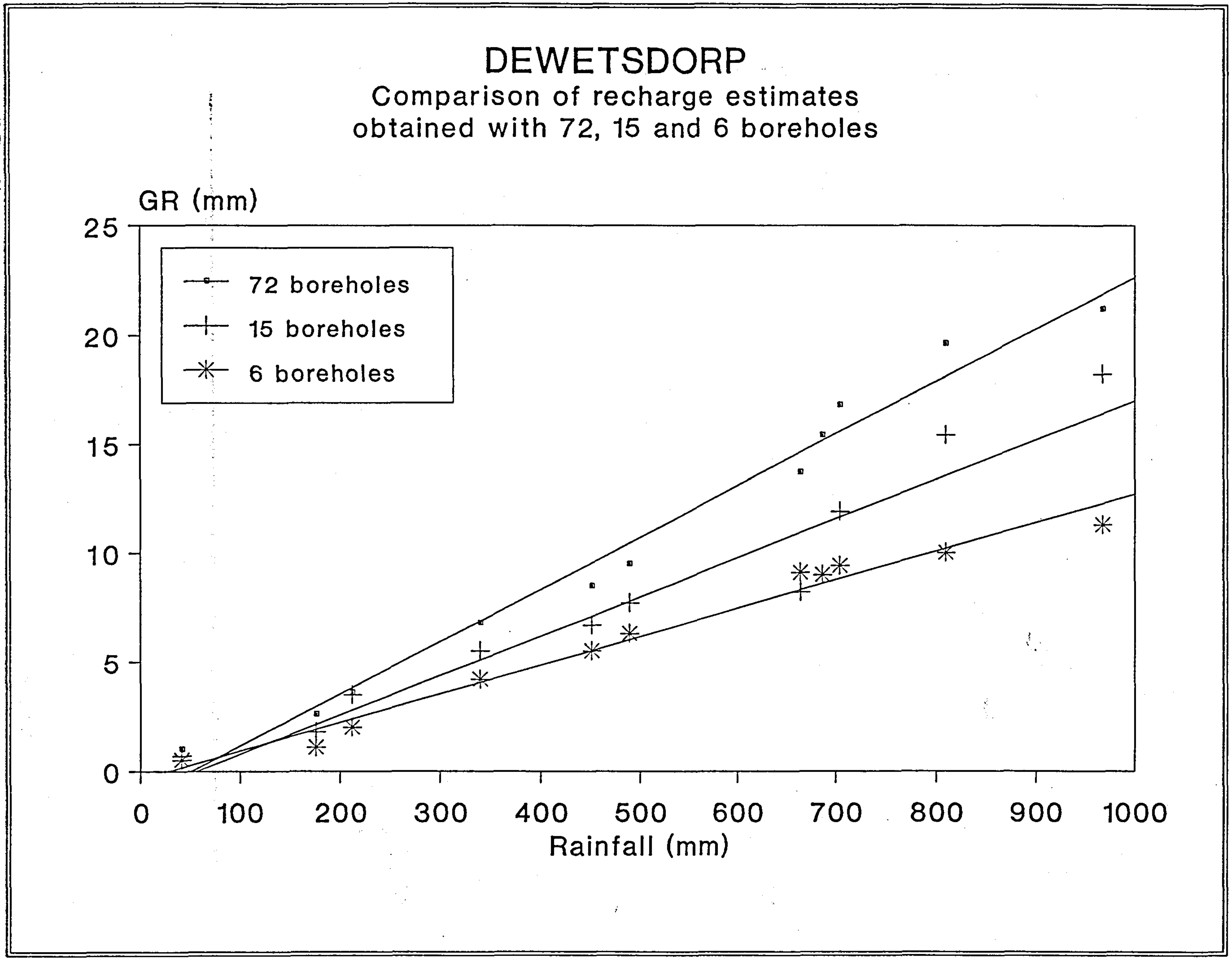


Figure 4.69 Comparison of recharge estimates obtained, using data from 72, 15 and 6 boreholes respectively.

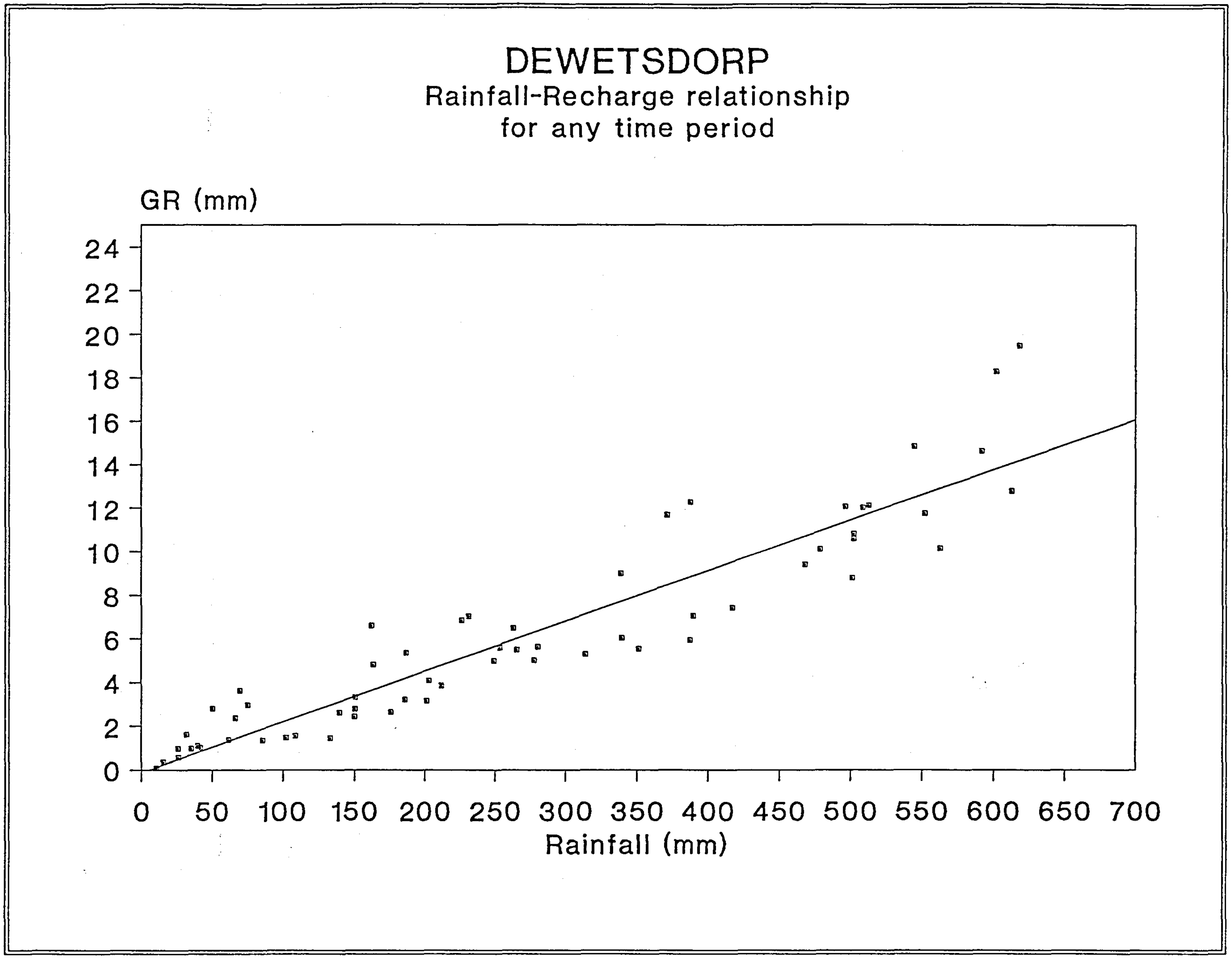


Figure 4.70 Rainfall-recharge relationship obtained with the SVF-method by combining the results of different Δt-values.

can be expected to have a recharge less than 7 mm (Figure 4.71). Although this recharge value may not cause any problems regarding the yielding of boreholes, a successive period of, say three years of a severe drought, may cause serious problems. However, it is not possible to predict when such a situation may arise.

Table 4.4 summarizes the discussed results obtained via program SVF for the Dewetsdorp aquifer.

	Equation	Recharge (as % of rainfall)
Whole aquifer	1. GR = 0,024 (P - 51) [mm] 2. GR = 0,046 (P - 263) [mm/a] 3. GR = 0,023 (P - 9,5) [mm]	2,2% 2,5% 2,2%
Northern aquifer (deeper soils)	GR = 0,035 (P - 324) [mm/a]	1,5 - 2,1%
Southern aquifer (hilly area)	GR = 0,058 (P - 132) [mm/a]	4,5 - 7,6%

Table 4.4. Recharge estimates for the Dewetsdorp aquifer as obtained with program SVF.

4.1.6.2 THE GROUND-WATER LEVEL FLUCTUATION METHOD

If the S-value of a borehole is known, ground-water level fluctuations can easily be transformed into ground-water recharge via Equation (3.4.3). A ground-water level of a borehole which remains constant, implies that recharge takes place, because ground water is always in some state of recession. For the use of the GLF-method, it is preferable that the no-recharge recession curve is known, otherwise an underestimation of the recharge is obtained. By incorporating the no-recharge recession, the difference between the actual water level and the level to which it would have declined, had no recharge taken place, is assumed to represent the recharge.

Figure 4.72 shows an example of how the ground-water recharge is calculated with the GLF-method, for borehole G36431, for the February 1988 flood. For all GLF-calculations, a S-value of 0,004 was assumed. Program GLF was used with $\Delta t = 12$ months to calculate the relationship between the annual rainfall and recharge (see Figure 4.73) for the whole aquifer:

$$GR = 0,035\ (P - 181), \qquad (r = 0,82)$$

where all values are in mm/a. An annual rainfall of 587 mm yields a GR = 14,2 mm or 2,4% of the annual rainfall. This value compares well with the annual recharge estimation of 2,5% which was obtained with the SVF-method in the previous section

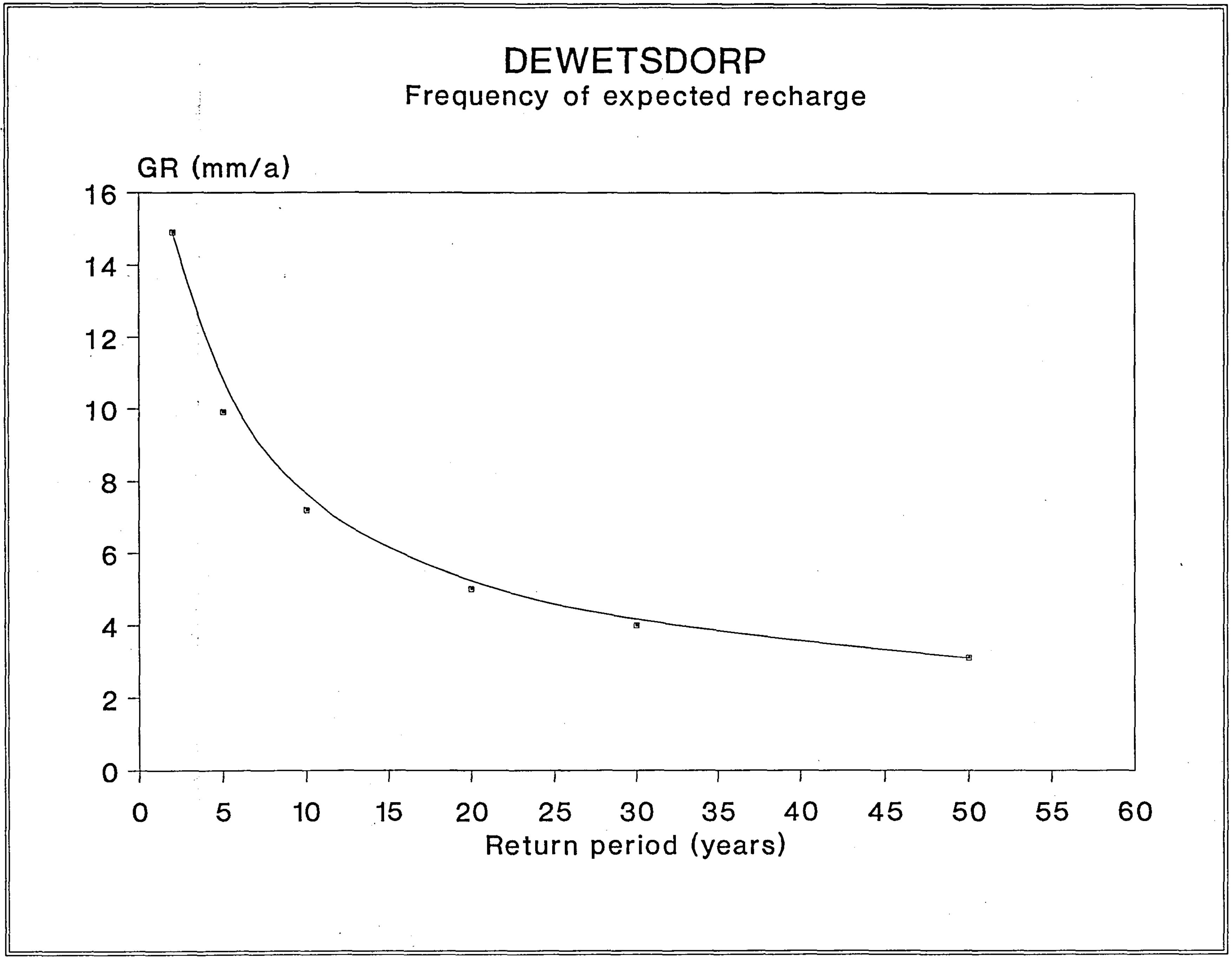


Figure 4.71 Frequency of expected annual ground-water recharge of the Dewetsdorp aquifer (SVF-method).

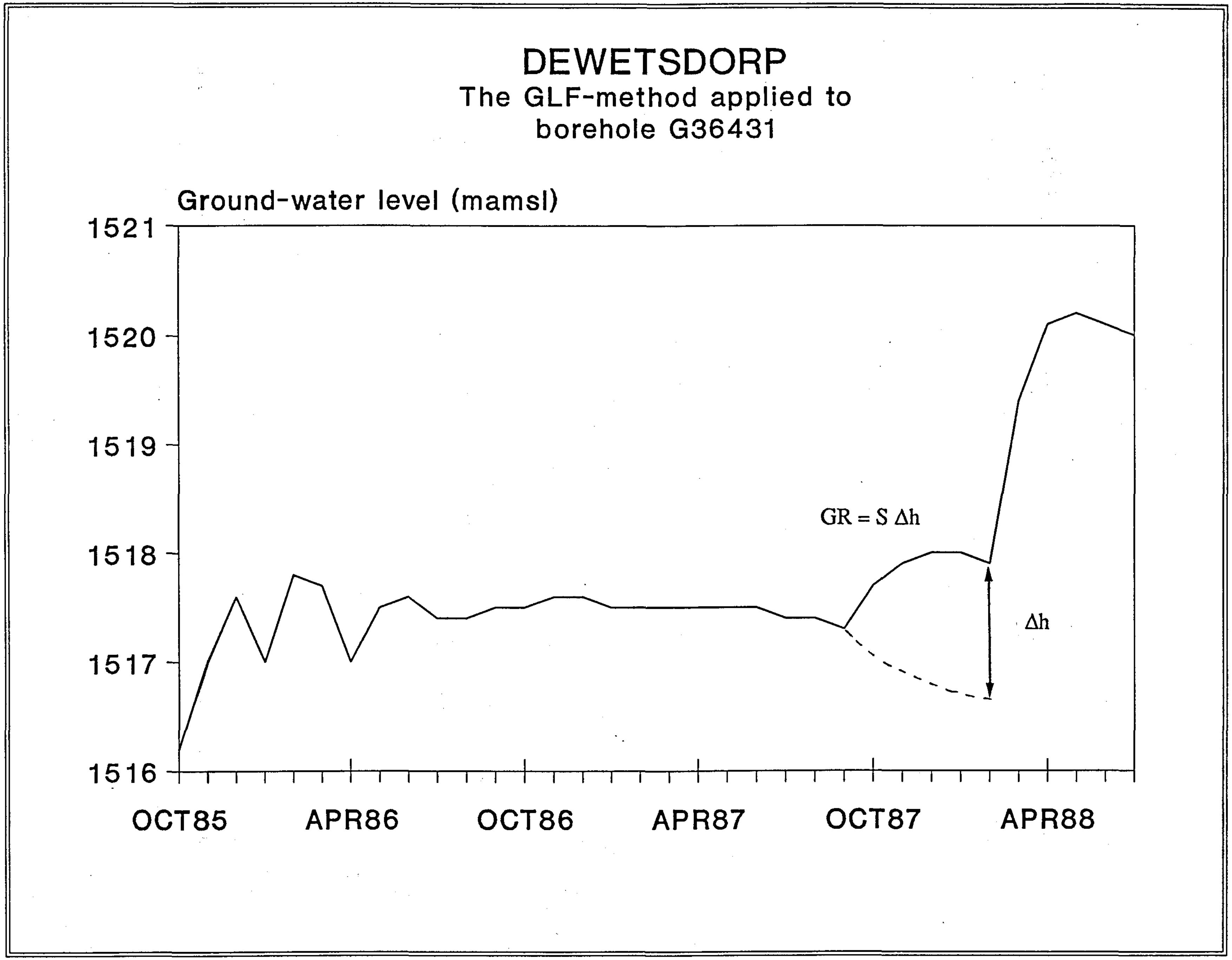


Figure 4.72 The point GLF-method applied to borehole G 36431.

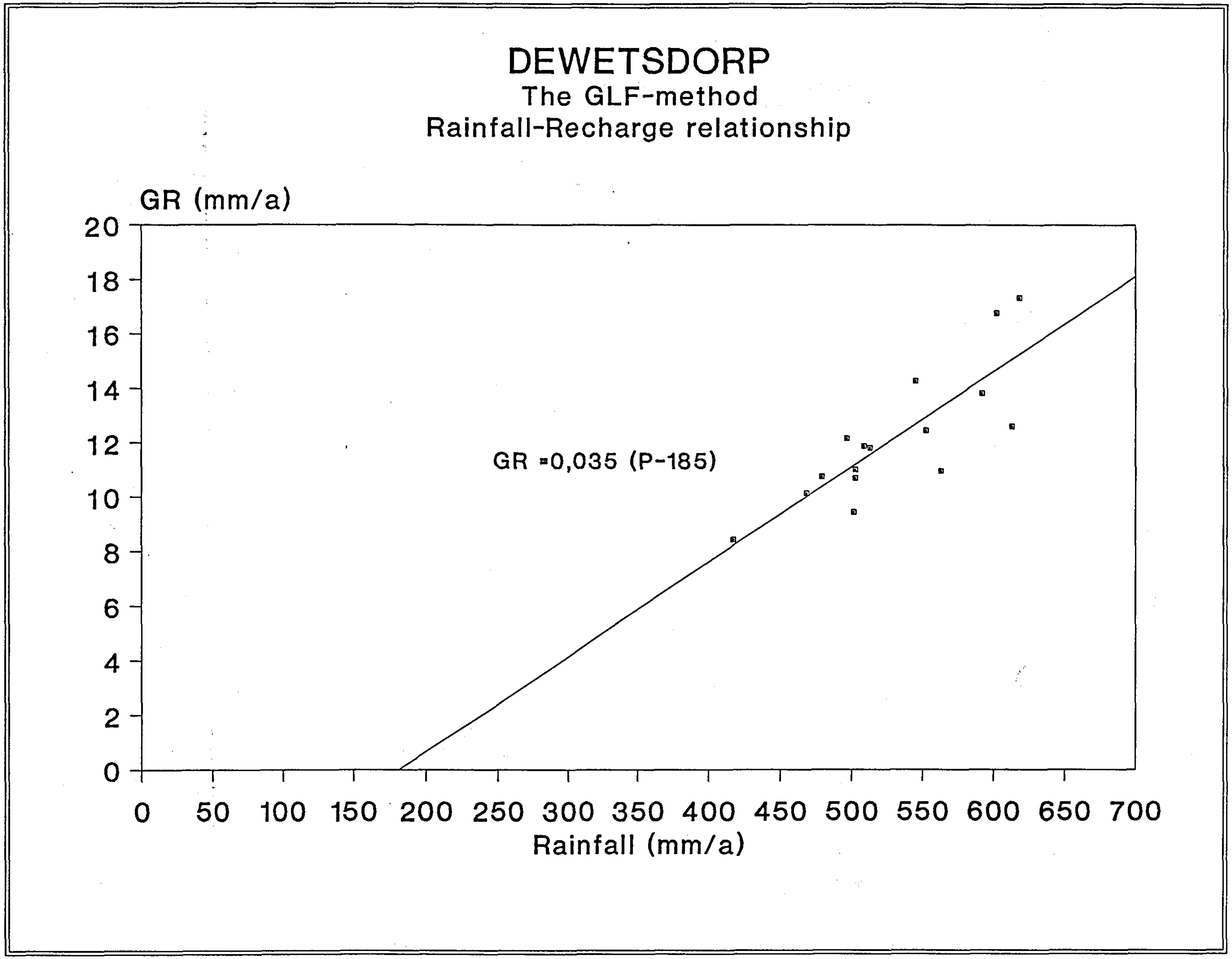


Figure 4.73 The annual ground-water recharge for the Dewetsdorp aquifer as estimated with the GLF-method.

(Equation (4.4)). The reason for this can be seen in Figure 4.74, which shows graphs of the saturated aquifer volume of Dewetsdorp for the 34 months, as calculated with both the SVF- and GLF-methods. Both graphs show an excellent correlation with each another. The GLF-method will thus yield the same results as the SVF-method for Case 1 (Equation (4.2)). The reader may ask why actually perform the SVF-method, which is much more time-consuming, if the same results can be obtained with the GLF-method. It must, however, be stressed that this will not always be the case, because it is only true for the situation where the average ground-water level times the area of the aquifer is a good approximation of the actual saturated volume of the aquifer. The SVF-method's estimate of the saturated volume will always be nearer to the exact value as the value calculated by the GLF-method.

Figure 4.75 shows contours of the percentage ground-water recharge during the flood based interpolated water-level changes. The contours of the mean recharge estimates in mm for the first 29 months of the study period previous to the flood, as obtained by the GLF-method, are shown in Figure 4.76. This figure strengthens the belief that the recharge is between two and four times higher in the hilly areas than in the northern parts, which have a deeper soil cover.

In an attempt to correlate the precipitation and ground-water levels on Long Island, Jacob (1943) demonstrated that the recession of water levels can be approximated by:

$$h = h_0 \exp(-\pi^2 Tt / 4r^2S)$$

where h_0 is the original water level, h the height after time t, and r the half-width of the aquifer, between the boundaries. This equation may be applied to any hydrological recession, where the recession represents the decline of storage in the absence of replenishment. For Dewetsdorp, the maximum mean recession during 30 days was 0,26 m. For r = 4 000 m (half the width of the aquifer in the flow direction) and an aquifer thickness of 30 m, it was calculated that the ratio T/S = 2791. Thus, for S = 0,004, T = 11 m^2/d, which is of the same order as previous calculations.

4.1.6.3 TWO-DIMENSIONAL FINITE ELEMENT MODEL (INVERSE MODELLING)

The finite element program AQUAMOD (Van Tonder and Cogho, 1987) was used for this study. This program allows for varying recharge inputs in different regions of the aquifer. A triangular finite element mesh (Figure 4.77) was constructed. The boundaries for the Dewetsdorp aquifer can be treated as ground-water divides (no-flow boundaries), except in the north, where a specified flux boundary exists. No difficulties were therefore encountered with the incorporation of these boundaries in the model.

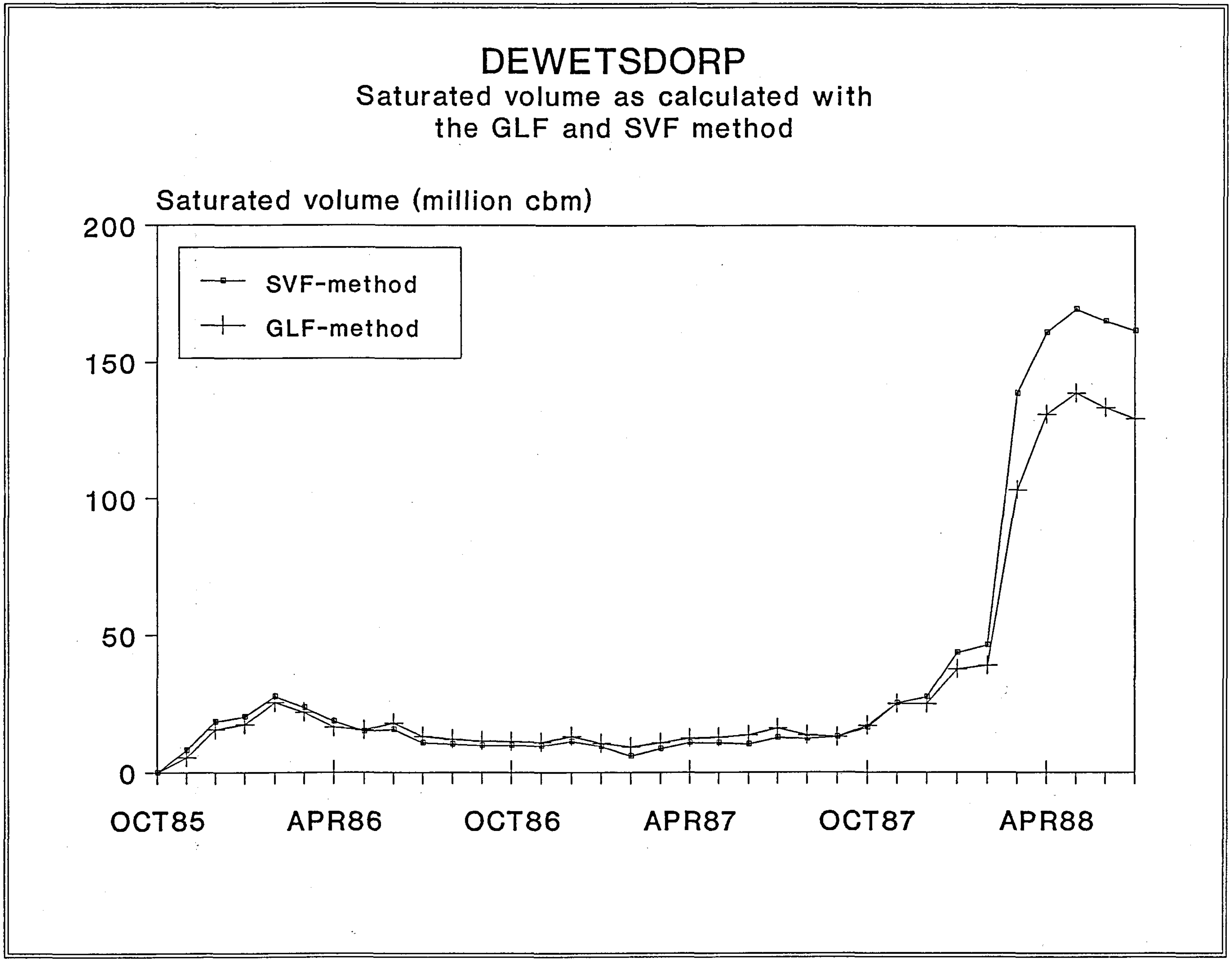


Figure 4.74 Comparison between the saturated volume as calculated with the GLF- and SVF-methods.

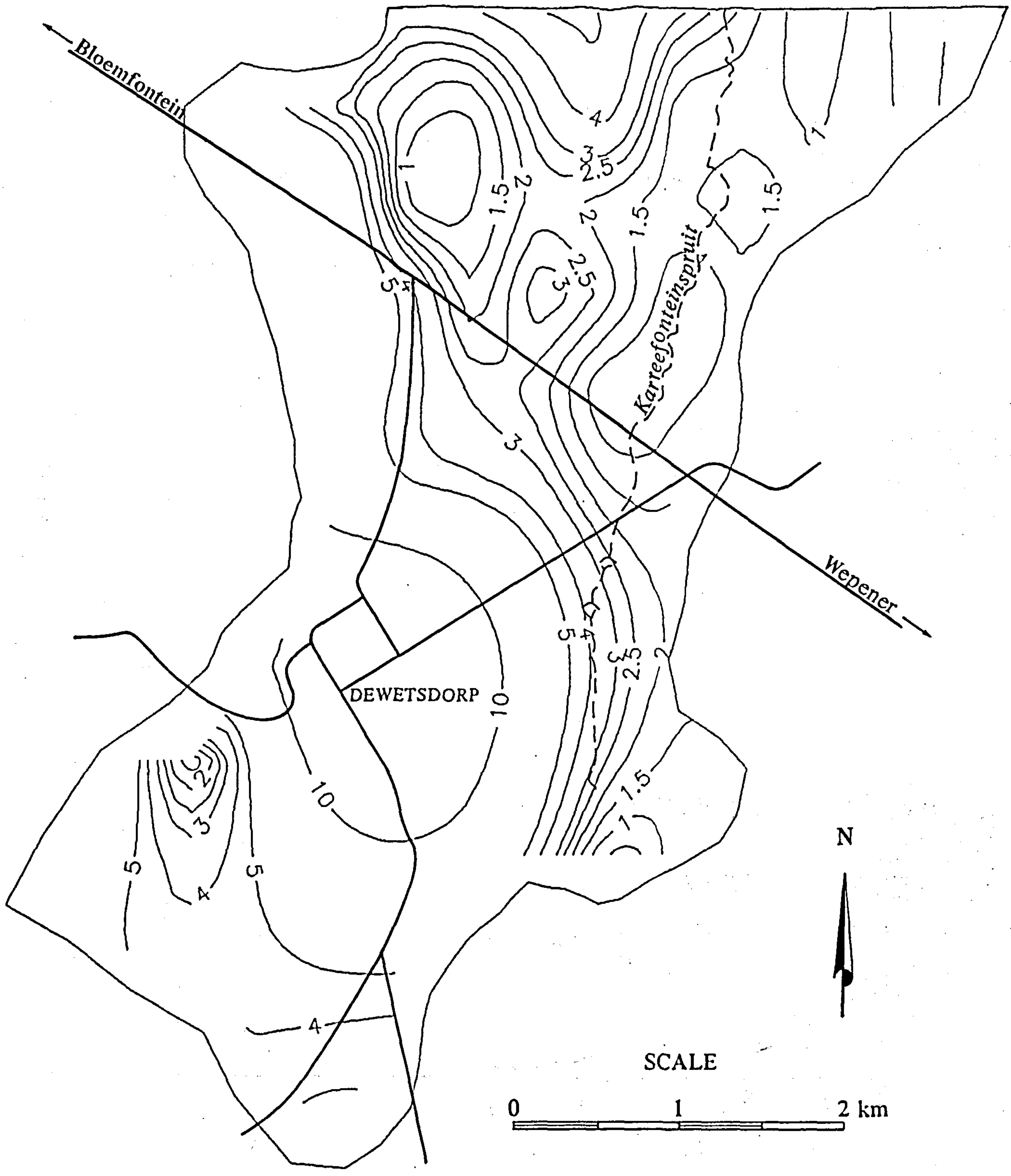


Figure 4.75 Contours of percentage recharge after the first 30 days of the flood.

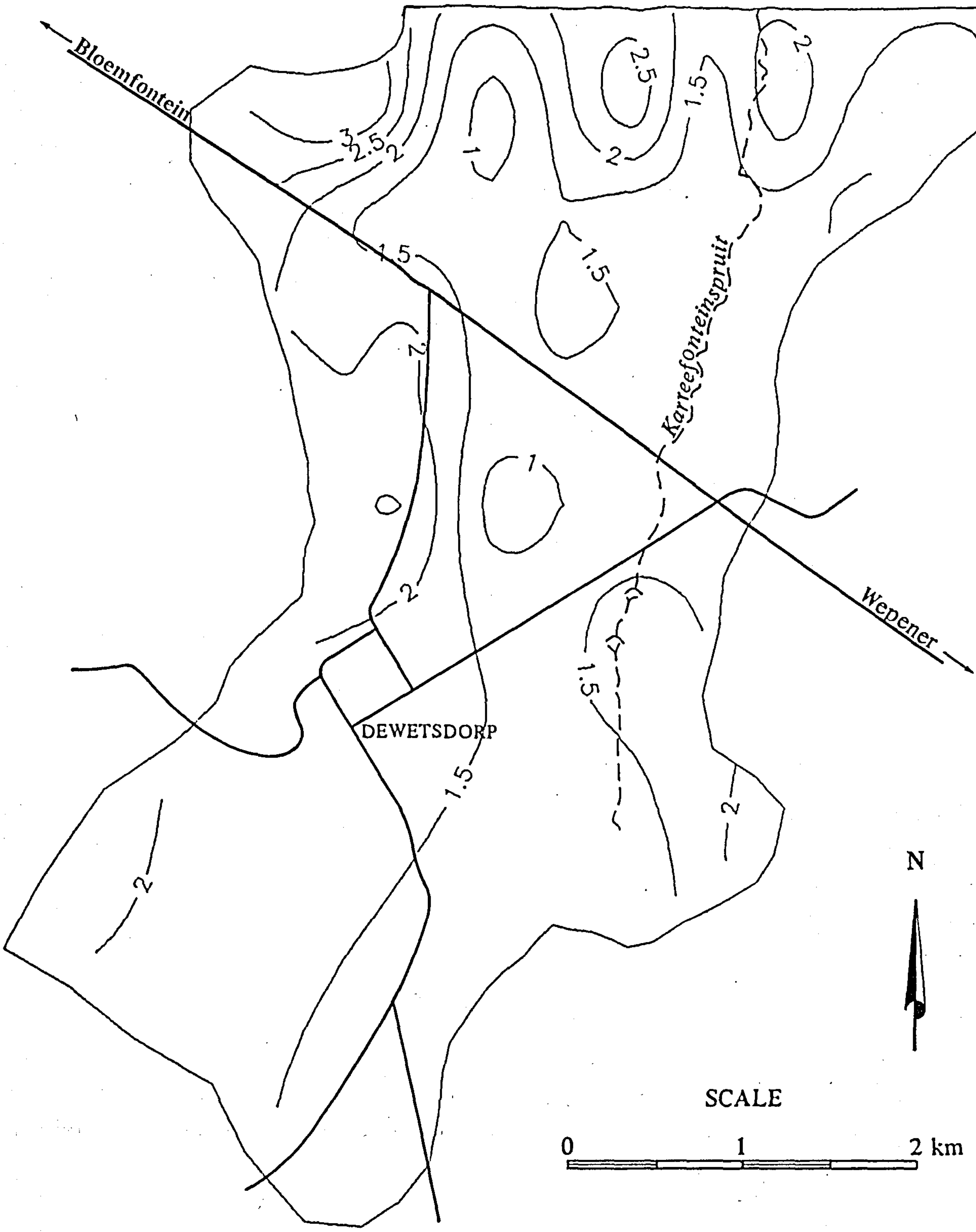


Figure 4.76 The mean ground-water recharge for Dewetsdorp expressed in mm/month.

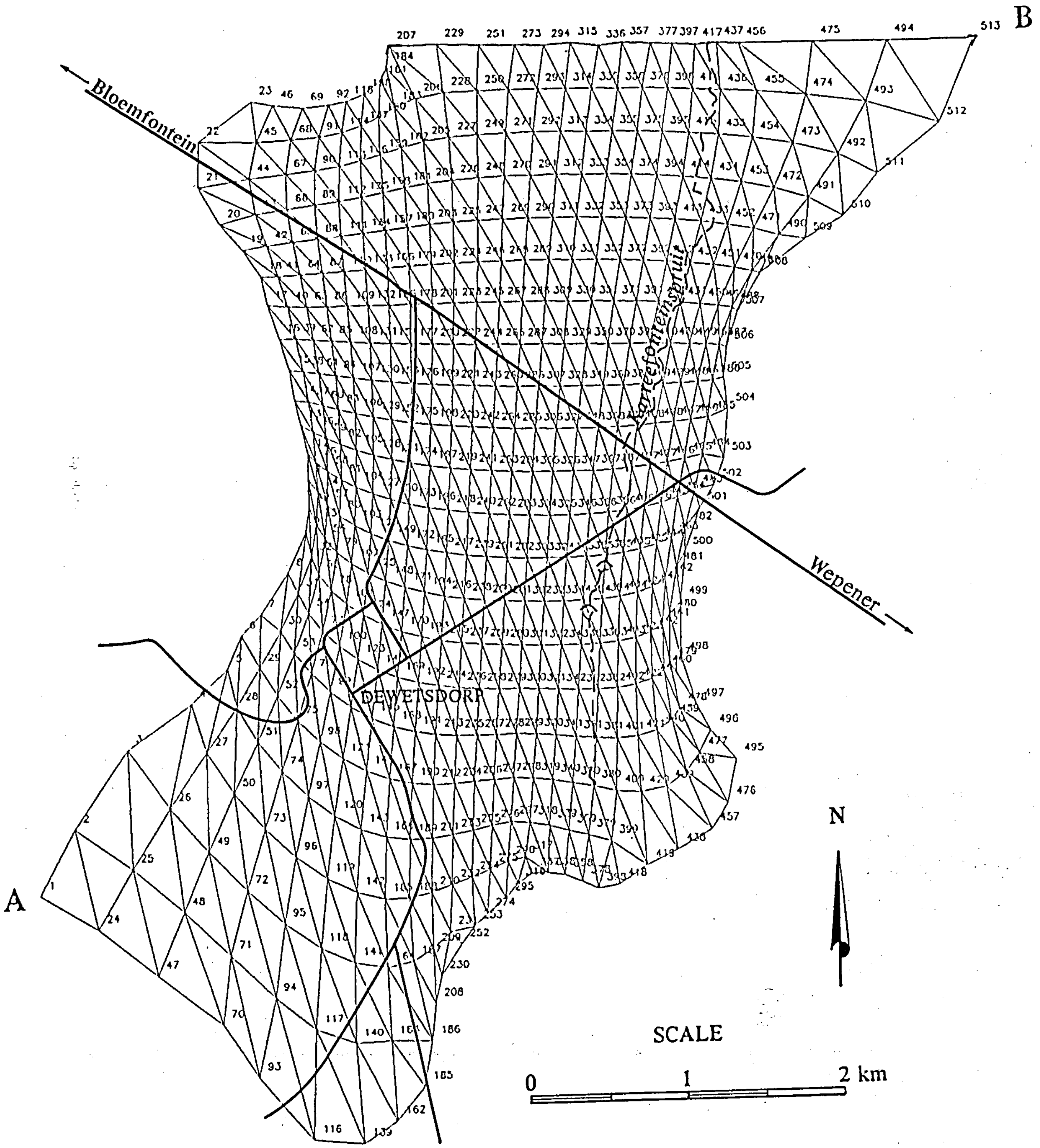


Figure 4.77 Triangular finite element mesh as used by AQUAMOD for the two-dimensional ground-water flow model. A-B is the location of a section through the aquifer.

The next step was to calibrate the model. The method proposed by Van Tonder (1988) was used to find the transmissivity distribution in the aquifer. T-values of between one and twenty were found to give a good description of the steady state solution of the Laplace equation. A S-value of 0,004 was found to simulate the physical behaviour of the aquifer well. A S-value of 0,001 was used to calibrate the model, but without success, as will be shown in the next section.

4.1.6.3.1 The February 1988 flood.

An attempt was made to try and simulate the February flood between the period 23rd February and the end of March, 1988. For a first run, a S-value of 0,004 was used with a ground-water recharge value of 18 mm (= 4,5% of the rainfall) over the whole aquifer. Figure 4.78 shows the results in terms of the difference in metres between the actual and the simulated water levels, which were not, as can be seen, very encouraging. A constant recharge value for the whole aquifer thus tends to overestimate the water table (in mamsl) in the northern part, and underestimate it in the southern part of the aquifer.

The following step was to apply varying recharge rates and S-values through the aquifer on a trial and error basis. Storativity values of between 0,001 and 0,008 were used. Figure 4.79 shows the final results obtained. A good fit was obtained, except at one or two boreholes, which experienced recharge behaviours which could not be simulated by the model. Figure 4.80 shows the distribution of the recharge percentages as estimated by AQUAMOD. Again, it is clear that the recharge is higher in the hilly areas.

To obtain a clearer view of the behaviour of the aquifer, a section through the points A and B in Figure 4.77 was constructed. Figure 4.81 shows the ground-water level behaviour for a recharge value of 18 mm, which is evenly distributed over the area. It is clearly noticeable that this recharge value is too high for the northern part and too small for the southern part. Figure 4.82 shows the results obtained for varying recharge over the aquifer.

The behaviour of the aquifer, due to the flood, is illustrated in Figure 4.83 for a S-value of 0,001 and a recharge value of 18 mm. It is clear that this combination implicated that the whole of the aquifer would have become a swamp. Thus, either the S-value must be much higher or the recharge value much lower (= 1,1% for a S of 0,001). It is felt that a recharge value of 1,1% is totally unacceptable for the February 1988 flood. For this reason, it is felt that the S-value for the Dewetsdorp aquifer is approximately 0,004.

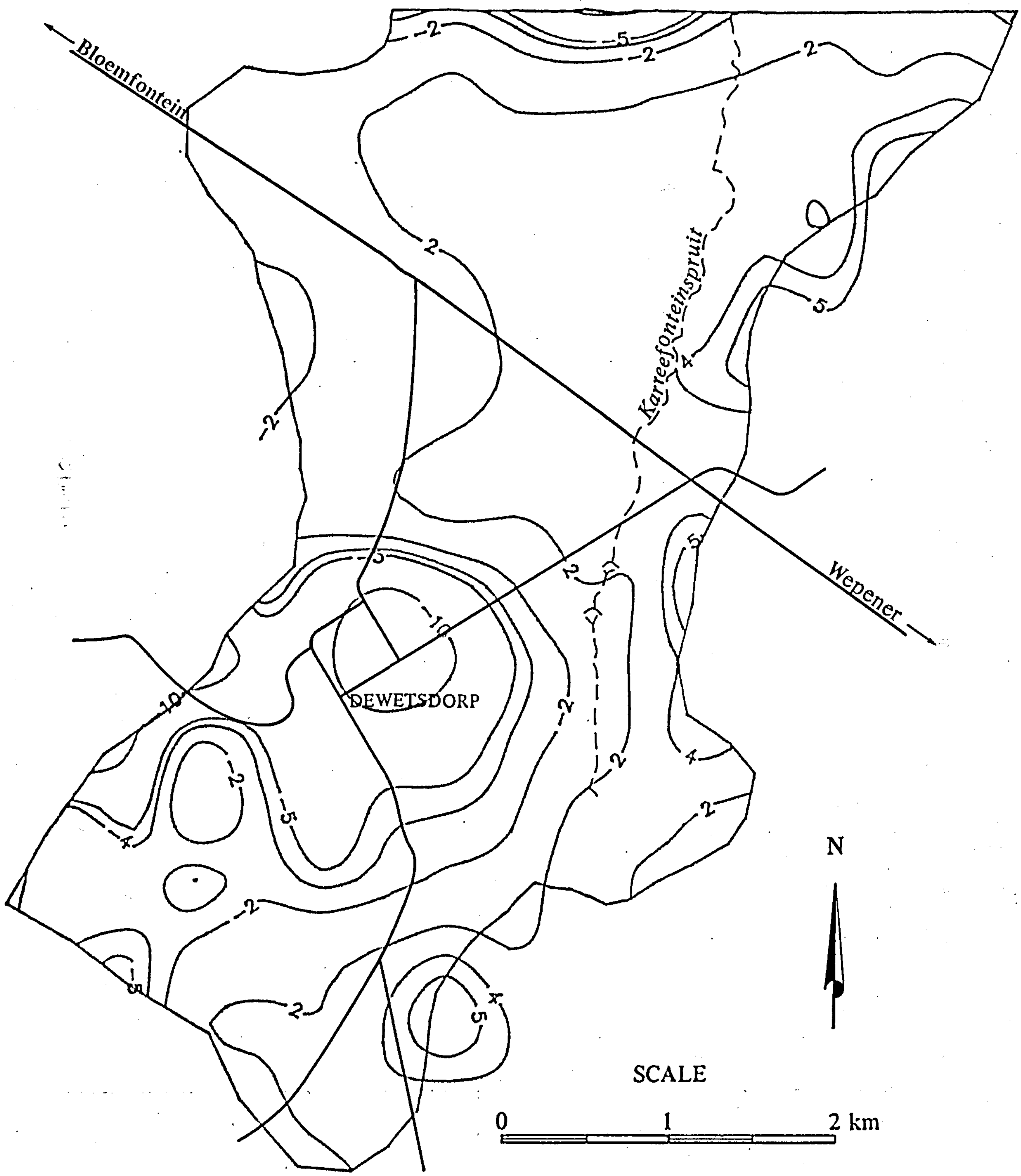


Figure 4.78 Difference between finite element simulated and actual ground-water levels in the Dewetsdorp aquifer, 30 days after the February flood by applying a constant S-value of 0,004 and a recharge value of 18 mm/month to the whole aquifer.

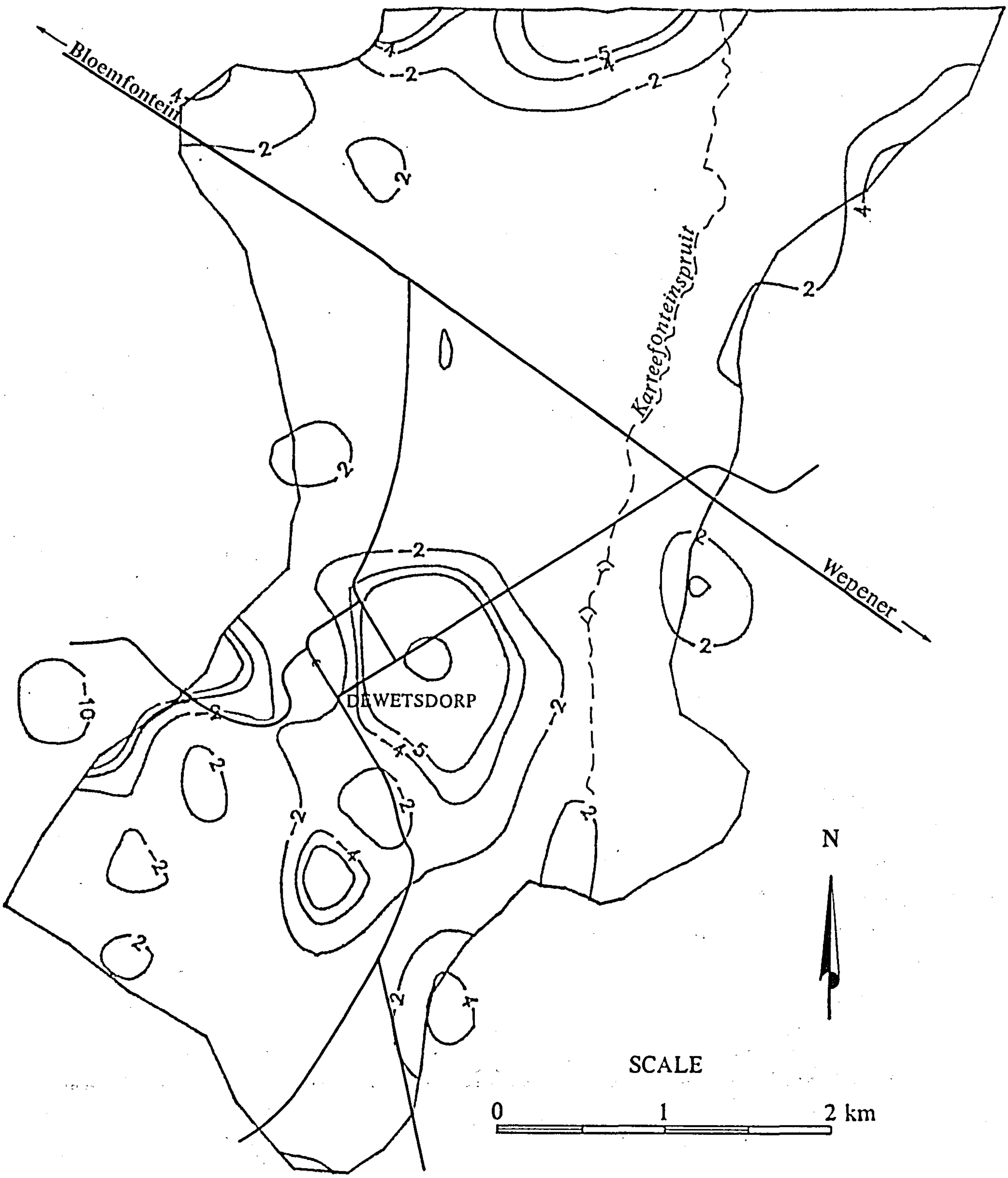


Figure 4.79 Difference between simulated and actual ground-water levels 30 days after the flood, by applying a varying S- and recharge values concept.

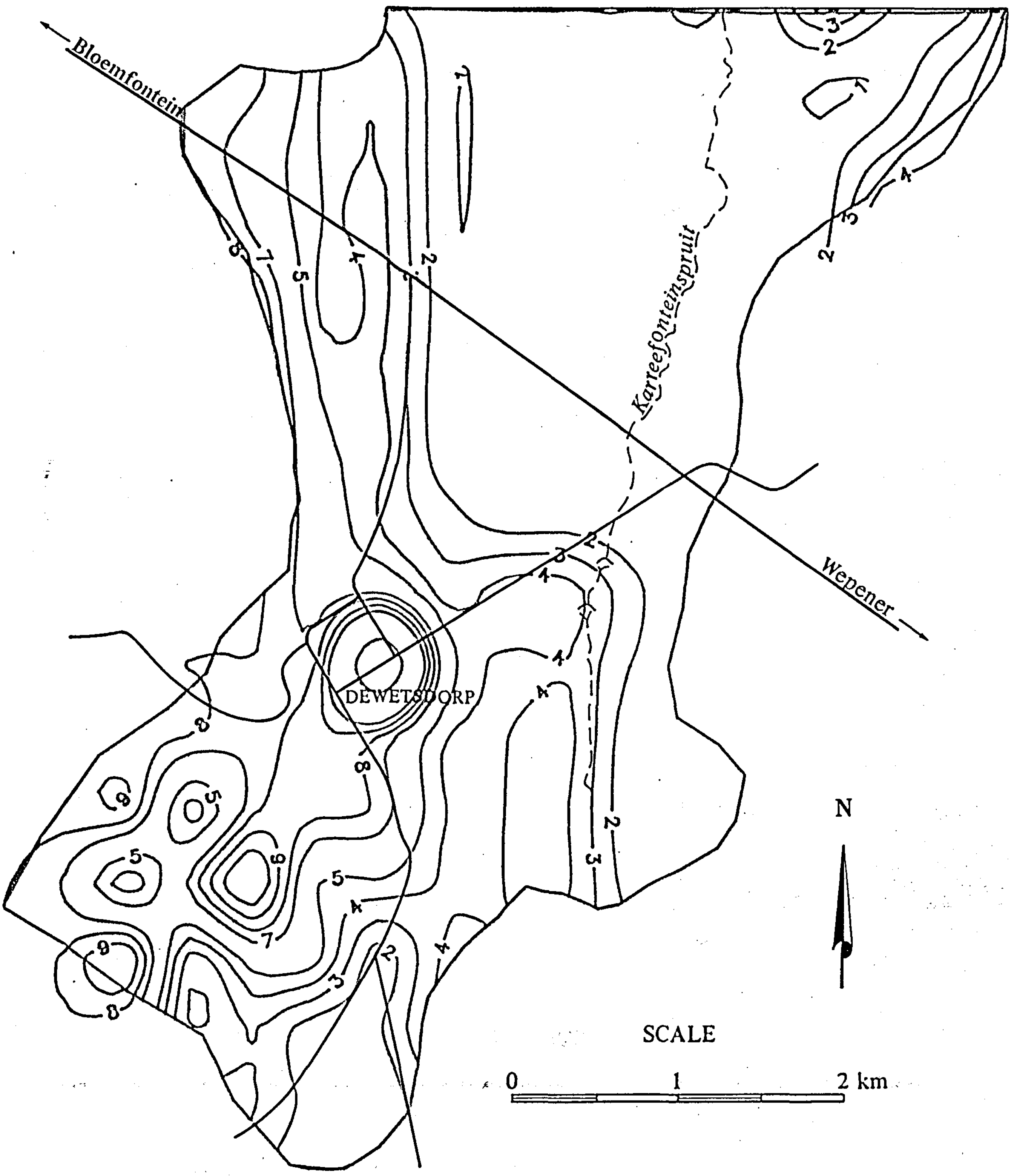


Figure 4.80 Percentage recharge of the flood for the first 30 days after the flood.

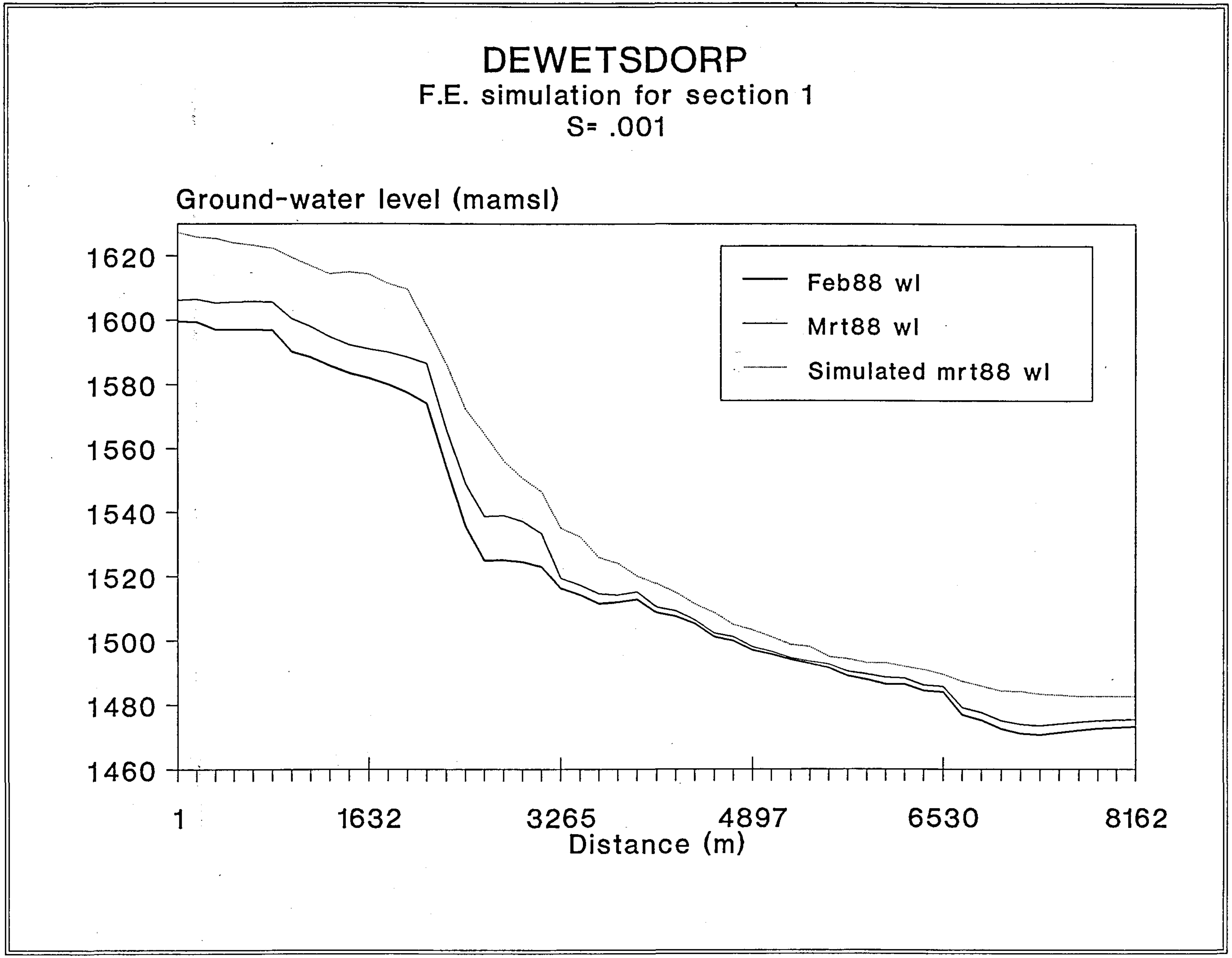


Figure 4.81 Finite element simulation for section A-B in Dewetsdorp by applying a S-value of 0,001 and a recharge of 18 mm/month for the flood. The results are not very promising, as can be seen.

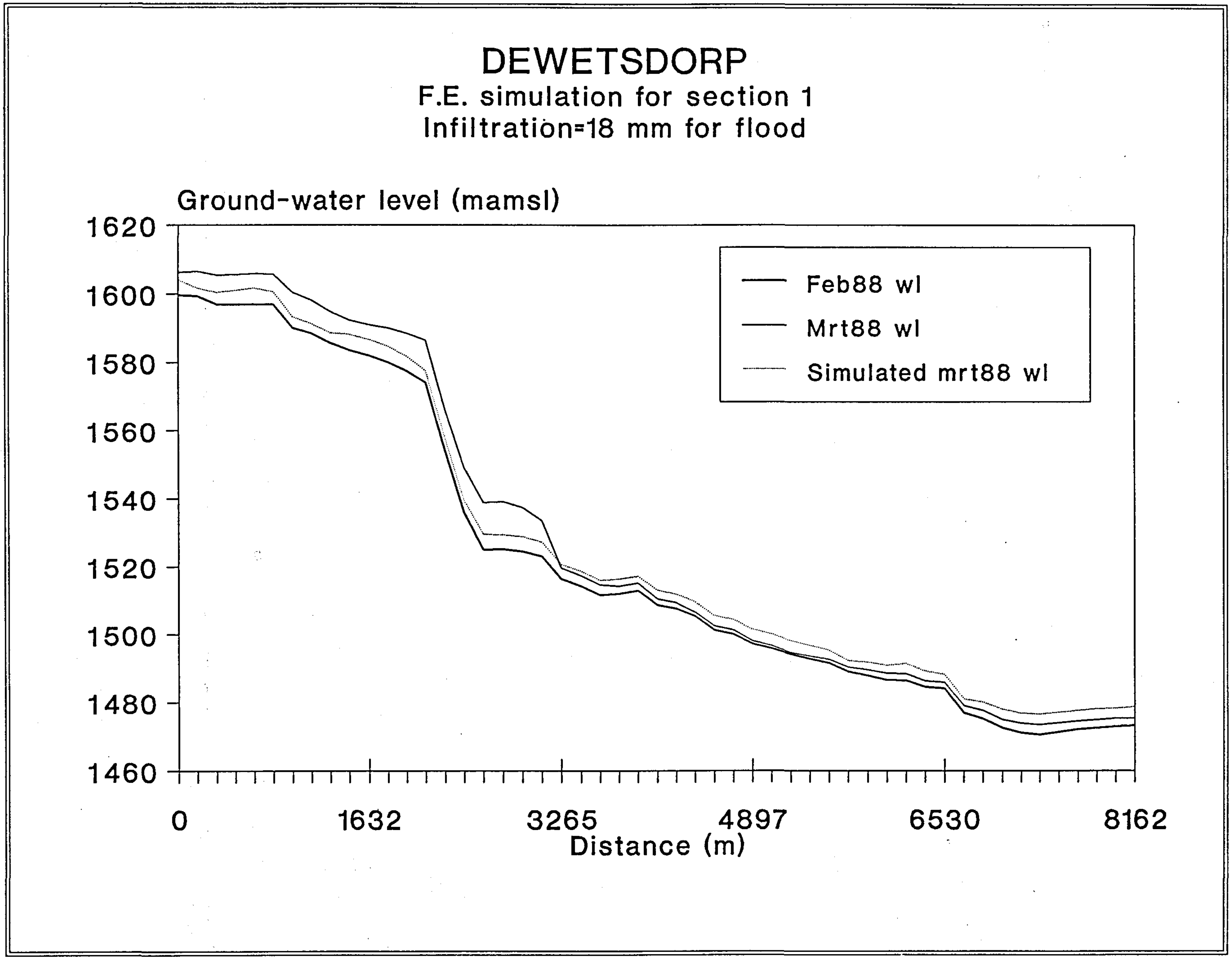


Figure 4.82 Finite element simulation for section A-B in Dewetsdorp by applying a S-value of 0,004 and a recharge of 18 mm/month for the flood. It can be seen that the recharge must be higher in the southern part of the aquifer (left part of figure).

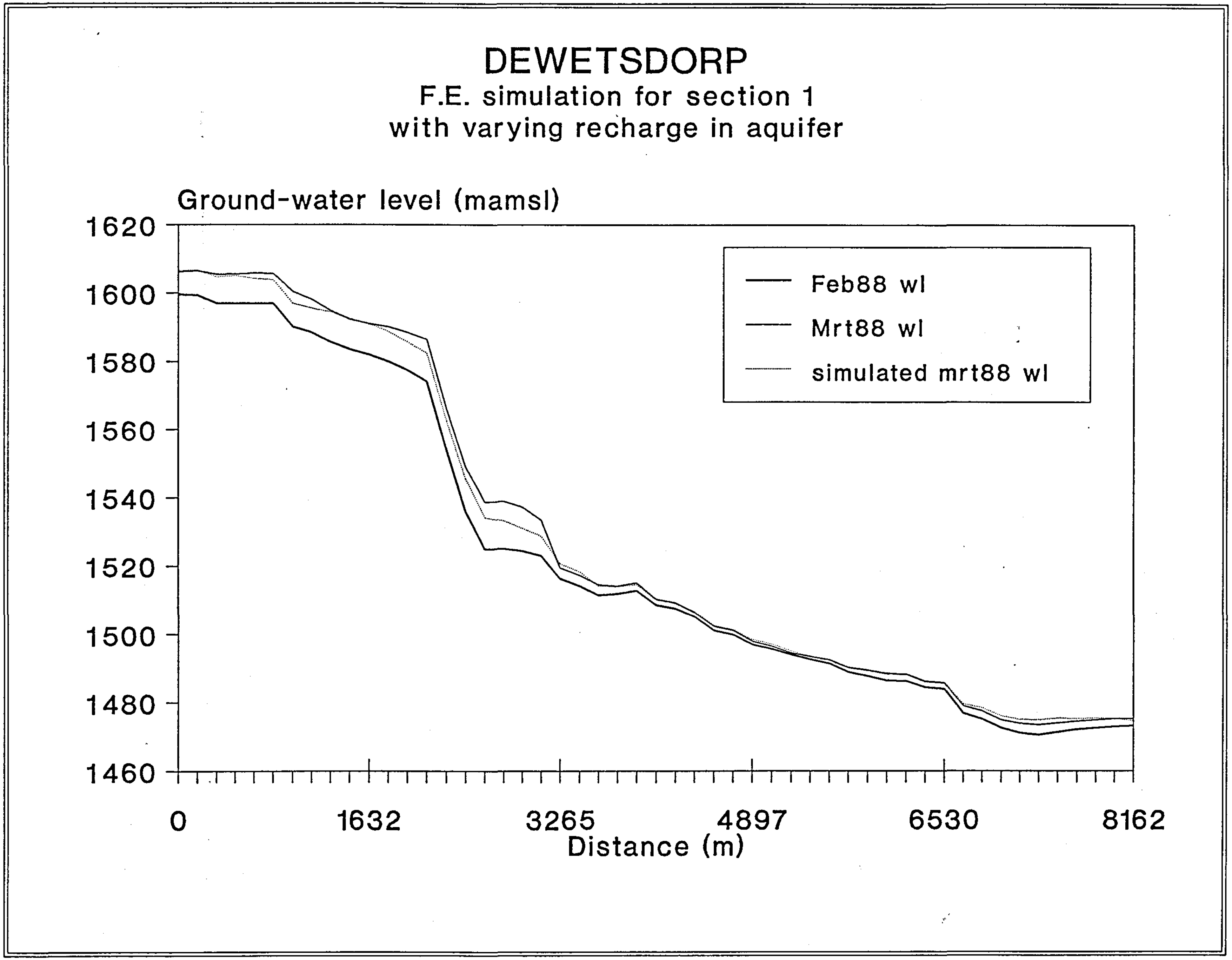


Figure 4.83 Final simulated and actual ground-water levels for the finite element simulation of section A-B after the recharge values were increased in the southern part of the Dewetsdorp aquifer.

4.1.6.4 RECHARGE VIA THE UNSATURATED ZONE

In this section, the estimation of the recharge of the Dewetsdorp aquifer with the direct methods which used soil water balance models of the unsaturated zone, shall be discussed.

4.1.6.4.1 The Zero Flux Plane (ZFP) method

Neutron counts were used to determine the water content up to a depth of 1,7 m. To calibrate the neutron counts N, the following equation was used:

$$\text{Soil moisture } \theta\ (\%) = (N + \gamma.\rho_B)/(\alpha.\rho_B + \beta)$$

where N = normalized neutron counts (e.g. 0,250)

ρ_B = Dry bulk density (e.g. 1,7)

and $\alpha = 0{,}00239$; $\beta = -0{,}01826$; $\gamma = -0{,}07453$, as determined with the Flether-Powell algorithm.

The neutron counts of sixteen out of the twenty tubes in the Dewetsdorp area, showed a similar behaviour during the study period. Because of this, only the analysis of one site (site 1) will be discussed here. The volumetric water content at this site with depth is shown in Figure 4.84 for certain selected times. The total water potential is obtained by the equation $\phi = z + \psi$, where z is the depth below the ground surface (a negative value) and ψ is the matric potential (negative) derived from the retention curve (Figure4.85) for site 1. Figure 4.84 shows the total water potential.

The ZFP-method is based on locating the point of zero hydraulic gradient and thus the zero reference flux in the soil profile, and summing up the changes in water content below that point. From Figure 4.86, it is clearly recognizable that the ZFP lies constantly 0,8 m below the surface for the whole of the study period. It is striking to observe that even during the February 1988 flood, a zero flux plane still exists, which implies that there was no water movement through that plane. By summing up the water changes beneath the ZFP between 9/2/88 and 9/3/88, a negligible small recharge value was obtained. The only explanation for this behaviour of the unsaturated zone, is that recharge has occurred along preferred pathways, by-passing the soil matrix.

4.1.6.4.2 The Darcian approach for the unsaturated zone

From the total water potential graph and the retention curve, the unsaturated hydraulic conductivity for the flood at site 1 was determined (see Figure 4.87). A number of saturated K-values (K_S) on undisturbed soil samples were determined in the laboratory with a constant head permeameter. They vary between $2 \cdot 10^{-1}$ to $4 \cdot 10^{-5}$ m/d. It is interesting to note that, for two of the samples, no K_S-value could be calculated, due to

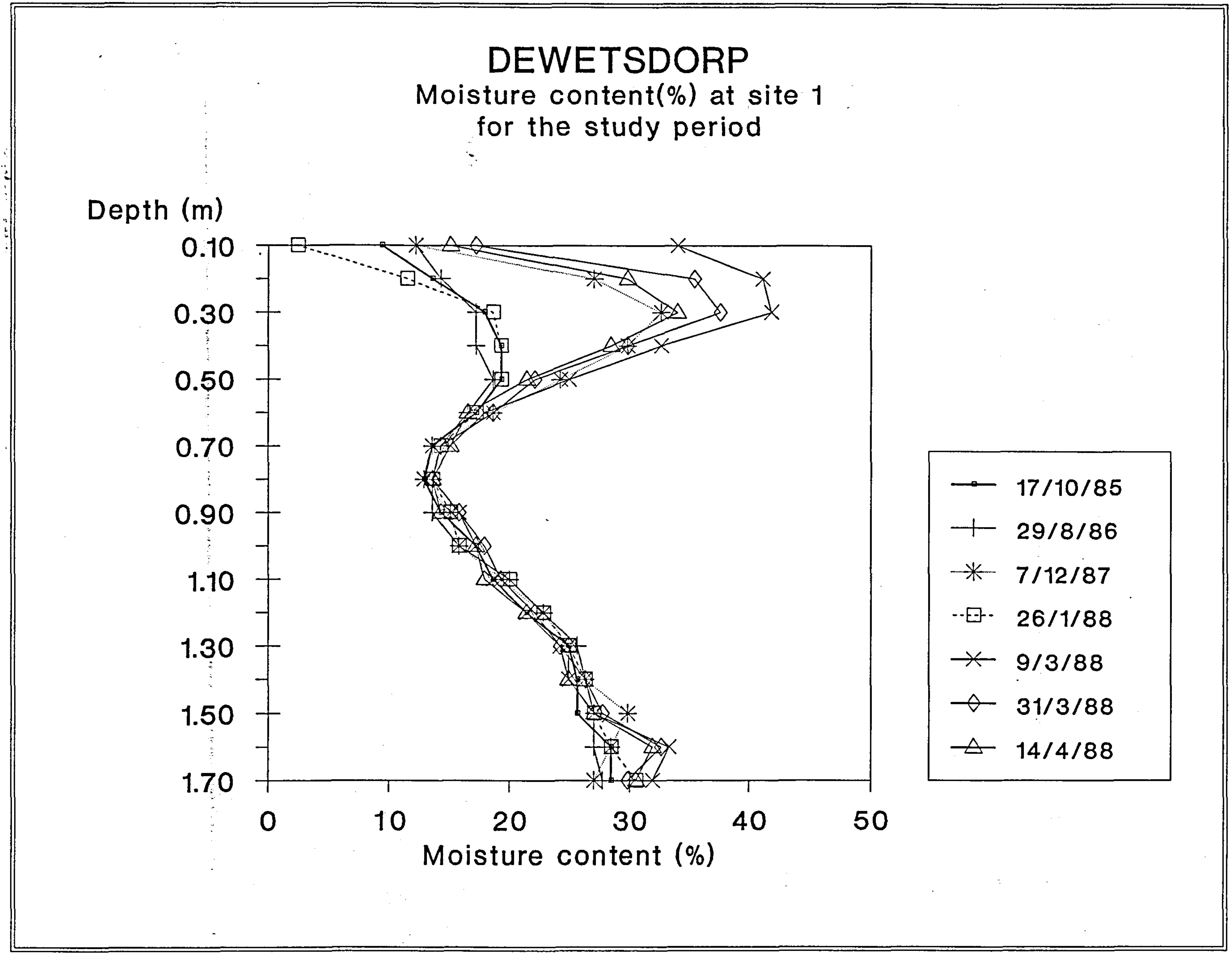


Figure 4.84 Water content at neutron access tube 1 during the study period.

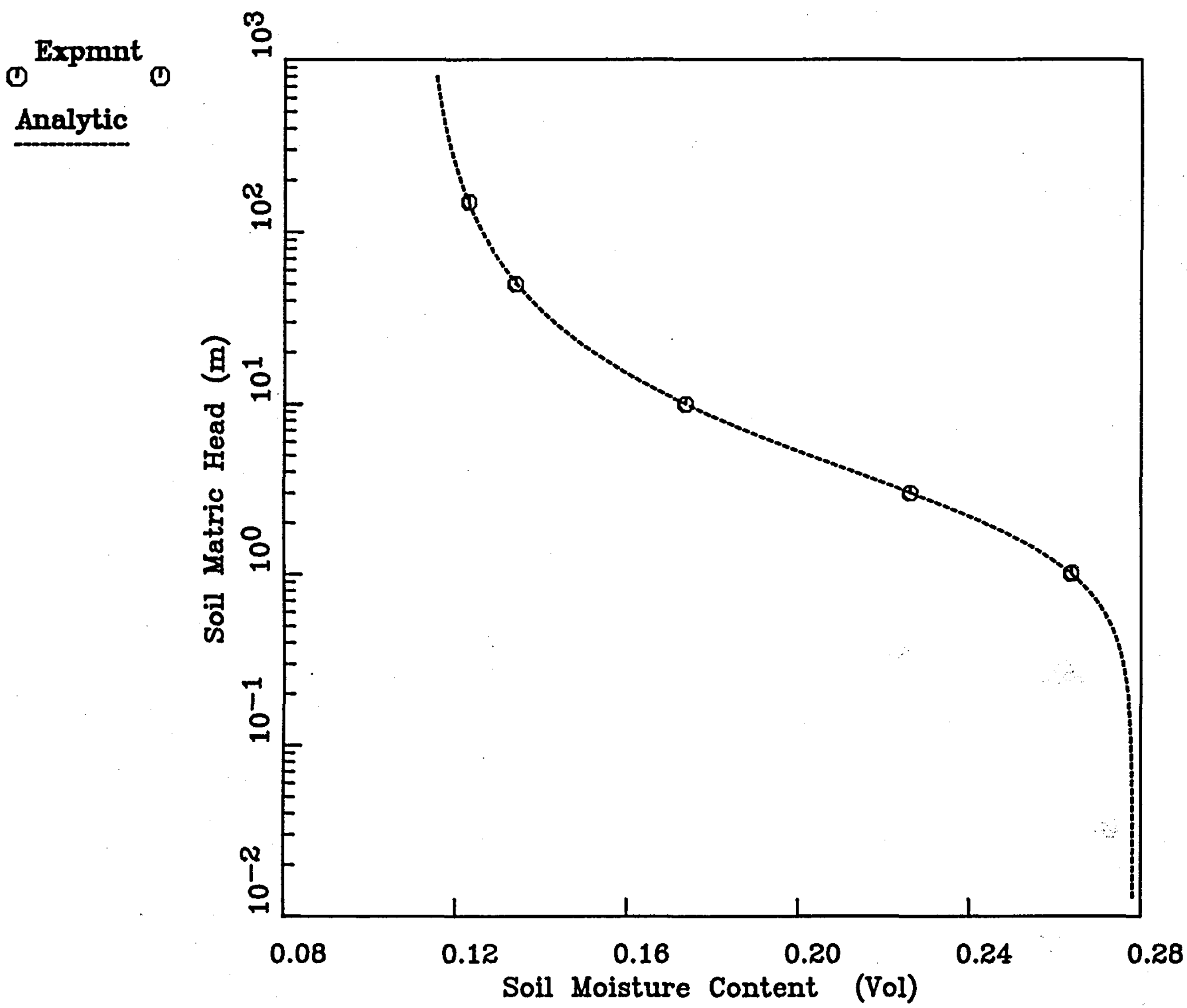


Figure 4.85 Mean retentivity values for the A-horizon at Dewetsdorp, together with the Van Genuchten analytical solution.

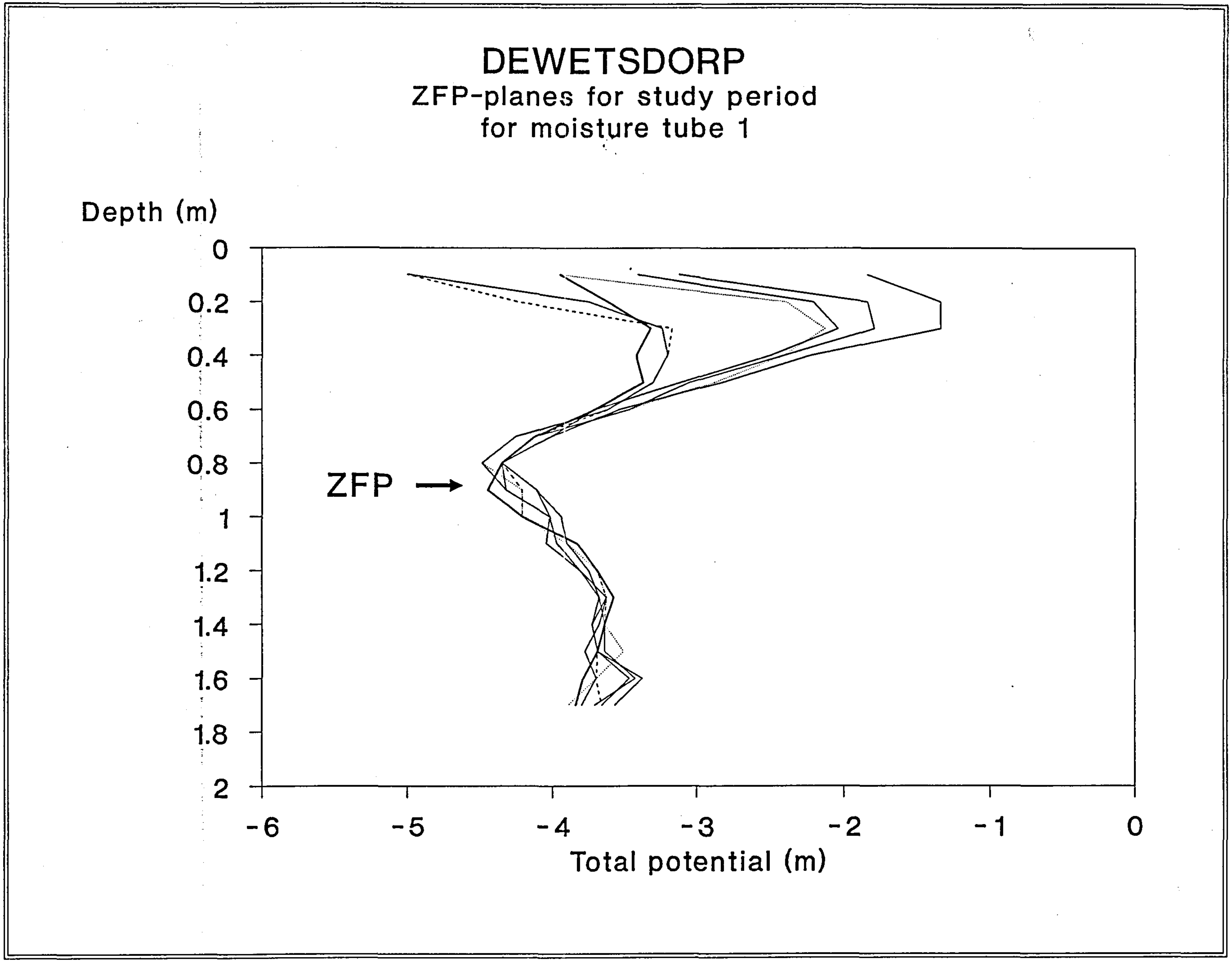


Figure 4.86 The total potential at neutron access tube 1 for the study period. The existence of a ZFP during the whole study period is obvious.

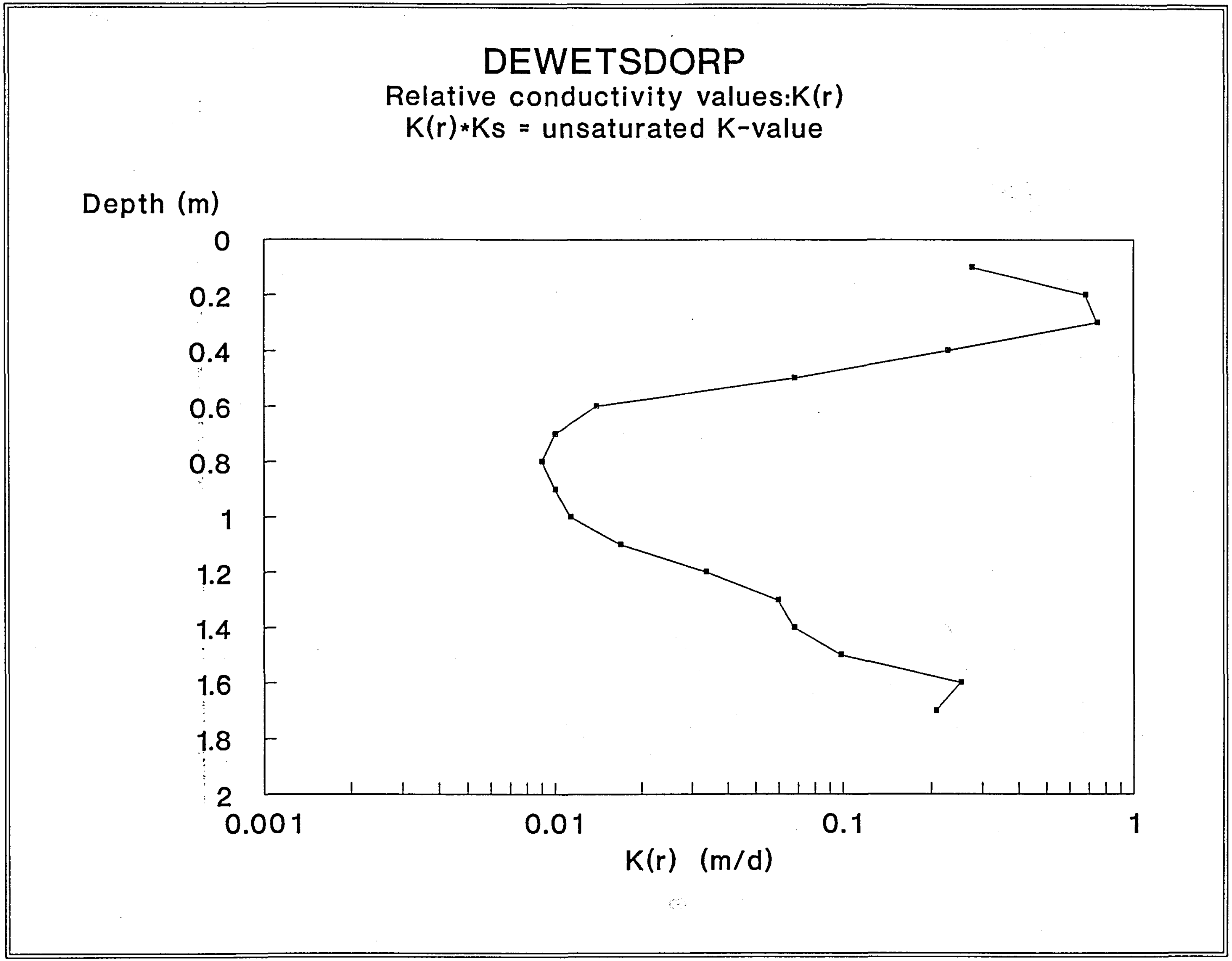


Figure 4.87 Relative hydraulic conductivity values at neutron access tube 1 (Dewetsdorp) after the February 1988 flood. Multiplication of these K_r-values by the saturated K-value at tube 1 yields the unsaturated K-values.

zero throughflow. Even after three weeks of applying a constant pressure head of 1,1 m (i.e. a gradient of 20) to the column of soil, there was still no flow that could be measured. For site 1, the representative saturated K-value was determined as $1{,}5 \cdot 10^{-4}$ m/d. The clay and silt fraction of a soil are also sometimes used for the calculation of K_s (Campbell, 1985). The equation reads:

$$Ks = 2 \cdot 10^{-3}\, e^{-4,26(ms+mc)}\ [kgs/m^3]$$

where m_s = silt content and m_c = clay content in fractions. For site 1, it follows that

$$\begin{aligned} K_s &= 2 \cdot 10^{-3}\, e^{-4,26(0,10+0,37)} \\ &= 2{,}7 \cdot 10^{-4}\ kgs/m^3 \\ &= 2{,}7 \cdot 10^{-4}\ (9{,}8 \cdot 10^{-3}) \cdot (86400)\ m/d \\ &= 0{,}22\ m/d \end{aligned}$$

which is not in agreement with the laboratory K_s. Naney (Naney *et al.*, 1983) proposed the following equation for the determination of K_s values:

$$K_s = 110{,}16 \cdot \rho_B^{-22,2}\ m/d$$

where ρ_B = dry bulk density. By using the ρ_B of site 1 of 1,86, the Naney equation yielded a K_s value of $1{,}1 \cdot 10^{-4}$ m/d which compared excellent with the laboratory K_s.

The potential gradient at a depth of 1 m below the ground surface was 0,81 m for the flood period and the unsaturated hydraulic conductivity was calculated as:

$$K_u = 1{,}5 \cdot 10^{-4}\ (0{,}011) = 1{,}6 \cdot 10^{-6}\ m/d$$

The Darcian flux was calculated as:

$$\begin{aligned} q &= -K_u d\phi/dz \\ &= -1{,}6 \cdot 10^{-6}\ (0{,}81) \\ &= 1{,}33 \cdot 10^{-7}\ m/d \end{aligned}$$

which implies that the recharge at this site was $1{,}33 \cdot 10^{-7}$ m^3/d or 0,05 mm/a, showing that no measurable soil matrix flow occurs. Because the vertical finite element soil water flux model (SUFF) is only applicable for unsaturated soil matrix flow and not for flow along certain pathways, it was decided that such an exercise would have been meaningless.

The only explanation for the fact that all the boreholes showed a rise after the flood, is that recharge has taken place along preferred pathways (e.g. cracks), by-passing the soil matrix. The common mechanism of recharge through the soil matrix of the unsaturated zone does thus not seem to be applicable for the Dewetsdorp aquifer. At this stage, we believe that infiltration does take place through the soil matrix of the upper (say 30 cm) soil. Measurable downward fluxes occur, until a relative heavy clay layer is reached, from where the water movement is predominantly horizontal. At specific locations, this clayey layer may have wide cracks, leaving a pathway for water to travel along. Dieleman and De Ridder (1963) found similar results when studying the recharge near Lake Chad in Central Africa.

4.1.6.4.3 The soil water balance method

The July 1987 version of ACRU (Schulze, 1984) was used for this study. The model has daily time steps and as such uses daily input of climatic data (rainfall and potential evaporation). Certain variables regarding land-use/vegetation characteristics are input at monthly level and then reduced to daily values. The partitioning and redistribution of water in the ACRU model are depicted diagrammatically in Figure 4.88.

The soil water retention curves are an important tool in ACRU. Factors which have an influence on these curves are, for example, particle shape, bulk density, clay mineralogy, organic matter content, structure and degree of aggregation. Some of these factors defy precise quantitative description; consequently their effect on retentivity may be assessed in qualitative terms only (Hutson, 1983). Using data from a wide spectrum of physical environments, Hutson (1983) developed a general model, in equation form, based on percentages of clay and silt together with bulk density, which may be applied in South African soils.

The model, applicable to stable soils, takes the form:

$$\theta = a + b.Cl + c.Si + d.\rho_B$$

where

a,b,c,d = regression constants
Cl = percentage clay
Si = percentage silt
ρ_B = bulk density

Values for the regression constants are given in Schulze (1984, p. 157). (For example: for field capacity (FC) calculations, a = 0,00558; b = 0,00365; c = 0,00554 and d = 0,0303, while for the wilting point (WP) calculations, a = 0,0602; b = 0,00322; c = 0,00308 and d = -0,0260). The method proposed by Bennie *et al.* (1988) can also be used to calculate the retentivity values.

By using this equation for the A and B-horizons at the 20 sites at Dewetsdorp, the results, as depicted in Figures 4.89 and 4.90, were found. The field capacity, FC, varies between 0,24 and 0,38, while the wilting point, WP, lies between 0,24 and 0,9 mm/mm, with a mean FC = 0,3 and WP = 0,15. (The Bennie equation yielded more or less the same WP value, but a higher FC value).

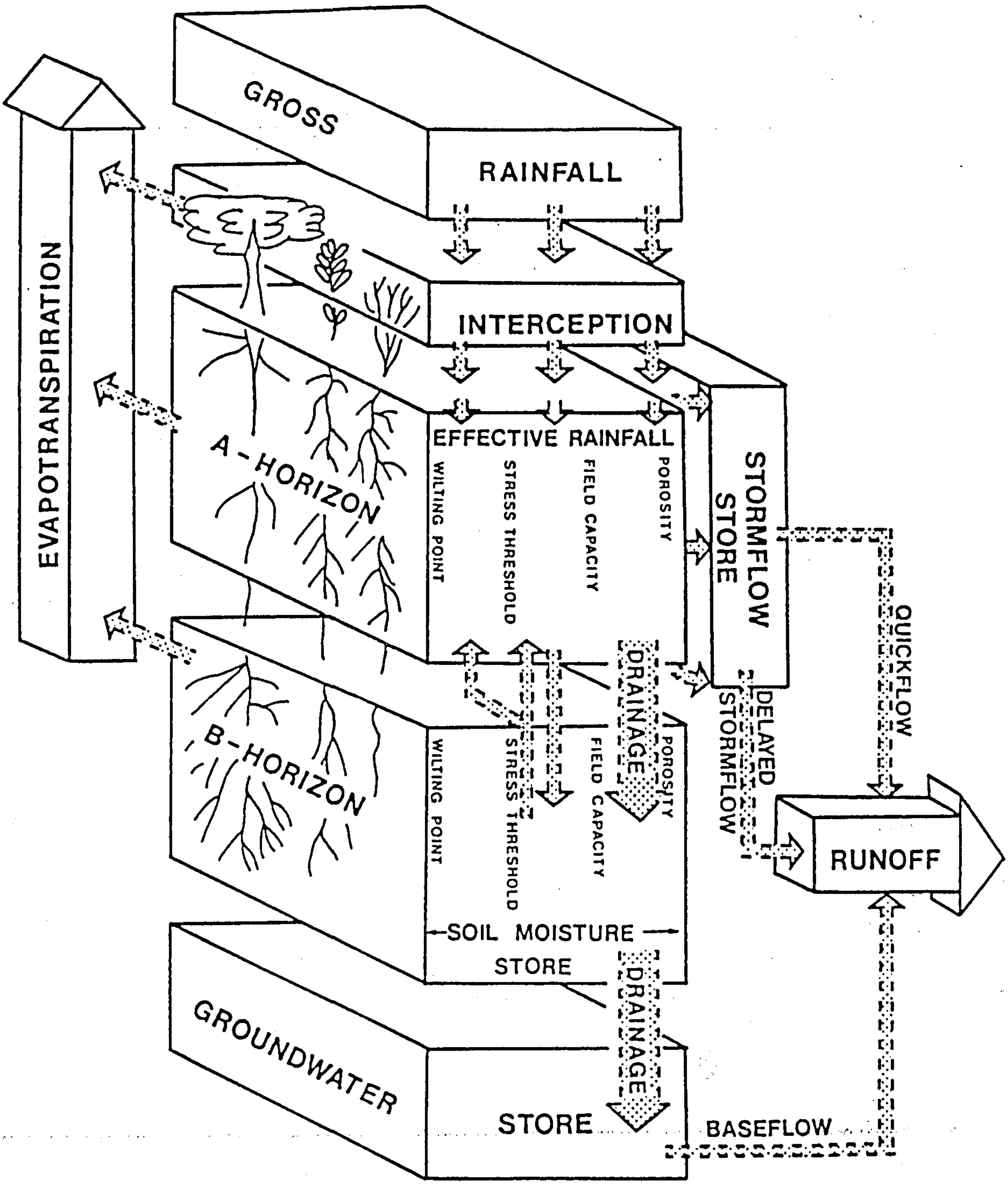


Figure 4.88 Partitioning and redistribution of water in the ACRU-model (after Schulze, 1984).

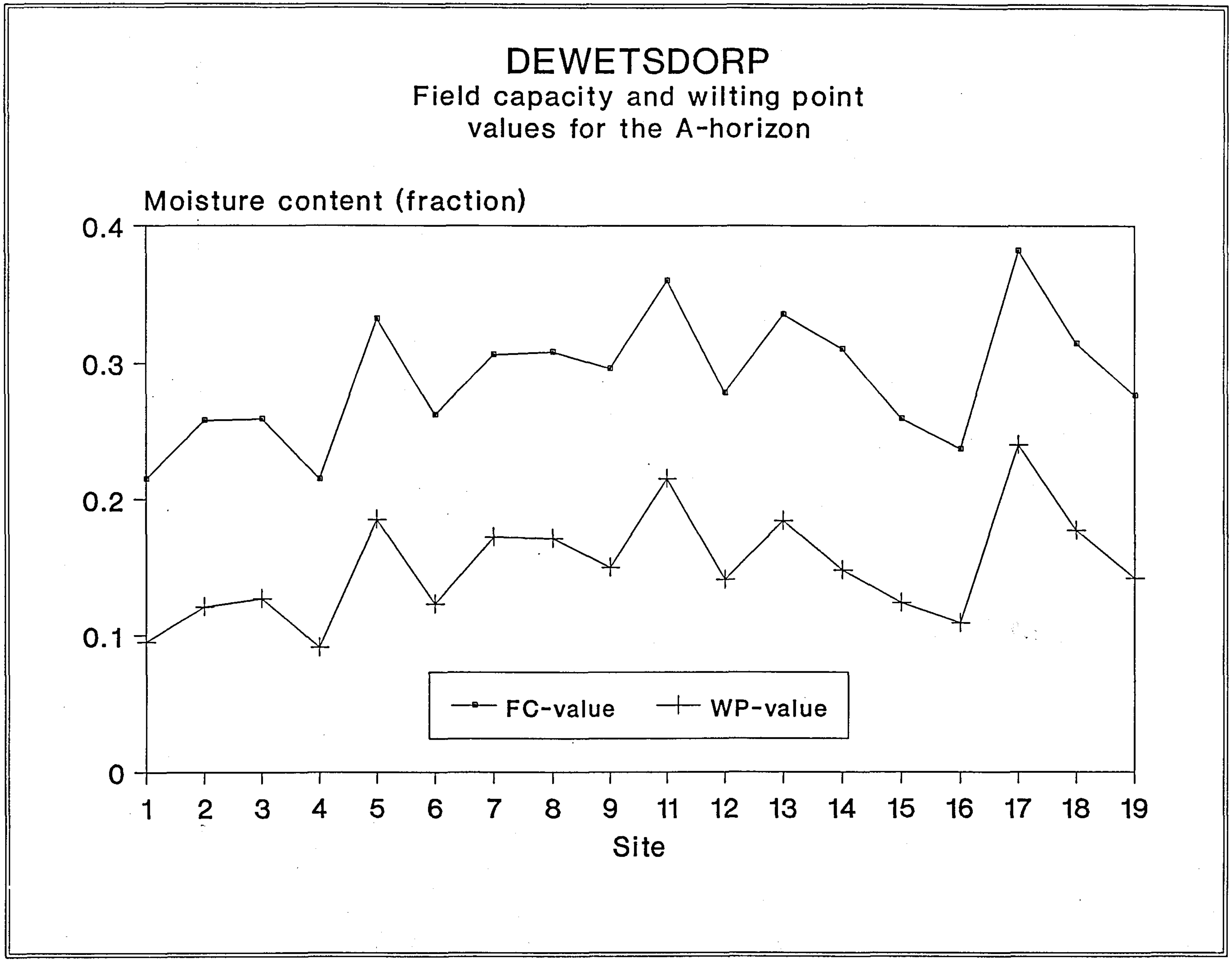


Figure 4.89 Field capacity and wilting points (used by the ACRU-model) of the A-horizon for different sites at Dewetsdorp.

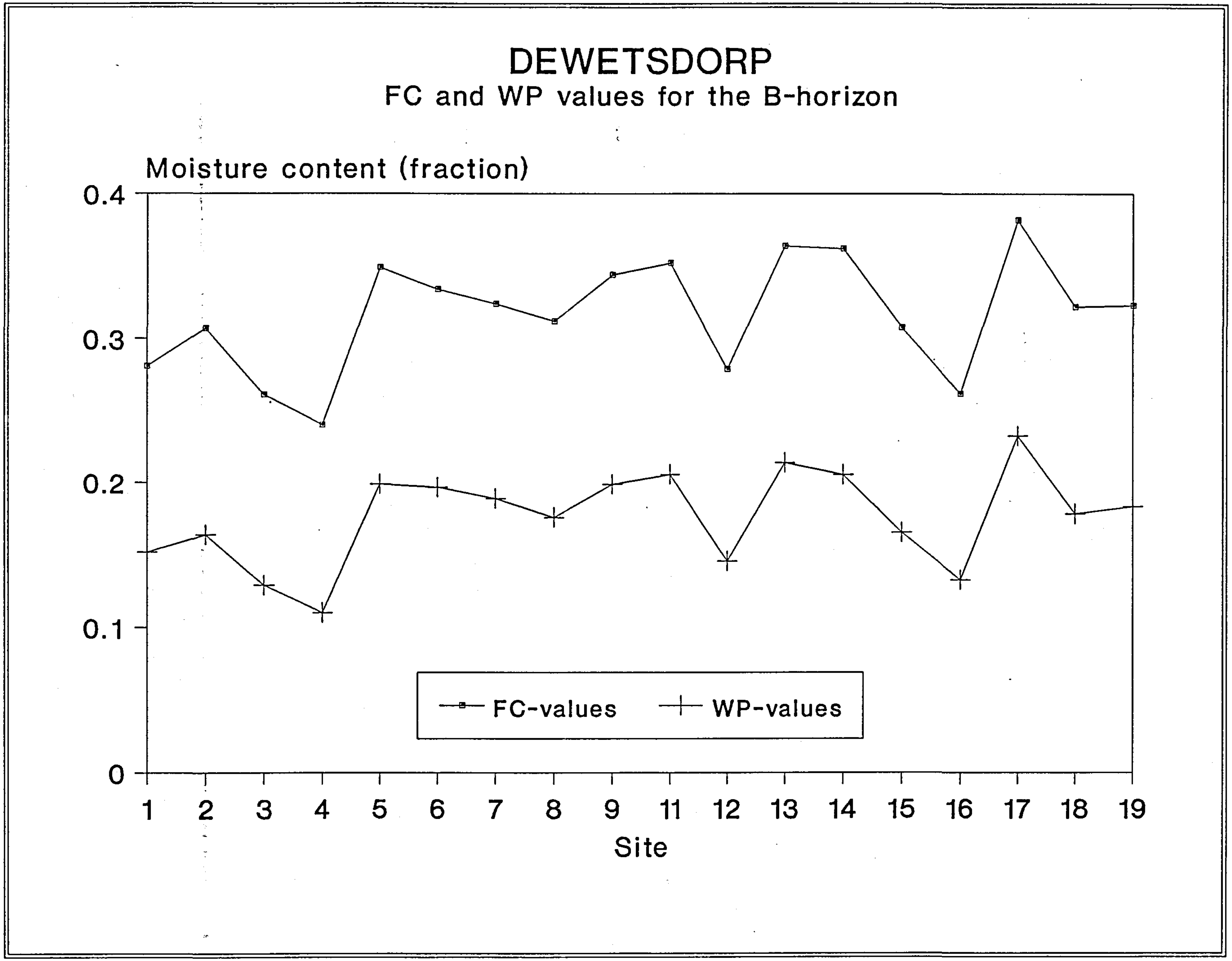


Figure 4.90 Field capacity and wilting points of the B-horizon for different sites at Dewetsdorp.

Daily meteorological data from the automated weather station were used for the application of the ACRU model, but the results were not, however, very encouraging. For the range of FP and WP values above, as well as the other parameters (e.g. porosity, vegetation information, latitude, longitude, root constant, etc.), ACRU yielded results which show that there was zero recharge during the flood period. The behaviour of the ACRU model is also very critical, dependent on the value of the porosity in the A and B horizons, as Table 4.5 indeed shows. From Table 4.5, it is evident that it is very difficult to calibrate the ACRU model. The main problem is that ACRU required a very experienced and specialized person to get meaningful results.

Porosity A-horizon	Porosity B-horizon	Total Recharge [mm]	Recharge for flood
0,29	0,34	0	0
0,30	0,35	3,4	0
0,31	0,36	40	0
0,32	0,37	73	0
0,33	0,38	100	0
0,34	0,39	248	0

Table 4.5. Recharge as estimated with the ACRU model for different porosity values and with a FC = 0,290 and WP = 0,152 for the A-horizon and a FC = 0,332 and WP= 0,192 for the B-horizon.

The recharge for the porosity values which yielded a recharge of 40 mm during the study period, is shown in Figure 4.91, whereas the soil water deficit for this period is shown in Figure 4.92. Contradictory to the ACRU model which shows a soil water deficit for the whole of 1987, the saturated volume graph reveals that recharge definitely has taken place during this period.

What can be said in this concluding paragraph, is that recharge estimates in which the unsaturated zone is used, are not reliable in the Dewetsdorp aquifer, because the recharge mechanism (which is mainly flow along preferred pathways) does not favour models that use soil water matrix flow.

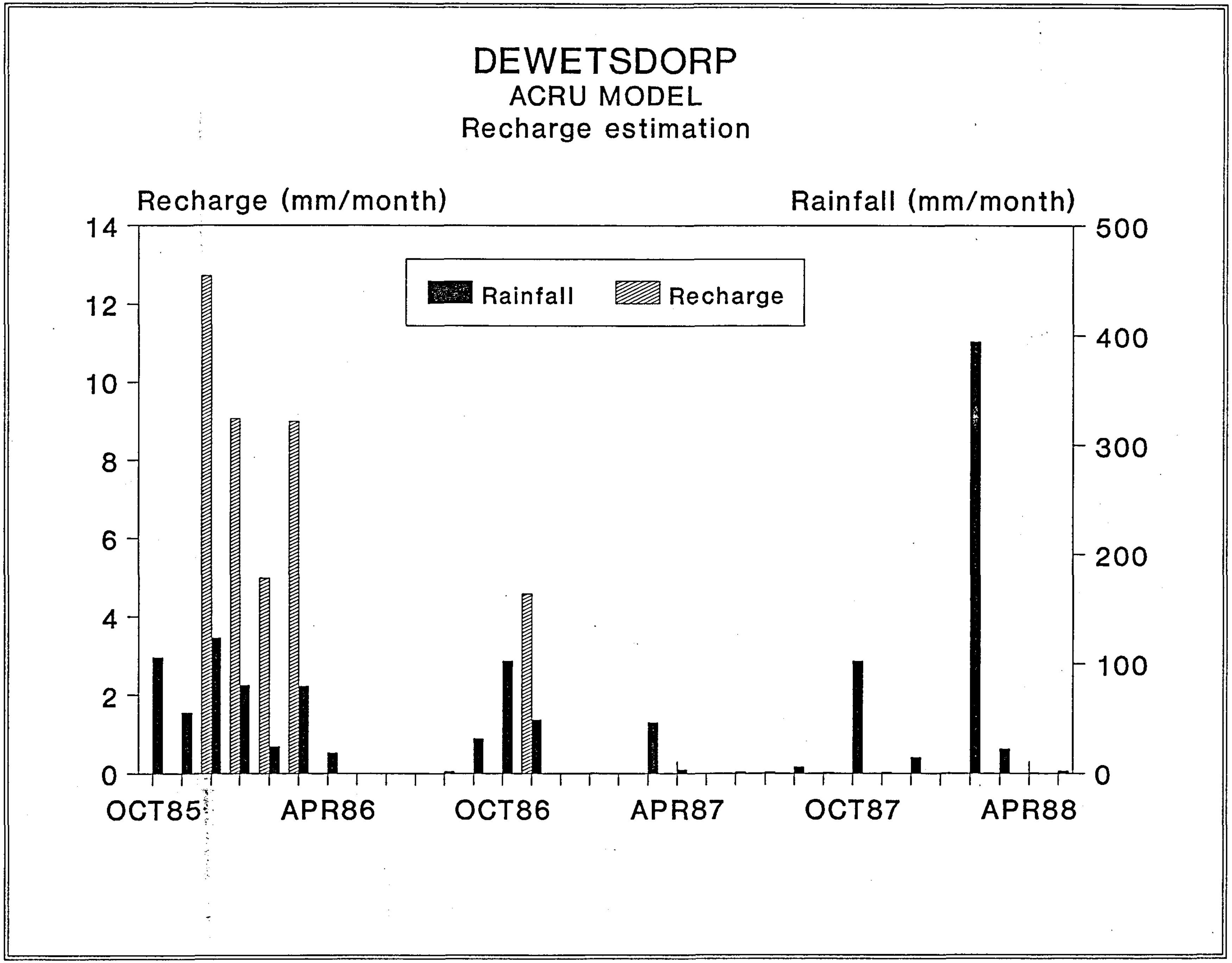


Figure 4.91 Ground-water recharge estimations as obtained with the ACRU-model for the Dewetsdorp aquifer during the study period.

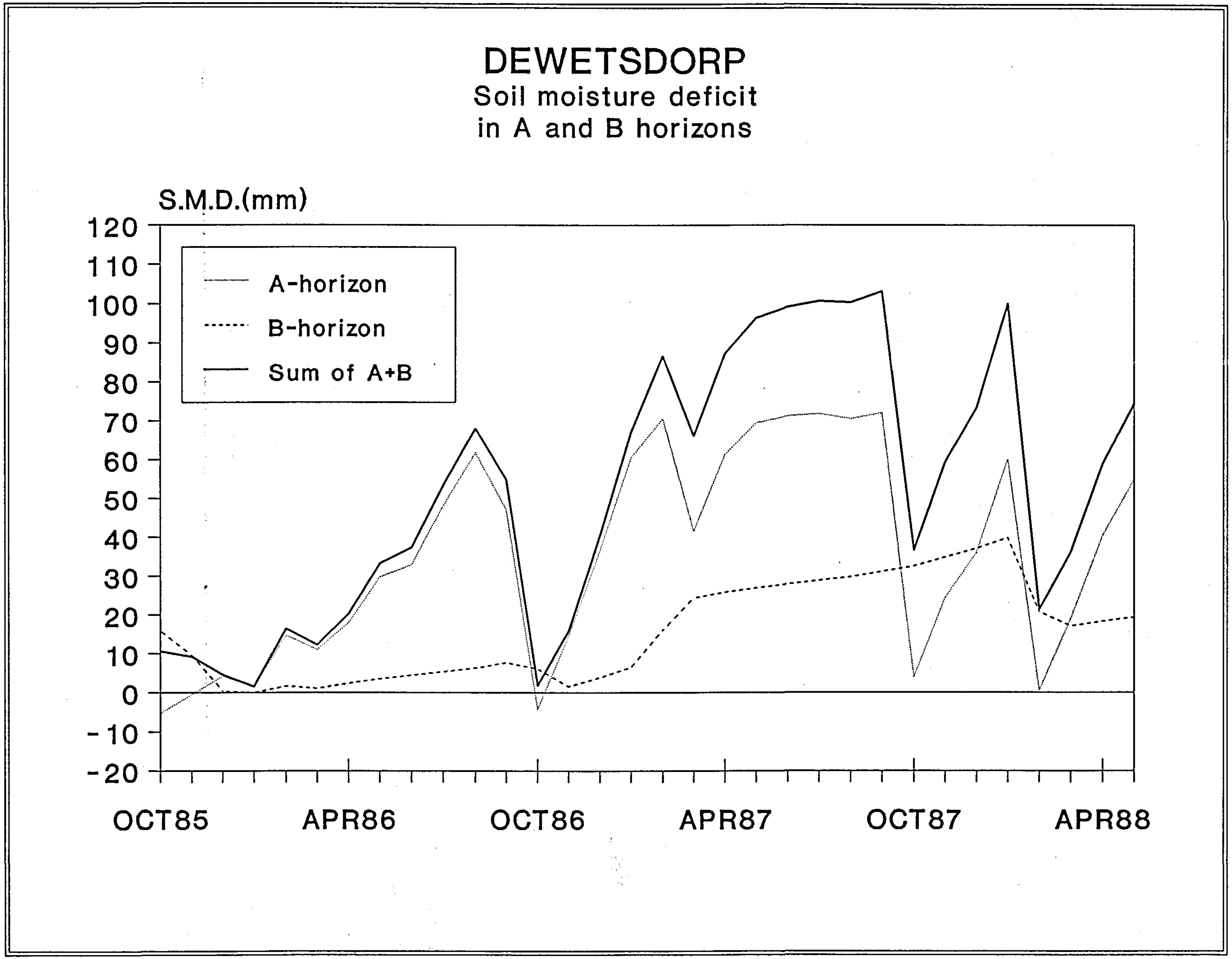


Figure 4.92 Soil water deficits in the A- and B-horizons during the study period as estimated with the ACRU-model.

4.2 DE AAR

4.2.1 PHYSIOGRAPHY

LOCATION - TOPOGRAPHY - DRAINAGE

De Aar lies approximately 100 km south of the Orange River in the north-western part, of what is called the Great Karoo, at an elevation of 1240 m. Here, the railway lines of the Western and the Eastern Cape, the Transvaal and South West Africa meet. The town plays a major role in the transport system of South Africa and is, after Germiston, the second largest railway junction of the Republic. In a surrounding where surface water is rare, the town's name refers to the ground-water resources found there.

Like Dewetsdorp, the surrounding of De Aar resorts, according to the terrain morphological classification (Kruger, 1979), under lowlands with hills. The landscape is relatively flat with inselbergs, where dolerite sills have protected the underlying shales from weathering. Dykes and "ring dykes" are also typical features in the surrounding of the town which form natural walls, often for considerable distances. The botanical classification is *False upper Karoo* (Acocks, undated).

The research area lies just outside the town within such a "ring dyke" which has a diameter of 12 km. It comprises parts of the farms Renosterpoort and Klein Brandfontein (both part of the original farm Brandfontein 87), Vaalbank and Vaalkop (parts of the former Zwartekopjes 131) and Sinclairs's Dam 133. The (western) Brak River,which is also called the Elandsfontein River, enters the area in the south in a rather narrow channel between the Swartkoppies and the Kasamberge. In the west where the river has cut its bed into dolerite of the Renosterpoort, it is dammed up. A tributary of the Brak River, which originates within the area under consideration, leaves it in the north-west at Brandfontein.

The elevation of the plain in the area is between 1200 - 1300 m. The surrounding hills are approximately 70 to 110 m higher. The highest point lies in the Swartkoppies with 1513 m.

4.2.2 CLIMATE

In the Köppen classification (see Figure 4.2), De Aar falls in the boundary area between the *Southern Steppe* SS and *Desert* W. The Southern Steppe is a semi-arid region with rainfall between 250 mm/a in the west and up to 500 mm in the east. Precipitation is mainly due to convectional showers and thunderstorms which fall between October and March and reach peak values in February or March. Unsettled weather conditions in the winter may occur on one or two occasions per month (Schulze, 1986).

Heating of the land during the day and cooling at night due to radiation cause large temperature variations. They display large variations both on a diurnal and a seasonal scale as can be seen from the statistics for the period 1941 to 1984 hereunder (after Weather Bureau, 1986).

	January [° C]	July [° C]
Maximum	40,6	25,0
Mean daily maximum	32,2	16,8
Mean daily temperature	23,9	8,6
Mean daily minimum	15,6	0,5
Minimum	4,4	-10,0

The annual temperature march for De Aar is shown in Figure 4.93.

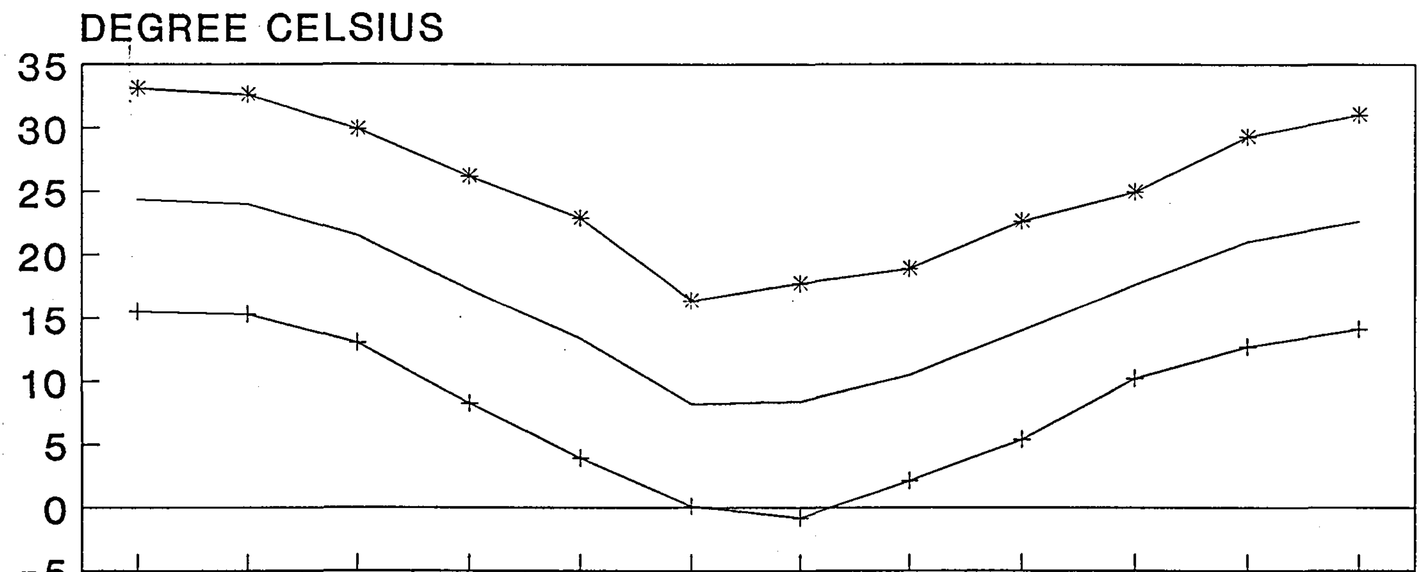
Winds are normally out of a north-westerly direction and attain maximum strength in the afternoon. In the winter months, occasionally cold southerly winds do occur.

4.2.2.1 RAINFALL

The mean annual rainfall of De Aar amounts to 287 mm (median = 265 and standard deviation = 114) with a C.V. of 40%. The mean monthly rainfall distribution shows a maximum in March and a secondary one in November (see Figure 4.94). The inherent variability of the rainfall at De Aar is evident from this large C.V. value. Since records started in 1888, the minimum annual rainfall was 89 mm in 1951 and the maximum 638 mm in 1976. For the period 1930 to 1988, the annual rainfall and the deviation from the mean are displayed in Figure 4.95. The graph demonstrates the high variability. From October, 1985, to September, 1986, the rainfall was 217 mm, and for the consecutive year, it was 144 mm. This low rainfall in 1986/87 was accountable for the lowest mean ground-water level for the study period of 1215,45 m in September, 1987.

A probability analysis was performed on the maximum 30-day rainfall amount of each year. From Figure 4.96, it can be seen that the 250 mm of the February 1988 flood has a return period of 200 years. The non-exceedence probability for a specific annual rainfall is given in Figure 4.97, while the return period for maximum annual rainfall is given in Figure 4.98. Figure 4.99 can be used for the purpose to obtain the return

DE AAR
DAILY MAX AND MIN TEMPERATURE



	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
MAX	33.13	32.6	29.93	26.17	22.88	16.37	17.69	18.91	22.63	24.94	29.28	31.05
MIN	15.54	15.34	13.05	8.28	3.96	0.09	-0.79	2.13	5.41	10.24	12.76	14.12
MEAN	24.34	23.97	21.49	17.23	13.42	8.23	8.45	10.52	14.02	17.59	21.02	22.59

—*— MAX —+— MIN —— MEAN

Figure 4.93 Daily maximum and minimum temperatures for De Aar.

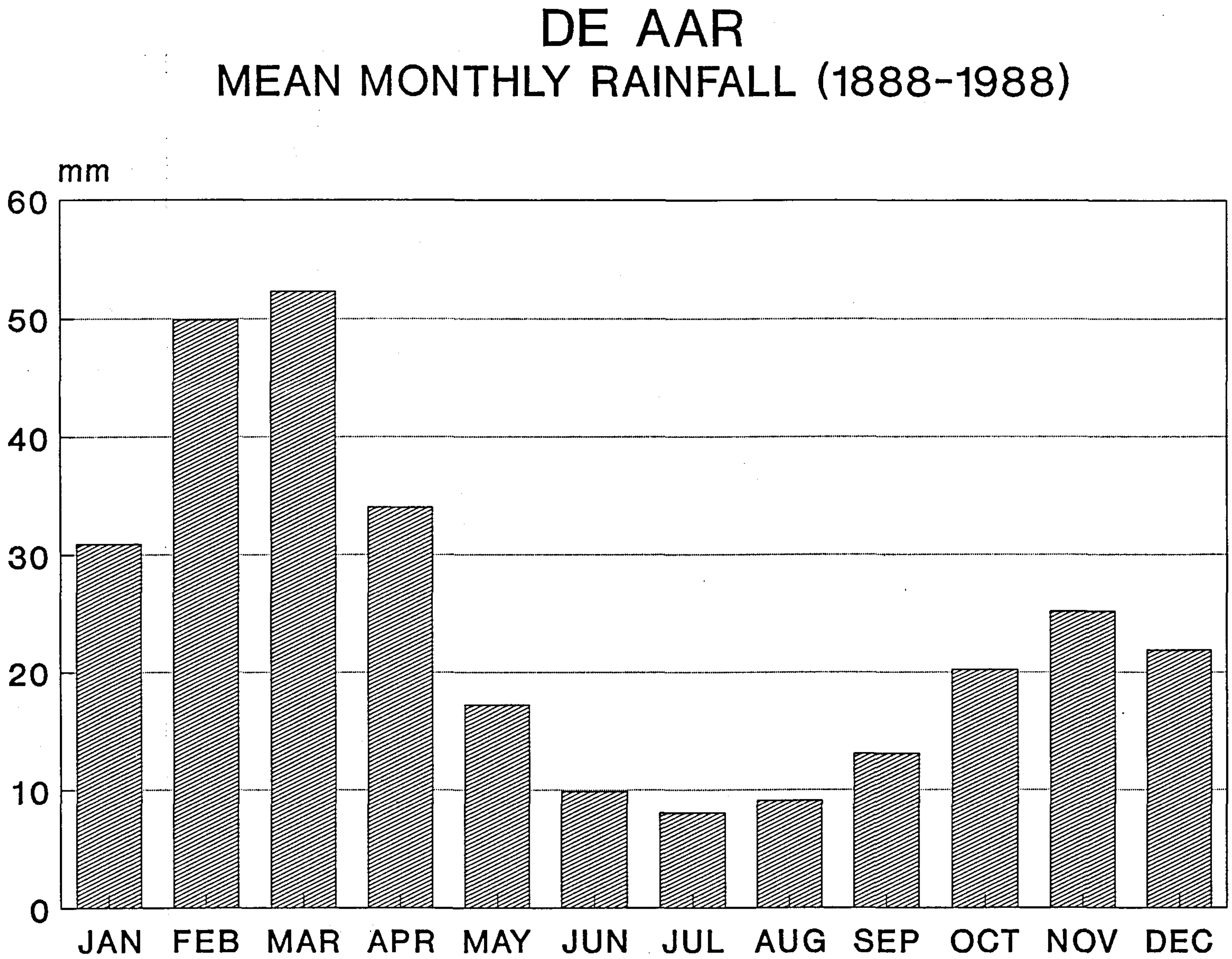


Figure 4.94 Mean monthly rainfall for De Aar.

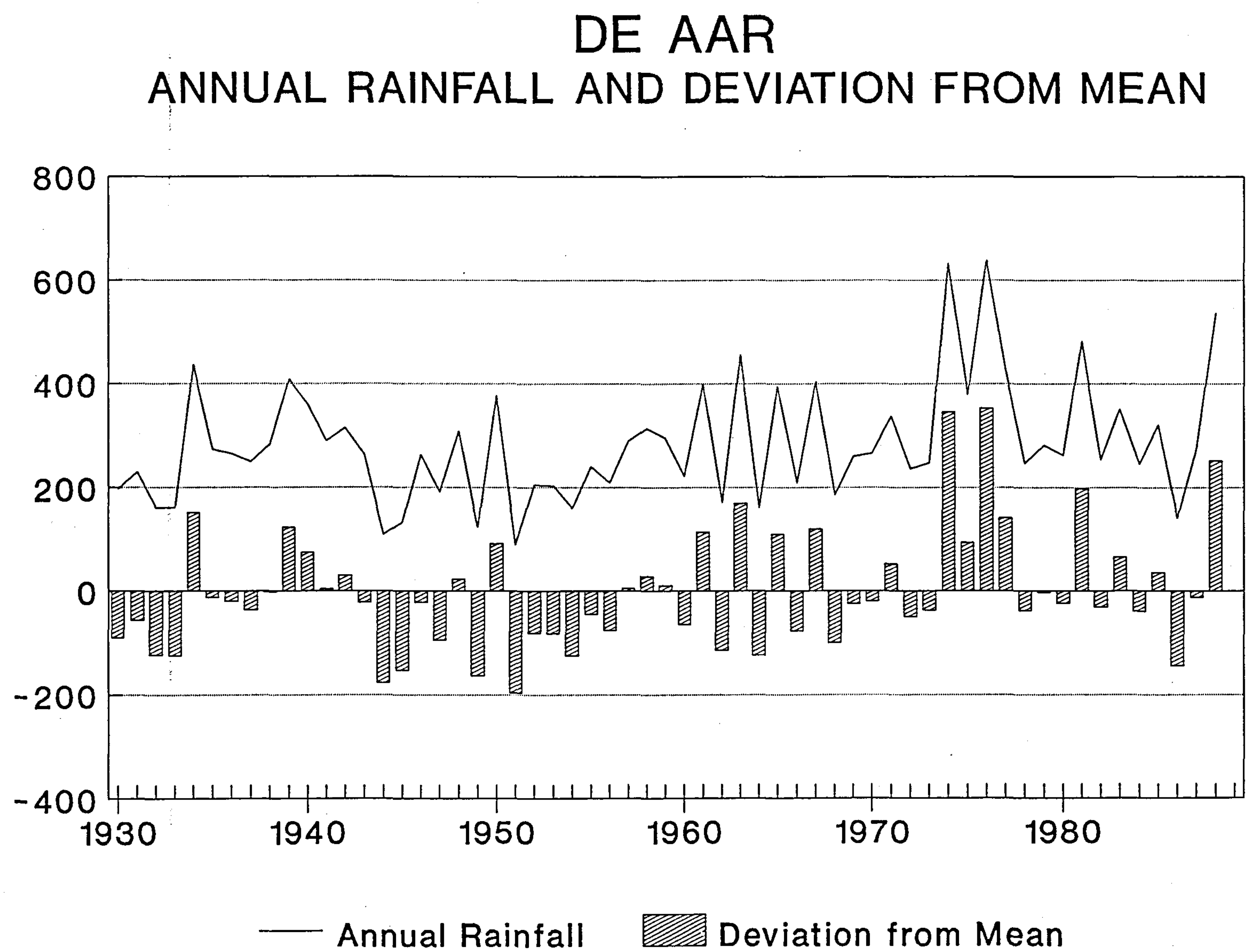


Figure 4.95 Annual rainfall and deviation from the mean.

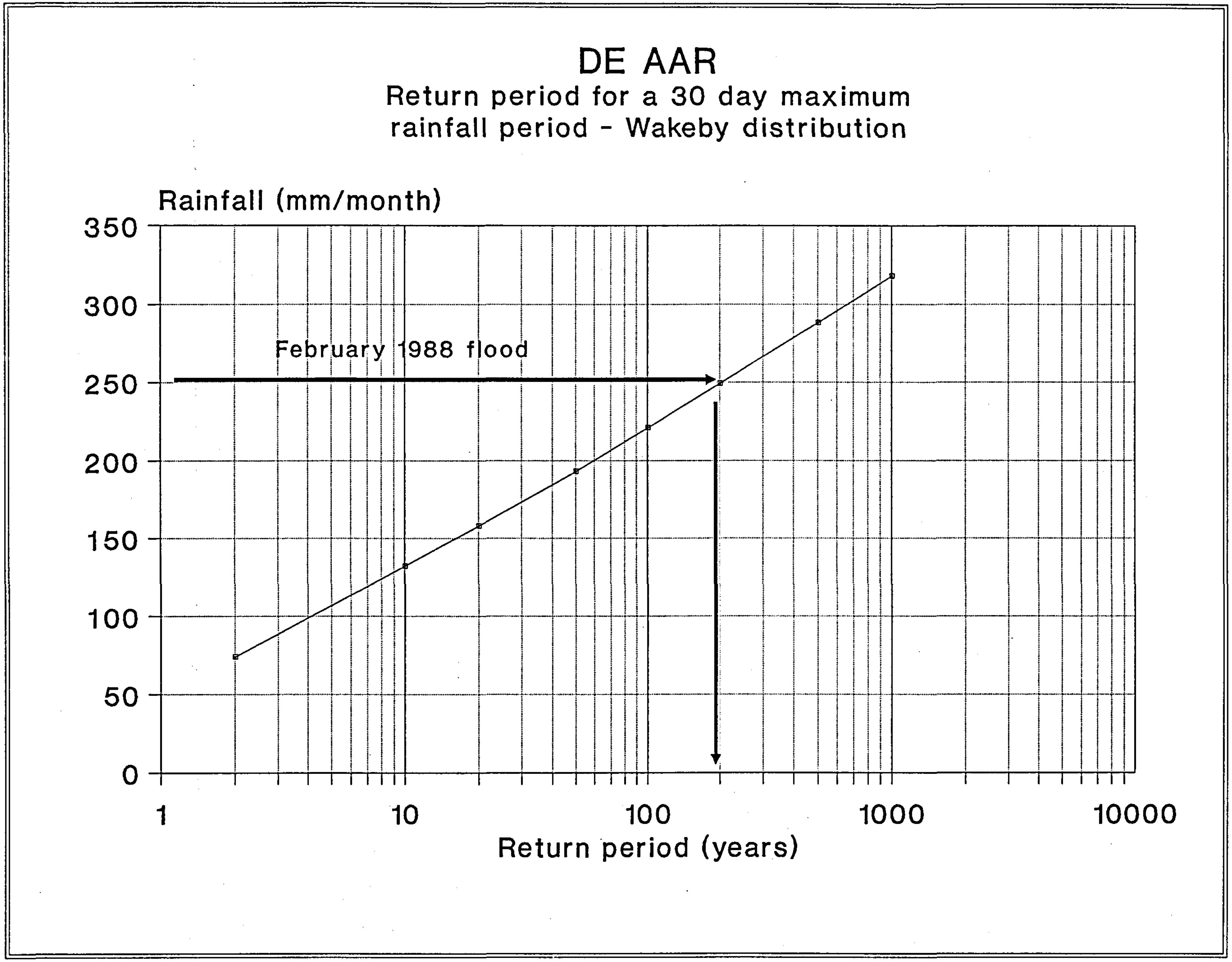


Figure 4.96 Return period of rainfall for a 30-day period at De Aar. It can be seen that the February 1988 flood has a return period of 200 years.

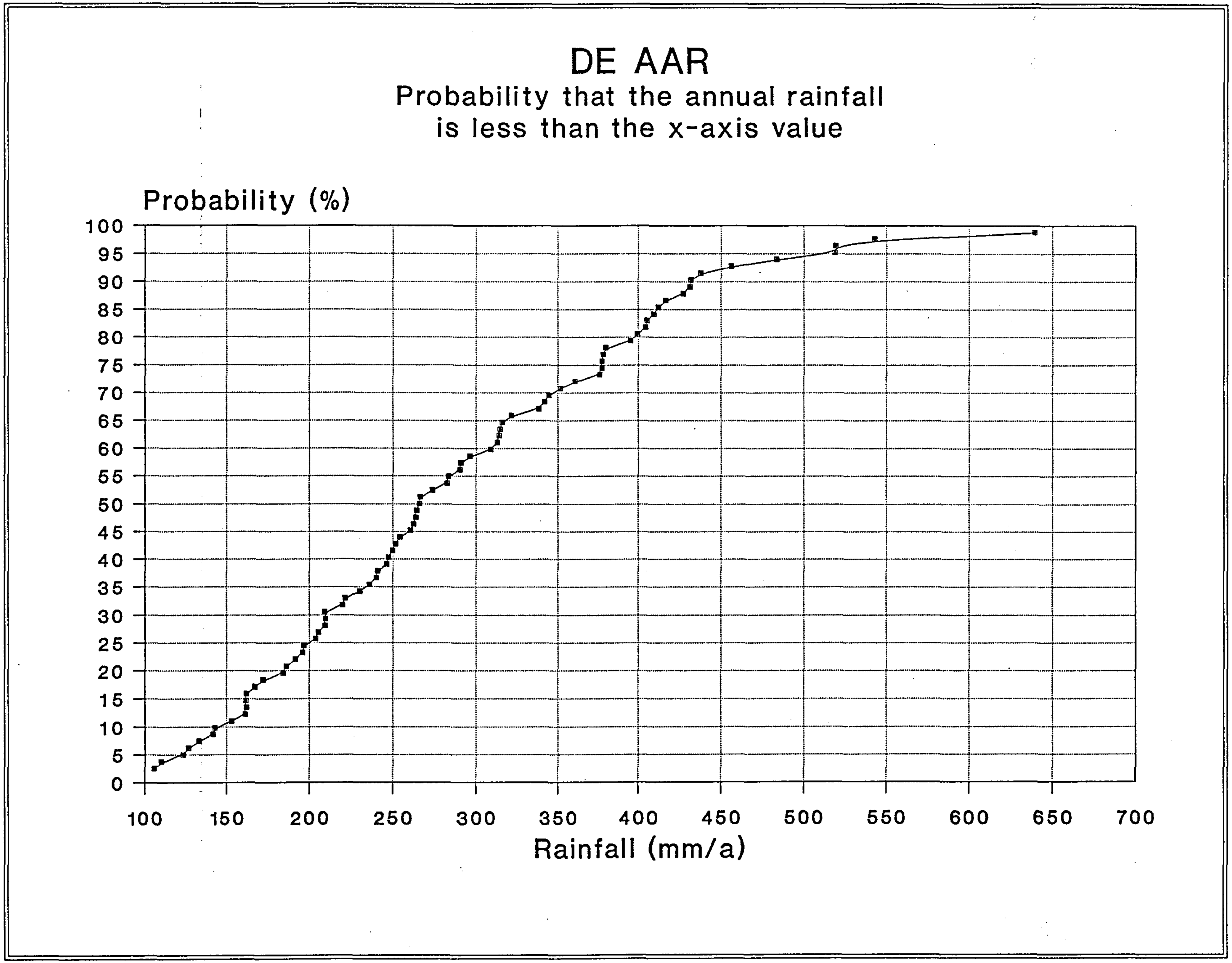


Figure 4.97 Probability that the annual rainfall in De Aar is less than the specified value.

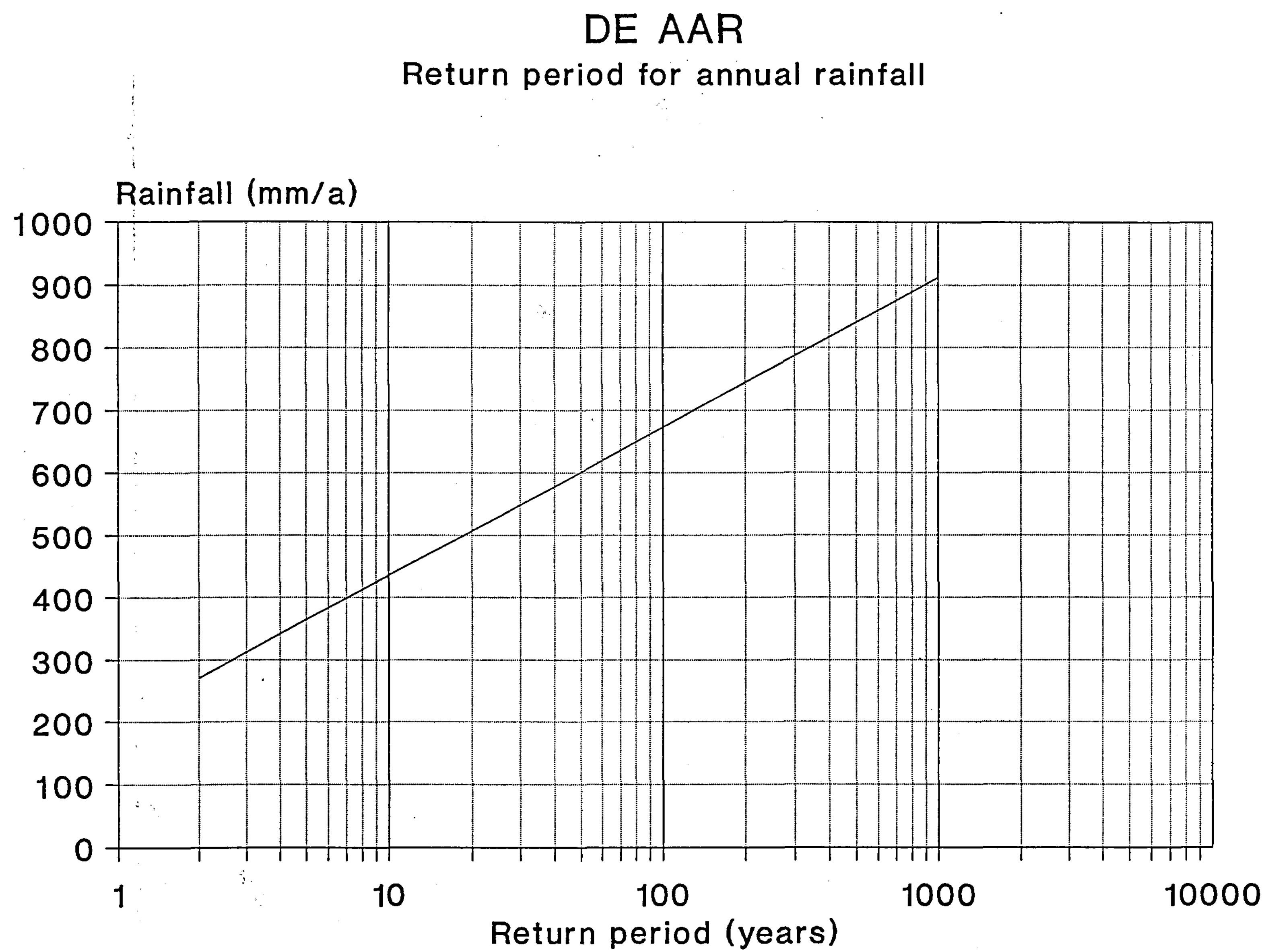


Figure 4.98 Return period for above average rainfall in De Aar.

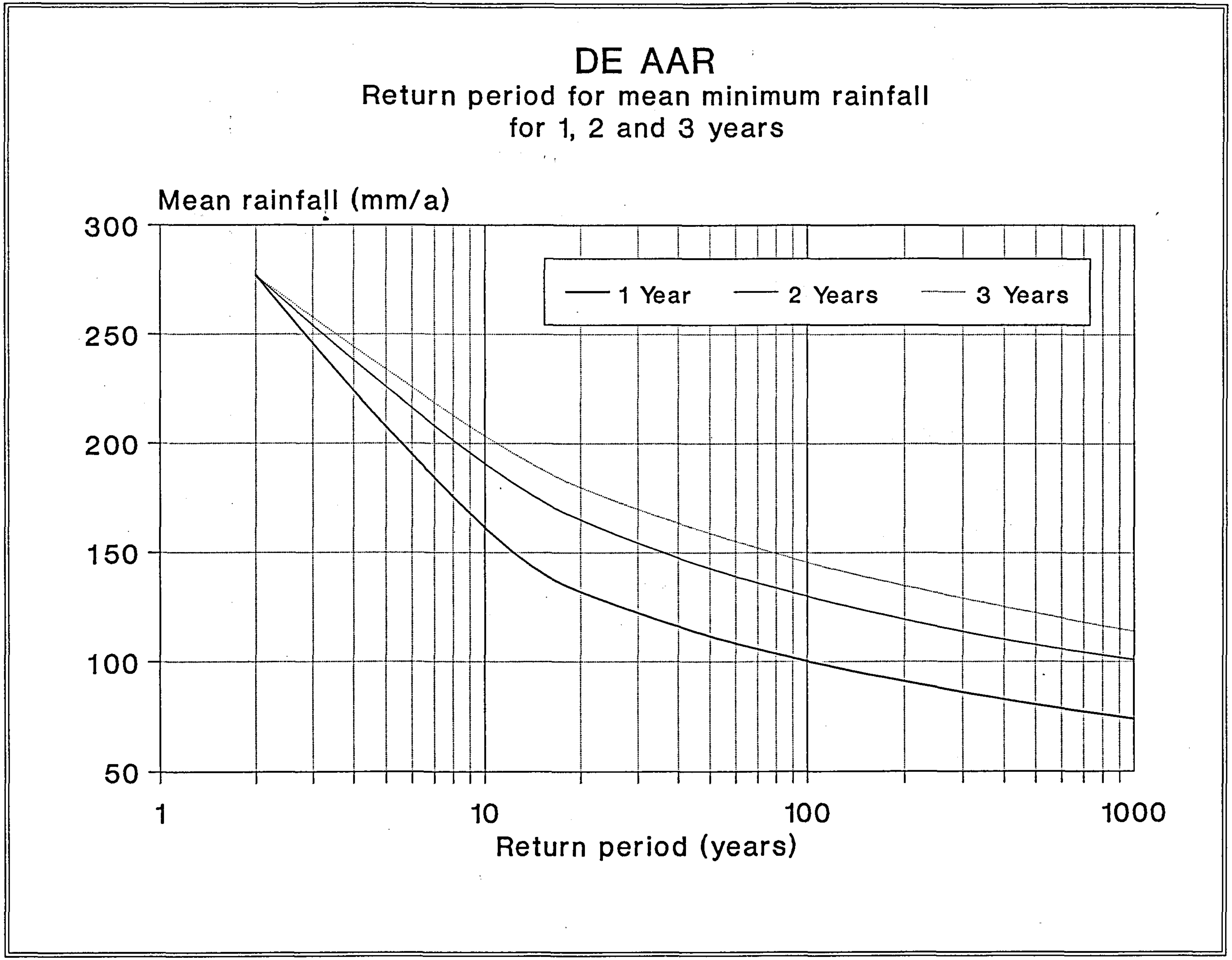


Figure 4.99 Return periods for minimum mean annual rainfall for 1-, 2- and 3-year periods.

period for a specific minimum annual rainfall. For example, it can be seen that De Aar can expect to have an annual rainfall of less than 160 mm on average, once every ten years. To conclude with the probability analyses, the rainfall events of each year were divided into three four-months periods, namely the January-April, May-August and September-December period, and a probability analyses were performed on these data. The results in Figure 4.100 show that the period January to April has a higher recurrence rainfall value than the other periods.

To see if there are any wet and dry cycles in the rainfall pattern of De Aar, the Dyer filter was applied to the annual rainfall values and is shown in Figure 4.101. If compared with the apparently random deviation of rainfall from the mean (Figure 4.95), a much more regular succession of dry and wet cycles is seen. Although not so obvious, a more or less 20-year cycle seems to exist.

4.2.2.2 EVAPOTRANSPIRATION

The potential evapotranspiration was calculated for De Aar according to the Linacre formula described in section 4.1.2.2. Figure 4.102 shows the results. During the duration of the project the potential evapotranspiration varied between approximately 50 and 250 mm/month.

4.2.3 GEOLOGY

4.2.3.1 SOIL

The following soil forms were encountered in the research area:

MAPPING UNIT	SOIL FORM	SOIL SERIES
O	Oakleaf	Letaba
S	Sterkspruit	Swaerskloof
H	Hutton	Shigalo
M	Mispah	Mispah
MK	Mispah	Kalkbank
SE	Sterkspruit	Swaerskloof (eroded phase)

The results of the soil survey are shown on a map at a scale of 1 : 20 000 (see Figure 4.103). The mapped area is in the land type Ae, i.e. red-yellow apedal, freely drained soils, Fb (e.g. Mispah) and Ib (rock areas with miscellaneous soils). The mountains and higher lying parts are covered with a thin layer of soil of the MISPAH form.

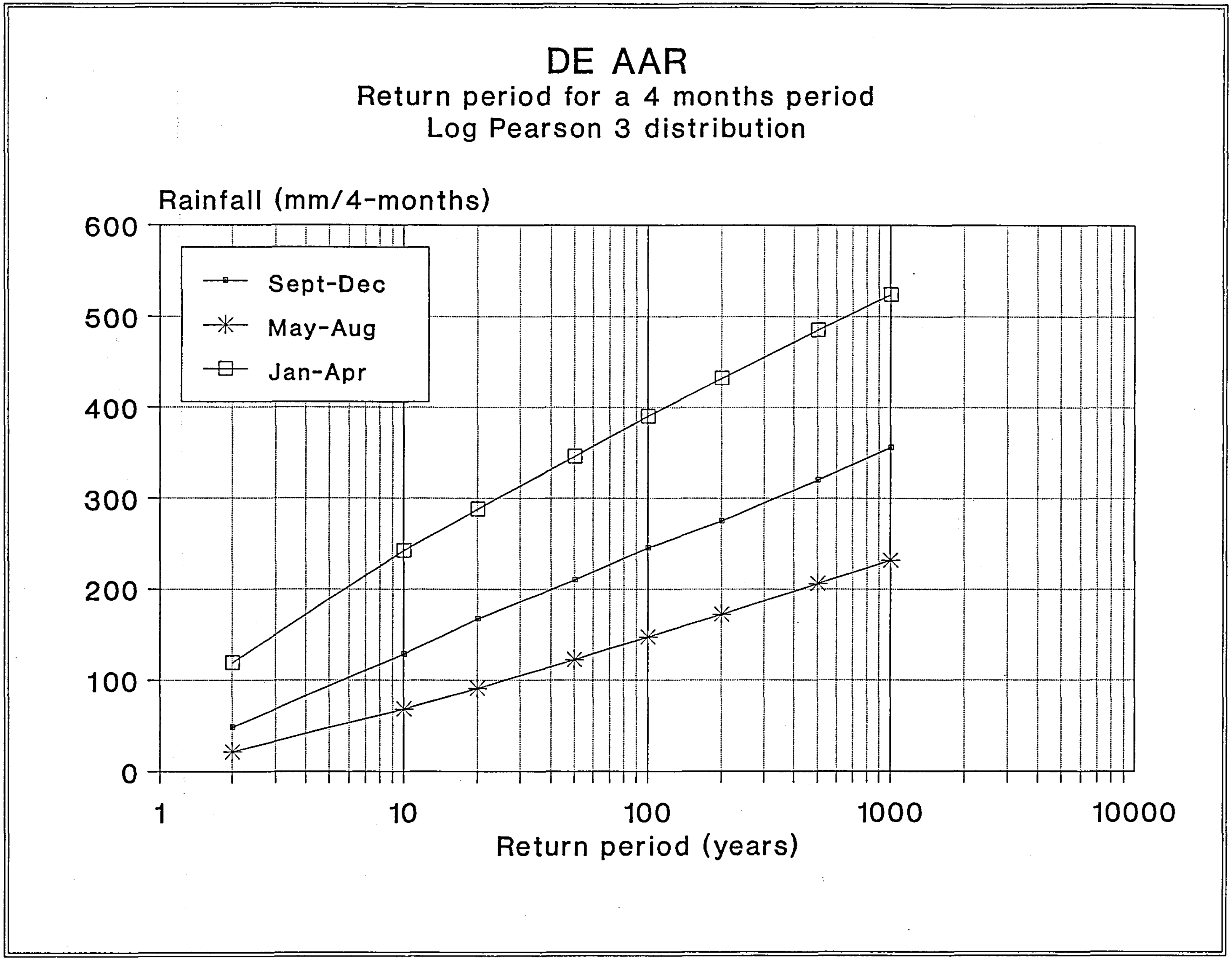


Figure 4.100 Return period for four-months' rainfall in De Aar.

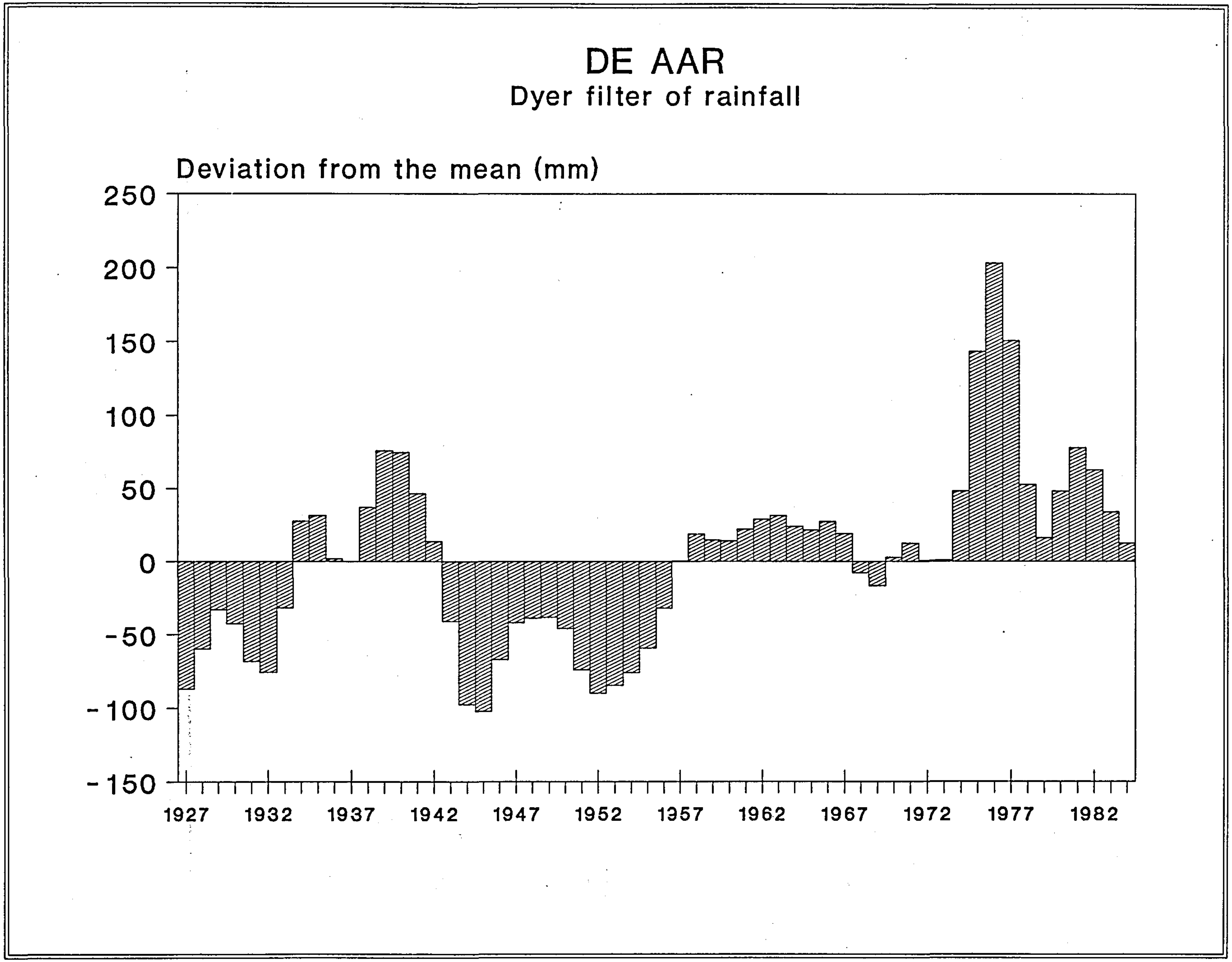


Figure 4.101 Dyer formula applied to annual rainfall of De Aar.

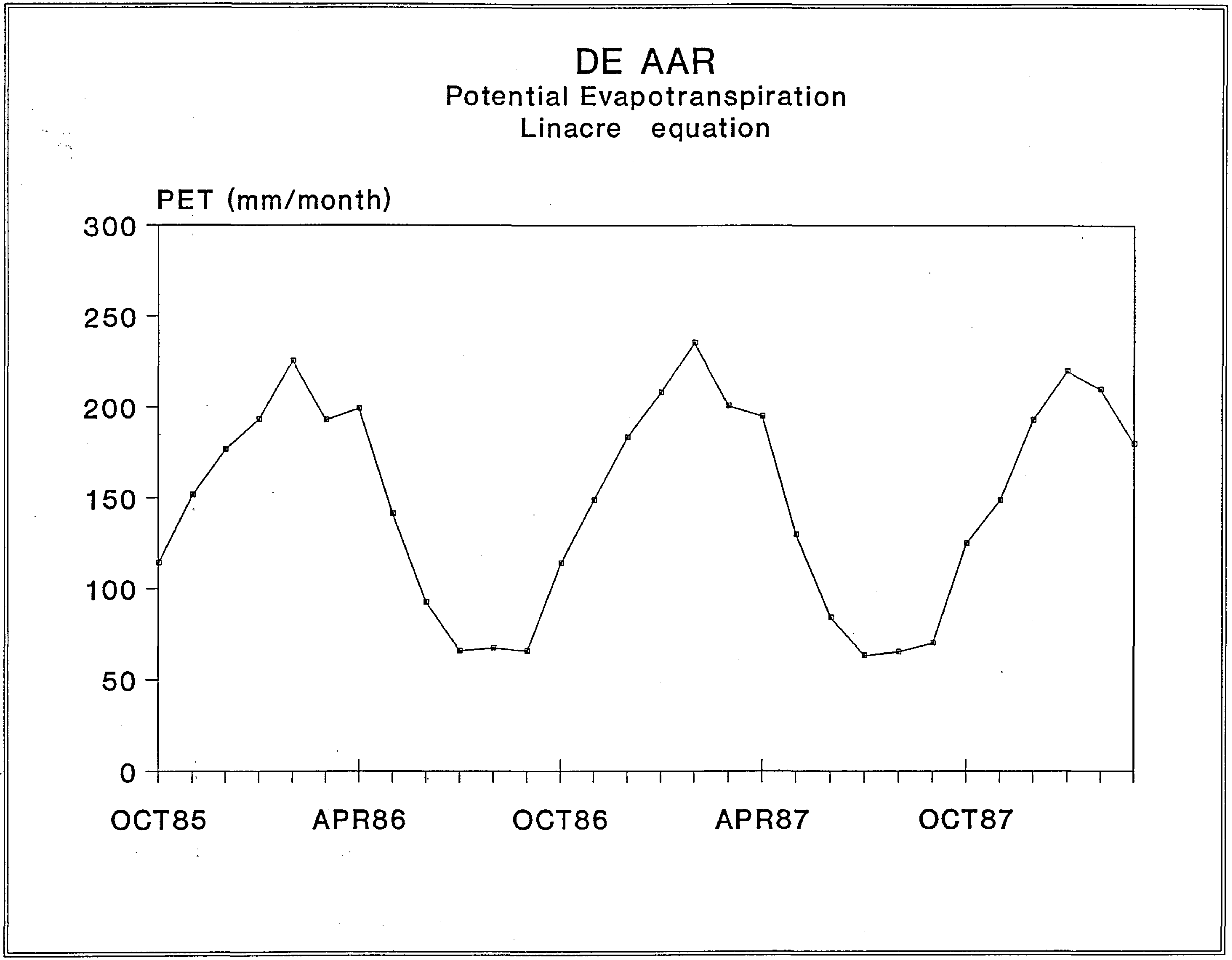


Figure 4.102 Potential evaporation of the De Aär aquifer for the study period.

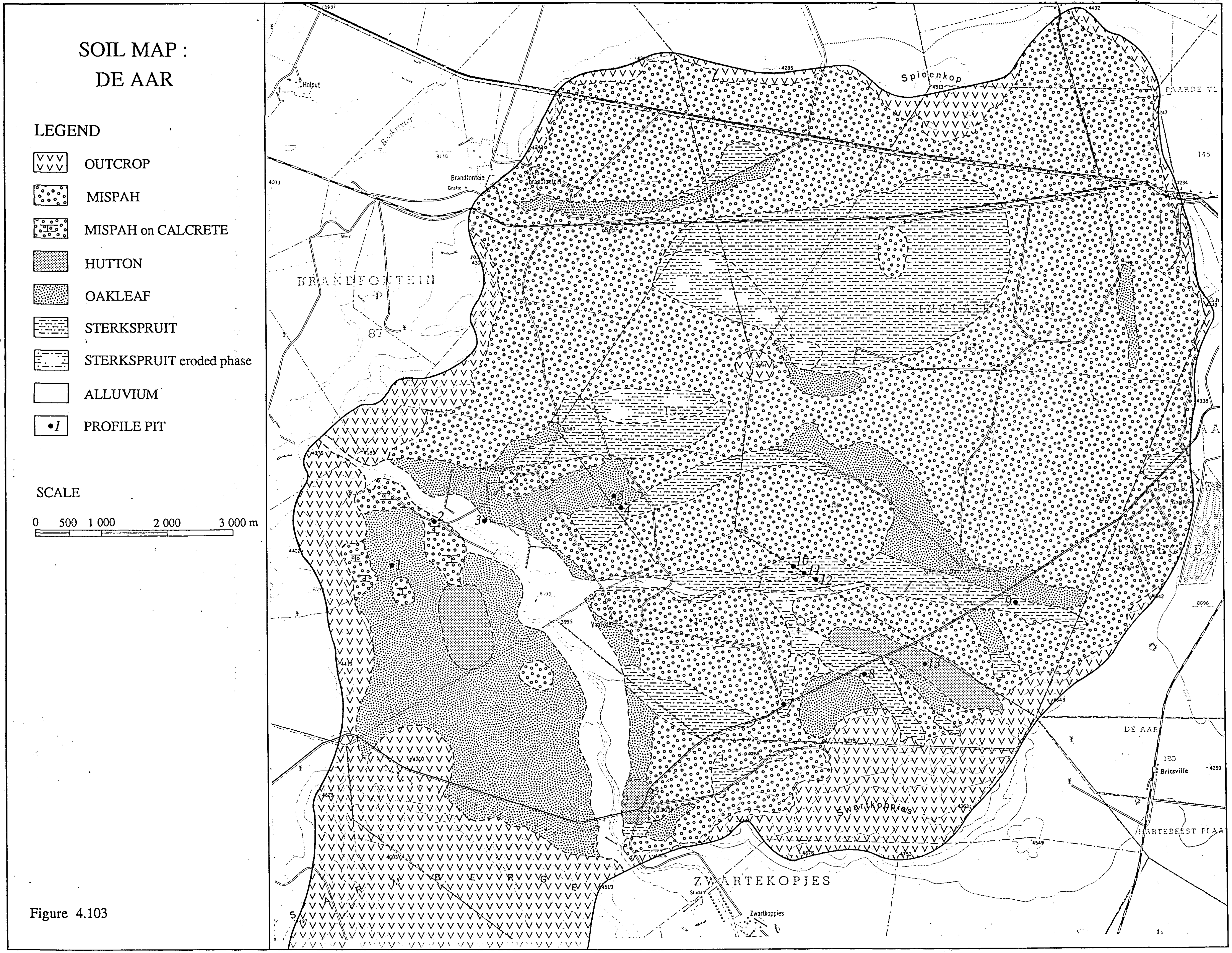

SOIL MAP :
DE AAR
LEGEND
OUTCROP
MISPAH
MISPAH on CALCRETE
HUTTON
OAKLEAF
STERKSPRUIT
STERKSPRUIT eroded phase
ALLUVIUM
•1 PROFILE PIT
SCALE
0 500 1 000 2 000 3 000 m
Holput
Spioenkop
Brandfontein
BRANDFONTEIN
87
DE AAR
Britsville
ZWARTEKOPJES
Zwartkoppies

Figure 4.103

ANNEXURE E

Analysis method for Total Fe, SG and Free Acid

ANNEXURE E

ANALYSIS METHOD FOR FERRIC CHLORIDE/SULPHATE SOLUTION

1. PROCEDURE FOR TOTAL Fe CONTENT

A. REAGENTS

(i) 0,10N POTASSIUMDICHROMATE SOLUTION

Dissolve 4,903 grammes of A.R. potassium dichromate in 200ml of distilled water and dilute to one litre with distilled water in a volumetric flask.

(ii) CONCENTRATED HYDROCHLORIC ACID C.P. (300 g/l HCI min.)

(iii) MIXED ACIDS

To 700 ml of cold distilled water add 100ml of concentrated sulphuric acid with caution. When ambient temperature add 100ml of concentrated phosphoric acid and dilute the mixture to one litre.

(iv) STANNOUS CHLORIDE SOLUTION 6% m/v

Dissolve 60g of stannous chloride dihydrate C.P. in 300ml of warm concentrated hydrochloric acid. Cool and dilute to one litre with distilled water.

(v) SATURATED MERCURIC CHLORIDE SOLUTION C.P. 7%

Dissolve 70g of mercuric chloride in 1 000ml of hot distilled water.

(vi) SODIUM DIPHENYLAMINE - 4 - SULPHONATE 2g/100ml, (SAARCHEM CAT. NO. 126200)

B. PROCEDURE

Weigh out 25ml of ferric solution accurately and transfer to a 500ml volumetric flask. Dilute to the mark with water. Invert sufficiently to ensure thorough mixing.

Pipette a 25ml aliquot into a 500ml wide mouth Erlenmeyer flask. Add 25ml of concentrated hydrochloric acid and heat to boil. Add 6% stannous chloride solution dropwise until the yellow colour is discharged. Add a few drops in excess. Cool rapidly and add 10ml saturated mercuric chloride solution and let stand for 10 minutes. Add 30ml of mixed acids and dilute the contents of the flask to 200-300ml with distilled water. Add 2,0ml of Sodium diphenylamine sulphonate indicator solution and titrate to a permanent purple end point with 0,10N Po tassium dichromate solution.

STERKSPRUIT form is found along the Brak River and its tributaries, alternating with OAKLEAF form along the river. Small patches of DUNDEE are found as well as soil of HUTTON form. The texture of the alluvial soil varies considerably.

The soil profile descriptions and analytical data are found in the Appendix.

4.2.3.2 HARD ROCK

The area has, in the past, been geologically mapped. In the early seventies, De Bruin did a photogeological study of the surrounding of De Aar during a hydrogeological study of the Geological Survey. On contract for the Geological Survey, Lemmer has mapped the Britstown area, including the research area, in 1976-77. The information has been transferred to 1:50 000 topographical maps and is available on Open File from the Geological Survey.

The study area lies next to the boundary between the Beaufort- and Ecca Group which is found slightly south of De Aar and has a WSW strike. It is enclosed by a prominent ring-shaped dolerite intrusion, which has been cut through by the Brak River where it enters the area in the south, where it leaves it in the west and in the north-west by a small tributary of the Brak River (see Figure 4.104). The ring dyke has a diameter of approximately 12 km and dips towards the centre at an angle varying between 20 an 60°. In the south-west, it becomes an extended sill with a thickness of approximately 40 m. Sills within the ring structure have caused an undulating relief, rising slightly above the surrounding plain underlain by Karoo sediments. A few jointed dolerite hillocks have originated because of differentiated weathering. At a number of places, the coarse-grained sills have been intersected by a few metres wide fine-grained dykes.

The sediments encountered belong to the Tierberg Shale Formation of the Ecca Group. This formation consists essentially of carbonaceous black to dark grey shale which indicates a marine environment. Outcrops are noticed in numerous dongas and shale can be seen in many *meerkat* and other animal burrows, proving that over large areas the soil cover is thin. Along dolerite intrusions, the shale is baked. Quite frequently, well-developed calcite-filled joint systems can be observed and over wide areas calcrete develops near the surface along joints and bedding planes of the shale.

In the higher lying areas near Zwartekopjes and Brandfontein, grey shale and light sandstone of the Lower Beaufort occur.

In the flood plain of the Brak River, alluvial sediments and calcrete which may exceed 10 m in thickness do occur. The alluvium consists of unconsolidated lenses of comparatively small horizontal and vertical extension of clay, silt, sand, gravel and boulders, deposited in a ± 700 m wide Pleistocene channel. Below the alluvium, the Tierberg Shale is weathered to a depth of between 1 and 7 m.

4.2.4 GEOHYDROLOGY

The success rate of boreholes drilled in the proximity of De Aar is high. De Bruin & Vegter (1971) report yields in excess of 0,45 m^3/h for 91,5 per cent of the holes drilled. Ground water is encountered in the alluvial deposits and secondarily in fractures or weathered shale and dolerite. Coarse sand and gravel lenses are the main aquifers.

A slightly different picture is obtained, if the 56 boreholes drilled for the present study are considered. These were selected outside the alluvium and rather randomly to obtain a representative overview over the waterbearing properties of the Karoo strata in the study area. The yields of these 56 boreholes varied between 0 and 60 m^3/h with 25 per cent of the holes dry and 52 per cent yielding more than 0,45 m^3/h. The median yield was 0,16 l/s (0,576 m^3/h). Nine holes (16 %) had yields of more than 3,6 m^3/h.

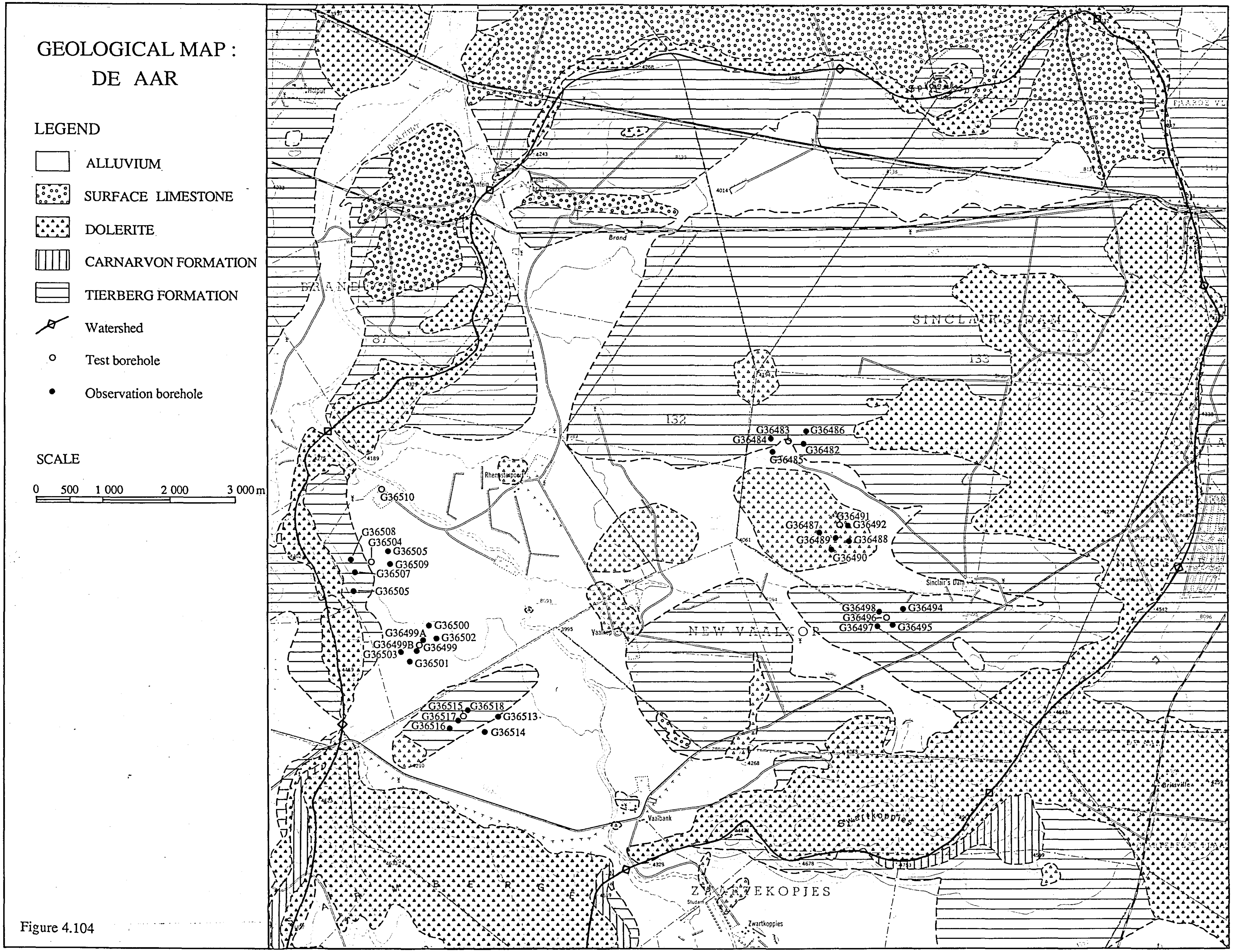
Previous geohydrological investigations included a survey by De Bruin & Vegter (1971), which was followed by a geophysical investigation of the alluvium of the Brak River with approximately 1 000 resistivity depth probes. Subsequently, 107 boreholes were drilled in 1972 on the farms Renosterpoort, Vaalbank and Brandfontein, followed by test pumping the year thereafter.

4.2.5 MEASURING NETWORK

4.2.5.1 UNSATURATED ZONE

Over much of the area, the thickness of the soil cover is considerably less than in Dewetsdorp. Outcrops of (weathered) Karoo sediments are common. For the determination of the recharge through the soil cover, twenty-two 0,7 to 2,0 m deep neutron probe access tubes were installed in the different soils, mainly in the centre of the study area, allowing measurements to a depth of between 0,5 and 1,7 m. The position of the neutron probe access tubes is shown in Figure 4.105.

During the investigation, neutron readings were taken in these tubes more than 20 times. Graphs, showing the results are contained in the Appendix. From these graphs, it can be deducted that during normal rainfall years, no or very little water reaches the water table via the soil matrix: only neutron probe access tubes 4, 14 and 15 responded



GEOLOGICAL MAP : DE AAR
LEGEND
ALLUVIUM
SURFACE LIMESTONE
DOLERITE
CARNARVON FORMATION
TIERBERG FORMATION
Watershed
Test borehole
Observation borehole
SCALE
0 500 1 000 2 000 3 000 m
SINCLAIR'S DAM
NEW VAALKOP
ZWARTKOPJES
Swartkoppies
Vaalbank
Brand
Sinclair's Dam
G36483
G36486
G36484
G36482
G36485
G36491
G36487
G36492
G36489
G36488
G36490
G36498
G36494
G36496
G36497
G36495
G36510
G36508
G36504
G36505
G36509
G36507
G36505
G36500
G36499A
G36502
G36499B
G36499
G36503
G36501
G36515
G36518
G36517
G36513
G36516
G36514

Figure 4.104

NEUTRON PROBE ACCESS TUBES : DE AAR

SCALE

0 500 1 000 2 000 3 000 m

LEGEND

- □ Neutron Probe Access Tube
- △ Rain Meter
- ■ Weather Station
- ▲ Rainfall Recorder

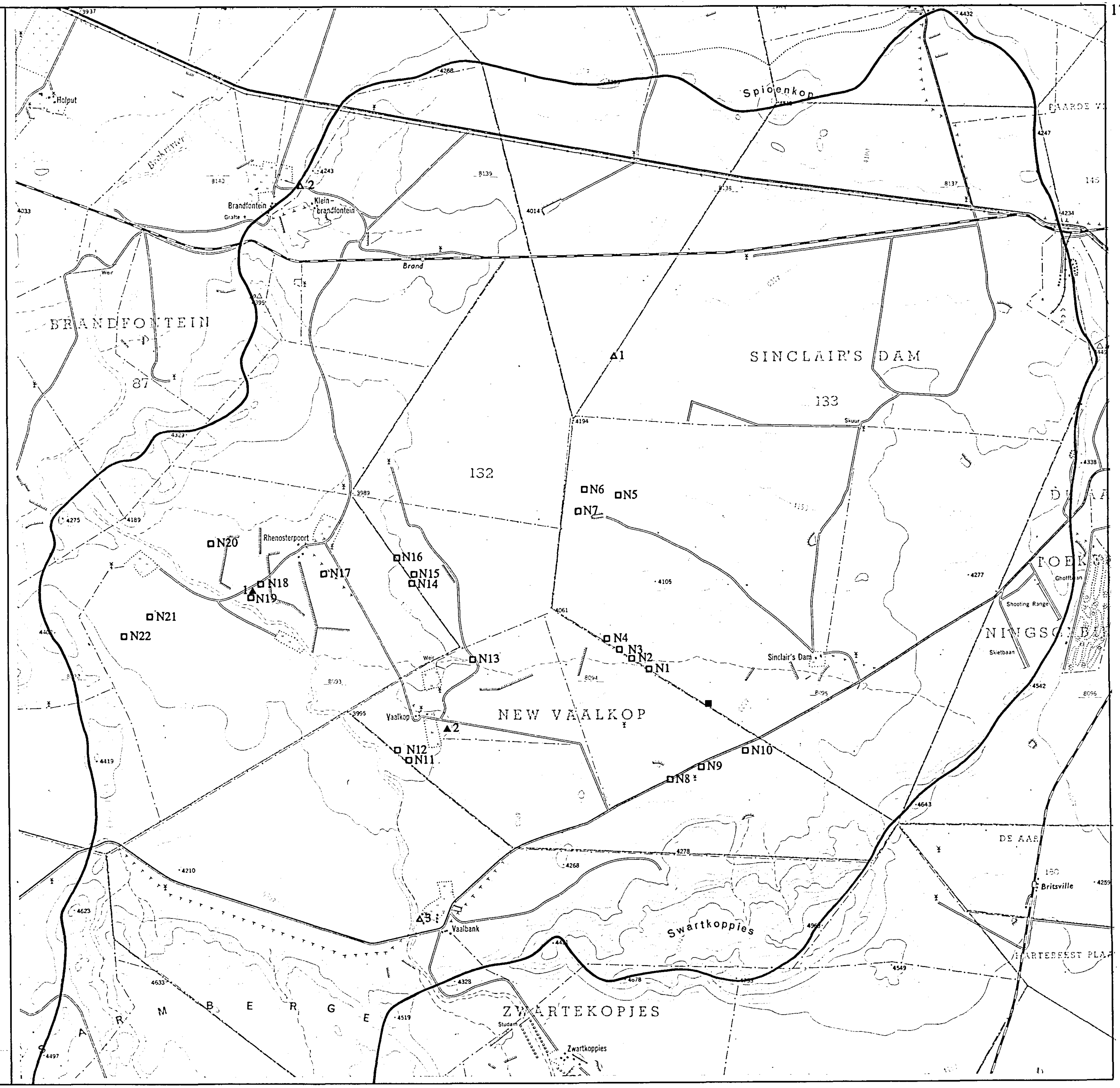


Figure 4.105

Soil	Neutron probe access tube no.
Alluvium	11, 12, 19
O	5, 8, 16, 17, 18, 20, 21, 22
S	1, 2, 3, 4, 9, 13, 14, 15
H	10
M	6, 7

to the 1985 rainfalls and none in 1986 and 1987. In 1988, the situation was different, however, when 11 of the 22 tubes displayed considerably higher neutron counts within 3 weeks after the rains started. The responding tubes were numbers 1, 3, 8, 10, 13, 14, 15, 17, 19, 21 and 22.

4.2.5.2 SATURATED ZONE

In order to detect and determine the depth of dolerite sills or possible waterbearing strata, approximately 50 electrical soundings, to a maximum depth of 200 m (O - A = 200 m), have been carried out in two areas; in the vicinity of the Army Camp and south-west of the Brak River. Guided by the results of these soundings, 56 borehole sites were selected. Exploration boreholes G 36469 to G 36518 (including G 36483 A to D and G36499 A and B) were drilled in the first quarter of 1985 by the Drilling Section of the Department of Water Affairs. The depth of the boreholes varied between 13 and 127 m with a total depth of 2932 m. The position of these boreholes is shown on Figure 4.106, while the logs of these boreholes are contained in the Appendix.

Ten of these boreholes were geophysically logged by Mr. Venter from the Directorate of Geohydrology: G 36474, 36483, 36489, 35491, 36496, 36499 and 36499A, 36504, 36510 and 36515.

Boreholes drilled during the investigation in the early seventies were also used for monitoring purposes. Some of them had been inundated during the 1974 floods and silt and sand were washed into them. They had to be cleaned. Some of the holes on Brandfontein have not been recovered. Private holes in the area were also observed.

Amongst others, basic information about 252 holes was entered into the data base, geological information about 101 holes, water levels from 97 holes and chemical data from 48 holes.

Yields

Yields of the forty successful boreholes in Karoo Formation drilled in 1985 are generally rather weak with a median yield of 0,34 l/s (see Figure 4.107). Twenty per

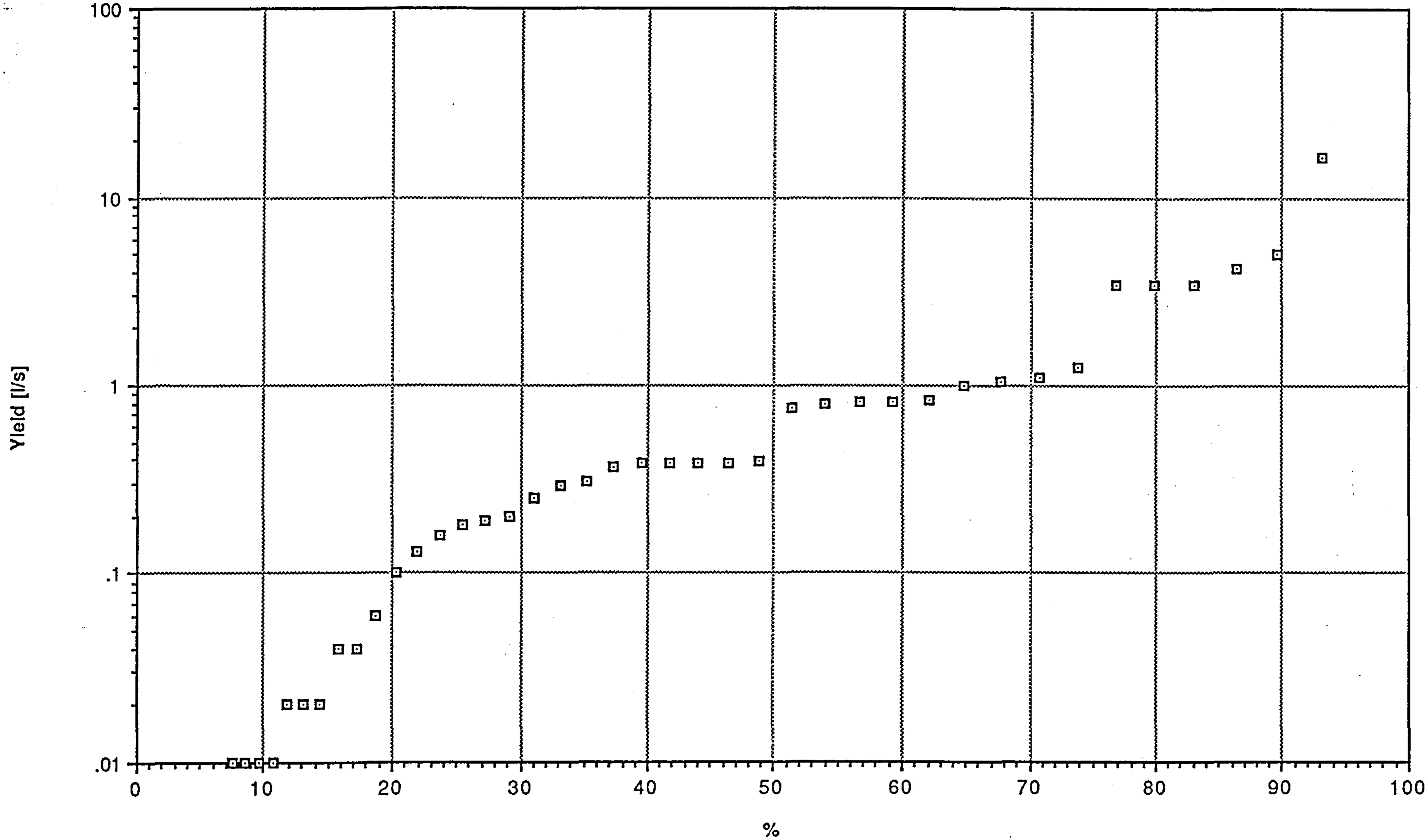


Figure 4.107 Yield distribution of De Aar exploration boreholes.

cent of these holes yield more than 1 l/s and only one hole yields more than 5 l/s (i.e. 16,5 l/s). Much higher yields are obtained from the production holes in the alluvium with constant yields ranging between 6,25 and 14,16 l/s (Von Hoyer 1976).

Water levels

The water-level observation network consisted of 97 boreholes. Minimum water levels observed during the study period ranged between 0,80 and 31,40 m, while the maxima varied between 2,60 and 56 m. The minimum and maximum water levels for the various boreholes and the dates on which these levels were observed, are shown in Table 4.6. They are graphically displayed in Figures 4.108 and 4.109. Data are sorted according to the date on which the minimum water level was observed. Table 4.6 also lists the time range for which observations are available. It is obvious that, with the exception of the first seven holes for which no observation is available after the heavy rains of February and March, 1988, the highest water levels were observed between approximately one week and four months after the heavy rainfalls. With hardly any exception, the lowest levels were recorded after the poor 1986/87 season and before the good rains of 1988.

Sixteen of the exploration holes were equipped with recorders and three of them were removed during the course of the project. The respective recorders and the type of response obtained are summed up hereunder:

G 36473: Good delayed response to major rainfall (range ± 7 m).
G 36474: Sharp response to rainfall, fast recession (range ± 2 m).
G 36480: Immediate sharp response to rainfall, v. fast recession (range ± 4 m; dry ?).
G 36483: Good delayed response to rainfall (range ± 10 m).
G 36489: Good delayed response to rainfall (range ± 10 m).
G 36496: Good delayed response to rainfall (range ± 10 m).
G 36499: Oscillating (tides), otherwise hardly any water-level change (range ± 0,5 m).
G 36501: Little delayed response to major rainfall (range ± 1,5 m). Poor record.
G 36502: Little response. Float not moving freely (range ± 1,5 m).
G 36504: Little response. Poor record (range ± 2 m).
G 36506: Good but little response to rainfall events (range ± 2 m).
G 36507: Good but little response to rainfall. Recorder removed (range ±2.5 m).
G 36509: Good but little response to rainfall events (range ± 2 m).
G 36514: Very slow response to major rainfall events only (range ± 0,5 m).
G 36515: Responds to exceptional rainfall. Recorder removed. (Range ± 1,5 m).
G 36516: Responds to exceptional rainfall. Recorder removed. (Range ± 2,5 m).

Number	Range	Date	Minimum	Date	Maximum	
36470	8502-8707	19860806	26.09	19870716	27.40	1985-1987
00054	8612-8612	19861210	4.51			Only 1 record
00063	8612-8703	19861210	6.86	19870304	7.51	1985-1987
00050	8612-8705	19861210	7.31	19870332	7.60	1986-1987
00080	8705	19870508	13.88			Only 1 record
36471	8705	19870508	27.30			Only 1 record
00075	8612-8710	19871006	11.30	19870902	11.82	1986-1987
36480	8502-8806	19880301	8.41	19880213	13.84	
36505	8503-8806	19880303	5.08	19870902	7.08	
00056	8612-8806	19880304	1.76	19871102	7.18	
00001	8612-8803	19880317	10.42	19870304	16.88	
00004	8612-8806	19880317	9.51	19870323	22.60	
36476	8511-8806	19880317	1.71	19870508	4.55	
00002	8612-8803	19880317	17.85	19870902	24.84	
00053	8701-8806	19880317	1.20	19871102	5.31	
00072	8701-8806	19880317	10.48	19871102	28.88	
00073	8612-8806	19880317	6.14	19880128	9.22	
36475	8502-8806	19880317	2.28	19880128	6.09	
00070	8612-8806	19880317	12.72	19880630	17.04	1986-1988
36512	8503-8806	19880323	2.03	19870508	4.64	
00007	8612-8803	19880323	3.85	19870804	4.95	
36472	8502-8806	19880323	16.10	19870820	24.53	
00034	8612-8806	19880323	3.81	19870902	5.69	
00036	8701-8803	19880323	18.98	19870902	21.44	
00077	8612-8806	19880323	15.55	19870902	50.32	
36490	8503-8807	19880323	10.12	19870916	20.38	
36487	8503-8807	19880323	15.66	19871006	22.96	
00076	8612-8806	19880323	20.02	19871006	56.00	
00037	8612-8806	19880323	7.23	19871130	9.86	
36518	8508-8803	19880323	11.87	19880128	13.34	
36473	8511-8710	19880419	31.40	19870916	37.66	
36474	8511-8806	19880419	2.76	19880212	5.37	
36504	8507-8806	19880424	4.69	19861020	6.70	
00061	8612-8806	19880429	12.48	19870508	17.32	
36479	8502-8806	19880429	5.91	19870804	10.53	
36511	8503-8806	19880429	1.54	19870916	3.32	
36510	8503-8806	19880429	2.00	19870916	3.65	
36508	8507-8806	19880429	4.59	19870916	6.37	
36478	8501-8806	19880429	2.59	19870916	6.69	
36507	8507-8806	19880429	5.76	19870916	7.54	
36513	8503-8806	19880429	8.94	19880128	11.06	
36515	8507-8806	19880429	12.20	19880129	13.73	
36493	8503-8807	19880502	10.11	19870902	16.65	
36509	8507-8806	19880504	6.34	19870917	7.95	
36506	8503-8806	19880511	17.65	19871015	20.68	
36502	8509-8806	19880530	8.51	19880306	13.23	
00051	8612-8806	19880603	0.80	19880118	2.60	
36514	8505-8806	19880603	12.49	19880128	13.24	
36496	8507-8807	19880606	13.78	19870907	22.58	

Table 4.6 Highest and lowest water levels recorded.

Number	Range	Date	Minimum	Date	Maximum	
36483	8505-8807	19880606	5.48	19871003	14.98	
36524	8508-8806	19880616	8.84	19870716	9.50	
36494	8509-8807	19880616	14.57	19870902	21.77	
36495	8507-8807	19880616	15.25	19870902	23.73	
36498	8503-8807	19880616	12.85	19870921	21.40	
36484	8503-8807	19880616	6.49	19871006	13.87	
36521	8812-8807	19880616	5.45	19871006	14.87	
36519	8612-8806	19880616	5.76	19871006	15.17	
36486	8503-8807	19880616	9.25	19871006	19.06	
36522	8505-8807	19880616	5.56	19871006	14.93	
36485	8503-8807	19880616	5.53	19880104	21.85	
36489	8509-8807	19880621	10.84	19871016	21.00	
36499	8508-8806	19880627	8.26	19870527	9.12	
00069	8701-8806	19880630	4.02	19870106	9.01	
00078	8612-8806	19880630	11.54	19870204	24.41	
00039	8612-8806	19880630	7.46	19870304	9.13	4 records
00032	8612-8806	19880630	2.76	19870508	4.80	
00059	8612-8806	19880630	2.91	19870508	10.60	
00047	8612-8705	19880630	5.95	19870508	15.65	
36523	8508-8806	19880630	8.18	19870716	8.77	
00058	8612-8806	19880630	2.90	19870804	10.34	
36469	8502-8806	19880630	21 29	19870820	26.95	
00074	8612-8806	19880630	5.12	19870902	6.42	
00003	8612-8806	19880630	7.45	19870902	15.91	
00045	8612-8806	19880630	18.57	19870902	30.77	
36501	8509-8806	19880630	9.80	19871004	11.01	
36500	8612-8806	19880630	10.19	19871006	10.69	
00079	8612-8806	19880630	7.63	19871102	15.57	
00066	8612-8806	19880630	8.37	19871102	17.94	
00057	8612-8806	19880630	12.07	19871130	12.92	
00055	8612-8806	19880630	9.19	19871130	13.22	
36516	8511-8806	19880630	14.38	19871130	16.51	
36477	8502-8806	19880630	2.13	19871218	5.25	
36503	8511-8806	19880630	9.67	19880104	10.64	
00067	8612-8806	19880630	7.46	19880104	11.06	
00048	8612-8806	19880630	10.95	19880104	17.82	
00038	8612-8806	19880630	6.74	19880128	8.01	
36517	8503-8806	19880630	12.48	19880128	14.19	
00006	8612-8806	19880630	8.58	19880128	16.63	
00071	8612-8806	19880630	21.65	19880128	27.54	
00046	8612-8806	19880630	5.22	19880129	8.41	
36488	8503-8807	19880701	12.87	19870902	20.02	
36492	8503-8807	19880701	13.92	19871006	22.48	
36491	8503-8807	19880711	13.72	19870716	22.50	
36481	8501-8807	19880711	6.70	19870916	17.97	
36497	8507-8807	19880711	13.14	19870916	20.73	
36482	8501-8807	19880711	7.39	19871006	17.29	
36520	8805-8807	19880711	5.98	19871009	15.27	

Table 4.6. (continued): Highest and lowest water levels recorded.

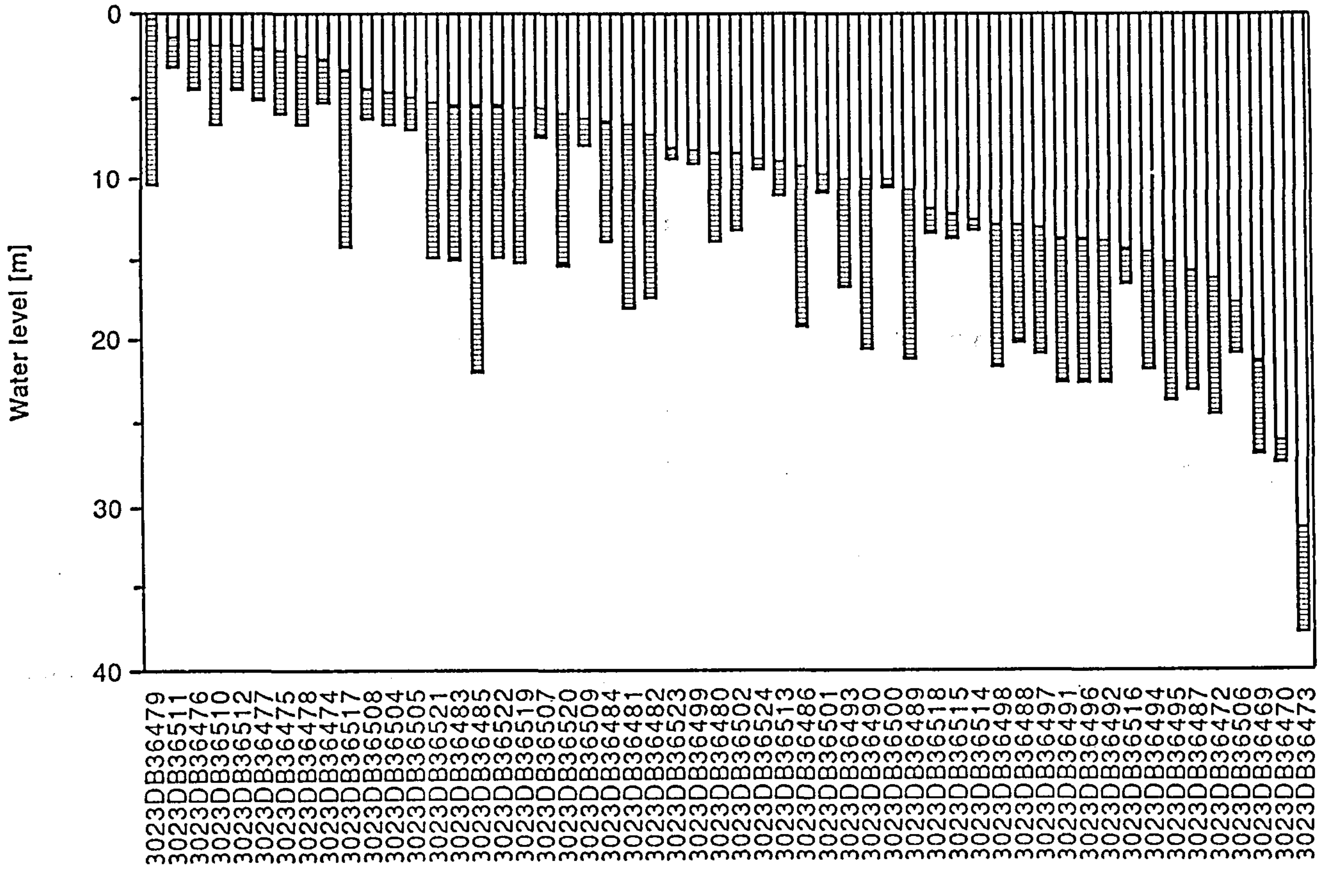


Figure 4.108 Highest and lowest water levels of G-number holes.

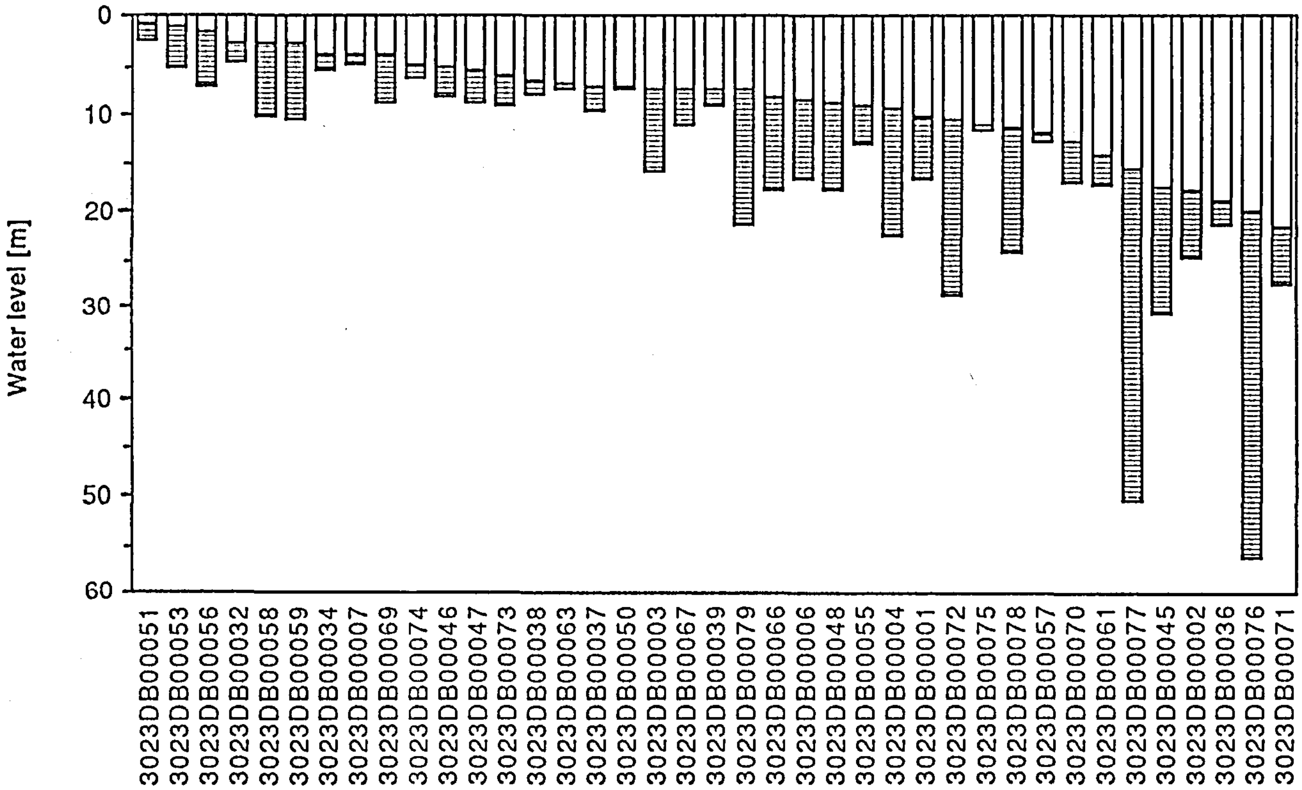


Figure 4.109 Highest and lowest water levels of ID-code holes.

The water levels of these boreholes are displayed, together with the rainfall records of the Weather Station in De Aar, in Figures 4.110 to 4.125.

Water-level contours for November 1985 are shown on Figure 4.126. Two water-sheds exist within the area, one between the northern and souther part and one near the eastern boundary. Contours for other times are shown in the Appendix.

Hydraulic parameters

Pumping tests

Six pumping tests were carried out in the De Aar research area. The following holes were tested: G 36483, 36491, 36496, 36499, 36504 and 36515. The graphs for these pumping tests evaluations are shown in the Appendix.

Pumping test G 36483

A step drawdown test comprising four steps, with pumping rates of 0.91, 2.4, 4.0 and 6.2 l/s, followed by a one-hour recovery period, was carried out on 24 May 1985. On the following day, a 72-hour pumping test, at a constant pumping rate of 6.25 l/s, commenced. Water levels were monitored in the test borehole and the following observation holes: G 36482, G 36484, G 36483 A, G 36483 B, G 36483 C, G 36483 D, G 36485, and G 36486.

Borehole	Method	Transmissivity [m2/d]	Storativity
G36482	Cooper-Jacob I	193	0.0002
G36484	Theis	126	0.003
	Cooper-Jacob I	168	0.002
G36485	Theis	99	0.002
	Cooper-Jacob I	173	0.0009
G36486	Theis	286	0.002
	Cooper-Jacob I	165	0.001
G36483A	Theis	134	0.003
	Cooper-Jacob I	110	0.009
G36483B	Theis	147	0.0007
	Cooper-Jacob I	110	0.002
G36483C	Theis	166	0.002
	Cooper-Jacob I	91	0.005
G36483D	Theis	176	0.0002
	Cooper-Jacob I	92	0.008

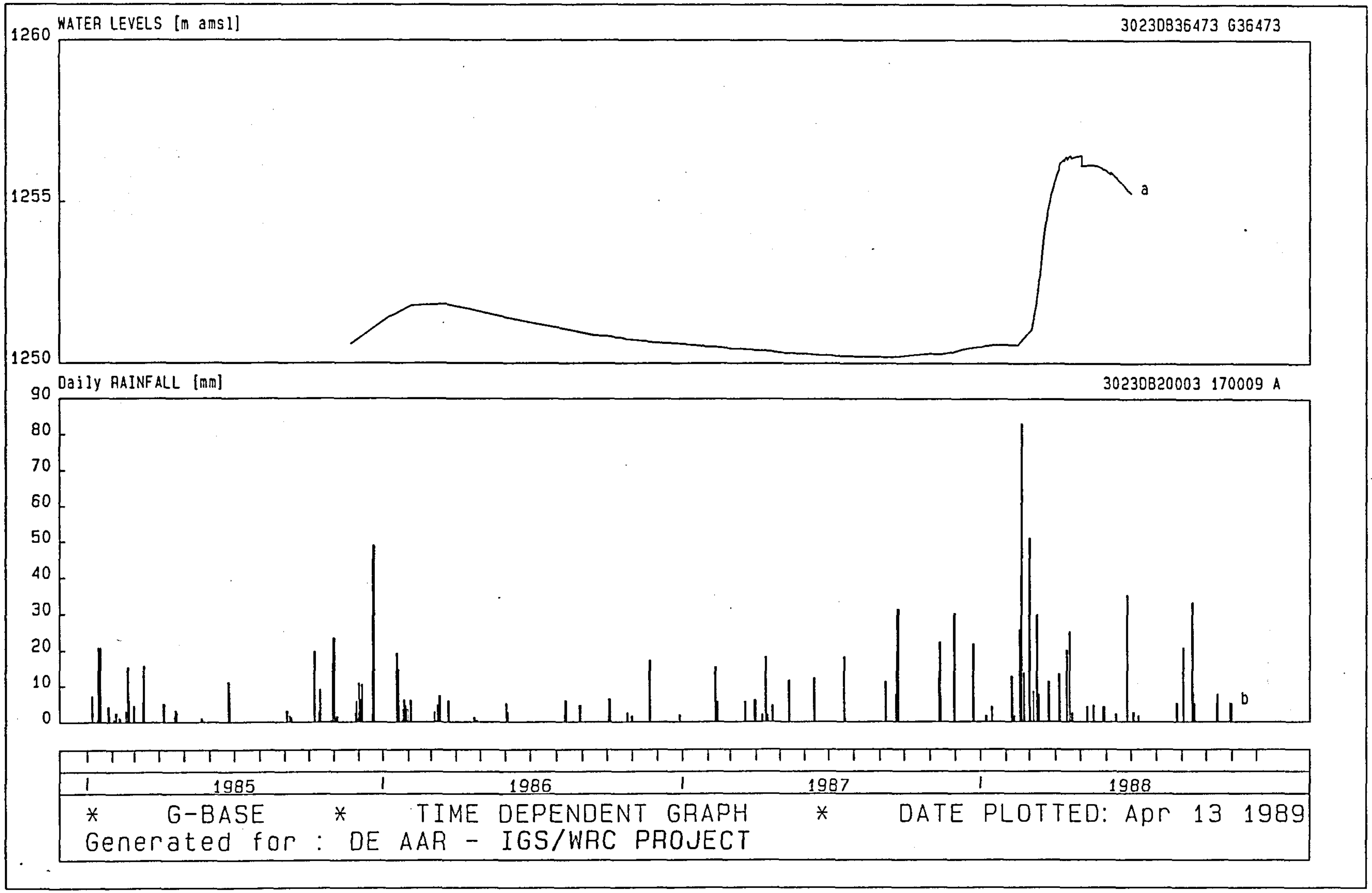


Figure 4.110 Water-level graph for borehole G 36473 and rainfall measured at the De Aar Weather Station.

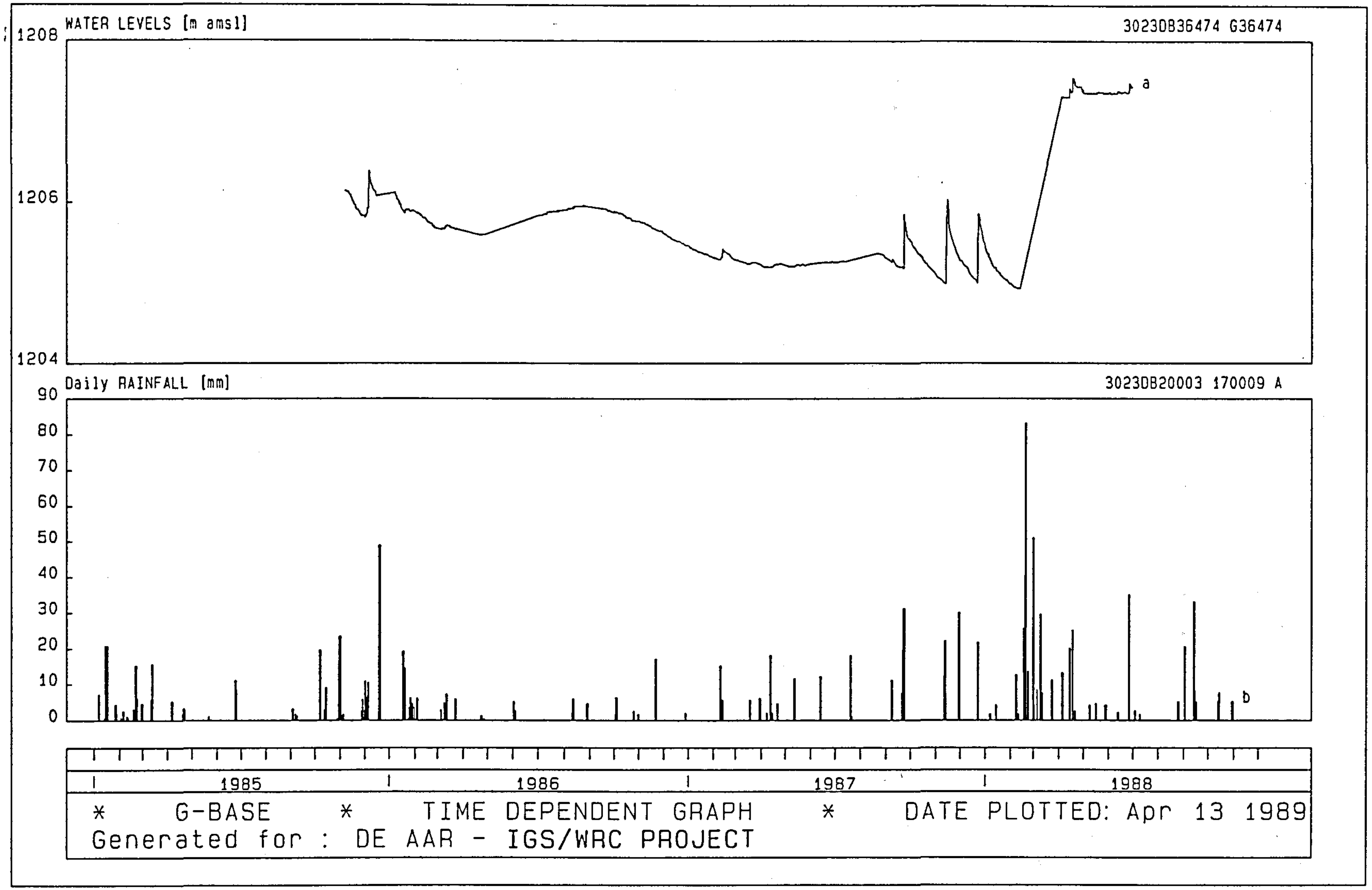


Figure 4.111 Water-level graph for borehole G 36474 and rainfall measured at the De Aar Weather Station.

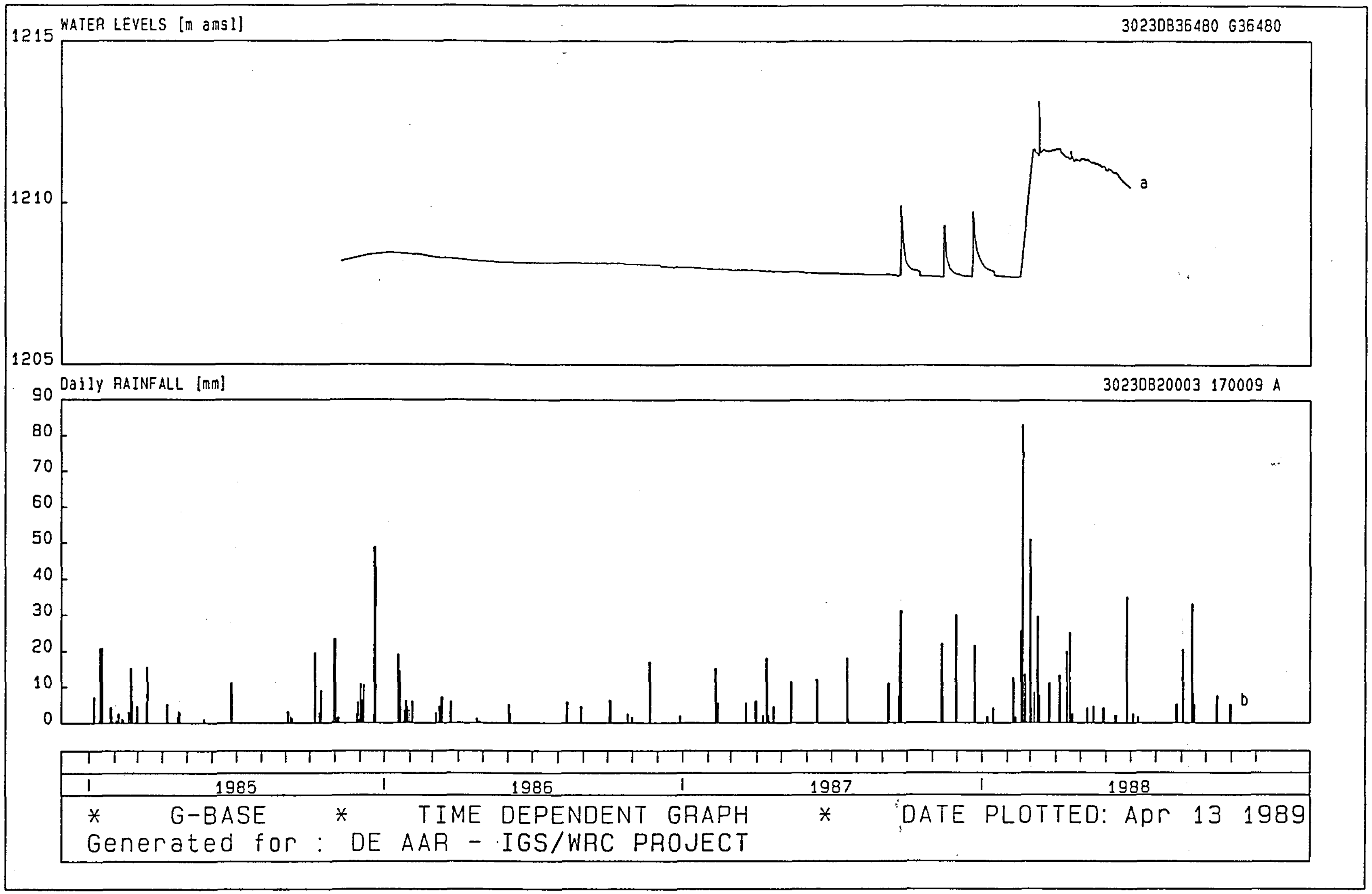


Figure 4.112 Water-level graph for borehole G 36480 and rainfall measured at the De Aar Weather Station.

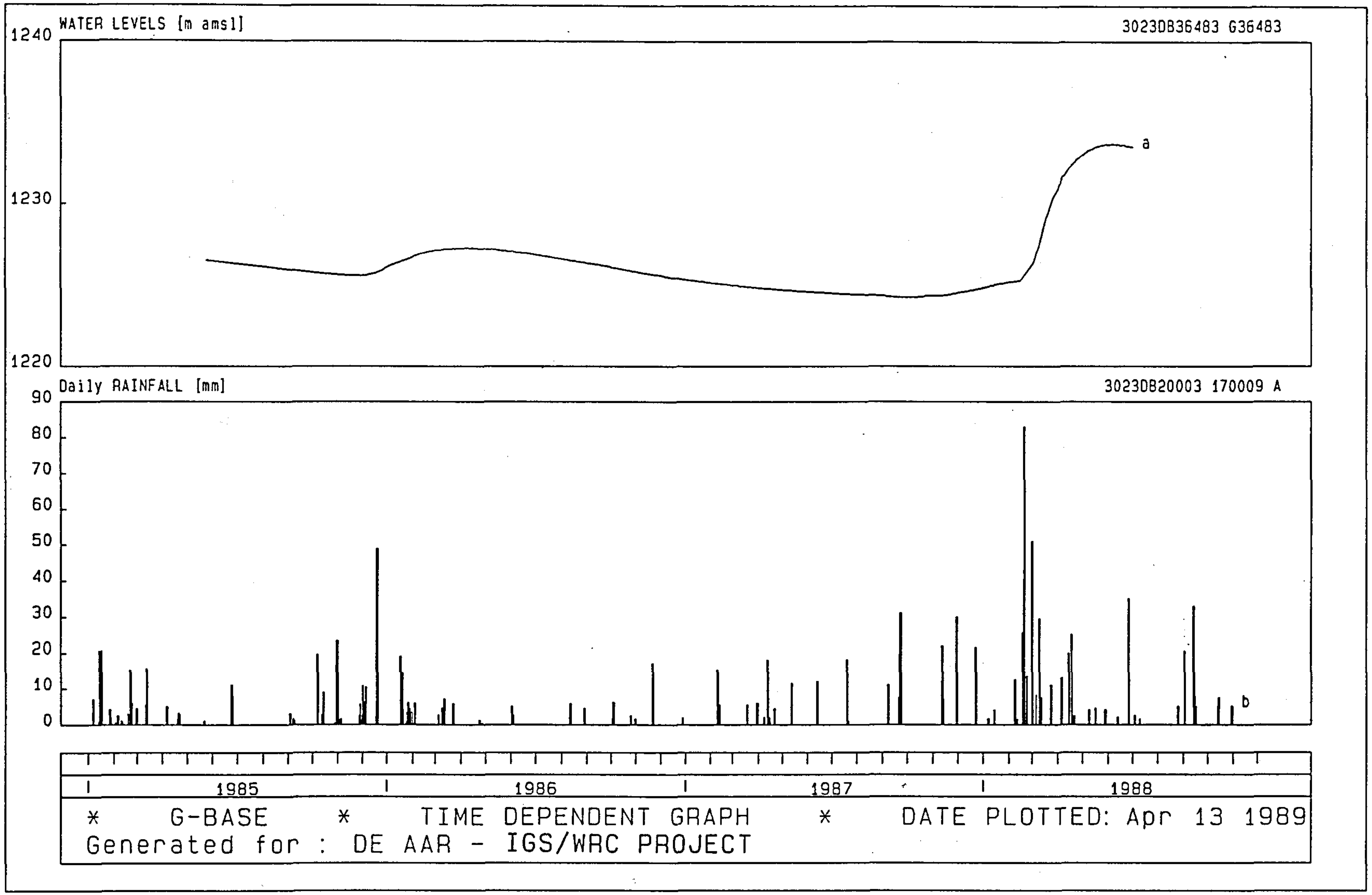


Figure 4.113 Water-level graph for borehole G 36483 and rainfall measured at the De Aar Weather Station.

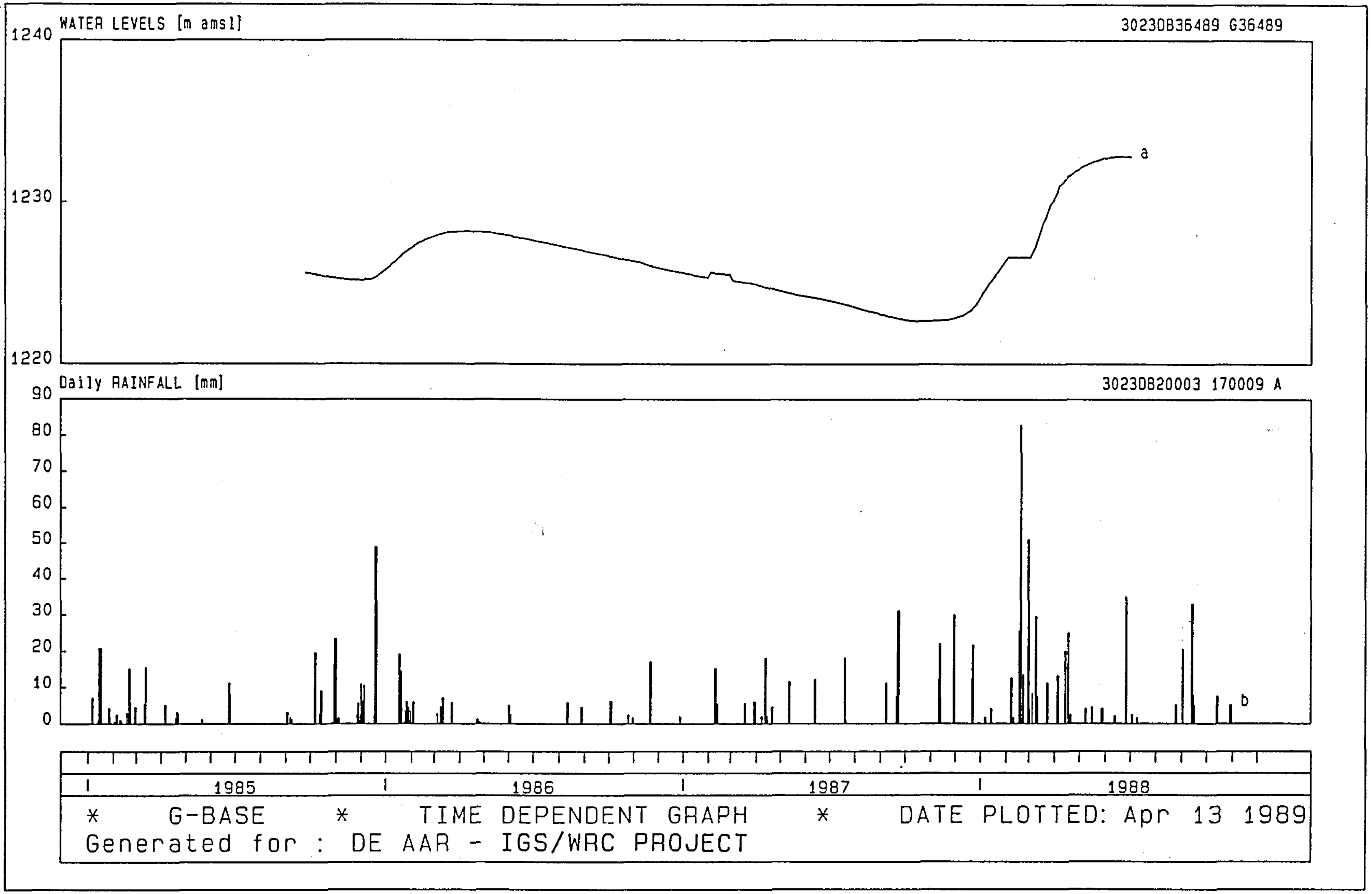


Figure 4.114 Water-level graph for borehole G 36489 and rainfall measured at the De Aar Weather Station.

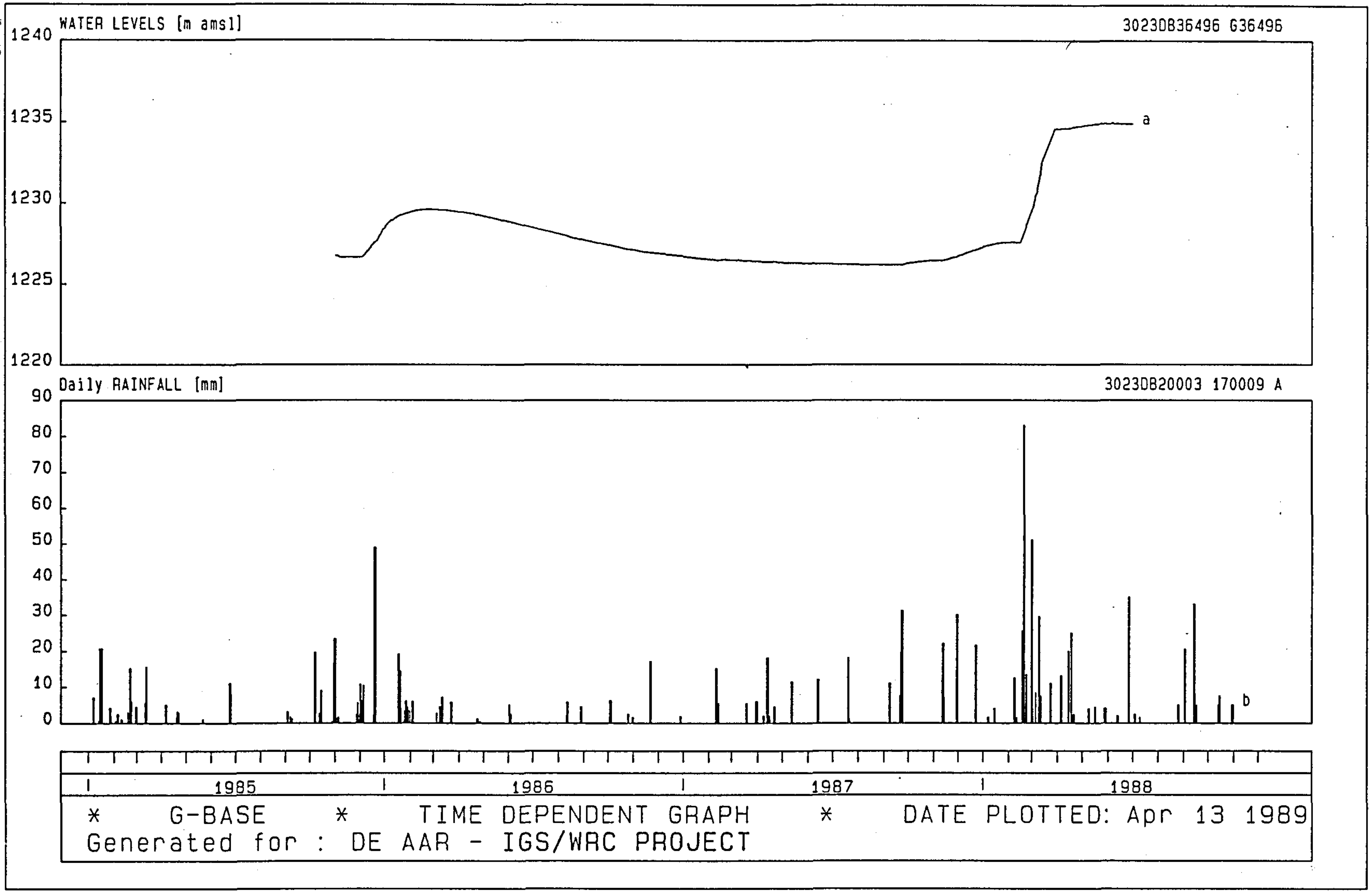


Figure 4.115 Water-level graph for borehole G 36496 and rainfall measured at the De Aar Weather Station.

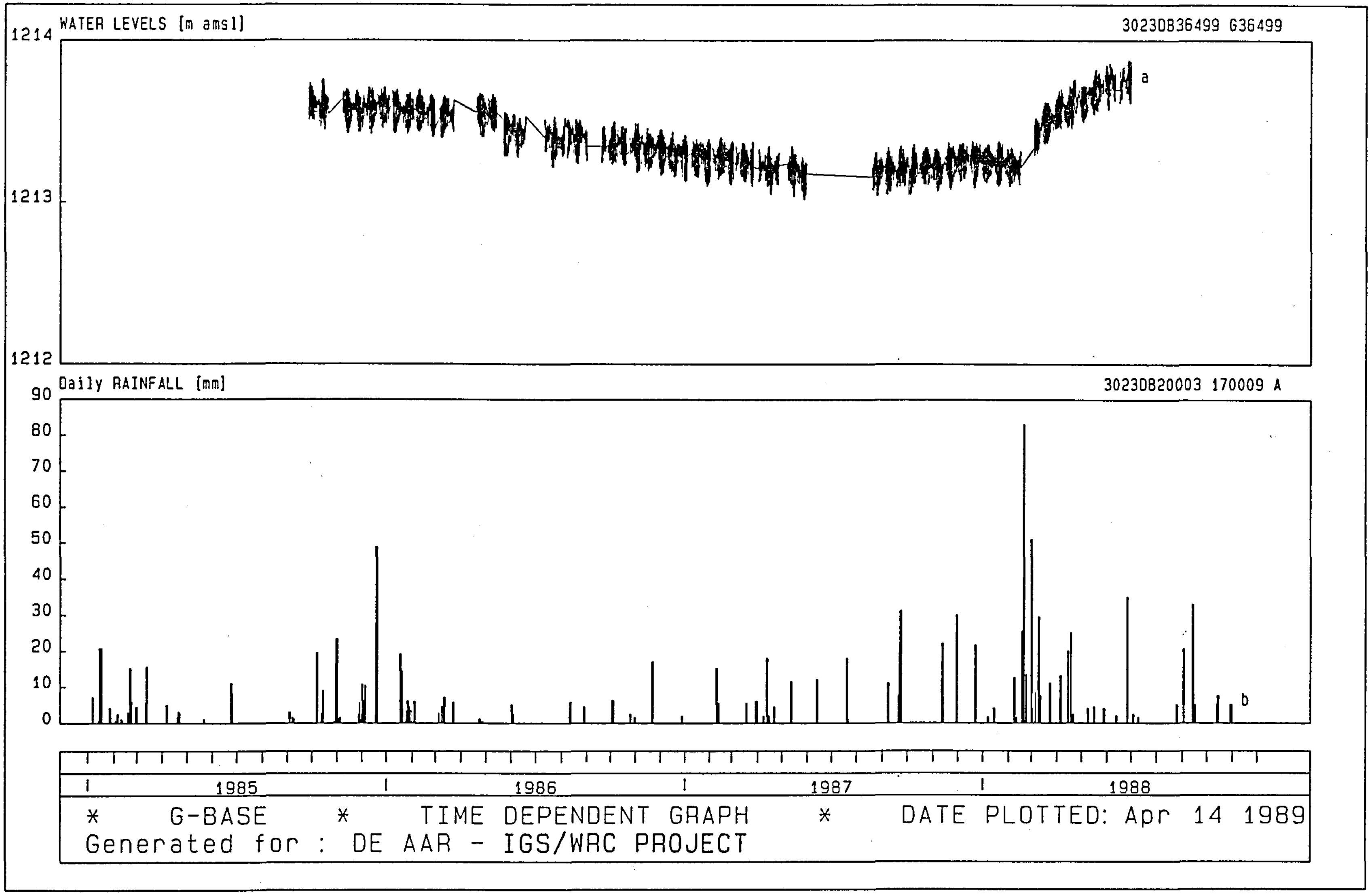


Figure 4.116 Water-level graph for borehole G 36499 and rainfall measured at the De Aar Weather Station.

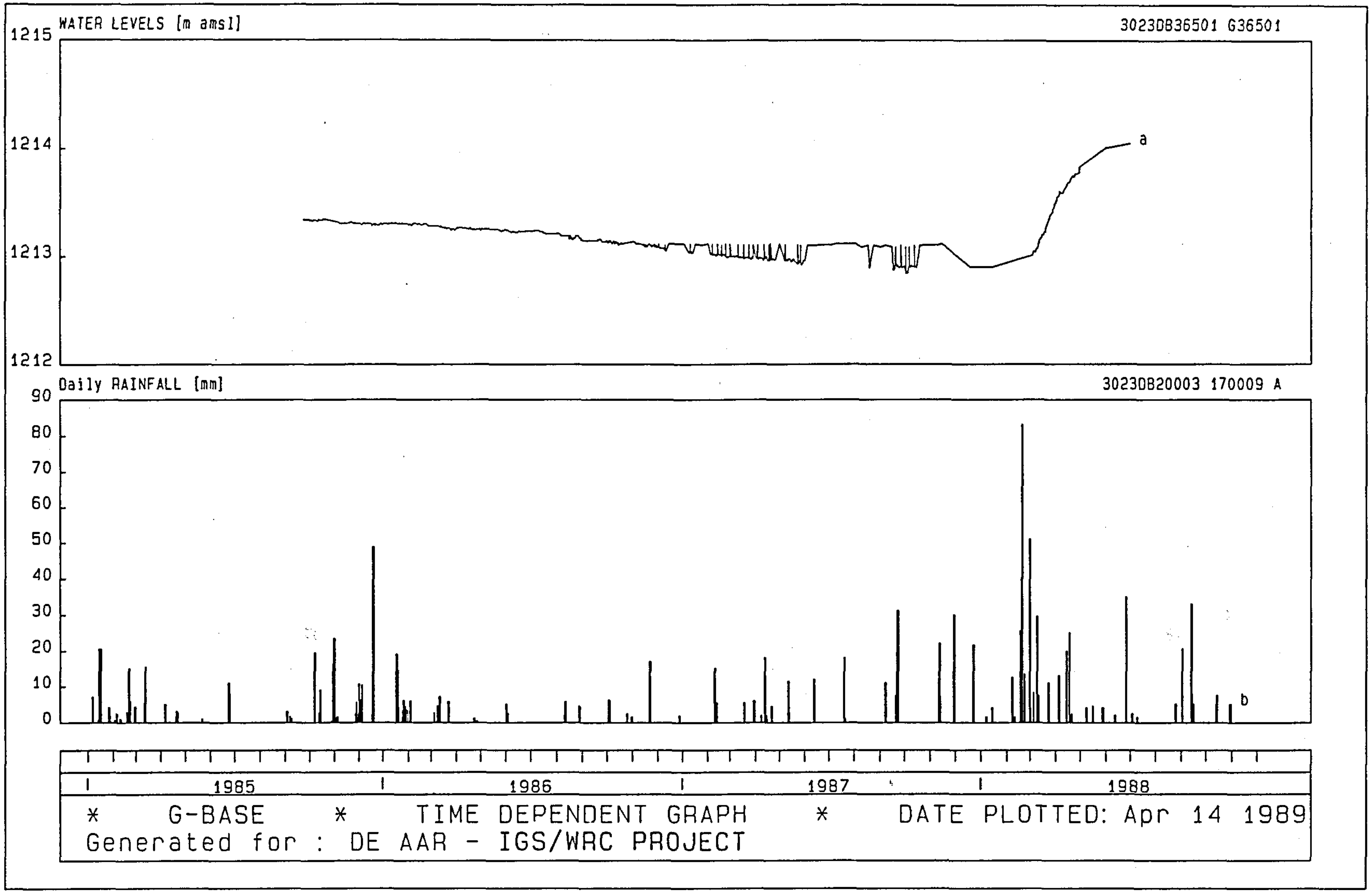


Figure 4.117 Water-level graph for borehole G 36501 and rainfall measured at the De Aar Weather Station.

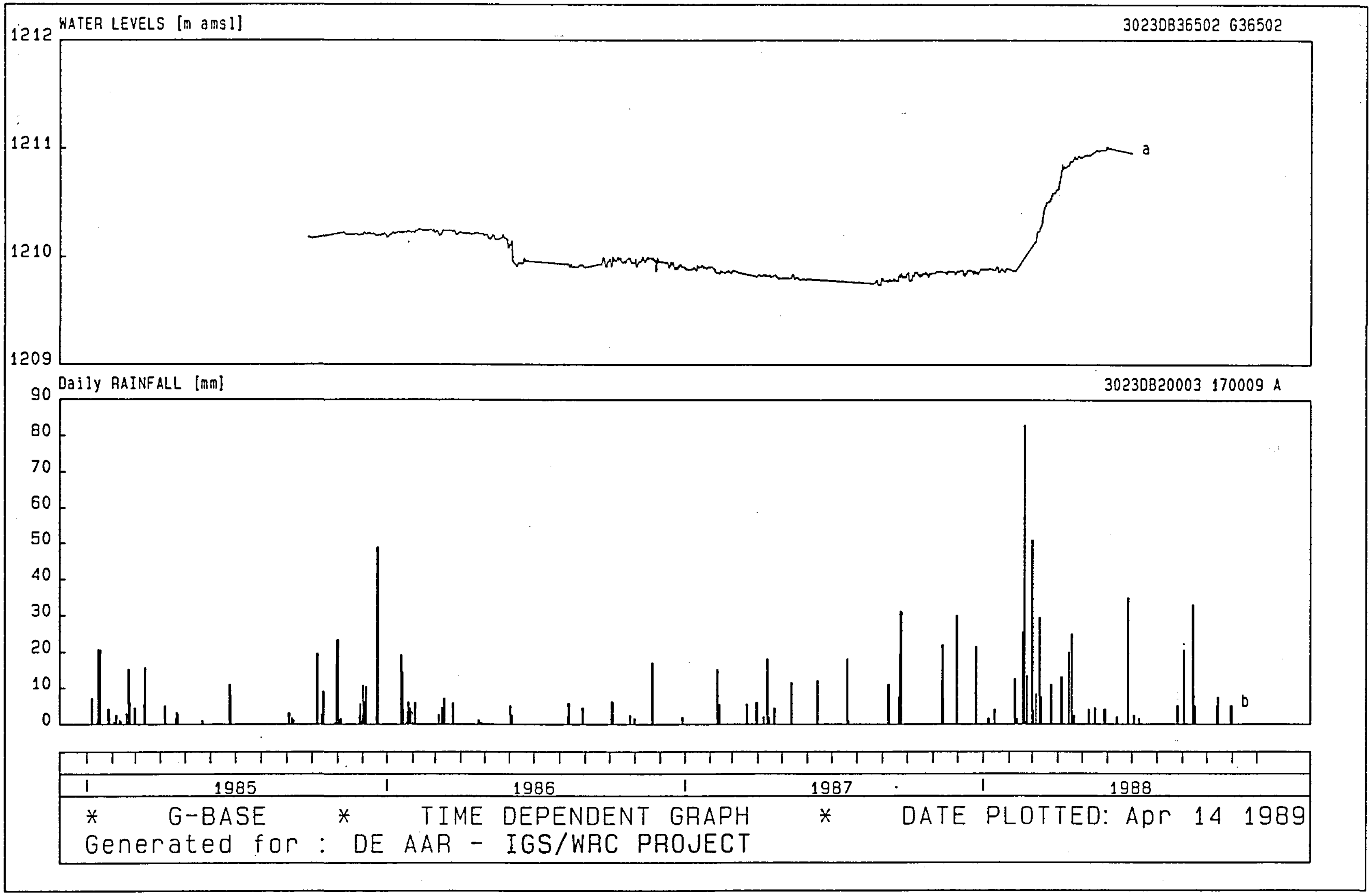


Figure 4.118 Water-level graph for borehole G 36502 and rainfall measured at the De Aar Weather Station.

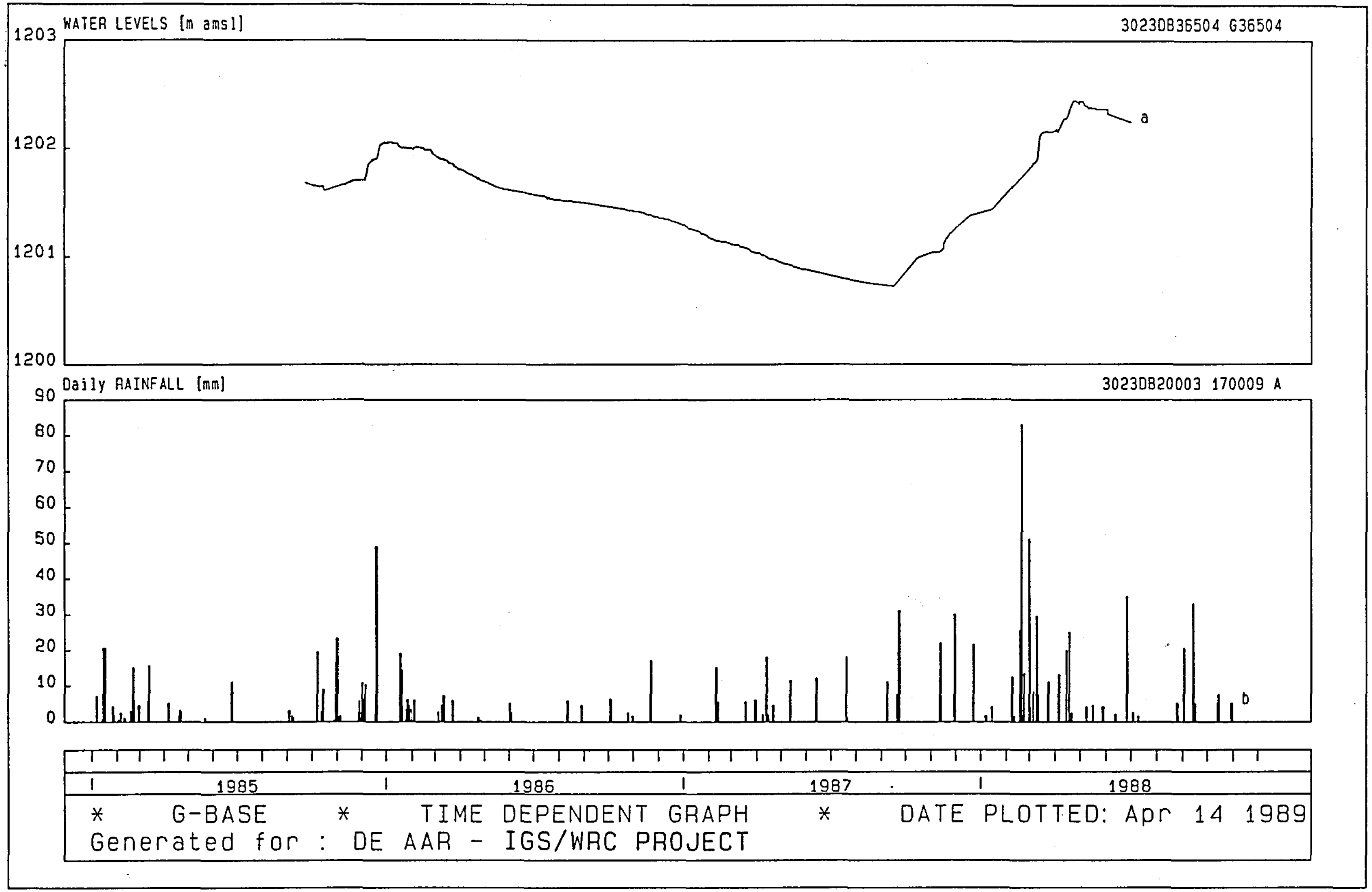


Figure 4.119 Water-level graph for borehole G 36504 and rainfall measured at the De Aar Weather Station.

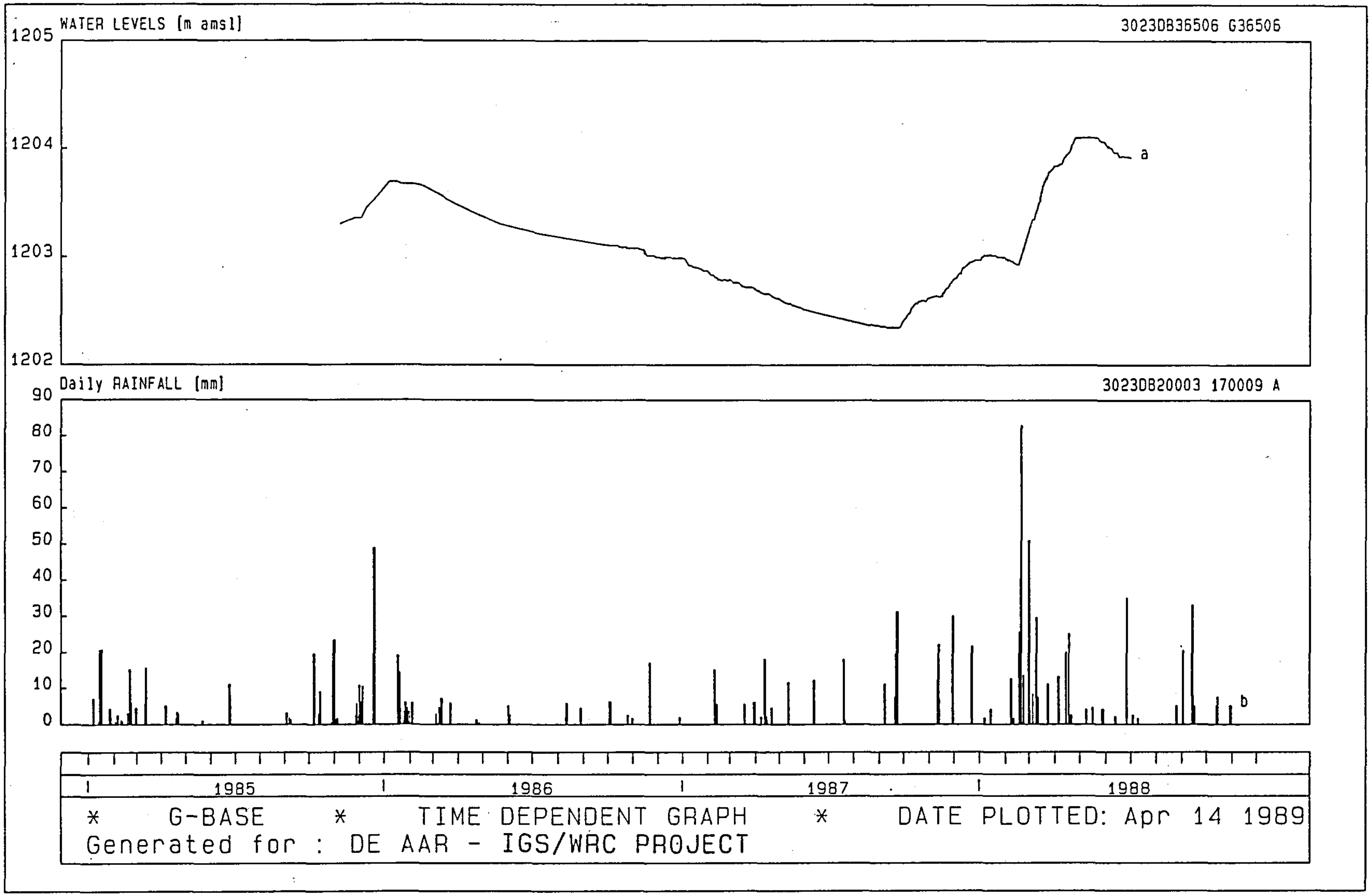


Figure 4.120 Water-level graph for borehole G 36506 and rainfall measured at the De Aar Weather Station.

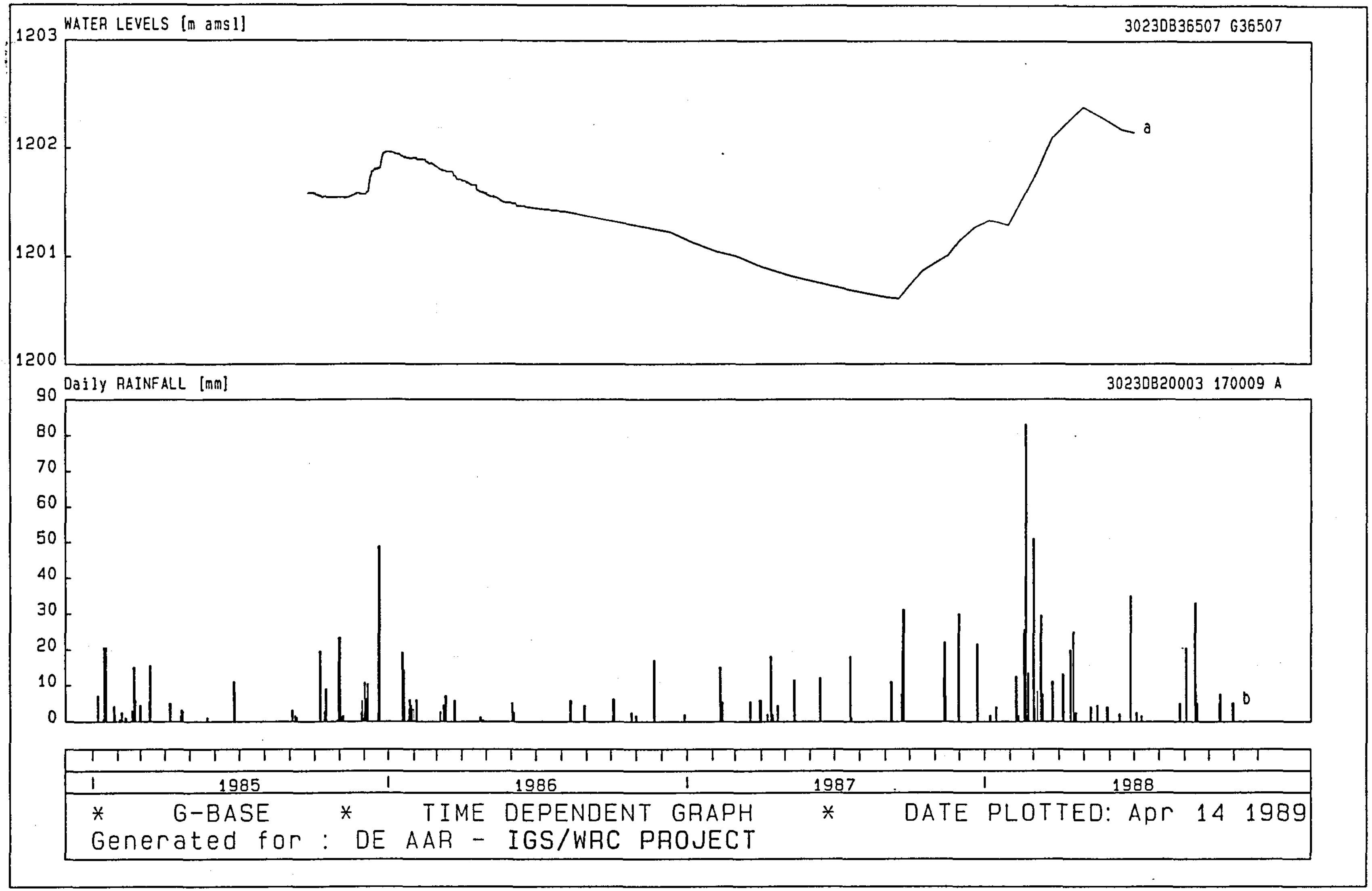


Figure 4.121 Water-level graph for borehole G 36507 and rainfall measured at the De Aar Weather Station.

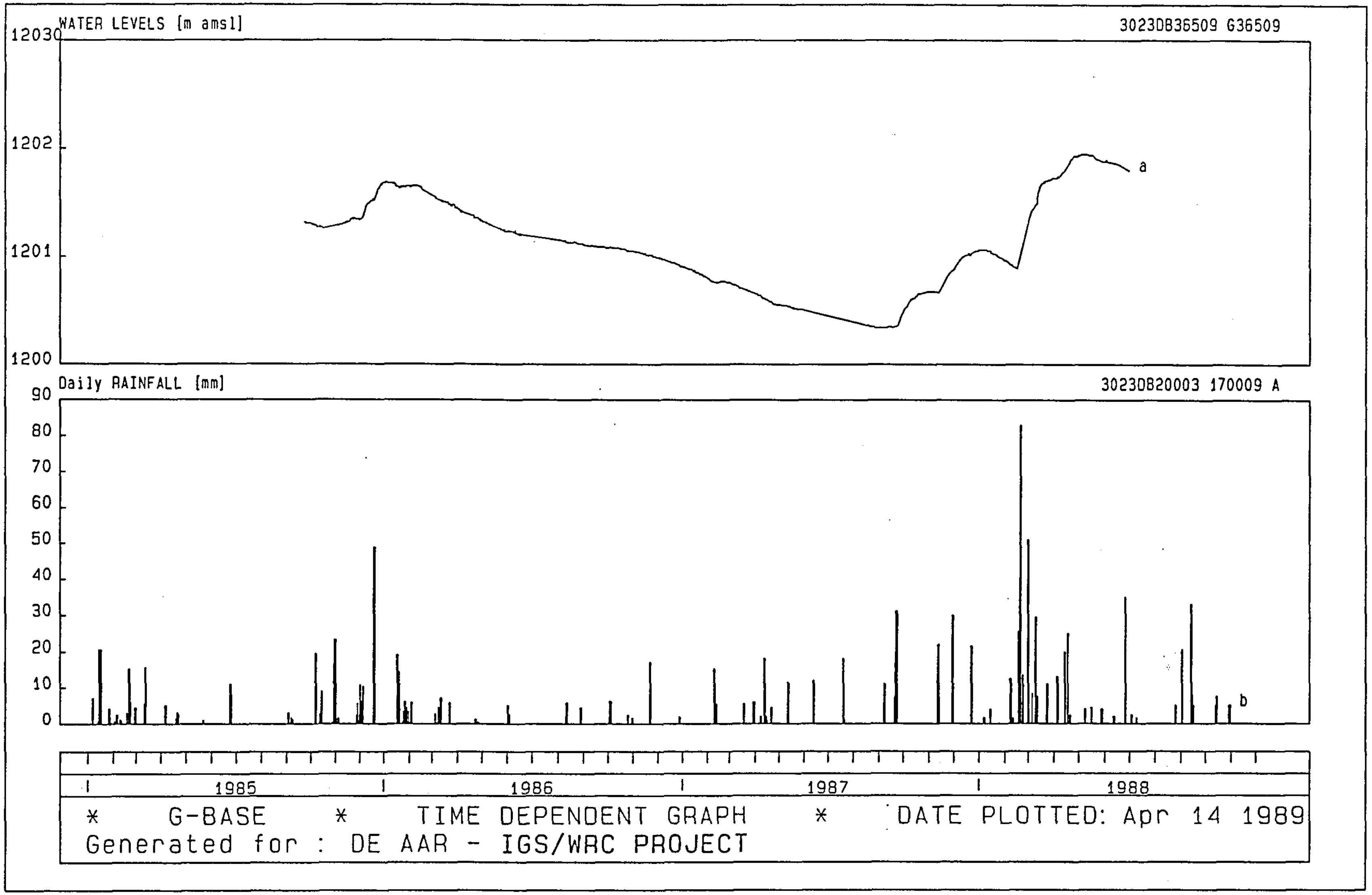


Figure 4.122 Water-level graph for borehole G 36509 and rainfall measured at the De Aar Weather Station.

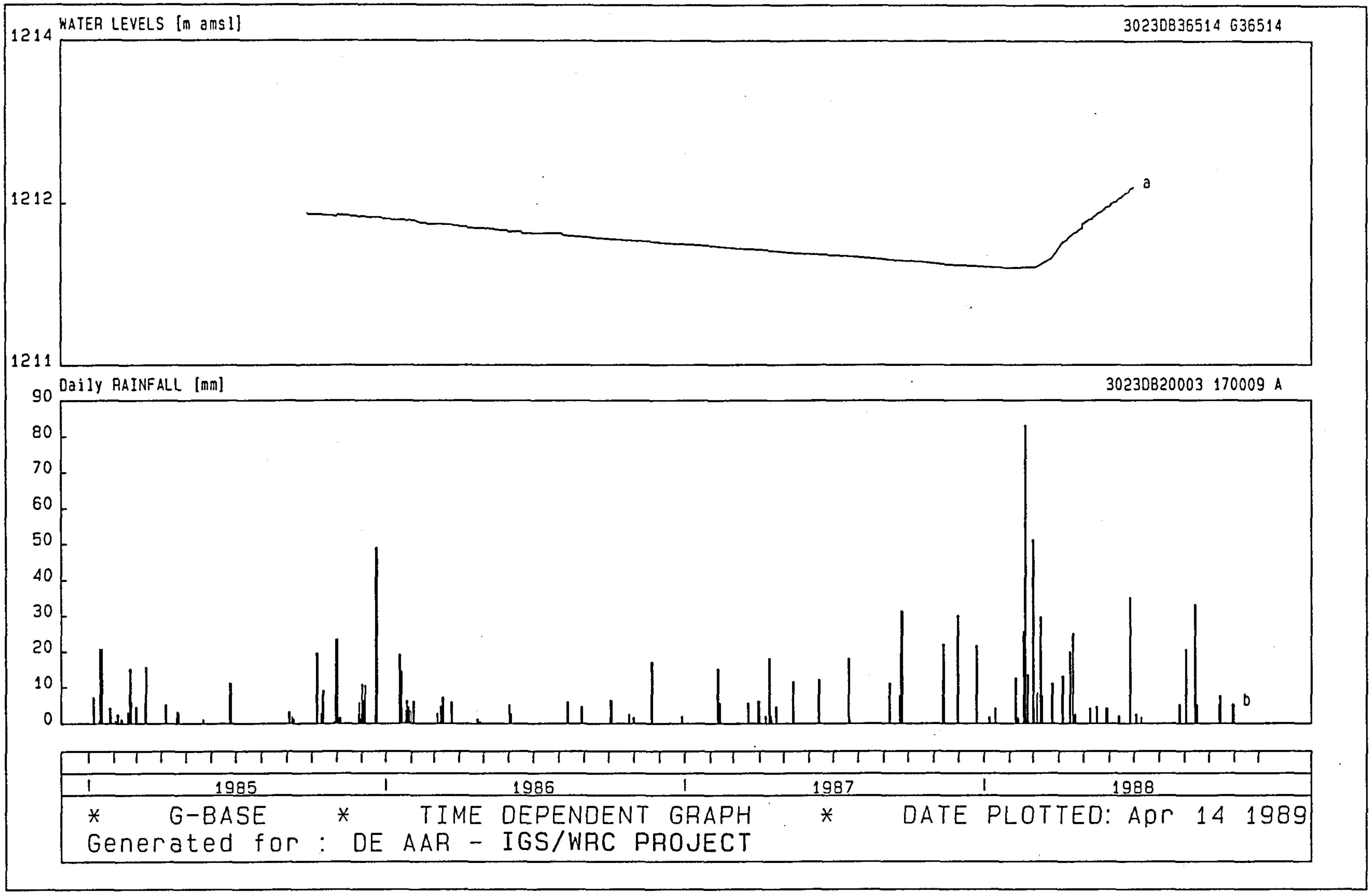


Figure 4.123 Water-level graph for borehole G 36514 and rainfall measured at the De Aar Weather Station.

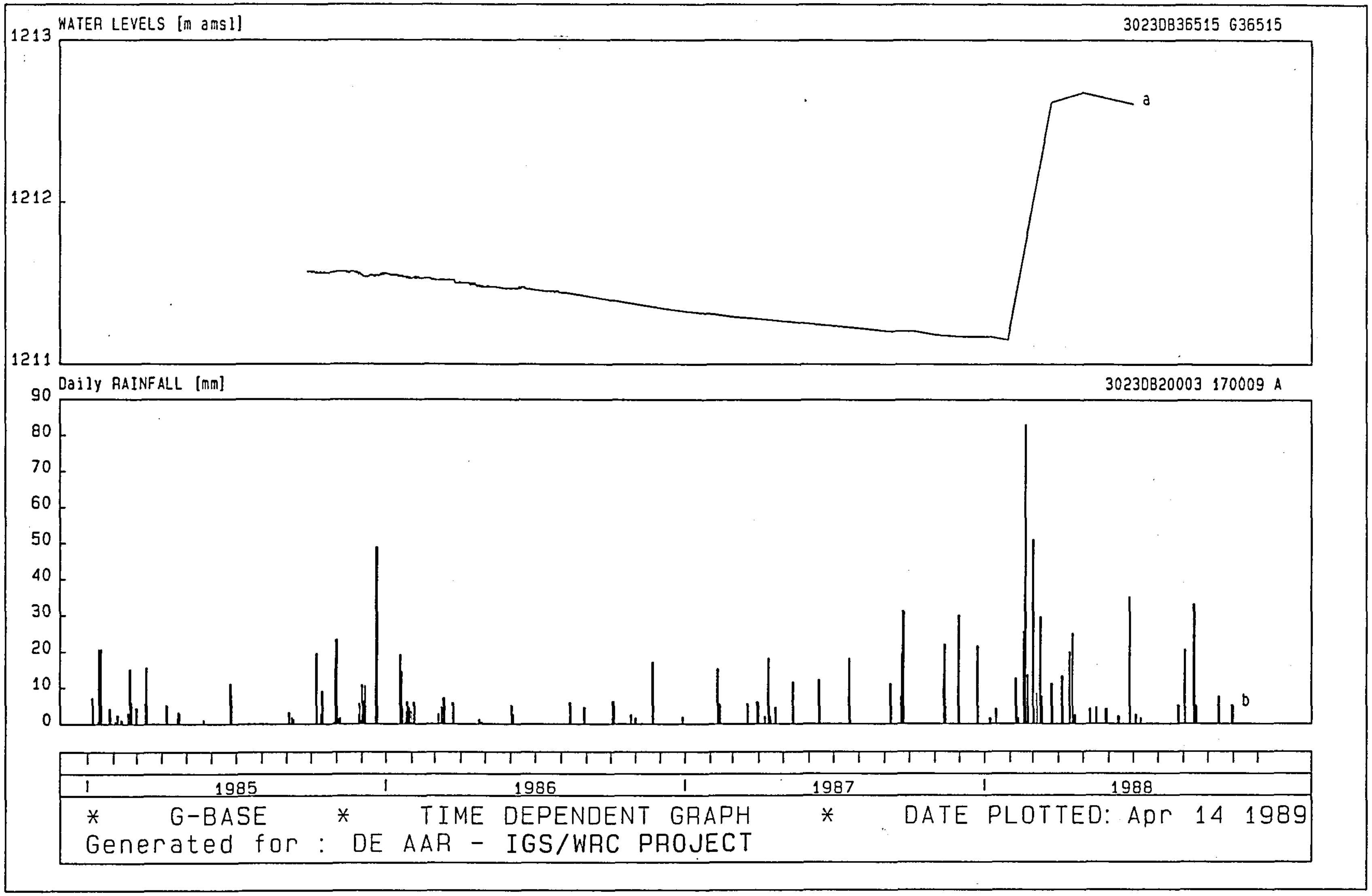


Figure 4.124 Water-level graph for borehole G 36515 and rainfall measured at the De Aar Weather Station.

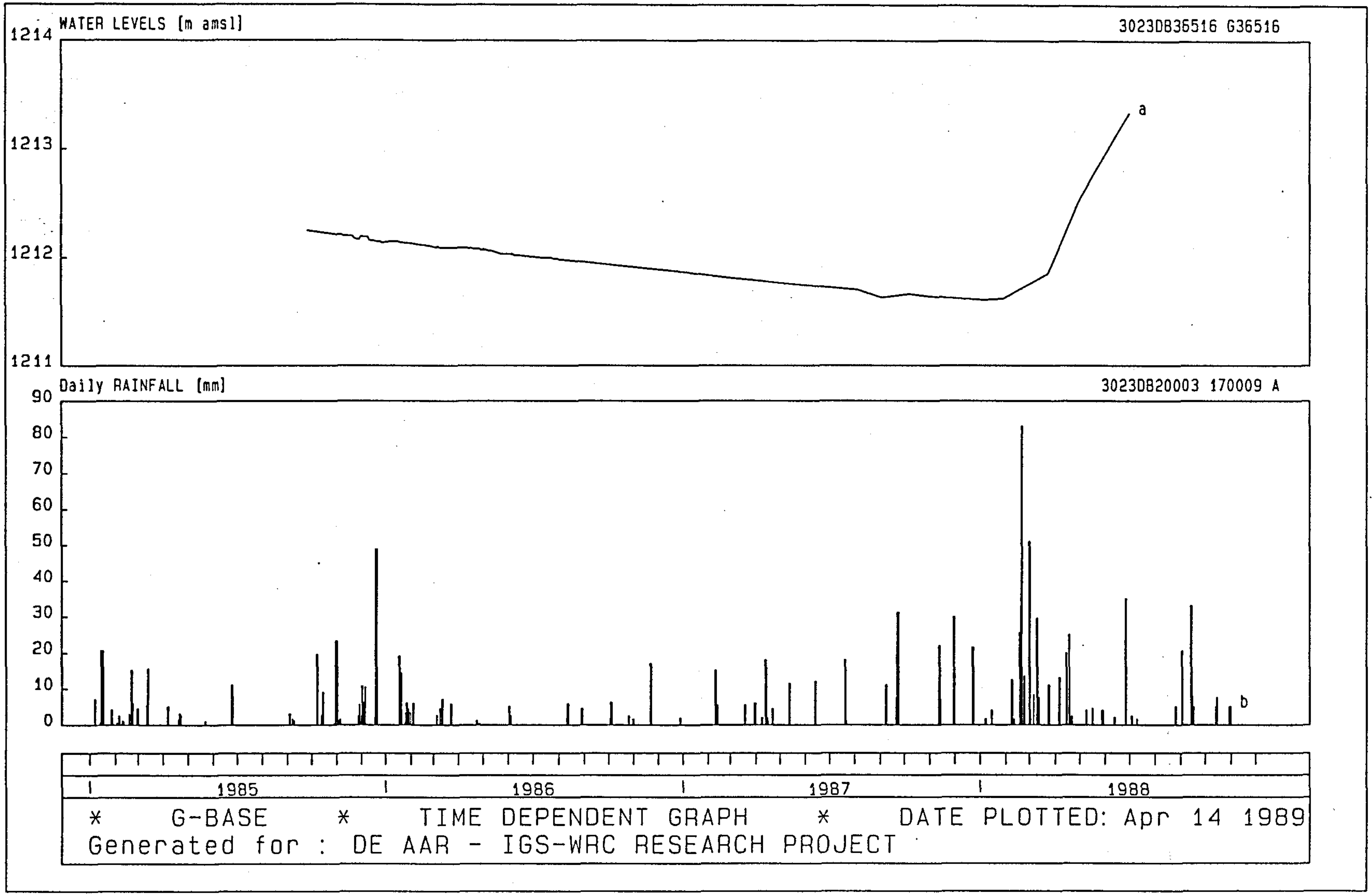


Figure 4.125 Water-level graph for borehole G 36516 and rainfall measured at the De Aar Weather Station.

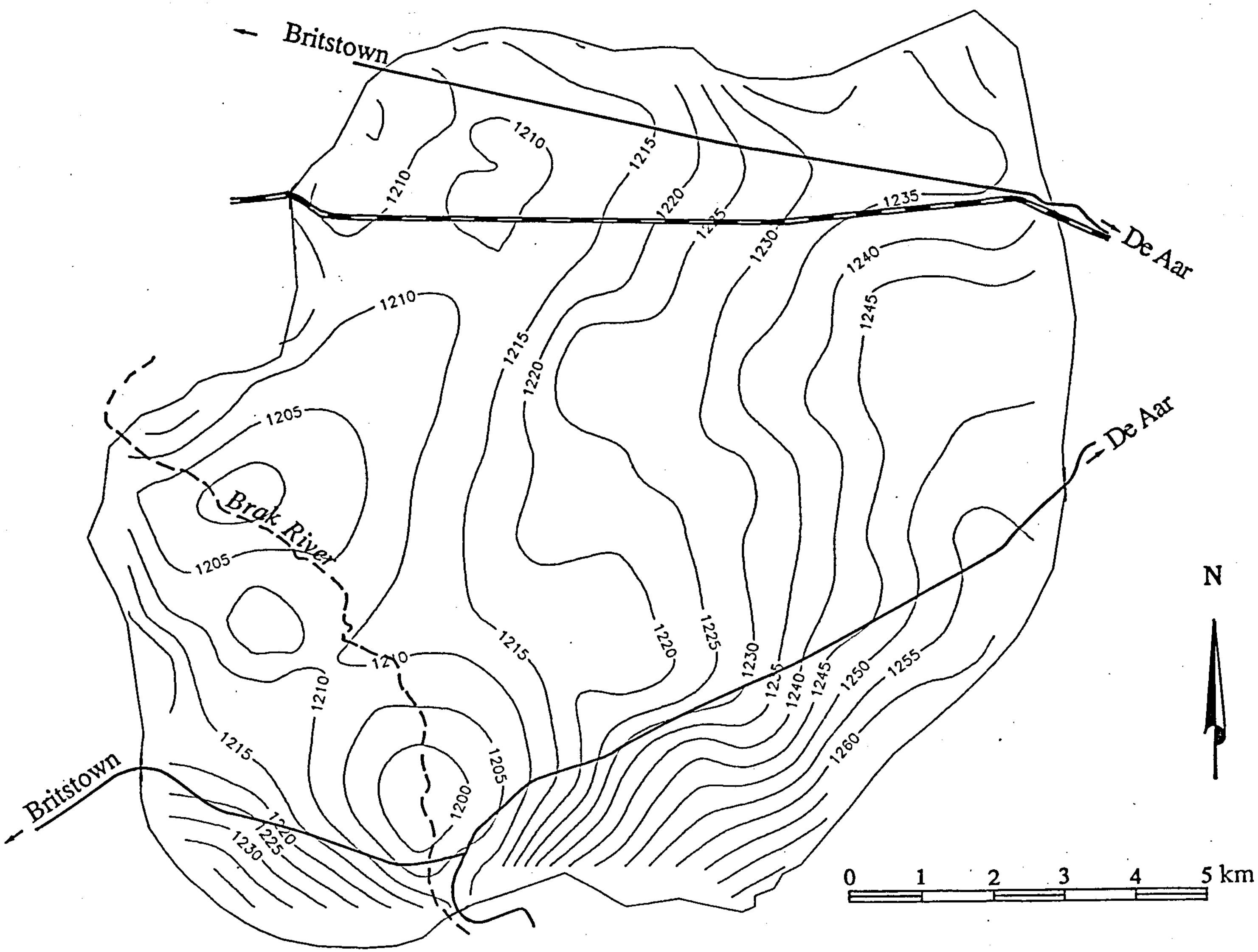


Figure 4.126 Boundaries of the De Aar aquifer showing the ground-water contours for October 1985.

Pumping test G 36491

A step drawdown test comprising three steps with pumping rates of 0.96, 1.8 and 3.0 l/s, followed by a one-hour recovery period, was carried out on 29th May, 1985. On 3rd June, 1985, a 32-hour pumping test, at a constant pumping rate of 6.25 l/s, commenced. Water levels were monitored in the test borehole and the following observation holes: G 36482, G 36487, G 36488, G 36489, G 36490 and G 36492.

Borehole	Method	Transmissivity [m2/d]	Storativity
G36491	Theis	7	-
G36489	Theis	20	0.003
	Cooper-Jacob I	66	0.002
G36492	Theis	34	0.0001
	Cooper-Jacob I	37	0.00007

Pumping test G 36496

A step drawdown test comprising three steps with pumping rates of 0.86, 1.85 and 3.05 l/s, followed by a one-hour recovery period was carried out on 6th June, 1985. On 17th July, a 72-hour pumping test, at a constant pumping rate of 1.27 l/s, commenced. Water levels were monitored in the test borehole and the following observation holes: G 36494, G 36495, G 36497 and G 36498.

Borehole	Method	Transmissivity [m2/d]	Storativity
G36496	Theis	191	-
G36495	Theis	87	0.02
	Cooper-Jacob I	193	0.02
G36497	Theis	70	0.005
	Cooper-Jacob I	170	0.004
G36498	Theis	59	0.005
	Cooper-Jacob I	61	0.004

Pumping test G 36499

A step drawdown test comprising three steps with pumping rates of 1.2, 2.6 and 4.2 l/s, followed by a one-hour recovery period, was carried out on 10th August, 1985. On the following day, a 72-hour pumping test, at a constant pumping rate of

4.13 l/s, commenced. Water levels were monitored in the test borehole and the following observation holes: G 36499A, G 36499B, G 364500, G 36501, G 36502 and G 36503. The last four holes were not significantly influenced during the pumping test.

Borehole	Method	Transmissivity [m2/d]	Storativity
G36499A	Theis	44	0.007
	Cooper-Jacob I	51	0.005
G36499B	Theis	53	0.0007
	Cooper-Jacob I	69	0.0004

Pumping test G 36504

A step drawdown test comprising four steps with pumping rates of 1.1, 2.5, 4.3 and 6.25 l/s, followed by a one-hour recovery period, was carried out on 23rd July, 1985. On the following day, a 72-hour pumping test, at a constant pumping rate of 6.25 l/s, commenced. Water levels were monitored in the test borehole and the following observation holes: G 36505, G 36506, G 36507, G 36508 and G 36509.

Borehole	Method	Transmissivity [m2/d]	Storativity
G36506	Theis	214	0.0008
	Cooper-Jacob I	258	0.0006
G36507	Theis	128	0.0008
	Cooper-Jacob I	135	0.0006
G36508	Theis	173	0.001
	Cooper-Jacob I	180	0.001
G36509	Theis	191	0.002
	Cooper-Jacob I	168	0.002

Pumping test G 36510

A step drawdown test comprising one step with pumping rates of 0.6 l/s, followed by a one-hour recovery period was carried out on 28th July, 1985.

Pumping test G 36515

A step drawdown test comprising four steps with pumping rates of 0.85, 2.4, 4.2 and 6.25 l/s, followed by a one-hour recovery period, was carried out on 28th July, 1985.

Borehole	Method	Transmissivity [m2/d]	Storativity
G36513	Theis	80	0.0003
	Cooper-Jacob I	89	0.0002
G36517	Theis	28	0.0005
	Cooper-Jacob I	95	0.000002
G36518	Theis	139	0.00005
	Cooper-Jacob I	156	0.00003

On 8th August, 1985, a 72-hour pumping test, at a constant pumping rate of 4.5 l/s, commenced. Water levels were monitored in the test borehole and the following observation holes: G 36513, G 36514, G 36516, G 36517 and G 36518.

Packer tests

Double packer tests have been carried out in nine boreholes. As in the case of Dewetsdorp, the tested sections were 5 m long and flow rates were observed for free flow and at 1, 2 and 3 bar. The holes tested were G 36474, G 36483, G 36489, G 36491, G 36496, G 36499, G 36504, G 36510 and G 36515.

Similar to the observations in Dewetsdorp, a trend of declining injection rates with depth was observed, i.e. the permeability decreased with depth and below depths varying between 30 and 45 m only comparatively small amounts (≈ 4 m^3/d or less) could be injected (e.g. Figure 4.127 for the injection rates of borehole G 36496). For some of the holes the calculated transmissivity correlated well with the results of the pumping tests, e.g. the Theis analysis for G 36491 yielded a transmissivity of 7 m^2/d while the packer test result was 6,2 m^2/d. Other holes like G 36483 with a calculated transmissivity of 7,02 m^2/d and G 36496 (T = 2,81 m^2/d), gave unexplainably low transmissivities for the packer tests.

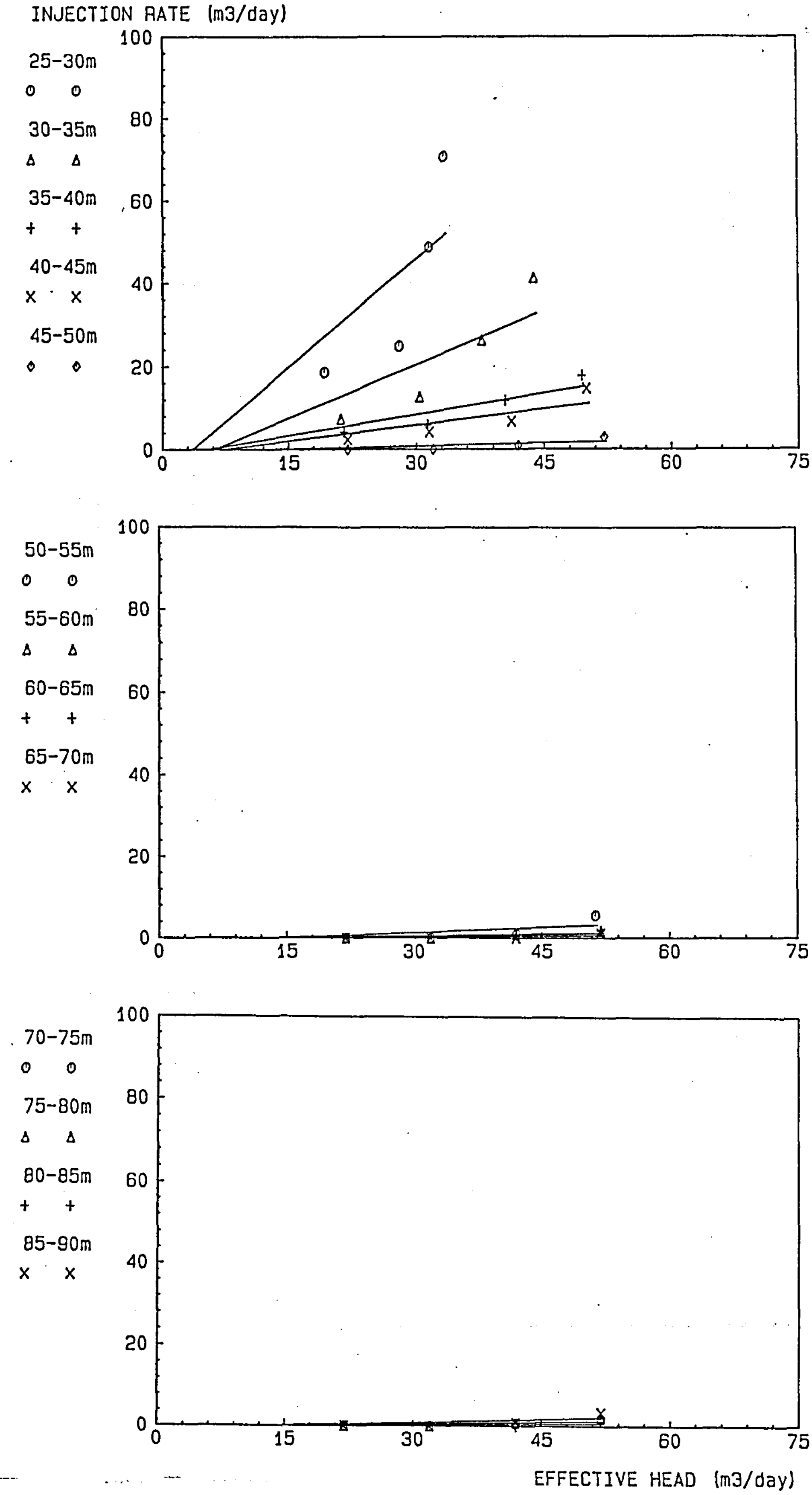


Figure 4.127 Packer test results of borehole G 36496.

Quality

Seventy-seven analyses from boreholes in the research area are available. They include samples from 28 G-numbered and 11 private holes drilled before the present investigation and 38 samples taken in 1985. All analyses are listed in the Appendix.

Alkalinity: It seems that pre-1985 samples have, in general, slightly higher values of between 200 and 400 mg/l. Very few samples exceed the 400 mg/l. The range of the remaining samples which were taken later, is between 150 and 250 mg/l. One value exceeds this range and four are below.

Calcium values tend to be around 50 mg/l, with a normal range of between 20 and 100 mg/l. Few samples fall out of this range.

Magnesium values are variable. The majority is about 50 mg/l or slightly below. A few samples exceed 100 mg/l, two have concentrations of over 150 mg/l. Holes SW of the Brak River have slightly higher values of between 50 and 100 mg/l.

Sodium: Concentrations of pre-1985 samples are scattered with minimum values, normally around 50 - 100 mg/l. A number of samples have concentrations of up to 200 mg/l, while eleven samples show values of above 300 to just below 500 mg/l. The younger samples show concentrations of <50 mg/l, except for the holes SW of the Brak River, where the sodium values lie between 50 and ±100 mg/l.

Potassium values, prior to 1985, lay between 0 and 5 mg/l, except for G 28402 (14,4 mg/l). Later analyses show slightly higher values which are generally between 5 and 10 mg/l. Five samples had concentrations of above 15 mg/l: G 36484 (19,1 mg/l), G 36496 at a depth of 94 m (37.8 mg/l), G 36501 (24.3 mg/l) and G 36502 (18,3 and 23.5 mg/l).

Chloride values show the same pattern as sodium: concentrations of pre-1985 samples are scattered with minimum values normally around 100 mg/l. A number of samples have concentrations of ± 200 mg/l, while ten samples show values of above 300 to just over 600 mg/l. The younger samples show concentrations around 50 mg/l, except for the holes SW of the Brak River, where the chloride values lie between 100 and 250 mg/l.

Fluoride: With the exception of six samples, the values range between 0 and 2 mg/l. The majority has concentrations of just below 1 mg/l. The samples which exceed the 2 mg/l limit came from G 36482 (6,7 mg/l), G 36499 (8,1 and 7,9 mg/l), G 36502 (5,0 and 5,0 mg/l) and G 36511 (4,2 mg/l).

Sulphate: Most of the samples have values of between 50 and 120 mg/l. A number of holes sampled in 1973 have concentrations between 150 and 250 mg/l. Four 1973-

samples contained between 350 and 450 mg/l. Again, it is obvious that holes SW of the Brak River differ from the remainder of the 1985 samples; sulphate is higher with values between 100 and 150 mg/l.

Nitrate levels are low, generally below 10 and more often below 5 mg/l. In three samples, higher values were encountered: BI 3 (>40), G 27202G (>30) and VP 1 (>15 mg/l). Holes SW of the Brak River have low values.

Silica minimum values are about 4 mg/l. Maximum values are between 20 and 30 mg/l for the pre-1985 samples and 15 to 20 for the later ones. Holes SW of the Brak River have in general slightly higher values.

From the aforesaid, it is obvious that ground water encountered SW of the Brak River has a chemically different signature. As regards differences between the pre-1985 and the remaining samples, it seems that between 1973 and 1985:

> alkalinity, sodium and chloride have decreased (possibly sulphate and silica also), and
>
> potassium has increased.

Piper diagram:

The majority of the samples plots in the upper half of the diamond (see Figure 4.128). In the lower half lie: G36499 (near the right corner), G36496 and G36497 (left of G36499) and BN 2, G23202A and G36482 (with HCO_3 increasing in this order along the lower right margin of the diamond). These holes have rather low electrical conductivities i.e. they are fresh. Their sulphate concentration is generally low while some of these samples, but not all, exhibit high fluoride concentrations. G 36499 is the highest yielding exploration hole drilled for this investigation.

Nearest to the recharge corner lie G36497, G36496, G23205F and G36479.

Expressed as a percentage, magnesium, chloride and, to a lesser extent, sulphate concentrations are relatively high in the investigation area, especially SW of the Brak River.

Natural Isotopes

Thirteen water samples from 11 boreholes (G 36483, 36491, 36496, 36499 and 36499A, 36502, 36504 at 18 m and 100 m, 36505 at 12 m and 42 m, 36508, 36515 and 36516) taken at the end of 1985, were analysed for $\delta\,^{18}O$, deuterium and tritium. Oxygen-18 varied between -3,29 and -4,98, deuterium between -25,1 and -37,1 and the tritium concentration was between 0 and 2 tritium units. The deuterium/oxygen-18 correlation of the De Aar data is with R = 0,66 poorer than that for the Dewetsdorp data (see Figure 4.129). The data points for boreholes G 36499 and G 36496 in the

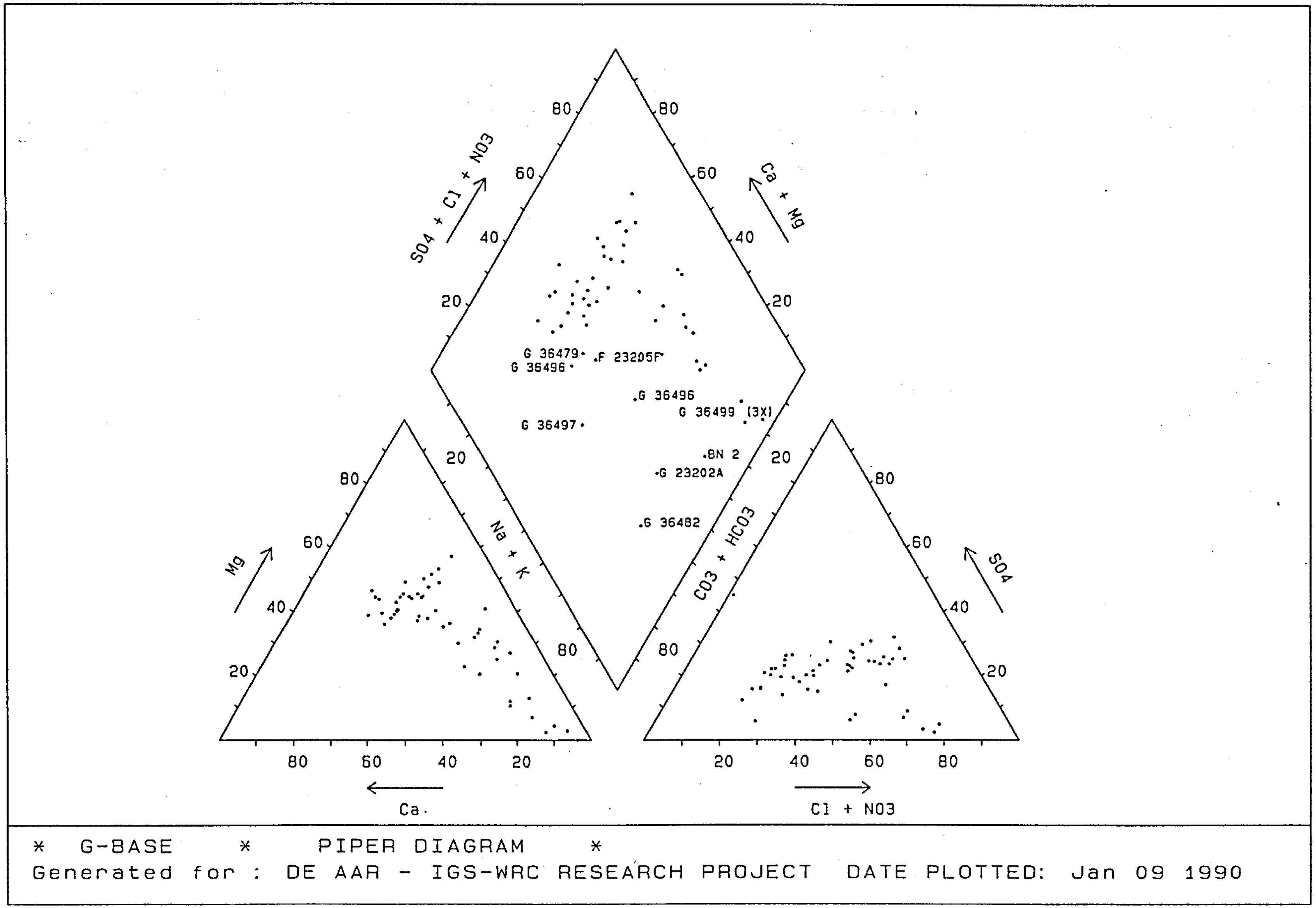


Figure 4.128 Piper Diagram : De Aar analyses.

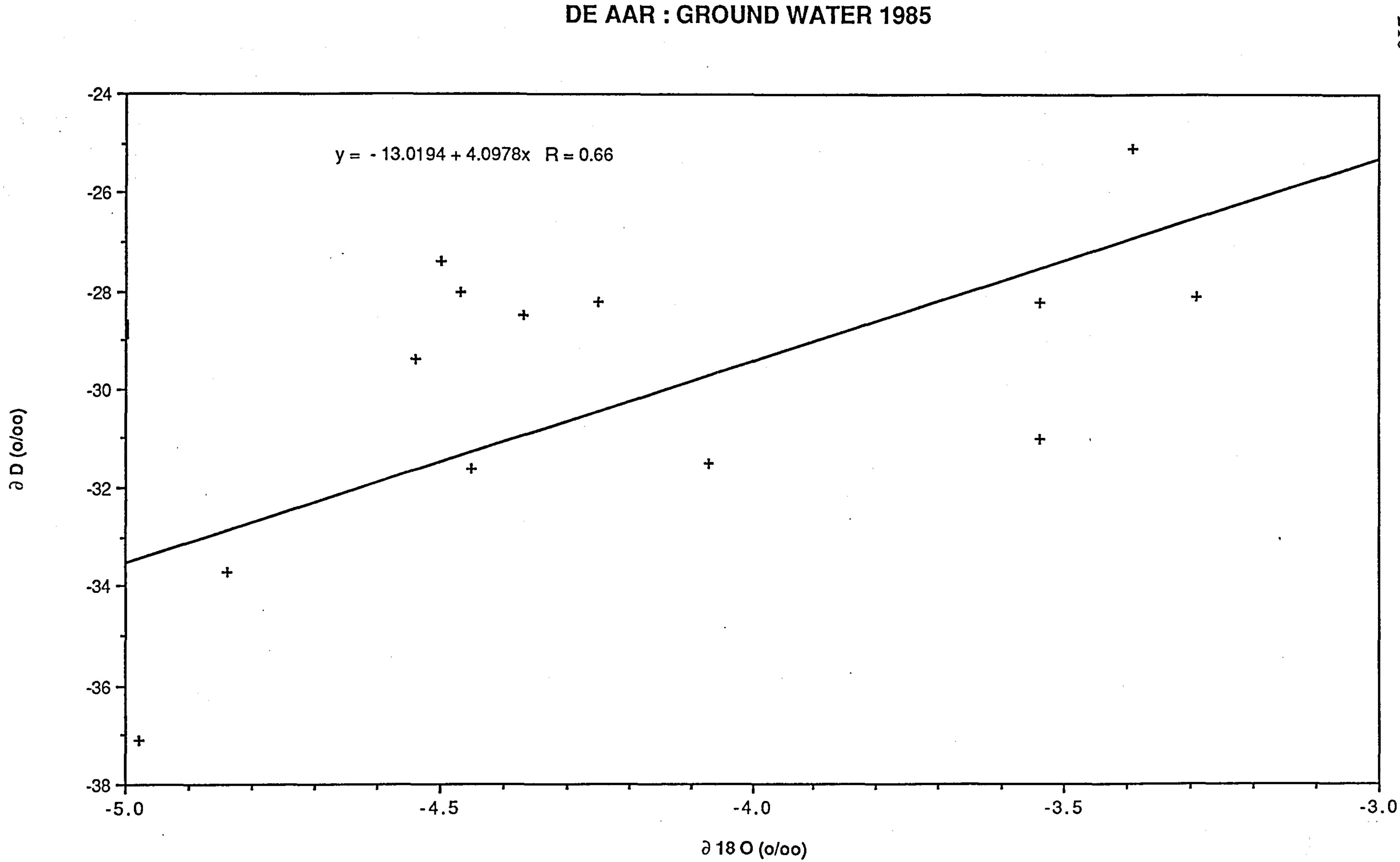


Figure 4.129 Oxygen-18 vs. deuterium concentrations of ground water from De Aar.

right corner of the diamond of the Piper diagram, plot comparatively far below the regression on the deuterium/oxygen-18 graph. Tidal variations of the water level in G 36499 indicate confined conditions (see Figure 4.116). Only small differences exist between the lowest level of 9,12 m (May, 1987) and the highest level of 8,26 m (June, 1988). These observations, as well as the time lag between the high rainfall event in February, 1988, and the highest level, may indicate that the fresh water plotting in the stagnant water corner of the Piper diagram has its origin outside the study area. (The influence of G 36499 on the calculated recharge is, however, small due to the insignificant water-level change).

Seventeen rainfall samples collected on November 28 (1 sample), November 29 (1 sample), December 3 (11 samples), December 4 (2 samples) and December 6 (2 samples) were analysed for oxygen-18 and deuterium. These samples ranged between +4,48 and -10,18 for $\delta\ ^{18}O$ ‰ and +38,6 and -65,1 for δ D ‰ (see Figure 4.130). Eleven oxygen-18 samples of rainfall on 880216 and one on 880219 showed little variation (-7.7 to -12.5 $\delta\ ^{18}O$‰).

Oxygen-18 vs. deuterium plots for rainfall and for ground water are shown on Figures 4.131 and 4.132. There is little difference in the isotope relation of rainfall from De Aar (shown as black squares in Figure 4.131) and Dewetsdorp. Ground water from De Aar (represented by symbol x in Figure 4.132) contains, however, slightly more deuterium than that from Dewetsdorp.

Extraction

In the *Southwestern Scheme*, the Renosterpoort- and the Vaalbank wellfield comprises

	1985/86		1986/87		1987/88	
October	79 390	(49 223)	96 100	(64 902)	55 758	(43 543)
November	102 133	(63 148)	79 505	(51 977)	57 831	(42 587)
December	47 339	(31 984)	99 662	(79 807)	72 837	(55 358)
January	85 433	(57 161)	131 608	(93 831)	114 345	(83 341)
February	100 341	(64 360)	99 478	(66 537)	39 767	(76 266)
March	93 345	(63 877)	78 755	(57 537)	30 419	(22 200)
April	82 008	(55 034)	87 990	(63 949)	39 767	(32 240)
May	79 216	(52 714)	21 157	(40 581)	25 593	(19 439)
June	62 228	(43 582)	70 406	(54 674)	36 852	(25 728)
July	74 728	(52 230)	58 598	(47 245)	29 427	(22 777)
August	59 741	(40 282)	48 345	(37 233)		
September	85 269	(59 158)	36 758	(28 043)		
Total	947 000	(632 753)	908 362	(686 316)	502 596	(423 479)

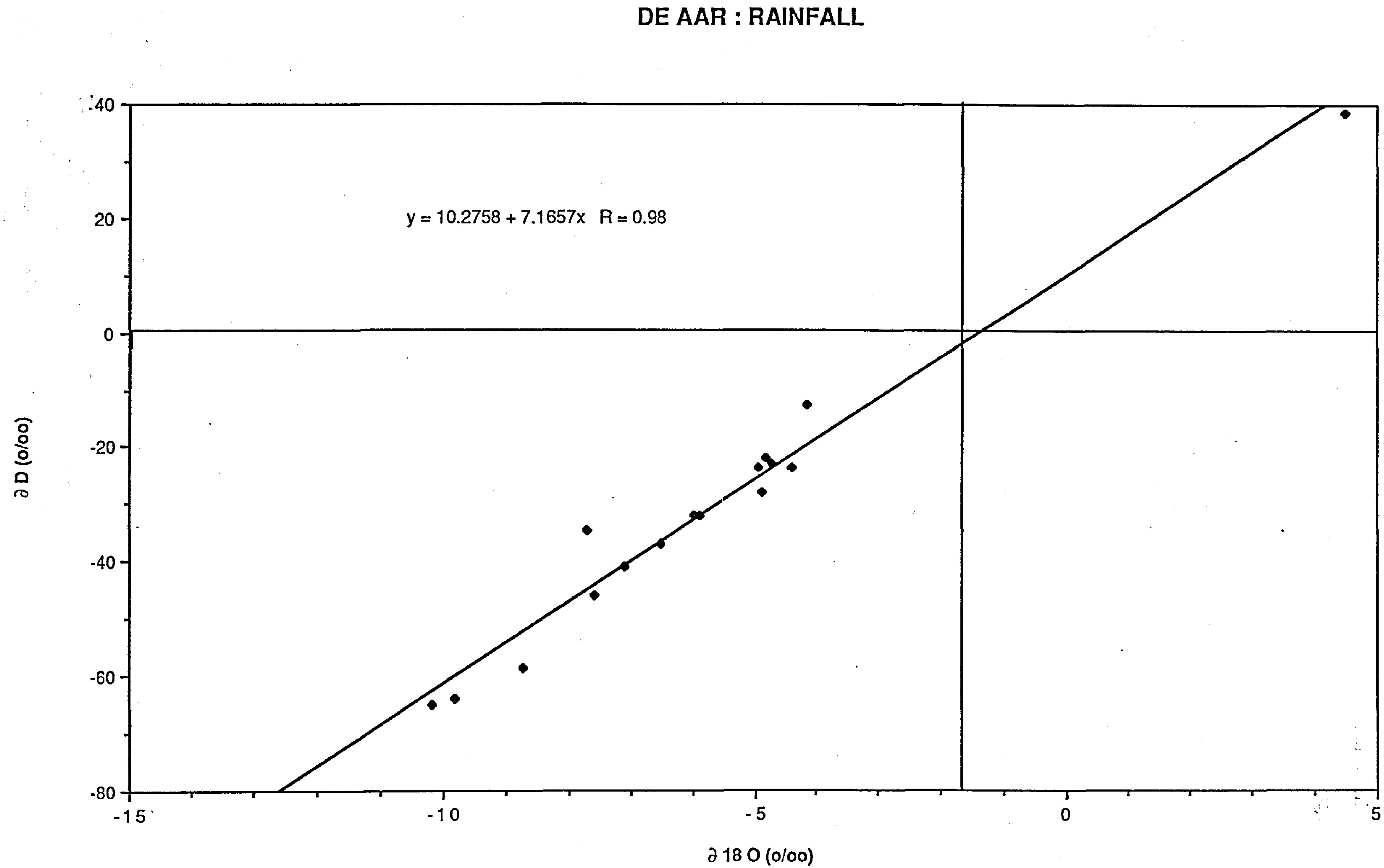


Figure 4.130 Oxygen-18 vs. deuterium concentrations of rain water from De Aar.

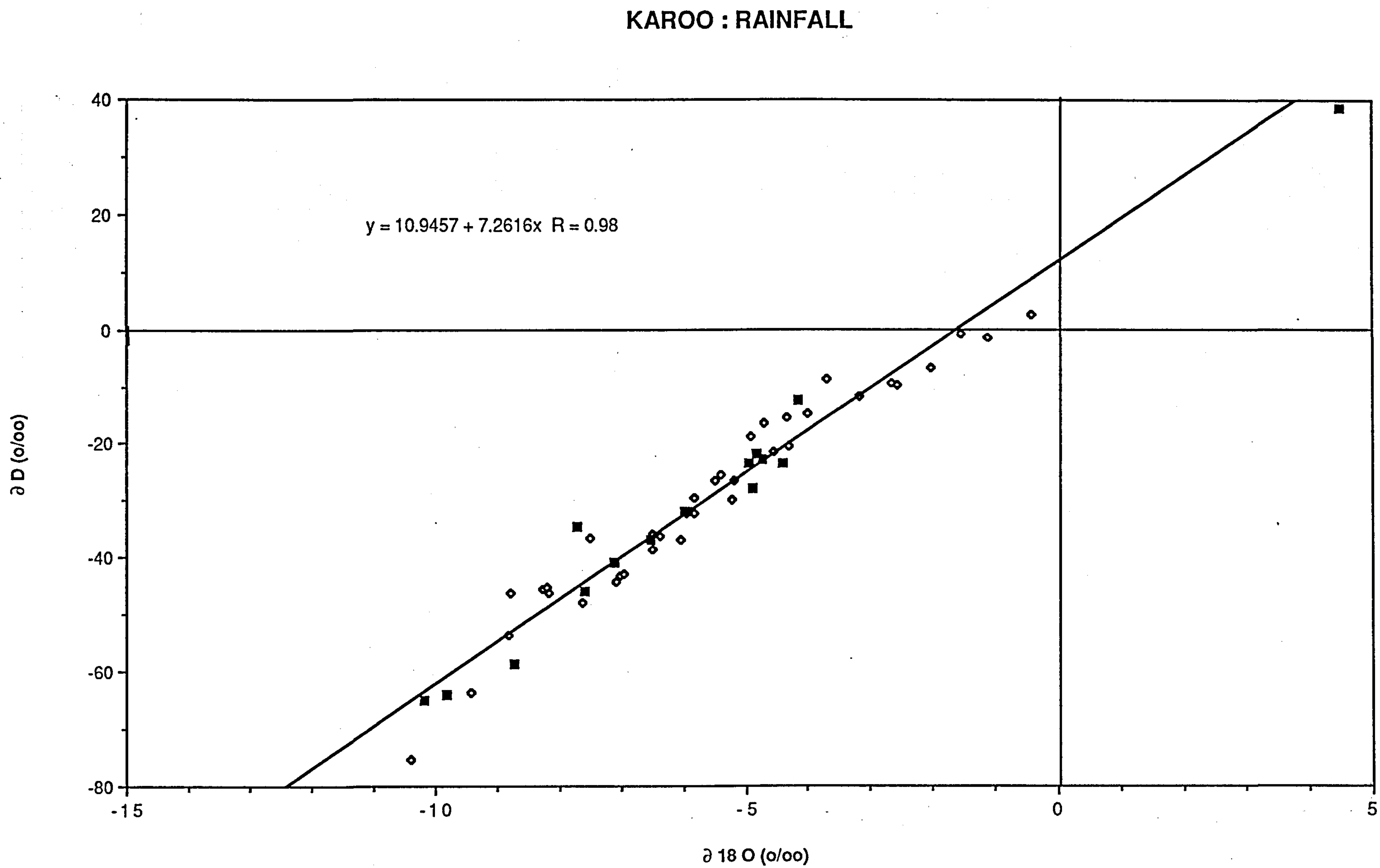


Figure 4.131 Oxygen-18 vs. deuterium concentrations of rain water from Dewetsdorp and De Aar.

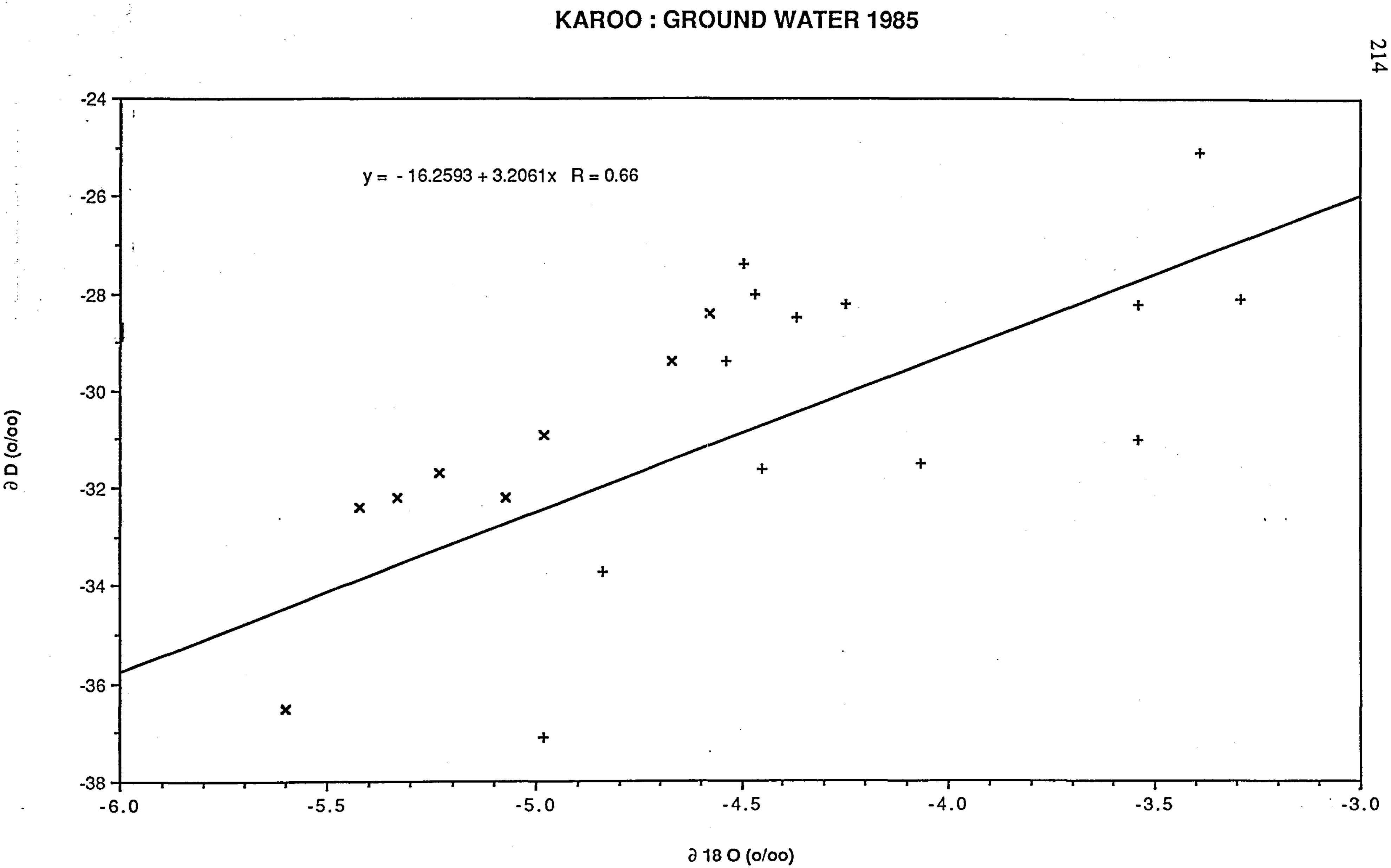


Figure 4.132. Oxygen-18 vs. deuterium concentrations of ground water from Dewetsdorp and De Aar.

four and eight production holes respectively. Four of the Vaalbank boreholes lie outside the study area, further upstream in the Western Brak River. Monthly extraction data for the individual boreholes have been used. The monthly totals for the two fields, as supplied by the municipality, are listed on the previous page. The approximated extraction data for the study area, which have been used in the following sections, are given in brackets. These values have been calculated by adding fifty per cent of the Vaalbank extraction to the Rhenosterpoort extraction, because four of the eight Vaalbank boreholes lie outside the study area.

No allowance has been made for extraction for farming purposes, as the assumed abstraction is comparatively small. Approximately 2/3 of the area is occupied by the Army where practically no ground water is pumped. The remainder of the area is used for sheep farming. With a carrying capacity of ± 2 ha/sheep and a consumption of 1 gallon/sheep/day the calculated requirement is in the order of 20 m^3/d. This is less than 1 per cent of the abstraction for De Aar and therefore negligible.

4.2.6 ESTIMATION OF RECHARGE

4.2.6.1 THE SVF-METHOD

The same sort of reasoning, as was the case with the Dewetsdorp aquifer, will be followed here. The procedure followed for the calculation of the saturated volume for the De Aar aquifer is outlined below:

(i) Firstly, the position of the topographic ground-water divide was deduced from reports. Ground water enters the aquifer in the south at Vaalbank Poort, where the Brak River enters the study area (Figure 4.133). Von Hoyer and Rinkel (1976) calculated the T-value at this inflow boundary as 1 000 m^2/d. The only lateral leakage of ground water out of the area is in the north, at the farm Brandfontein. This quantity is small (probably in the order of 100 m^3/d), compared to the abstraction rate of about 1 800 m^3/d by the municipality in the area. This value does not include the abstraction of the four municipality boreholes, just south of the Vaalbank Poort.

(ii) The aquifer outlined on Figure 4.133, was discretized into 762 triangles with 425 nodes (Figure 4.134) and comprises an area of 134 km^2.

(iii) Ground-water levels, on a monthly basis, at 71 boreholes were used by the program SVF to calculate the saturated volume of the aquifer. Figure 4.135 shows the saturated volume for the period October, 1985, to July, 1988, (a period of 34 months). Figure 4.135 was obtained by subtracting the saturated

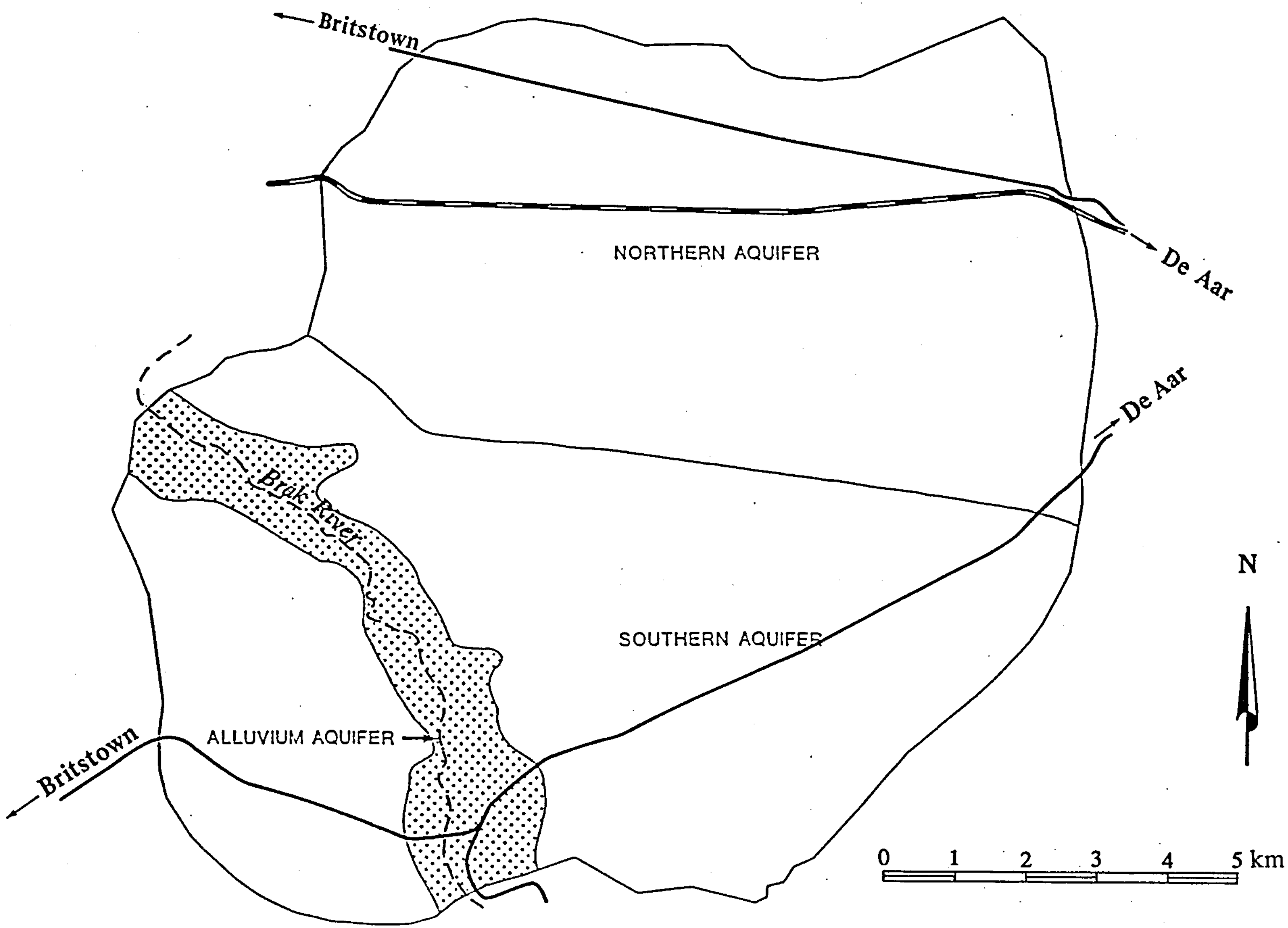


Figure 4.133 Position of the northern, southern and alluvium aquifers in De Aar.

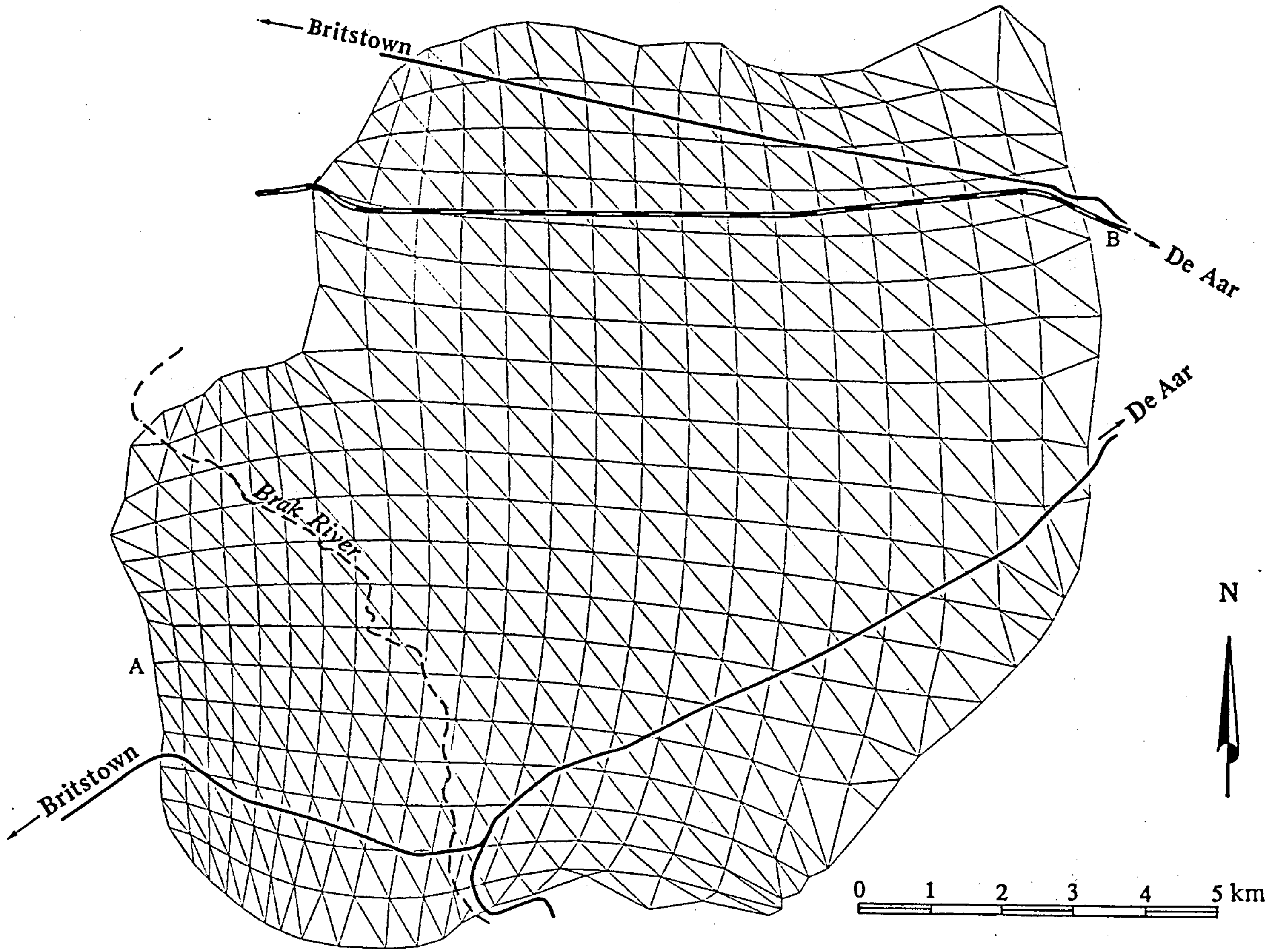


Figure 4.134 Triangular mesh used for the SVF-method for the De Aar aquifer.

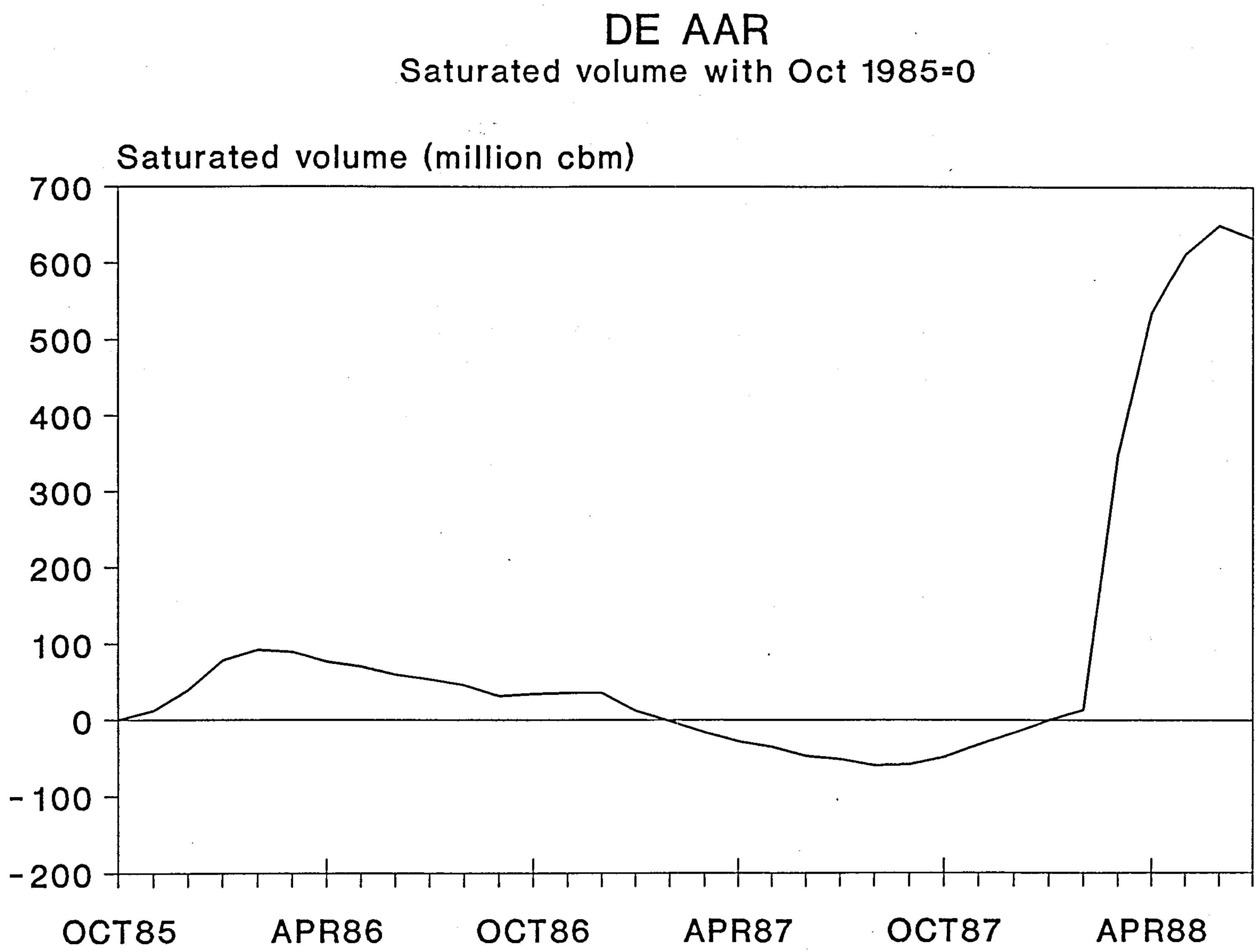


Figure 4.135 The saturated volume of the De Aar aquifer during the study period.

volume of each month from that of October, 1985. A shortened version of Figure 4.135 is given in Figure 4.136. Times where the saturated volumes were equal, can be obtained from this figure.

The SVF-method will now be applied to the De Aar aquifer. The study area, as depicted in Figure 4.133, was divided into three parts, namely (a) a northern aquifer, (b) a southern aquifer and (c) an alluvium aquifer. The ground-water divide between the northern and southern aquifers is evident from the ground-water level contour maps. The southern aquifer comprises 73,5 km^2 and the northern aquifer 62 km^2. Because of the importance of the alluvium surrounding the Brak River, this part of the aquifer (i.e. 8,5 km^2) was also selected for study purposes. The hydraulic parameters of this alluvium aquifer differ considerably from that of the other aquifers. Saturated volume graphs (see Figures 4.137 and 4.138) of the northern and alluvium aquifers, show a similar behaviour than that of the saturated volume graph of the whole aquifer.

Firstly, a specific yield (S) analysis was performed for all the different aquifers (Table 4.7) with program SVF and $\Delta t = 1$ month. It was assumed that a maximum decrease in saturated volume constitutes a "no recharge" period for the same abstraction rate. Minimum and maximum T-values were respectively assumed at the in- and outflow boundaries.

	Area [km^2]	$A_1T_I - A_2T_O$ [m^3/d]	S_{min}	$A_1T_I - A_2T_O$ [m^3/d]	S_{max}	Probable S
Whole aquifer	134	834	$2,8 \cdot 10^{-3}$	-0	$4,4 \cdot 10^{-3}$	$4,0 \cdot 0^{-3}$
Northern aquifer	62	-12	$3,0 \cdot 10^{-4}$	-2517	$5,0 \cdot 10^{-3}$	$1,0 \cdot 10^{-3}$
Southern aquifer	72	0	$5,0 \cdot 10^{-4}$	830	$9,0 \cdot 10^{-3}$	$7,5 \cdot 10^{-3}$
Alluvium aquifer	8,5	1410	$2,3 \cdot 10^{-2}$	0	$6,3 \cdot 10^{-2}$	$3,7 \cdot 10^{-2}$

Table 4.7 S-value sensitivity analysis for the different aquifers.

The alluvium aquifer was taken as part of the southern aquifer. If this alluvium part is omitted, the mean S-value for the southern aquifer becomes $3 \cdot 10^{-3}$. It is interesting to note that the S-value of $3,7 \cdot 10^{-2}$ compares favourably with the S-value for the alluvium of between $3,1 \cdot 10^{-2}$ and $9,5 \cdot 10^{-2}$, as reported by Vegter (1961).

Some special cases for the De Aar aquifer shall now be discussed.

Case 1

Equation (3.4.8) describes the case where the change in ground-water storage is zero (i.e. $\Delta V = 0$):

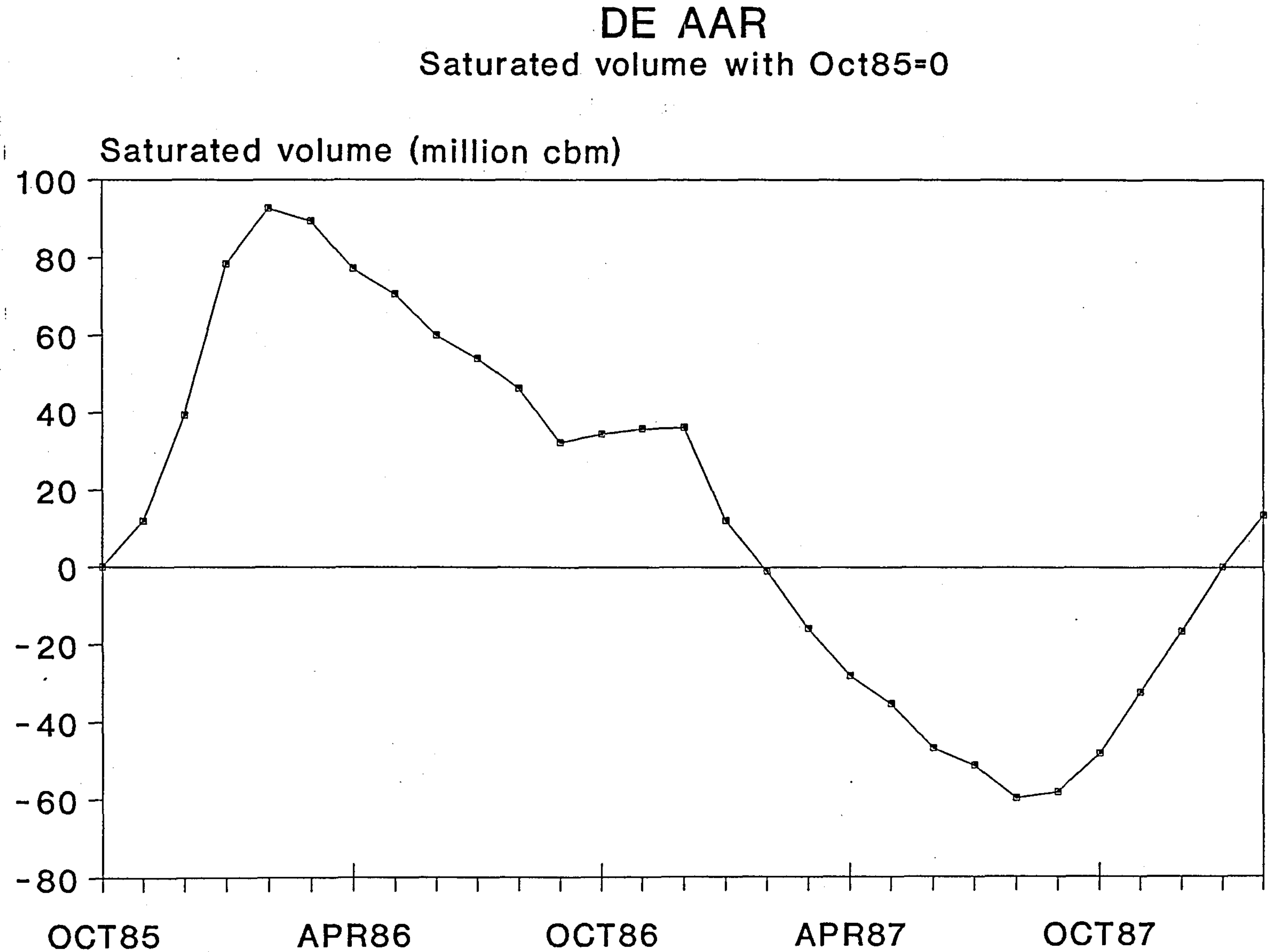


Figure 4.136 The saturated volume of the De Aar aquifer until the flood.

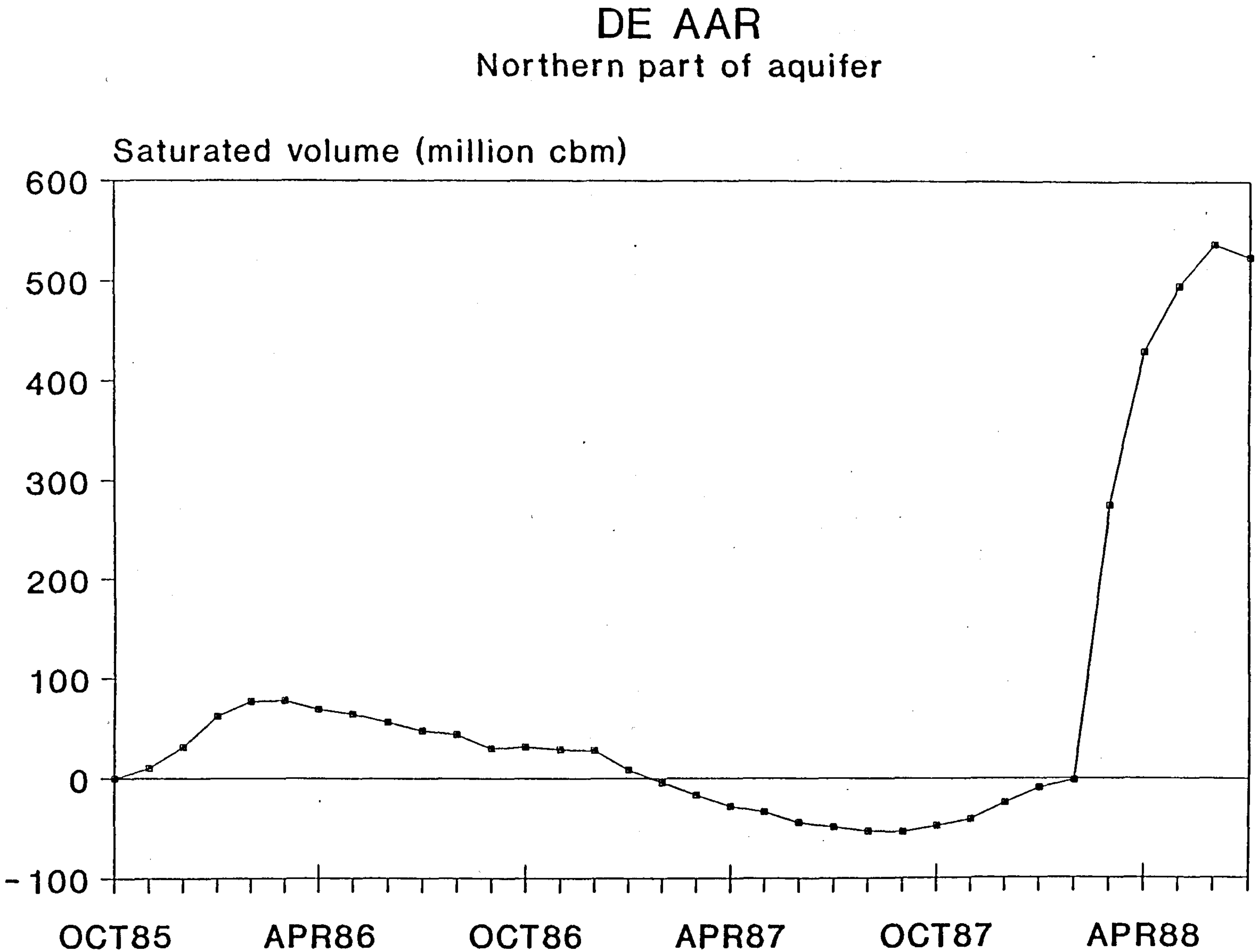


Figure 4.137 The saturated volume of the northern aquifer in De Aar.

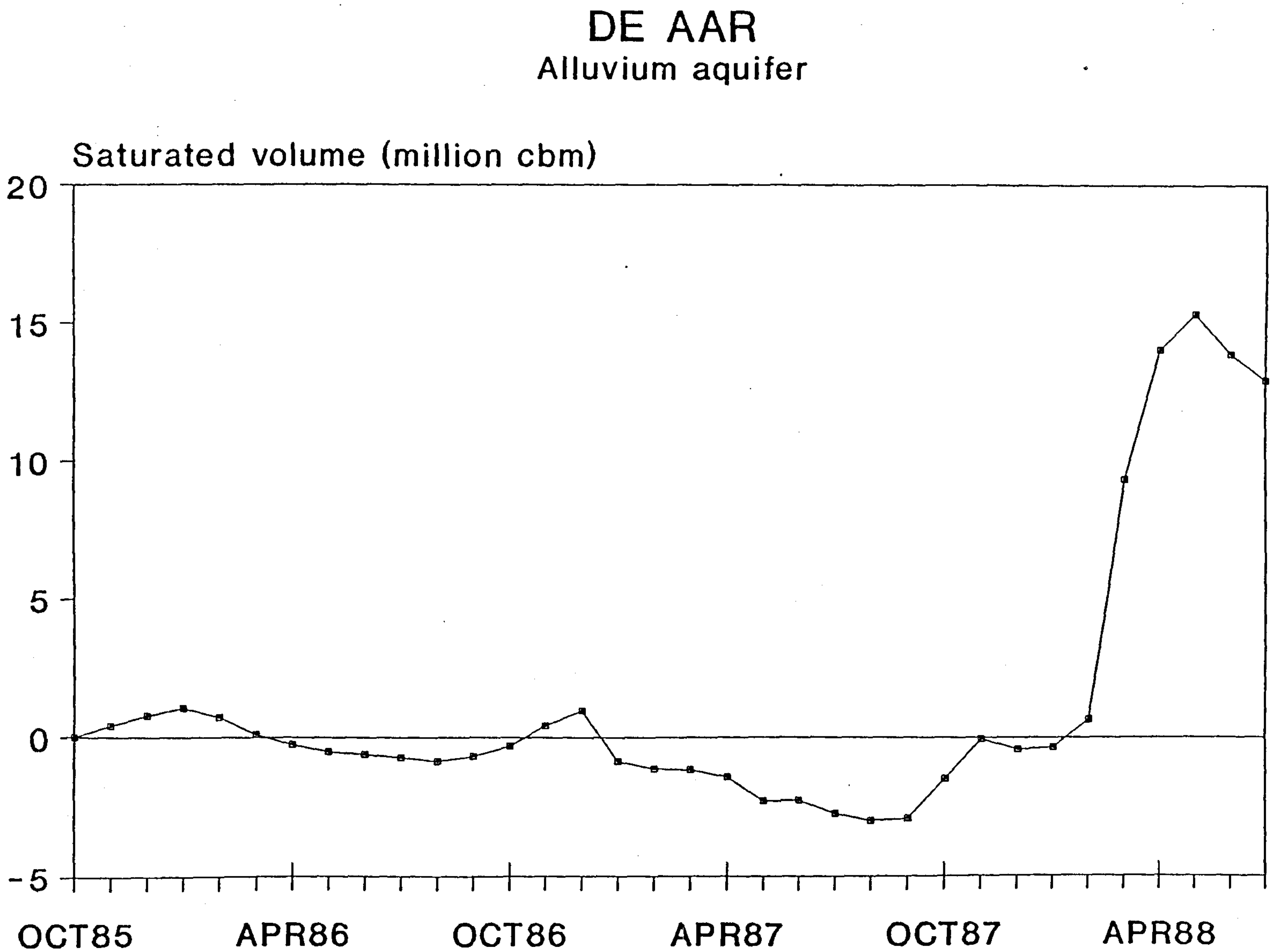


Figure 4.138 The saturated volume of the alluvium aquifer in De Aar.

$$GR = A_2T_O - A_1\ T_I + Q$$

Periods where $\Delta V = 0$ were calculated for all the aquifers. The results are illustrated in Figure 4.139. A regression analysis on these results yielded the following relationship between rainfall and recharge :

Whole aquifer:	$GR = 0{,}025\ (P - 18{,}5)$,	$(r = 0{,}94)$	
Northern aquifer:	$GR = 0{,}012\ (P - 4)$,	$(r = 0{,}96)$	
Southern aquifer:	$GR = 0{,}040\ (P - 25)$,	$(r = 0{,}94)$	(4.6)
Alluvium aquifer:	$GR = 0{,}15\ (P - 16)$,	$(r = 0{,}91)$	

where all values are in mm.

For the average rainfall of 180 mm/a between October, 1985, and October, 1987, the recharge values for the different aquifers are given below:

Whole aquifer:	2,0%
Northern aquifer:	1,2%
Southern aquifer:	3,4 % (i.e. 2,2%, if the alluvium part is excluded)
Alluvium aquifer:	13,6%

The recharge value of 13,6% for the alluvium aquifer may be largely in error, because it was difficult to estimate the amount of lateral inflow of ground water into this aquifer. However, program SVF estimated that the maximum amount of lateral inflow of ground water which could enter this aquifer, was 1 410 m^3/d. For this inflow amount, the recharge would be 8%. If the same calculations for the minimum inflow amount of zero m^3/d is repeated, a recharge value of 29% is obtained. It must, however, be kept in mind that the recharge values for the alluvium aquifer were calculated, by ignoring the surface inflow that occurred at the Vaalbank Poort, where the Brak River enters the alluvium aquifer. This inflow amount of surface water must be added to the rainfall amount, which will led to a lower recharge value relative to the amount of rainfall that occurred. The surface inflow was, however, not known for the study period.

Case 2

For the De Aar aquifer, Equation (3.4.7) can be used for recharge estimations, i.e :

$$GR = Q + A_2\ T_O - A_1\ T_I + A_3\ S$$

Table 4.8 shows that a lag of one month between rainfall and recharge yielded the highest cross-correlation coefficient (see Figure 4.140).

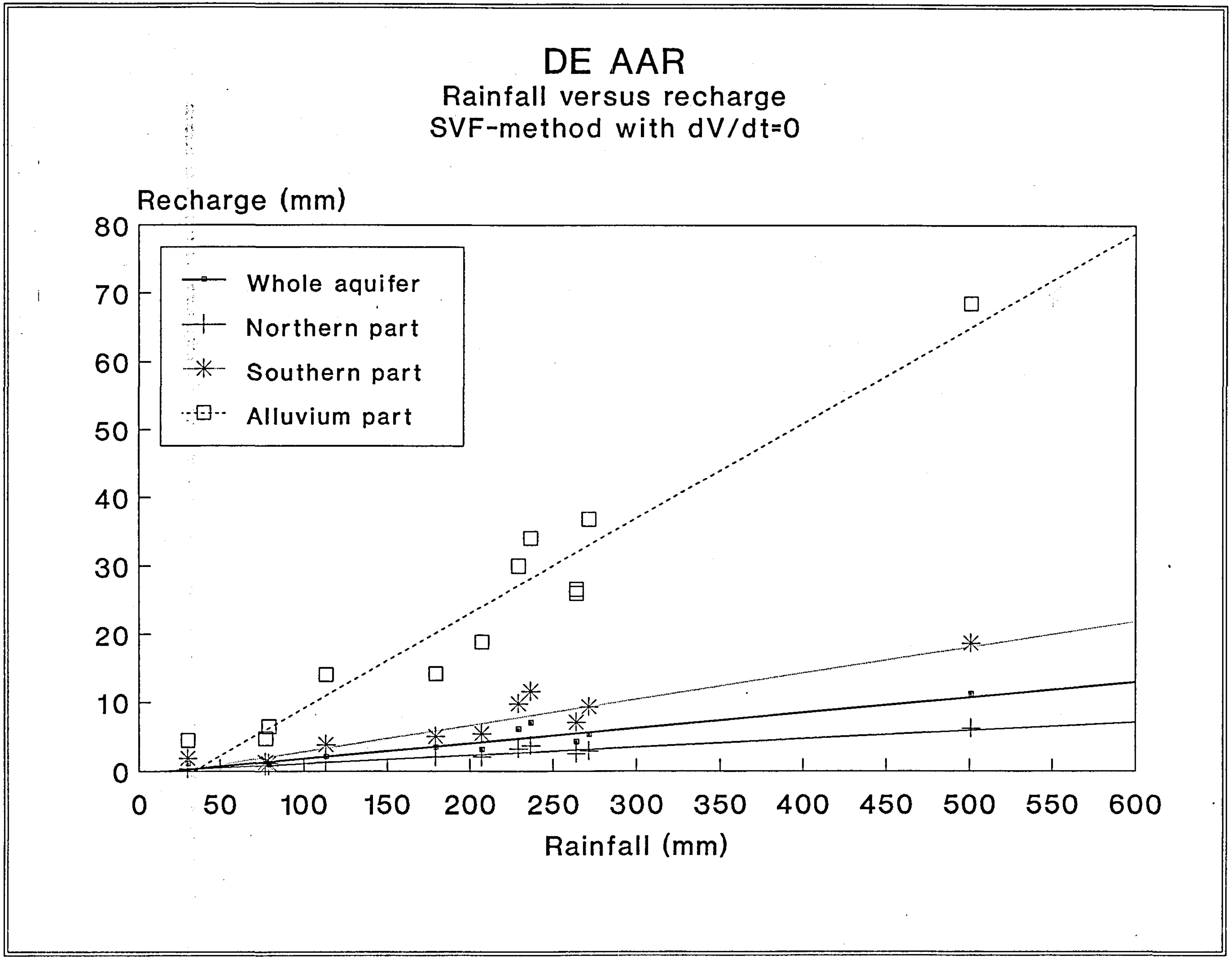


Figure 4.139 Graph showing the relationship between rainfall and ground-water recharge (SVF-method) for the case $\Delta V/\Delta t = 0$ and for different aquifers.

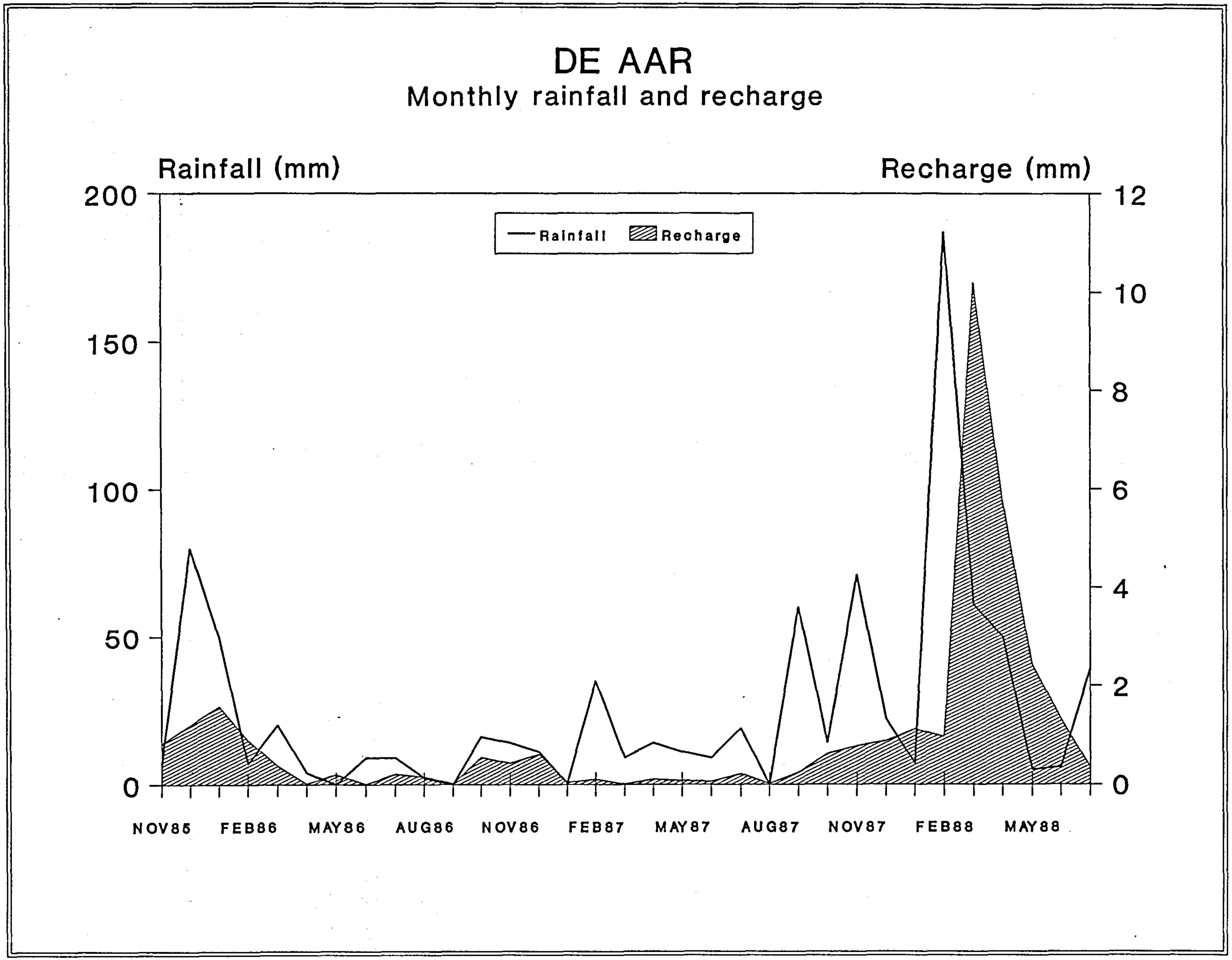


Figure 4.140 Graph showing the lag between rainfall and recharge. The recharge was calculated with the SVF-method with $\Delta t = 1$ month.

Lag (months)	Cross-correlation
0	0,31
1	0,85
2	0,30

Table 4.8. Cross-correlation of the lag between the rainfall and the ground-water recharge.

With a lag of one month, the following relationship between annual rainfall and recharge existed for the study period:

Whole aquifer:	GR = 0,014 (P + 49),	(r = 0,70)	
Northern aquifer:	GR = 0,006 (P + 176),	(r = 0,69)	
Southern aquifer:	GR = 0,035 (P - 13),	(r = 0,86)	(4.7)
Alluvium aquifer:	GR = 0,07 (P +187),	(r = 0,84)	

Figure 4.141 is a graphical representation of the results of Equation (4.7) for the different aquifers.

With Equation (4.7), the mean recharge, as a percentage of the rainfall for the study period, was estimated as:

Whole aquifer:	1,9%	
Northern aquifer:	1,1%	
Southern aquifer:	3,2%	(i.e. 1,3%, if the alluvium part is excluded)
Alluvium aquifer:	14,2%	

These values compared favourably with those obtained with Equation (4.6).

Figure 4.142 shows the cumulative rainfall and recharge as calculated with SVF and $\Delta t = 1$ month. To try to differentiate between recharge that occurred during different periods of the year, the program was run with $\Delta t = 3$ months. Figure 4.143 shows, however, that such a differentiation could not clearly be identified, except for the winter of 1986, when a higher percentage recharge was experienced.

From the rainfall figures in the De Aar region, it is clear that a severe drought has been experienced during the study period. It is therefore expected that the aquifer will replenish at a much higher rate during a normal rainfall year. To accomplish a relationship between the rainfall and the recharge during years of mean rainfall, it was

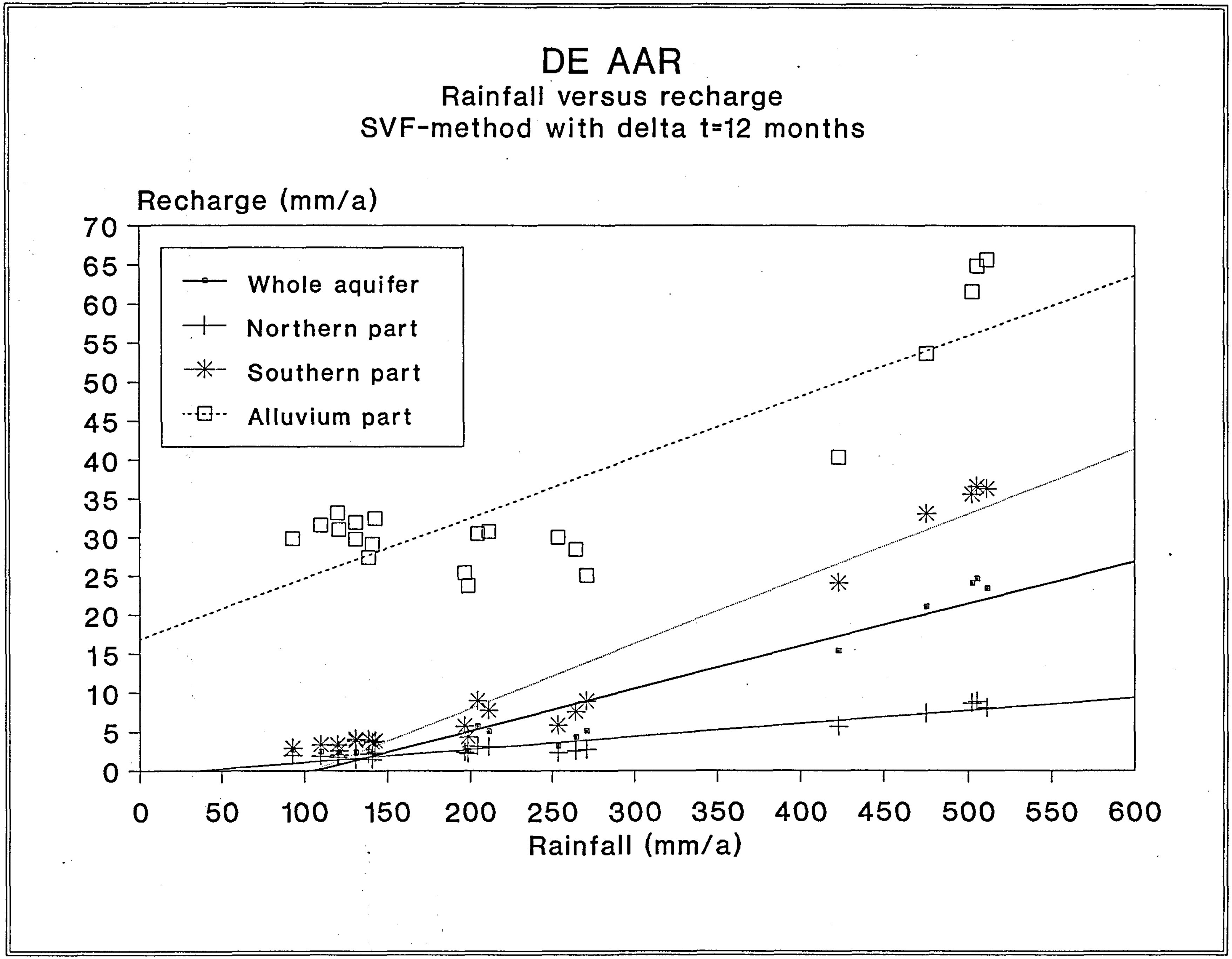


Figure 4.141 Annual recharge and rainfall for De Aar (SVF-method).

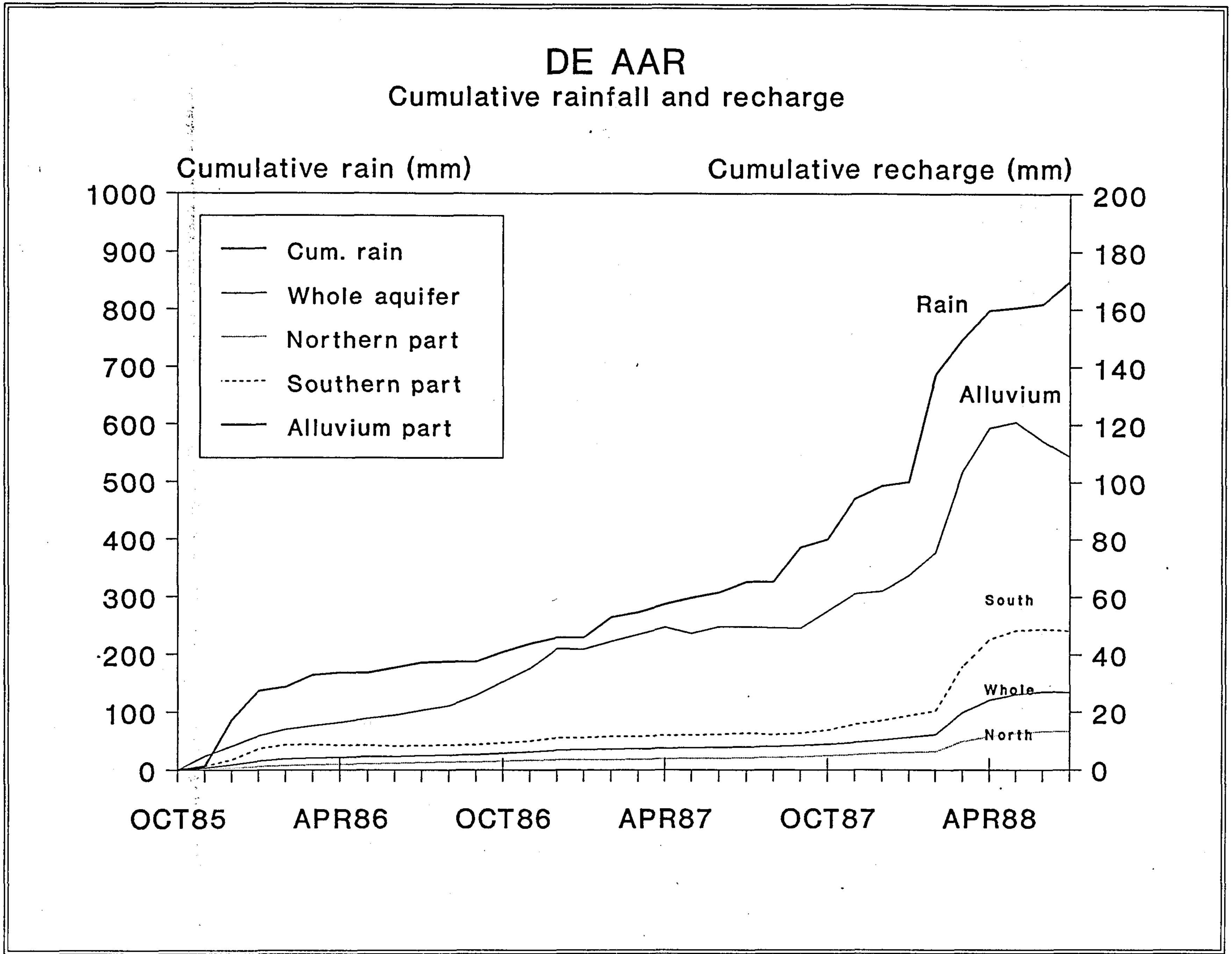


Figure 4.142 Cumulative ground-water recharge for different aquifers in De Aar (SVF-method).

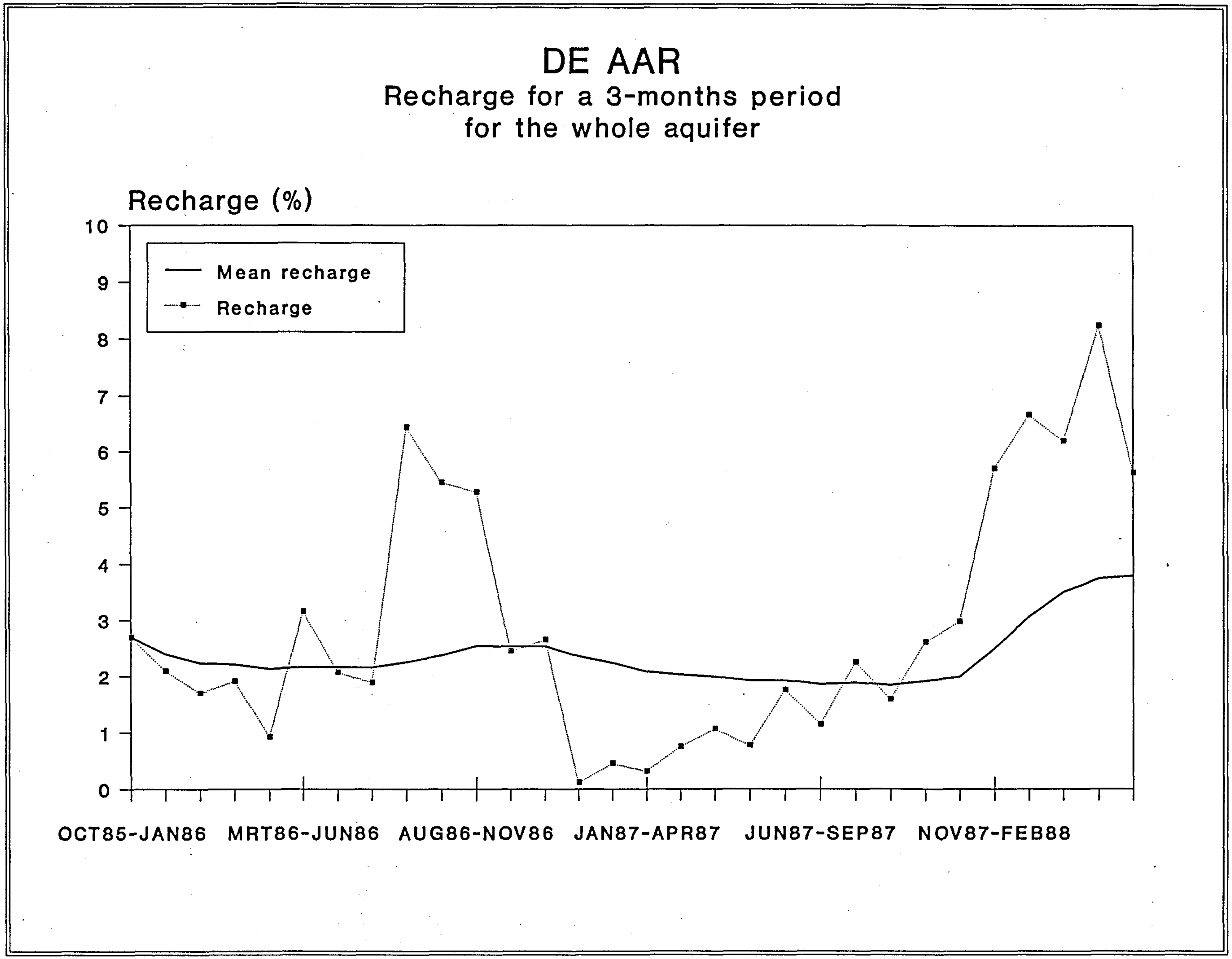


Figure 4.143 Percentage of the three-months' mean, and of the total recharge since the commencement of the project (SVF-method).

decided to include the flood values in the analyses. The following relationships were obtained:

Whole aquifer:	GR = 0,054 (P - 105) mm/a,	(r = 0,96)	
Northern aquifer:	GR = 0,016 (P - 35) mm/a,	(r = 0,96)	
Southern aquifer:	GR = 0,083 (P - 104) mm/a,	(r = 0,97)	(4.8)
Alluvium aquifer:	GR = 0,08 (P + 230) mm/a,	(r = 0,86)	

For an annual rainfall of 287 mm, the recharge will possibly be as follows:

Whole aquifer:	3,7%
Northern aquifer:	1,4%
Southern aquifer:	4,4% (i.e. 2,8%, if the alluvium part is excluded)
Alluvium aquifer:	14,4%

To obtain a recharge estimate for any required period of time, the program was run with different Δt values which included the flood period (Figure 4.144). The corresponding equation reads:

Whole aquifer: GR = 0,048 (P - 27) mm/a, (r = 0,88)

which implies an annual recharge of 4,3% for an average rainfall year.

4.2.6.2 THE GROUND-WATER LEVEL FLUCTUATION METHOD (THE GLF-METHOD)

The recession of ground-water levels can be approximated by:

$$h = h_0 \exp(-\pi^2 Tt / 4r^2S)$$

where h_0 is the original water level, h is the height after time t and r is the half-width between the boundaries. This equation may be applied to any hydrological recession, where the recession represents decline of storage in the absence of replenishment. For De Aar, the maximum mean recession during 30 days was 0,4 m. For r = 6 500 m (half the width of the aquifer in the flow direction) and an assumed aquifer thickness of 20 m, it was calculated that T / S = 11 648. Thus, for S = 0,004 , T = 46 m^2/d.

For the GLF-method, only an analysis on the whole of the aquifer was performed. It was found that the following relationship between rainfall and recharge exists:

With flood:	GR = 0,048 (P - 100) mm/a,	(r = 0,96)
and	GR = 0,016 (P + 13) mm/a,	(r = 0,78)

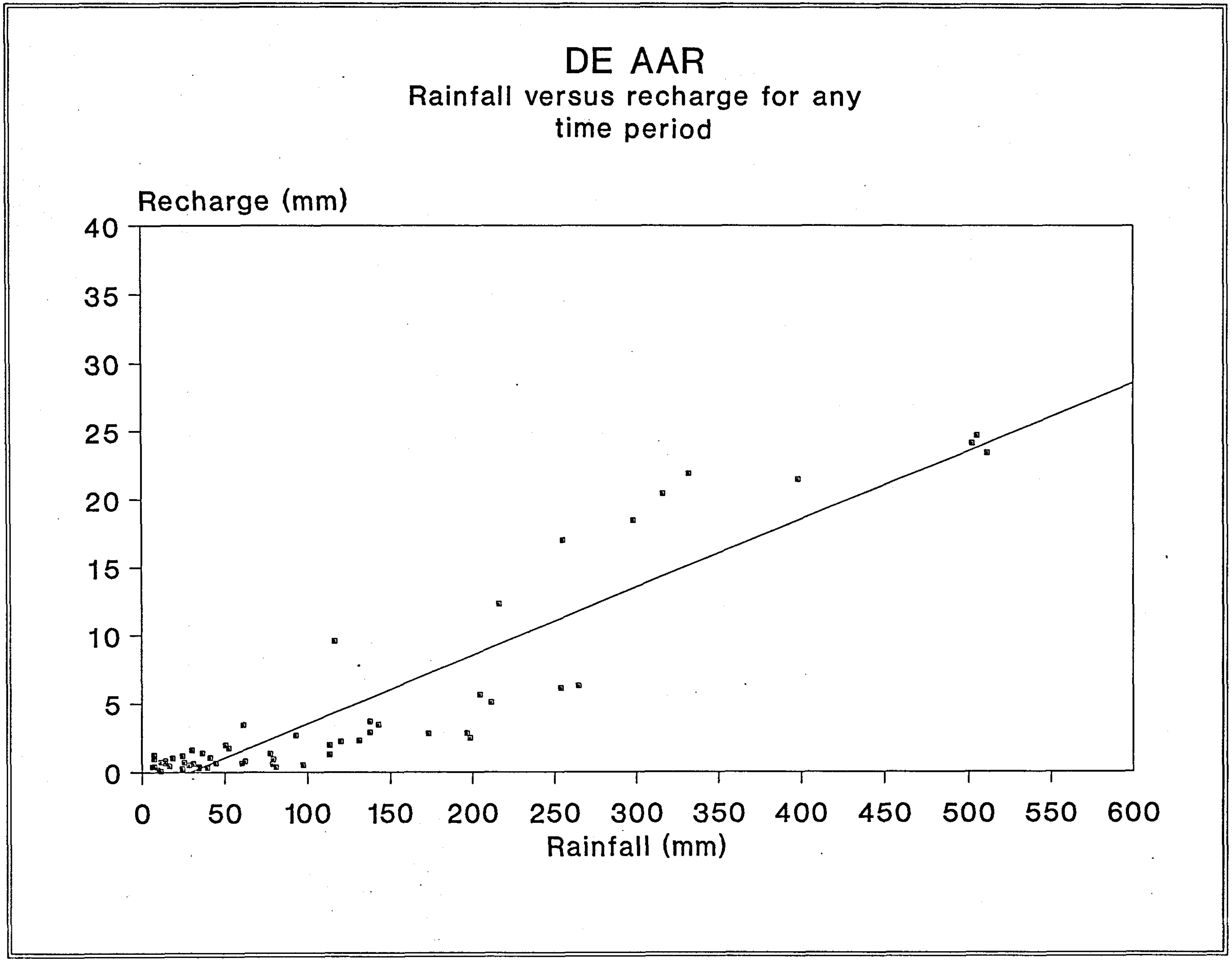


Figure 4.144 Ground-water recharge (SVF-method) obtained by combining different time intervals.

which implies that the mean recharge was 1,8% for the study period. For an annual rainfall of 287, the recharge is expected to be 3,1%, which is less than the average of 4%, as obtained with the SVF-method.

4.2.6.3 THE TWO-DIMENSIONAL FINITE ELEMENT MODEL (INVERSE MODELLING TECHNIQUE)

A triangular finite element mesh (see Figure 4.145) was constructed for the De Aar aquifer. The boundaries were treated as no-flow boundaries, except at two places, namely Vaalbank Poort (where the Brak River enters the aquifer) and at Brandfontein in the north, where ground water flows out of the system.

The first step in the modelling of the De Aar aquifer, was to calibrate the model. This implies that suitable values for S and T must be found which predict the ground-water level responses well. Figure 4.146 represents the final calibrated S-values.

4.2.6.3.1 The February 1988 flood

The February/March 1988 flood of approximately 250 mm caused a rise of the ground-water levels, as shown in Figure 4.147. A definite trend in the rise of the ground-water level in a north-easterly direction is obvious and is ascribed to the fact that the S-values of the aquifer show a descending behaviour in this direction.

The difference between the measured and simulated water levels of April 1988 (Figure 4.148) shows that the model was quite capable of predicting the behaviour of the aquifer, as a consequence of the flood event. The recharge of the aquifer during this period is depicted in Figure 4.149 and varies between 17% in the alluvium aquifer and up to less than 5% in the northern aquifer.

To obtain a clear view of the response of the aquifer to the flood event, a section through the points A and B in Figure 4.134 was constructed. Figure 4.150 shows the ground-water level behaviour, as simulated by the model, if a recharge value of 17 mm, which is evenly distributed over the aquifer with a S of 0,004, is applied. It is noticeable that this S-value, together with the recharge value, does not fit the observed ground-water levels. When a varying recharge and S-value concept were introduced into the model, the results were very promising, as Figure 4.151 indeed shows.

4.2.6.4 RECHARGE VIA THE UNSATURATED ZONE

4.2.6.4.1 The Darcian approach

The determination of the unsaturated K-value is an integral part of the calculations of the Darcy fluxes in the unsaturated zone. As can be seen in a previous chapter, this unsaturated K is related to the saturated K-value (K_S), which can be determined in the

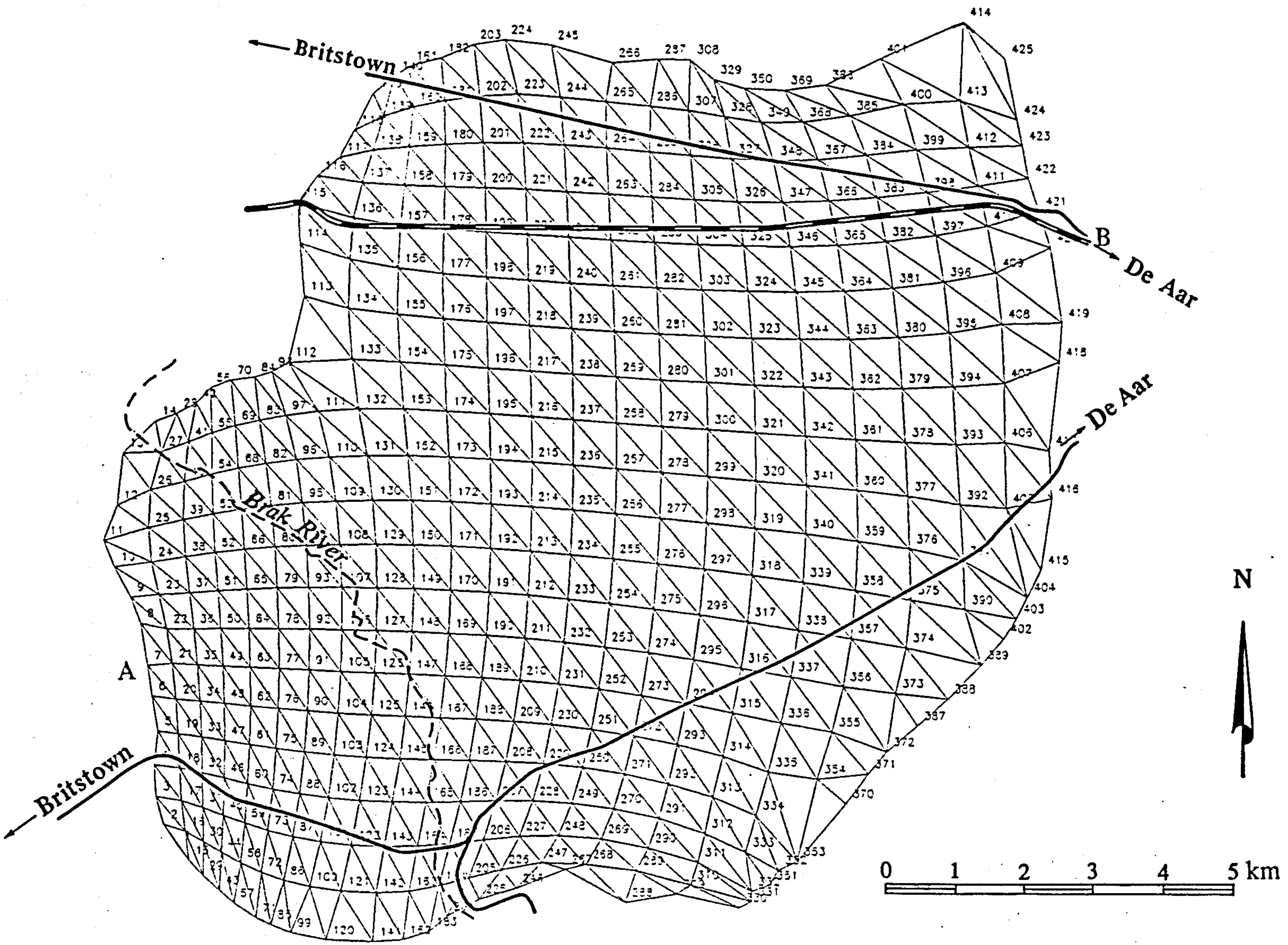


Figure 4.145 Triangular mesh constructed for the De Aar aquifer for the SVF-method. A - B is a section through the aquifer.

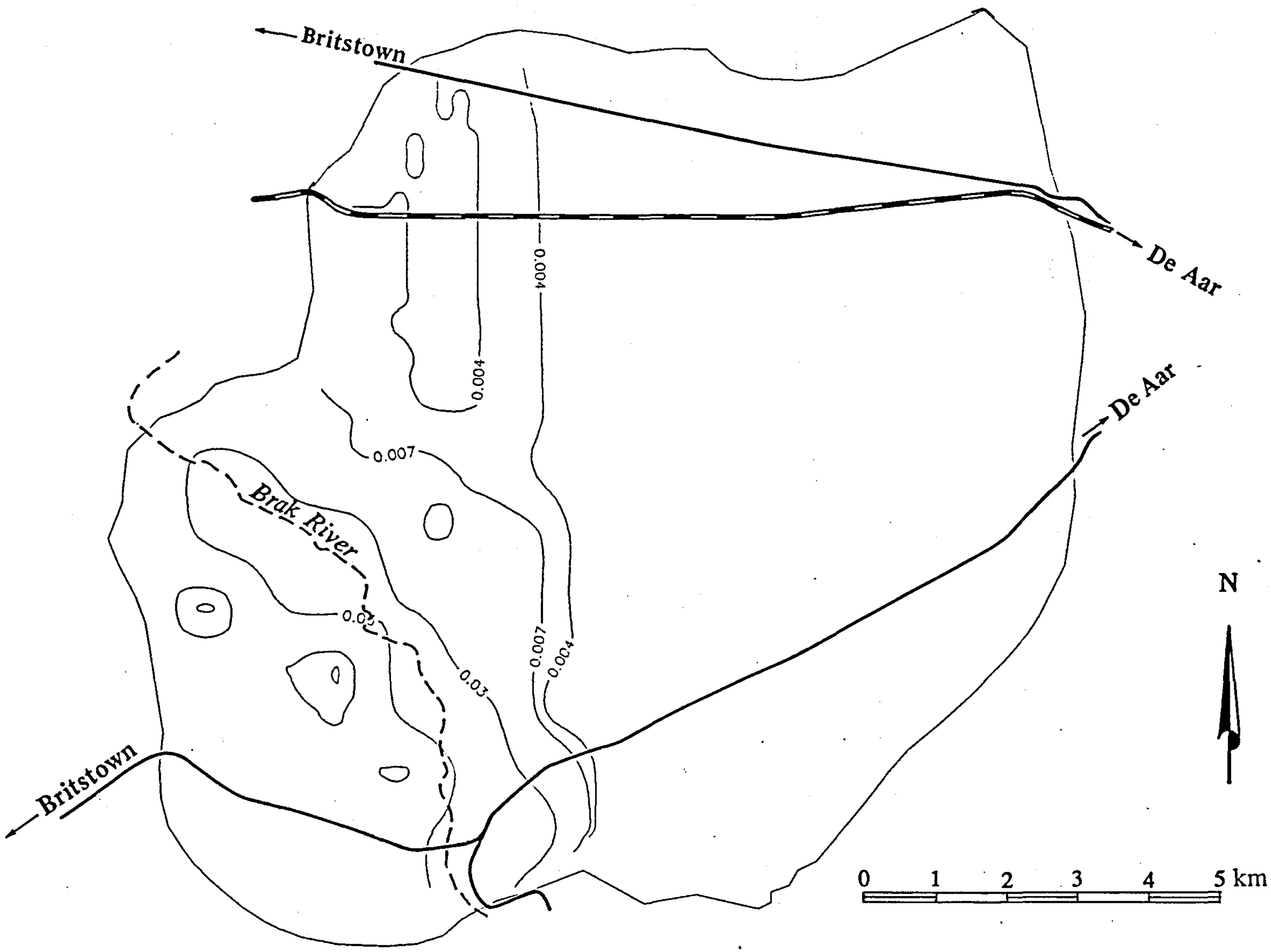


Figure 4.146 Final calibrated S-values obtained with the two-dimensional finite element model.

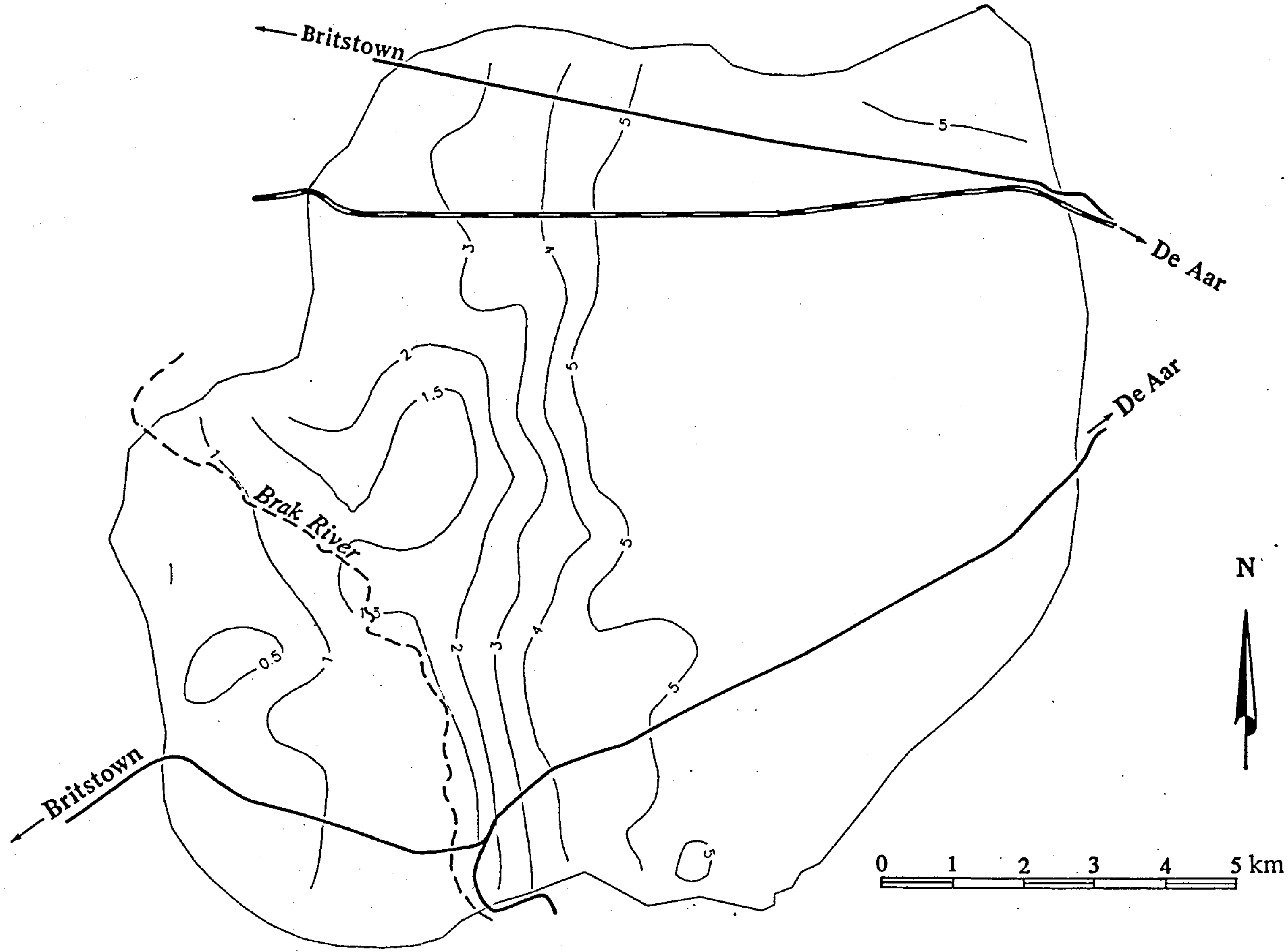


Figure 4.147 Rise of ground-water levels as measured 90 days after the flood.

Figure 4.148 Difference between simulated and actual ground-water levels as measured 90 days after the flood.

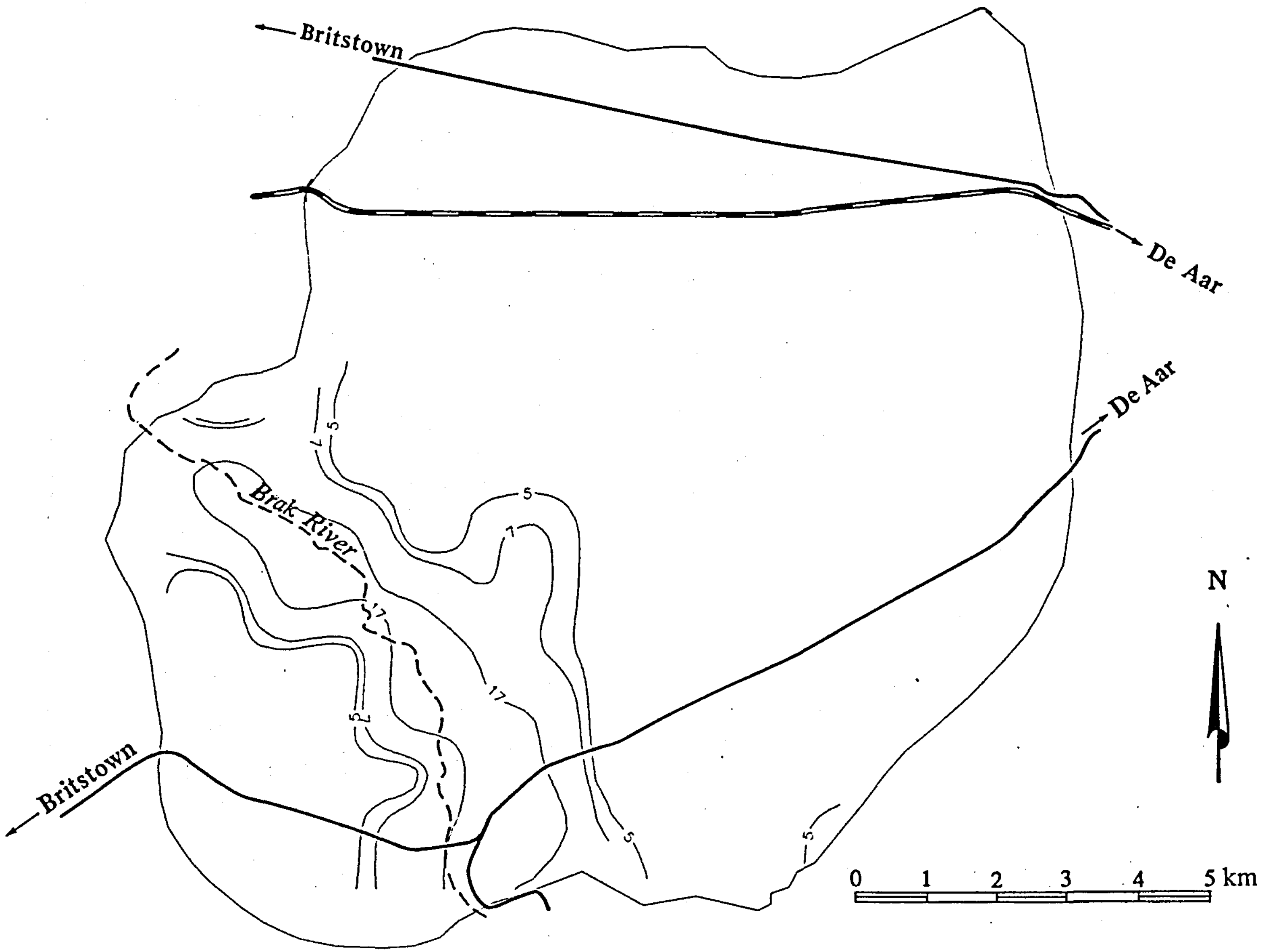


Figure 4.149 Percentage ground-water recharge as obtained with the finite element method for 90 days.

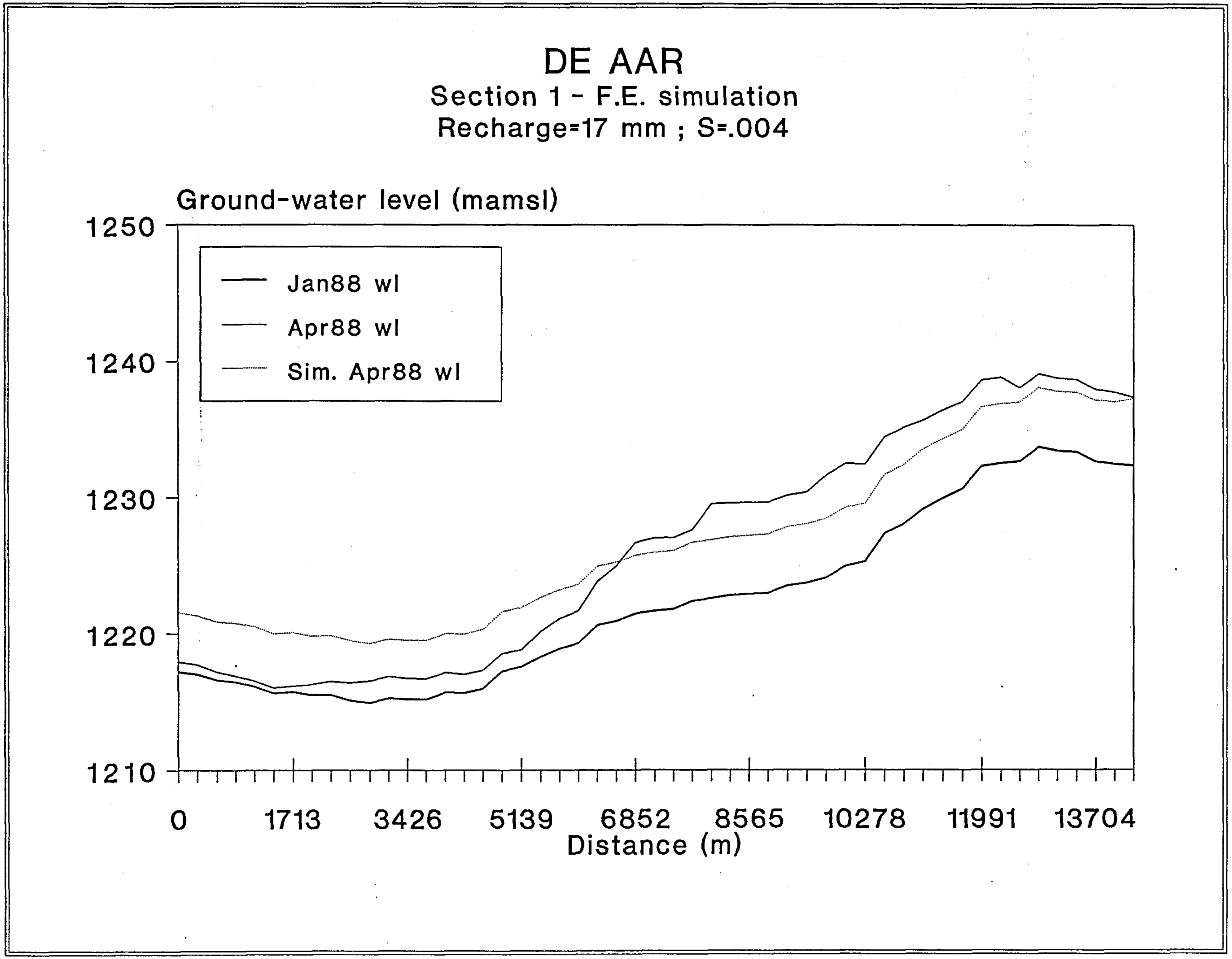


Figure 4.150 Finite element simulation for section A - B for a constant S-value of 0,004 over the whole aquifer and a constant recharge of 17 mm during 90 days.

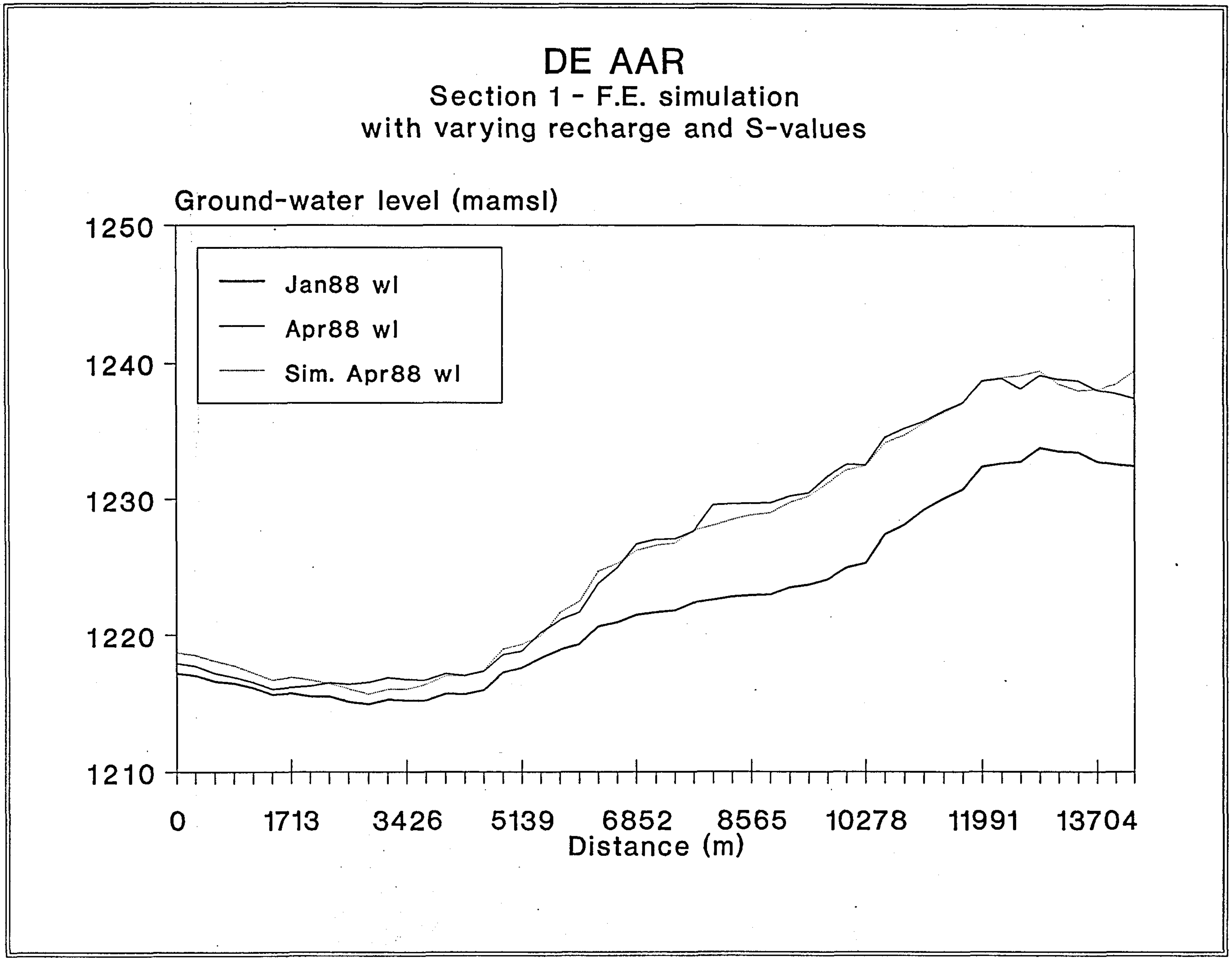


Figure 4.151 Final calibrated finite element simulation for section A - B after applying varying S- and recharge values.

laboratory (with a constant head permeameter) or in the field (e.g. the auger hole method, the piezometer method and the infiltrometer method (Kessler and Oosterbaan, 1974)).

Infiltration of water into a bare soil can be markedly reduced by a seal formed on the soil surface, as a result of aggregate breakdown and compaction by impact of raindrops. Hydraulic conductivity of the crust may be a few to 2 000 times smaller than that of the original soil (McIntyre, 1958). Rainfall on many types of soils causes gradual formation of a surface seal. The formation of the seal can occur within 15 minutes after rainfall started (Ahuja and Ross, 1983) and the permeability of the seal decreases exponentially with time.

At the start of infiltration, the length of the transmission zone is small. This means that at the beginning of infiltration, the gradient (i.e. $(d + \psi + z)/z$) is large which leads to a high infiltration rate. (The d in the gradient is the ponding depth and ψ is the suction at the bottom of the transmission zone). The value of ψ is equal to zero at the interface between the transmission and unsaturated zone. As the length of the transmission zone, z, increases, the gradient decreases up to a point where the length of the transmission zone is large, relative to ψ+d, and reaches a value of 1, and the infiltration becomes constant, attaining what is known as the final (or basic) infiltration rate, which is equal to the hydraulic conductivity of the transmission zone. For wet, medium and heavy textured soils in which the hydraulic conductivity of the transmission zone is approximately the same as in the saturated zone, one gets that $q = K_S$. For studies on the basic infiltration rate in most soils, infiltrometer measurements should extend for a long enough period to permit a constant infiltration rate to be obtained. This may take quite some time in dry clay soils, because a decrease in the infiltration rate may be caused by the decrease in the hydraulic gradient, as well as by a change in the hydraulic conductivity due to swelling. The use of an infiltrometer in layered soils where a specific layer is underlain by a soil layer with a lower hydraulic conductivity, may be in question,because lateral movement will become dominant on the contact of these two soils, and the calculated K-value must only be regarded as a fair approximation of the actual K-value. For ground-water recharge calculations, the use of an infiltrometer is not beyond any doubt, mainly because it is impossible to insert the infiltrometer to a depth below the root zone.

For the calculation of Darcian fluxes with the aim of recharge calculations, it is important that the different parameters (water content, hydraulic conductivity values and retentivity values) be measured at a point lower than the depth of the active root zone of the plants. To compromise for capillary rises, this point must be at least 200 mm below the end of the root zone. This point for the De Aar soils is in the order of 0,8 m. Figure 4.152 shows the water content at four sites for certain selected times during the study period. It is obvious from this figure that, while some of the tubes have shown an increase in water content below 0,8 m during the February 1988 flood, others showed no sign of a moisture increase.

Undisturbed soil samples were taken at this depth and the K_S values were determined. As was the case with the soils of Dewetsdorp, the K_S-values were found to be very small. A sample taken at neutron probe 22 in the alluvium aquifer, yielded a K_S value of $1 \cdot 10^{-4}$ m/d for the first seven days after the experiment has started. For the following 14 days, no measurable amount of water flow occurred. The use of the soil bulk density in the Naney equation, yielded a K_S value of $4{,}3 \cdot 10^{-3}$ m/d. The neutron reading at probe 22 shows an increase in water content during the flood (Figure 4.152), so that movement of water through the soil matrix actually has occurred. Figure 4.153, which is a graph of the total hydraulic potential just before and after the flood at this site, confirms this vertical movement of soil water. The Darcy approach yielded the following results:

$$q = -K_u d\phi/dz$$

where $K_u = 1{,}0 \times 10^{-4}\ (0{,}07) = 7{,}0 \times 10^{-6}$ m/d

then

$$q = -7{,}0 \times 10^{-6}(-6{,}2)$$
$$= 4{,}3 \times 10^{-5} \text{ m/d}$$

A summation of these fluxes, for 30 days, yielded a value of 1,3 mm recharge. The only explanation for this low value is that the calculated K_S-value is only representative of the soil matrix and not of the cracks that may be present in the layer. Discussion with personnel of the Department of Soil Science at the UOFS, strengthens the belief that recharge has been taken place along specific pathways. Neutron readings in a cluster of probes, situated in irrigation land near Hoopstad, showed a similar behaviour. While some sites showed an increase in water content, others, within a few metres, did not show any increase in water content at a certain depth. They found that downward infiltration occurred, until a layer with a higher clay content (i.e. a lower K_S) was reached, from where the movement of soil water is predominantly lateral. One can imagine that the soil water will continue to move lateral, until cracks in the specific soil layer occurred along which the water could flow further downwards.

The undisturbed soil sample taken at tube 20, yielded a K_S-value of $1{,}4 \cdot 10^{-4}$ m/d during the first seven days. After four weeks of applying a constant water pressure head of 20,1 m, the K_S was calculated as $4{,}2 \cdot 10^{-5}$ m/d.

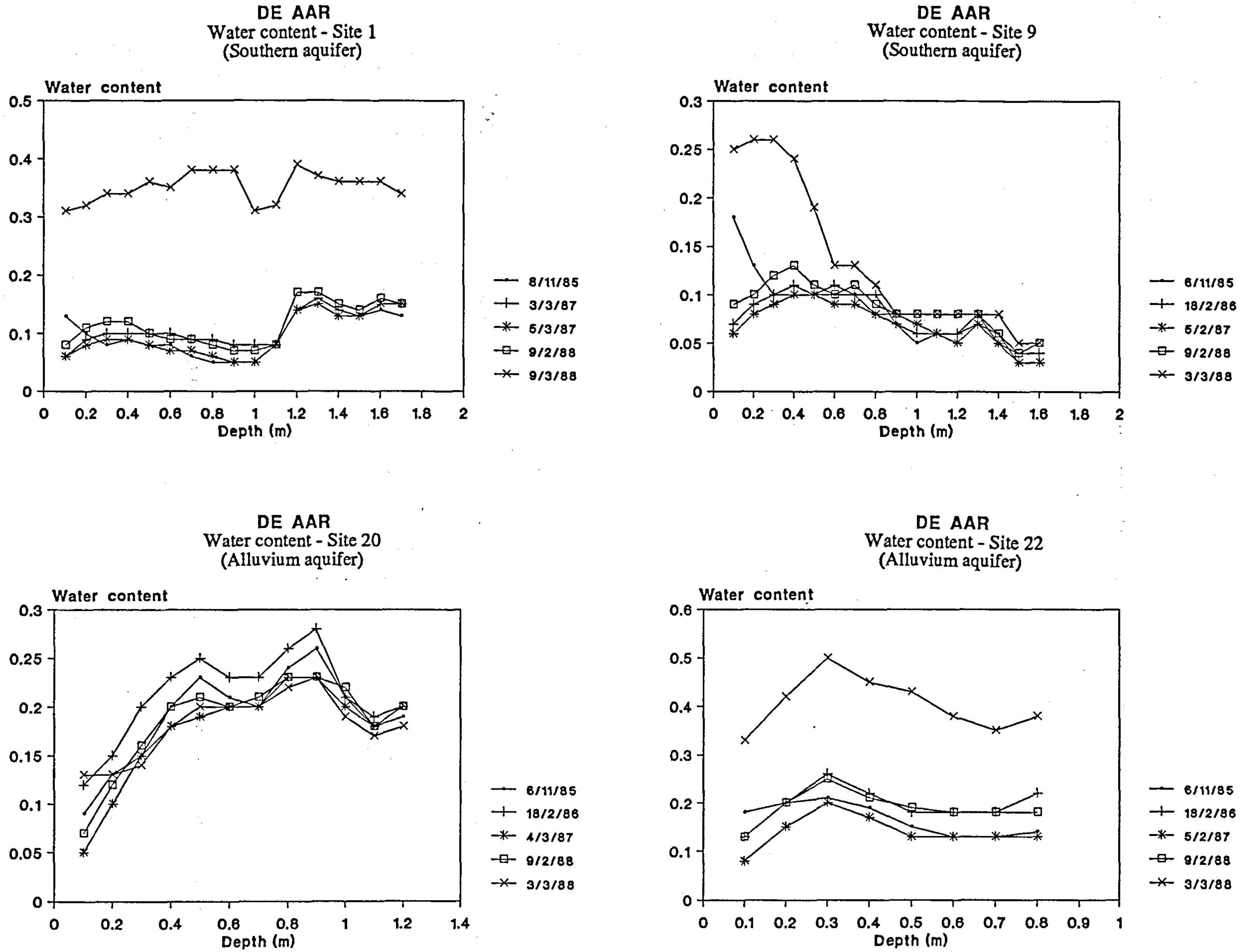


Figure 4.152 Water content at four neutron access tubes in De Aar.

DE AAR

Site 22 : Total potential before and after the flood

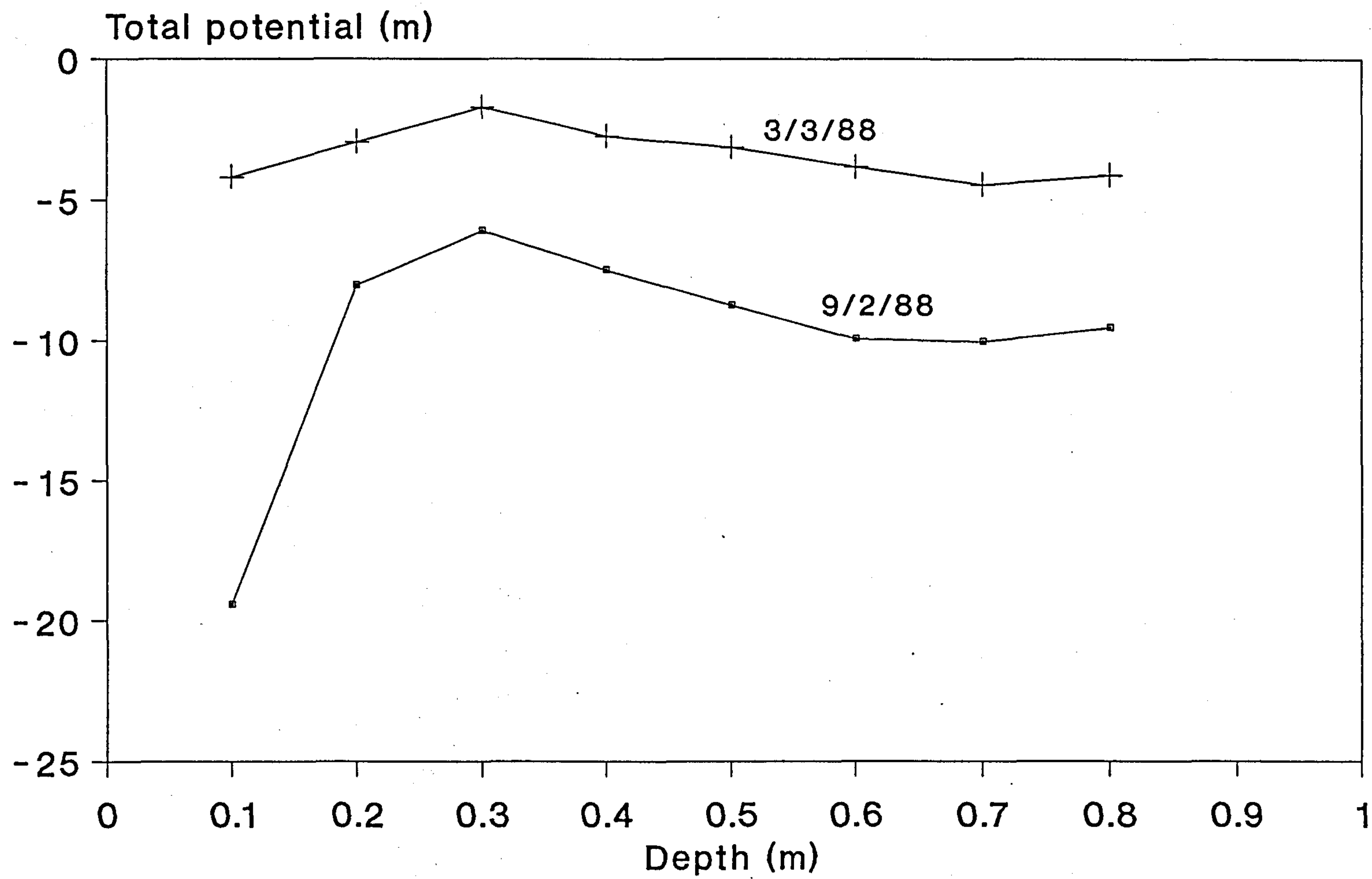


Figure 4.153 Total potential before and after the flood at site 22 showing that soil water flow had occurred.

Because it would be impossible to calibrate the vertical soil water flow model (SUFF) with these saturated K_S-values, as calculated in the laboratory, and the measured water content values, it was decided not to apply SUFF to the De Aar aquifer.

4.2.6.4.2 The soil water balance method

In the original report, Schulze (1984) stated that the ACRU-model, by virtue of its structure, is a small catchment lumped model for use on catchment areas under 10 km^2. The reason for this restriction, is that ACRU uses the SCS-method for the calculation of stormflow and discharge, which is only applicable for small catchments under 8 km^2.

Although the De Aar aquifer comprises of a catchment area of 134 km^2, the ACRU-model was, nevertheless, run for this aquifer. Figures 4.154 and 4.155 show the FC and WP values, as calculated with the texture equations. As was the case with the Dewetsdorp aquifer, we found it impossible to calibrate the ACRU model; the reason being that the main flow mechanism does not favour this model.

The MOUSE model was used but without any success. This model uses the saturated K_S-value of the unsaturated zone and because of the very low saturated K-values (see the previous section), the MOUSE model calculated that no water flow occurred at a depth of 0,8 m below the surface.

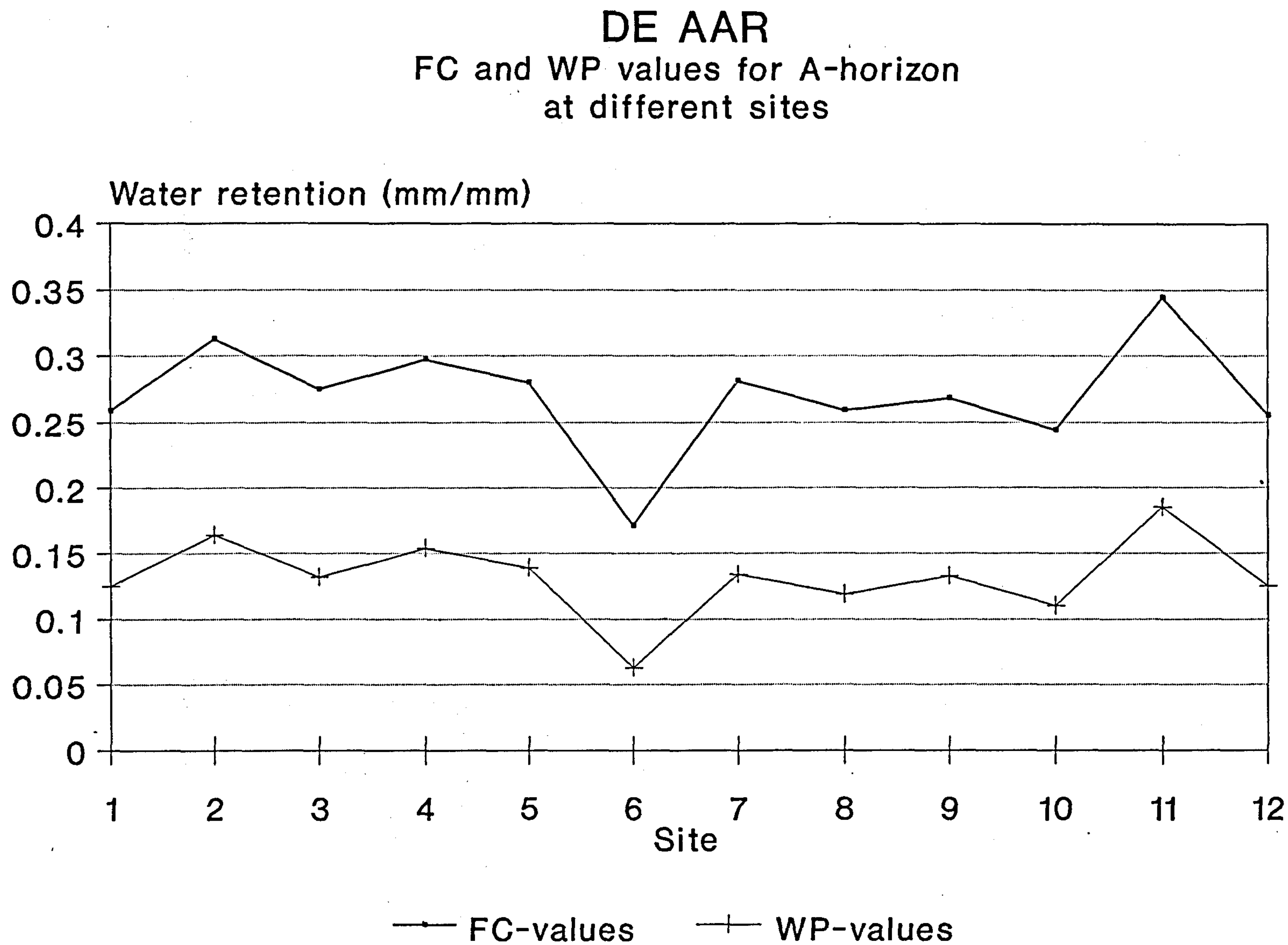


Figure 4.154 Field capacity and wilting point values for the A-horizon in De Aar.

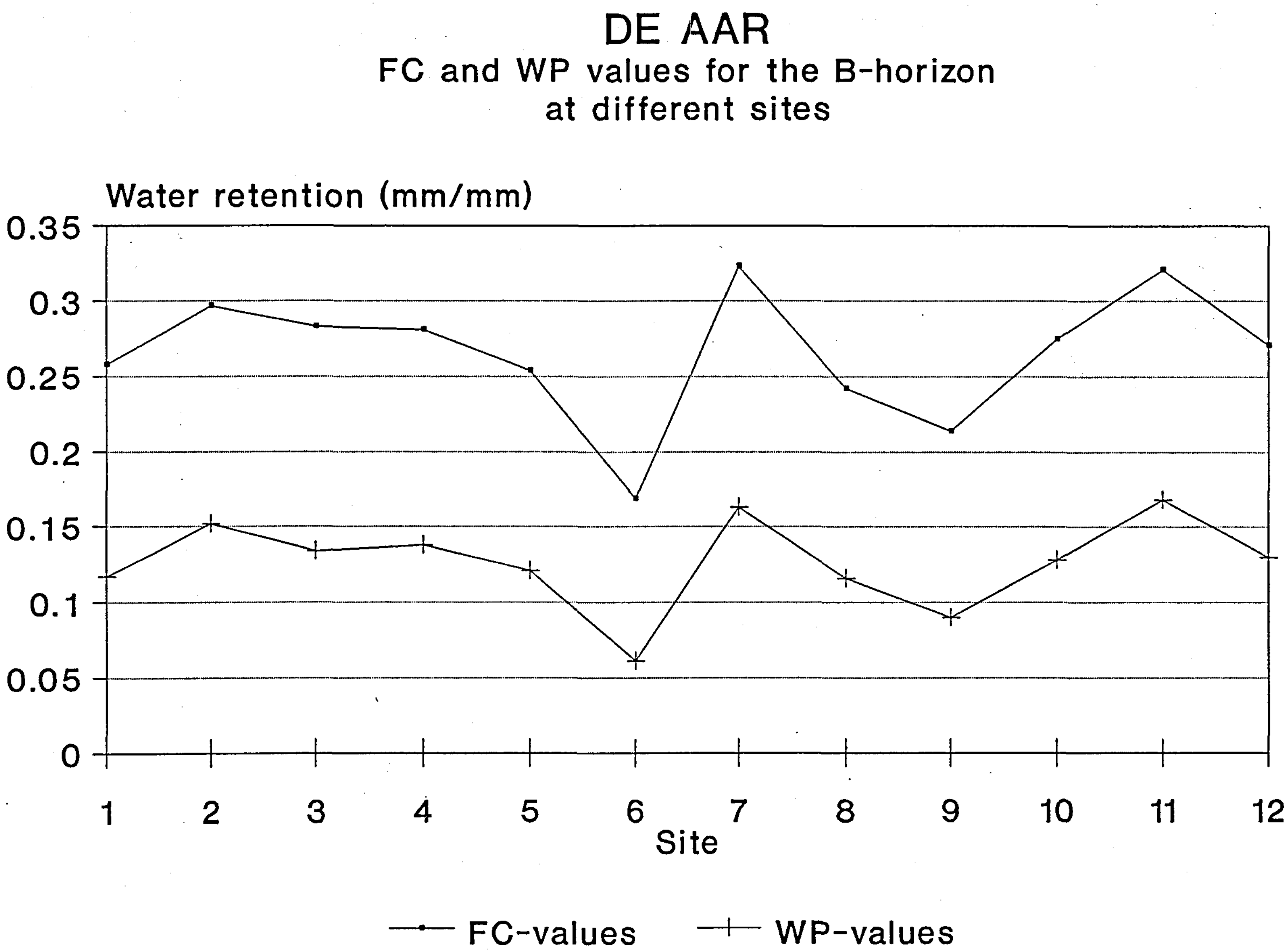


Figure 4.155 Field capacity and wilting point values for the B-horizon in De Aar.

5 ECONOMIC UTILIZATION

In the contract, the exploitation potential is defined as the long-term yield which depends on recharge due to rainfall, underground in- and out-flow and ground-water storage.

The exploitation potential of an aquifer depends, to a large extent, on the (average) recharge and storage capacity, as well as the quality of the water dealt with in the previous chapter. The geology of the area is also of importance, because it controls both the geohydrological exploration methods and the number of independent abstraction points that can be sunk. The exploitation potential is influenced further by the method used to abstract the water, e.g. boreholes, wells, galleries, kanats or well points, as well as by changes in the evapotranspiration which, to a certain extent, is or can be controlled by man. The abstraction method, together with the number of abstraction points, has an influence on the percentage of the available water that can be extracted, but it influences also the cost at which the water can be supplied.

5.1 EXPLOITATION POTENTIAL

The cost of a water-supply scheme depends on a number of factors. These are:

(i) Cost of the geohydrological exploration.
(ii) Drilling cost.
(iii) Construction cost (servitudes, access roads, pipe- and power lines, pumping equipment, reservoirs, booster stations, water treatment works, housing).
(iv) Operational cost (energy, chemicals, personnel, travel, maintenance).
(v) Overheads (administration, data capture, aquifer management).

The cost of the geohydrological investigation depends on the geological circumstances and may become expensive, if extensive geophysical surveys and exploration drilling are required. For single boreholes or very small schemes, they may become prohibitive. For medium size and large schemes, they constitute generally only a negligible fraction of the total cost.

Drilling and borehole development costs depend on the geological formation, the required depths of the holes and their yield, which determine both the diameter of the holes and their number, as well as necessary casing or filter requirements.

Construction costs are dependent on the distance of the well field from the consumer and the aquifer parameters which, together with the geology, determine the spacing of the holes. These factors have a bearing on the length of servitudes, access roads, pipe- and power lines. Geologically controlled circumstances, like accessibility, thickness of overburden and hardness of rock will also influence construction costs. Lastly, thicker pipeline walls or booster stations may be needed, because of the head between well field and reservoir and the chemical quality may require special pipeline materials to be used because of corrosiveness of the water. To some degree, the cost of the three items discussed above, may also depend on the remoteness of the area.

Operational costs consist of (a) energy costs which depend on pumping water levels, elevation differences, friction losses, efficiency of pumps and, to some degree, on the way the scheme is operated, (b) water treatment cost which depends on the quality and (c) miscellaneous costs, e.g. personnel, travelling, maintenance and size of scheme.

Overheads, like those mentioned above, are normally not calculated separately, but a fixed amount or a certain percentage is added to the calculated cost of the scheme.

It cannot be the task of the *Klein Akwifeerprojek* to make an evaluation of the economic feasibility of any envisaged scheme and include the results in a formula, which classifies the scheme as e.g. highly feasible, acceptable or totally unfeasible. This investigation has concentrated on the exploitation potential *sensu stricto*, i.e. on the determination of the average nett recharge. According to our results, it varies with rainfall. Depending on the acceptable percentage of time a proposed scheme fails, a certain minimum recharge can be calculated. This is done by inserting the minimum rainfall with the respective recurrence period into the recharge equations.

In the following sections of this chapter, a short description of the different aquifers found in the Karoo Basin is given. This will indicate whether the same exploration methods can be used throughout the Karoo. Together with a statistical evaluation of existing boreholes about depth and water level as well as the yield, this information can be used to judge the variability of drilling conditions in the Karoo. Taking the results obtained in the Dewetsdorp study area as an example, it will finally be shown how the utilization of an aquifer can be improved.

5.1.1 OCCURRENCE OF GROUND WATER IN KAROO ROCKS

The permeability of fresh, non-fractured, joint-free Karoo sedimentary rocks is generally in the order of 10^{-8} m/s and less. With the exception of a few sandstones, this applies to the whole range of argillaceous to arenaceous types, comprising the Karoo Sequence - diamictites, shales, mudstones, siltstones, arkoses, greywackes and sandstones.

Karoo strata owe their water-bearing properties to the fracturing which is generally limited to the top 30 metres or so below the surface. Weathering processes play an essential role in opening incipient fractures/joints and in converting the rocks into more porous media.

Except for a belt of gently folded strata in the far south, the main Karoo basin has been extensively intruded by dolerite dykes and inclined and undulating dolerite sheets or sills. These intrusives are hydrogeologically very important, as the more prolific water-yielding structures are associated with them. These are the following:

Weathered and jointed sedimentary rock alongside dykes.

Weathered and fractured dolerite dykes.

Weathered and fractured upper and lower contact zones of dolerite sheets.

Well-jointed zones of sedimentary rocks also occur in association with kimberlite or lamprophyre dykes and pseudo-coal veins in the south.

Gentle folding in the far south has resulted in more or less uniform well-jointed strata.

Karoo aquifers have small storage cavities, owing to the low storativity as well as the small thickness. The most productive aquifers are found where water-bearing alluvial deposits overlie the fractured and weathered strata.

5.1.2 STATISTICS

In his research regarding the National Ground-water Data Base, Cogho (1989) did a statistical analysis of geohydrological data of the Orange Free State. He reported the following statistical data:

	No. of observations	Mean	Standard deviation
Depth [m]	8219	44.62	24.57
Water level [m]	5152	12.75	11.65
Yield [l/s]	4988	1.72	

The respective statistics for Dewetsdorp yield:

	No. of observations	Mean	Standard deviation
Depth [m]	116	44.89	29.47
Water level [m]	158	10.36	6.47
Yield [l/s]	70	1.37	

For the De Aar research area, the following values were obtained:

	No. of observations	Mean	Standard deviation
Depth [m]	179	25.50	24.35
Water level [m]	50	16.00	8.17
Yield [l/s]	69	2.90	

Mean and standard deviation of the Dewetsdorp data compares well with the larger sample from the whole Orange Free State. The shallower mean depth and the higher mean yield of the De Aar boreholes can be ascribed to the fact that a considerable number of these boreholes are drilled in the alluvium of the Brak River, where stronger supplies are found at shallower depths.

If only the holes are taken into account which were drilled during this research program and outside the alluvium, the statistics hereunder are similar to those of the whole Orange Free State:

	No. of observations	Mean	Standard deviation
Depth [m]	54	51.98	25.98
Water level [m]	47	15.62	8.05
Yield [l/s]	54	1.08	

The statistical evaluation of these parameters indicates therefore that the geohydrological conditions in the two research areas are representative for, at least, a fairly large area in the Karoo, if not for most parts of the Karoo.

For square degree 3025 in the southern Orange Free State, sufficient geological information was available for statistical evaluation. Of 217 boreholes for which aquifers were identified, 119 were drilled in Beaufort sediments, 94 in dolerite and four encountered water in sediments and in dolerite. If this is compared with the geological and the geohydrological information on the Site-ID reports in the Appendix, a similar distribution between sedimentary and dolerite aquifers is obtained, showing again that the geohydrological conditions in the two research areas are representative for a larger area in the Karoo.

5.1.3 ECONOMICS

5.1.3.1 INTRODUCTION

The ground-water supply of aquifers in the Karoo is traditionally developed to satisfy fully, or in part, the municipal, industrial or agricultural water demands. The planning problems associated with the management of these ground-water supplies may include the following:

(i) the determination of the optimal pumping pattern, i.e. the correct spacing of all the pumping wells in the aquifer, together with the pumping rate and time necessary to satisfy the given water demand;

(ii) the staging of future well field development, if an increase in demand should occur; and

(iii) the design of a surface storage facility and pipe network to transport and distribute the ground-water supply to the various demand points.

To increase the economic efficiency of a ground-water supply project, all the abovementioned factors must be taken into account. These projects are often large and expensive and require long lead times. Once the facilities have been created, they are often operated for decades. Pipelines, storage reservoirs, water treatment facilities, electrical power lines and distribution networks are examples of such expensive long-lived investment projects. The uncertainty as to the level of service these facilities will need to provide in say 20 or even 50 years from when they are planned and implemented, makes the project evaluation and selection process very difficult.

During the last 20 years, the use of optimization techniques for the planning, design and management of ground-water supply schemes, came into practice. Such an analysis may involve many decision variables and constraints. Especially, the usage of computers together with an optimization technique and flow model, helps to provide decision alternatives which are optimal in some defined sense and which can be used by water resource managers to assist their decision-making.

Although the development and application of such management tools are an extensive study on its own and go beyond the scope of the present investigation, an application of such a management tool will nevertheless be demonstrated for the Dewetsdorp aquifer.

5.1.3.2 DEWETSDORP AQUIFER

The two-dimensional ground-water flow model (AQUAMOD) requires transmissivities, storativities, monthly rainfall, recharge percentage, as well as the boundary conditions as input. For the Dewetsdorp aquifer, the model has shown that about 470 m^3/d of ground water, leave the aquifer at the northern boundary (cf. Section 4.1.6.1). The model also calculates flow velocities which can be plotted with suitable software (e.g. Plot 88). Figure 5.1 shows the flow direction of the ground water in the aquifer for the amount of water at present withdrawn from the aquifer, i.e. $\pm 65\ 000\ m^3/a$. The length of the arrows in this figure represents the magnitude of the ground-water velocities (and thus the water-level gradients) at that point. The question arises whether it is possible to intercept the ground water which leaves the aquifer in one or more additional production boreholes.

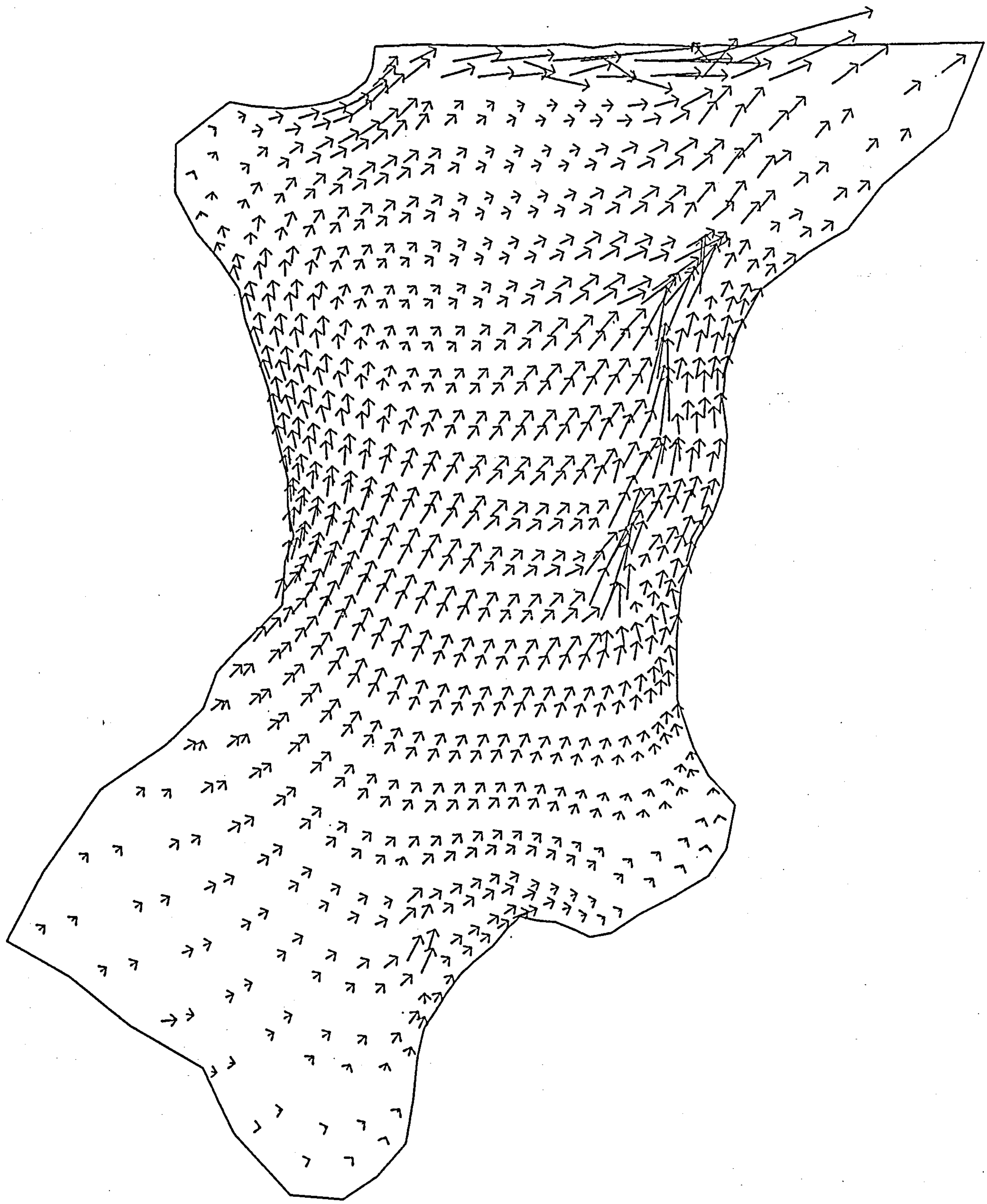


Figure 5.1 Steady-state ground-water flow direction at present withdrawal rate.

To be on the safe side, it was decided to incorporate extra boreholes in the aquifer with a total abstraction rate of only 300 m^3/d. In a first trial run, only one production borehole was added, just north of the existing production field. The model was run for a period of two years. The flow diagram (Figure 5.2), after only six months, shows that this was not a sound practice, because the abstraction from this borehole caused a considerable water-level drop in the boreholes in the production field and the resulting cone of depression became too steep.

In a second trial, three boreholes (with an abstraction rate of 100 m^3/d each) were spaced in a line parallel to the northern boundary. The results of the steady state solution of the flow model (see Figure 5.3) show that this option is indeed feasible and that the existing production field would not be influenced by the incorporation of the extra three production boreholes.

In practice, more than three boreholes would be necessary to withdraw 300 m^3/d, because the maximum capacity of boreholes in the Dewetsdorp area is only in the order of 40 to 50 m^3/d. However, six to eight or more successful holes in the area of the three proposed holes would have a similar effect on the water-level gradients.

At present, approximately 28% of the net mean annual recharge is abstracted. Additional production holes developed in the area indicated in Figure 5.3, would increase the utilization of the aquifer to about 70%.

The results of the model runs show

i) that additional water is available for abstraction during times of normal rainfall conditions and

ii) how the utilization rate of the Dewetsdorp aquifer can be improved.

It must be kept in mind, however, that it has not been part of this study to investigate whether suitable drilling sites can be found in the proposed area, and also that proposed additional abstraction (with the consequent improvement of the utilization rate) assumes a mean annual recharge as calculated with the SVF-method. During periods of drought, considerably less water may be available for abstraction. How much less is available, depends on the acceptable recurrence period. Further runs of the model with rainfall and recharge percentage values adjusted to the respective recurrence period, will show how much additional water can be abstracted in which area.

From his experience and the data an evaluation is based on, the geohydrologist carrying out the evaluation might be able to indicate which percentage of the available water could economically be abstracted. He should also be able to advise

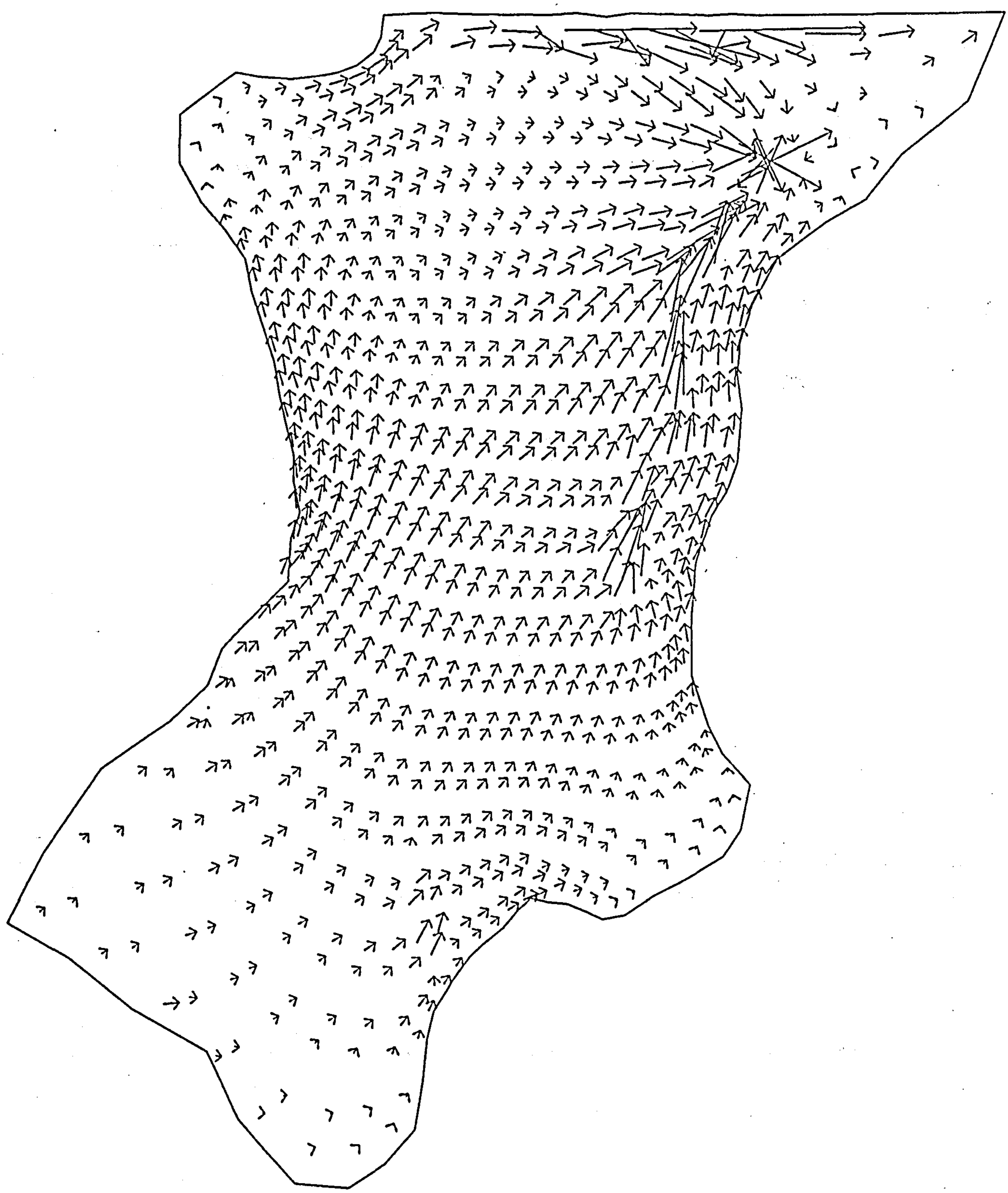


Figure 5.2 Ground-water flow direction after six months with one additional abstraction borehole pumped at a rate of 300 m^3/month.

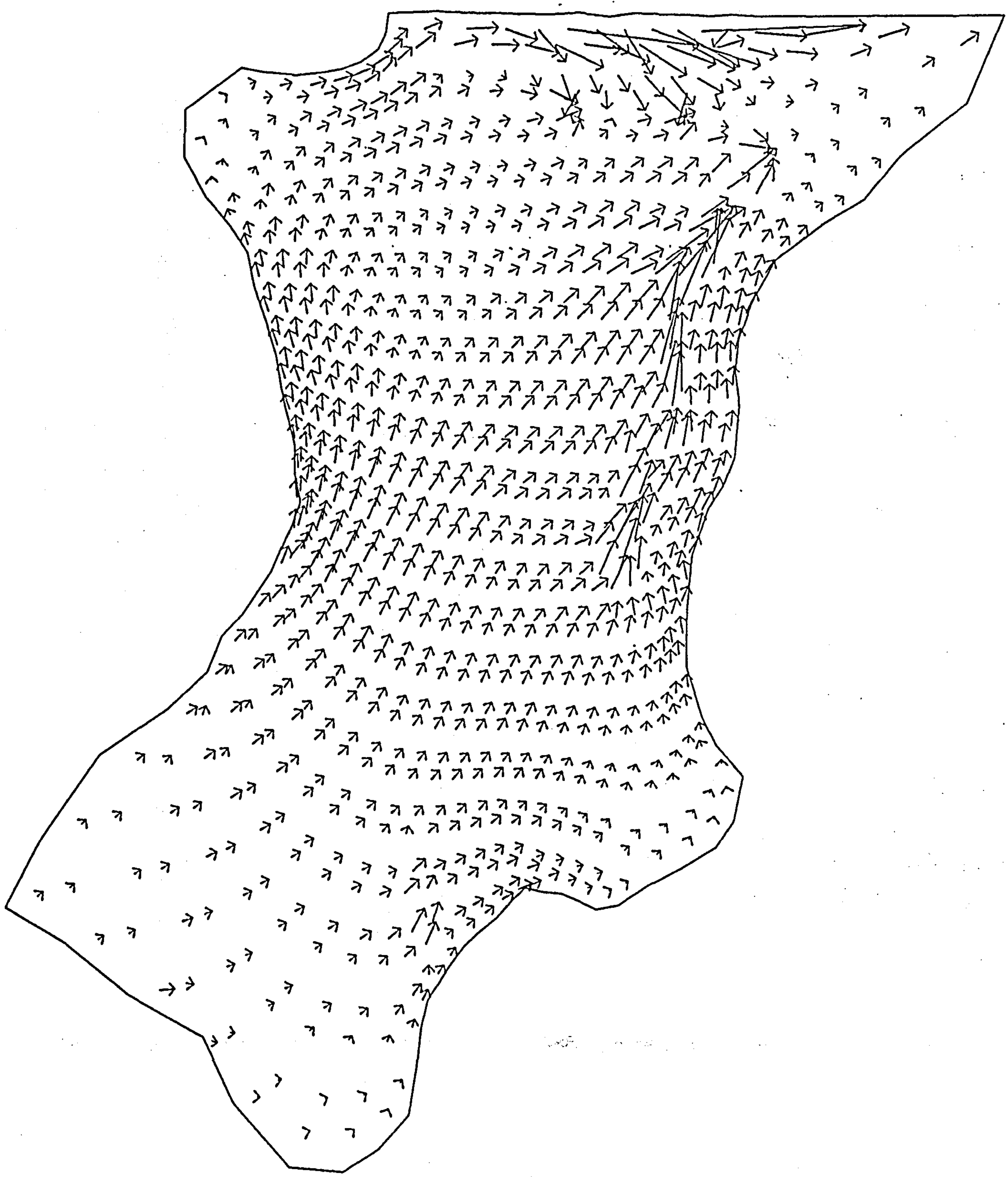


Figure 5.3 Steady-state ground-water flow direction with three additional abstraction boreholes pumped at a rate of of 100 m^3/month each.

the planning engineer on items like expected yield of holes, quality of the water, minimum distance between holes, required depth, diameter and construction of holes, expected pumping- and rest water levels; possibly, with the help of the ground-water data base, on items such as success rate, drilling cost and other ground-water related aspects. The calculation of the cost for the civil and mechanical works and, finally, the judgment regarding the feasibility of the whole scheme has to be left to the engineer.

6 DISCUSSION OF RESULTS

6.1 INTRODUCTION

Practical resource development and management decisions are subject to normal project constraints and data limitations under complex hydrological conditions. All these constraints and limitations pose a problem to the geohydrologist, from whom it is usually required to make an estimation of the exploitation potential of an aquifer. He must find a realistic way of dealing with this problem and usually a pragmatic approach, involving careful monitoring of aquifer response to abstraction and subsequent evaluation of these data is followed. Quantification of the rate of natural ground-water recharge is thus a prerequisite for sound management of a ground-water resource.

Given the increasing demand for water in South Africa and especially of the townships in the Karoo Basin, due to an increasing population, a knowledge of the rate at which these aquifers replenish, cannot be over-emphasized. Uncontrolled over-development of an aquifer may have serious consequences, such as yield reduction of boreholes with subsequent economic implications. Policy design must include a certain degree of flexibility to accommodate initial uncertainty about recharge estimations. To aid with such a design, the geohydrologist must be aware of the errors in key parameters which can affect the rate of aquifer replenishment.

In Section 6.2, a brief summary of the most important factors affecting recharge, together with the limitations of the recharge estimations techniques is given. Section 6.3 contains findings which have and have not been made during the present study.

6.2 LIMITATIONS AND POSSIBLE ERRORS OF RECHARGE ESTIMATION METHODS

Recharge estimates, by whichever method, are always subject to error. It is important in development and management decisions to have an idea of these errors.

At this point, it may be worthwhile to recapture briefly some of the factors affecting the replenishment of an aquifer:

A. RAINFALL

- magnitude .
- intensity
- duration
- spatial distribution

Although general practice, the achievement of a relationship between recharge and rainfall is in question, because the lag between rainfall and actual recharge may differ for different parts of an aquifer and is not known for each part.

B. GEOLOGICAL ENVIRONMENT

- boundaries
- hydraulic conductivity and storativity of formation
- total area which is covered by no or relative shallow soil cover

C. EVAPOTRANSPIRATION

- the vegetation system
- meteorological parameters (e.g. radiation, albedo, air temperature and windspeed)
- soil physical parameters (e.g. soil moisture capacity)
- soil moisture deficits in relation to soil moisture capacity

D. HYDROLOGY

- run-off (which depends mainly on slope and vegetation)
- rivers

E. UNSATURATED ZONE

- thickness of the unsaturated zone
- hydraulic properties of the different soil layers
    - the characteristic of the most non-porous layer is the limiting factor
- soil physical parameters:
    - retentivity values
    - porosity
    - saturated hydraulic conductivity
    - salt concentration
- swelling soils
- crust formation (forming of a thin non-porous layer)

- air entrapment

F. FLOW MECHANISM

- soil matrix flow or
- flow along preferred pathways

The present study has indicated that the main flow mechanism of ground-water recharge in Karoo formations, is along preferred pathways. According to J. Wagenet of Cornell University (U.S.A., personal communication), these pathways are also called macropores.

During rainfall showers of high intensity, the ground water moves in macropores in the unsaturated zone. During rainfall events of a low intensity, the water in the macropores is forced into the adjacent soil matrix, because of the higher negative matric potential in the soil matrix.

The ideal situation is that the joint impact of all the above parameters, influencing the recharge process, should be analysed.

In addition to the evaluation of the respective influences of these factors, we feel it is necessary to give a short summary of the recharge estimation techniques and the possible limitations of each method (see Table 6.1).

6.3 FINDINGS OF THE PRESENT RESEARCH PROJECT

6.3.1 NON-ACHIEVEMENTS

During the project it has not been possible to determine either the change of the storage coefficient with depth nor the "thickness" of a fractured aquifer. Better estimates of the storage coefficient than from pumping- or packer tests can be obtained with the SVF-method.

Without knowledge on the "thickness" of the aquifer, even with an estimated storage coefficient, it cannot be determined in a general way to what degree an aquifer can be used in excess of recharge during droughts. Based on previous droughts in an area to be investigated, an estimate of the aquifer behaviour could possibly be made by the geohydrologist.

Method	Name of computer code used at IGS	Minimum data requirements	Limitations	Possible errors
A. Direct methods:				
1. Soil water balances	ACRU(CCWR) and MOUSE	Daily rainfall Soil info Vegetation info	Data requirements are large Needs experienced user	Uncertainty about E_T Only applicable for soil matrix flow
2. ZFP-method		Water content with depth	Only applicable if ZFP exists Site specific	Uncertainties in water content
3. Darcian approach		Water content Saturated K	Site specific	Only for soil matrix flow
4. Soil water flow model	SUFF (unsaturated/saturated flow), LEACHM	Retentivity values Water content Saturated K	Vertical cross-section Needs specialized user Not for fractured flow	Wrong calibration Uncertainty about boundaries
B. Indirect methods:				
1. GLF-method	GLF	Water levels Abstraction S-value at borehole Lateral inflow and outflow	Horizontal cross-section	Uncertainties in S, lateral inflow and outflow Distribution of boreholes Fluctuations due to pumping
2. SVF-method	SVF	Water levels Abstraction Lateral inflow and outflow	Horizontal cross-section	Uncertainties in S, lateral inflow and outflow Distribution of boreholes Fluctuations due to pumping
3. Numerical modelling	AQUAMOD	Water-level information Abstraction rates	Need specialized user Horizontal cross-section	Calibration of model Uncertainty about boundaries
C. Chemical methods				
1. Cl profile method		Cl concentration in rainfall and ground water	Not applicable for fracture flow Yield long- term estimate	Uncertainties in wet and dry deposition

Table 6.1 Summary of the different recharge estimation techniques and the limitations of each method.

6.3.2 DATA CAPTURE

The study was based on

(i) geological, general geohydrological and some geophysical data which have been gathered during field surveys, during the drilling campaign and, to some degree, from previous reports

(ii) data from the unsaturated zone, i.e. soil analyses and neutron measurements

(iii) meteorological data which have been gathered at the erected weather stations and a number rainfall recorders and rainfall measuring stations, as well as data those withdrawn from the records of the Weather Bureaux

(iv) water-level data collected automatically at recorder stations and levels in other boreholes measured at intervals, which varied between a week and three months, depending mainly on weather circumstances, i.e. whether recharge was to be expected or not

(v) aquifer data calculated from pumping and packer tests, carried out during the course of the project, as well as those found in previous reports

(vi) abstraction data obtained from the respective municipalities and historical data from reports

(vii) stable isotope data from rainfall and ground-water samples collected at the beginning of the project and rainfall samples of the heavy falls in February, 1988

(viii) chemical analyses of ground-water samples collected during construction of the holes, as well as historical data found in previous reports

(ix) run-off data should have been collected, especially in the De Aar area. This has not been done, however, because of lacking measuring devices

Water-level recorders worked well, except for losses caused by occasional malfunction or when visits were late. This was no serious problem, however, since water levels, especially when receding, change slowly and at a relatively constant rate. Very good coverage of the water-level recovery after the heavy rains in February/March, 1988, was also obtained for the hand measured holes, especially in the Dewetsdorp area. The De Aar area was too large and the accessibility after the rain much worse to get equally frequent measurements.

Meteorological data and, in particular rainfall data, were less satisfactory. Rainfall occurs in discrete events and even short interruptions of the records may lead to irrecoverable data losses. It has been shown, as is generally known, that rainfall, even over short distances, is rather variable. When it is intended to correlate rainfall data with water-level rises on a local scale, missing data cannot be patched. The use of totaling meters, which allow, to some degree, controlled correction of data sequences, is recommended if staff is not permanently stationed in the field.

Isotope samples of both rainfall and ground water must be taken at much more frequent intervals (especially after a recharge event). Care must be taken that a representative number of holes in shallow and deeper lying aquifers, and also holes with varying time lags, are sampled if semi-quantitative results, regarding the local recharge, are expected from isotope studies.

If the chemical changes of the ground water, as a result of recharge events, were to be investigated in more detail, samples for chemical analyses should be taken according to the same rules as set out for isotope studies.

Abstraction data for the municipal boreholes were collected by the respective municipalities. Apart from the fact that the De Aar pumping figures were only available for borehole groups and not the individual boreholes, the data are quite satisfactory. More problematic are the figures for the private boreholes; water meters are never installed and the amounts abstracted have to be guessed. The relative error of private abstraction may be higher for Dewetsdorp because municipal abstraction is less and a rather large number of boreholes exist which are used for gardening. Gardening, of course, requires comparatively more water than sheep farming.

In the present study, the same experience was made as in the SRK project; more data were generated than were later effectively used. While this must be preferred to a situation where necessary data have not been collected, it still seems desirable not to unnecessary extend the measuring network, but rather improve on the quality of the data, i.e. have less gaps and sample or measure at higher frequency.

6.3.3 METHODOLOGY

Three different models to determine recharge from soil water movement in the unsaturated zone yielded zero recharge. This is in agreement with most of the neutron gauge measurements, but contradicts the observed water-level rises after rainfall events. This discrepancy is explained by the high clay content of the soils in the investigated area, which prohibits any noticeable matrix flow for which the models were designed. Recharge can therefore only take place along preferred pathways for which no models have yet been developed.

Two other approaches to determine ground-water recharge from water balance calculations seem possible: consideration of the whole system in which rainfall minus run-off and evapotranspiration must equal recharge; or of the ground-water component only where the nett recharge is equal to the increase in water stored minus abstraction minus subsurface inflow plus outflow.

The first alternative requires knowledge of the actual evapotranspiration which cannot be quantified with sufficient accuracy. The second approach avoids this problem, because only that part of the recharge is calculated which remains after actual evapotranspiration. It is only this nett recharge that constitutes the exploitation potential, unless management of the evapotranspiration is considered part of this potential.

Computer programs have been developed to construct a mesh for modelling of the aquifer and to calculate the recharge according to the SVF- and GLF-methods. A sufficient number of monthly water-level observation points and of the actual monthly abstraction are required as input. A sensitivity analysis carried out for the Dewetsdorp area, has shown that a reduction in observation points from 72 to 15 and to 6 boreholes resulted in a decrease of the calculated yield to approximately 67 per cent. This finding can, however, not be generalized, because the number of observation points required, depends on the geohydrological complexity of the area, e.g. an area with different water-level gradients in different parts requires more observation points.

Calculation of the nett recharge from water-level observations depends on the proper determination of aquifer constants. It has been shown that unrealistic modelling results are obtained if transmissivity and storativity values are used which were determined from pumping tests in fractured aquifers. However, if the transmissivity is assumed to be zero, a minimum storativity can be calculated and for periods of zero recharge, it is possible to determine the transmissivity/storativity relation. With these two values as guide-lines, the aquifer response can be modeled successfully. From the results obtained, it seems that transmissivity values determined by packer and pumping tests are in the right order, while the calculated storativity values are approximately by a factor 10 smaller than the correct values.

These results underline the importance not to neglect the conditions with which aquifers must comply when pumping test evaluations are done.

Slightly differing equations of the form

$$\text{Recharge} = A\ (\text{precipitation} - B)$$

have been obtained for the recharge estimate, depending on which method and which time period have been considered. Especially, the constant B can vary by more than a factor 10. Although the recharge estimates for calculated periods are in good agreement with each other, this situation remains unsatisfactory if, depending on the area and period investigated, large variations in the precipitation must be accommodated. The results of the SVF method based on periods between equal saturated volumes are supposed to yield the best results, because the storativity value is not considered in these calculations.

Recharge results obtained in this study are valid only for areas where no ground water enters the area without the inflow being accounted for in the water balance. If this is the case, e.g. ground water entering the area along major faults, the calculated recharge will be a minimum value. Similarly, if ground water is unaccountedly lost to deeper lying aquifers, the recharge calculation will lead to an overestimate. The same restriction applies to the recharge calculation, if surface water from outside the area under consideration recharges an aquifer or ground water is lost to a river leaving the area.

In the following chapter, a methodology is presented for the first recharge estimate of an aquifer, as well as an improved estimate which is based on a sufficient number of water-level and abstraction data.

7 DETERMINATION OF THE EXPLOITATION POTENTIAL

7.1 INTRODUCTION

To get such a reliable estimate of the exploitation potential of an aquifer, may require a number of years during which thorough and continuous observations are made, regarding the ground-water level fluctuations and abstraction in an aquifer.

If the case studies of the Dewetsdorp and De Aar aquifers are assumed to be representative of the general behaviour of a Karoo aquifer, it can be said that the mean recharge of an aquifer in the Karoo Basin is between 2% and 4% of the annual rainfall. These recharge values can, however, change drastically if certain favourable features occur in the area. One such feature is the presence of a river with its alluvial bed. For example, if it was not for the Brak River in the De Aar study area, the potential of this aquifer would have been at least four times lower. Another factor which has a great influence on the replenishment of an aquifer, is the occurrence of outcrops and areas with a relative shallow (<200 mm) soil cover. Recharge in these areas can be as much as three times higher than in areas with a thicker soil cover.

Experience gained on the Ecca Formation, to the west of Bloemfontein, showed that the occurrence of a calcrete layer improves the prospect of recharge. An important factor in determining the exploitation potential of an aquifer, is the presence of ground-water barriers, such as ground-water divides and no-flow boundaries. The degree to which an aquifer can be economical exploited depends also on the prevailing aquifer parameters T and S. These parameters have an influence on the quantity of water stored and the optimum spacing of production boreholes in the aquifer. Aquifer boundaries define the recharge area of the aquifer, and are thus an important factor to be considered.

The present study has shown that the recharge estimates in Karoo aquifers may be considerably lower than the usual quoted figure of between 5% and 8%. The estimates are, however, for nett recharge, i.e. the recharge remaining after evapotranspiration. On the other hand S-values of the Karoo Formations seem to be higher than usually anticipated and probably vary between 0,001 and 0,005, contrary to previous calculated S-values of even as low as 10^{-7}. The higher S-value implies that the Karoo Formations contain more ground water than generally assumed.

7.2 " THE FORMULA" = A METHODOLOGY

Although we could not deduce a simple equation for the determination of the exploitation potential of a Karoo aquifer, we are, however, able to advocate a certain methodology which can be used to obtain such an estimate. In this methodology, which is essentially an integrated approach, the use of three simple equations is recommended as a first step in obtaining such an estimate. The aim of this section is to supply the user with a methodology for the estimation of the exploitation potential of Karoo aquifers:

1. For a first approach, the following three simple equations can be used to get an idea of the exploitation potential of an aquifer:

Thin soil cover: $GR = 0{,}06\ (P - 120)$

Thick soil cover: $GR = 0{,}023\ (P - 51)$

Alluvium cover: $GR = 0{,}12\ (P - 20)$

where GR = annual ground-water recharge (mm) and P = mean annual precipitation (mm).

These equations are based on the experience gained in the Dewetsdorp- and De Aar aquifers.

The equations are coded into program FORMULA (see Appendix). FORMULA requires the annual rainfall data file of a nearby weather station (available from the CCWR) and the percentage area which comprises a thin soil cover (< say 200 mm), thick soil cover and alluvium cover in the area respectively. The annual record of rainfall is ranked by the program from the largest to the smallest value, whereafter the mean recharge for the aquifer is calculated for a required return period, as specified by the user. It is foreseen that the program FORMULA will be updated in the future as more information becomes available on Karoo aquifers. It must be stressed that the use of FORMULA is recommended only as a first step in obtaining an estimate of the exploitation potential of an aquifer. A more reliable estimate may only be obtained, after the following steps were followed.........

Select existing geohydrological information which may be available from the National Ground-water Data Base or other records. The information must preferably include the following:

- data about the geological environment (boundaries, outcrops, etc.)
- positions where lateral inflow and outflow can occur
- ground-water level information which can be used for contour maps
- rainfall values

- abstraction from the aquifer
- position of existing boreholes

A crucial step is to determine if the existing boreholes are well-distributed over the area and are suitable to produce reliable ground-water level contour maps. If not, additional exploration boreholes must be drilled at suitable sites to fill the gaps. Because the ground-water level tend to follow the topography, boreholes must be sited where a ground-water gradient change is expected to occur. The installation of at least five ground-water level recorders is recommended. For the proposed methodology, it is important to be able to quantify the lateral in- and outflow from the aquifer. For this purpose, it is necessary for that the T-values are known at these points.

2. The use of the direct methods (methods which use data of the unsaturated zone) in the estimation of ground-water recharge was found to yield unreliable results, because the main flow mechanism in the Karoo Basin does not support the underlying physical concepts of matrix flow in the unsaturated zone. The ground-water balance approach and, especially the saturated volume fluctuation method (SVF-method), are to be used in the calculation of the replenishment of an aquifer.

3. The boundaries (preferably no-flow boundaries) of the aquifer are assumed to be a known parameter. If not, they must be deduced from geological-, ground-water contour maps or geophysical investigations.

4. For the sake of clearness, suppose the boundaries of the aquifer are as depicted in Figure 7.1.

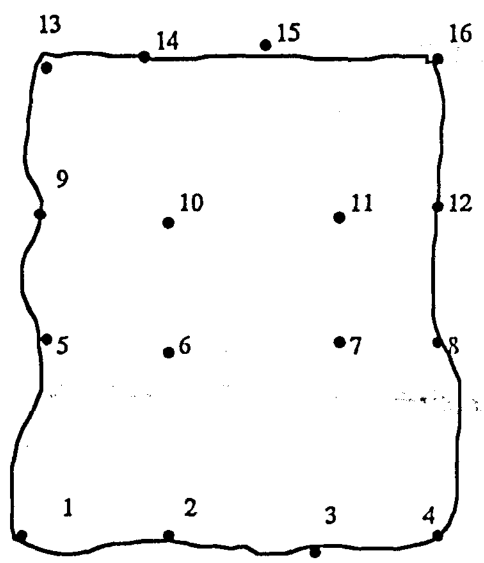


Figure 7.1. Construction of a triangular mesh: Selection and numbering of nodes.

The first step is to construct a triangular mesh over the aquifer. One may proceed as follows:

- Select points on the map which will be regarded as nodes of the triangles
- number each of these nodes, starting from one up to a maximum of 700, number in the shortest direction, e.g. from left to right, and repeat this sequence.
- Digitize these nodes exactly in ascending order.
- Construct the triangular mesh automatically from the digitized (x,y)-nodal coordinates by running program TFEM (Triangular Finite Element Method). (TFEM is available from IGS on floppy disk).

5. Program SVF is used for the solution of the SVF-method.

The data requirements of program SVF are as follows:

- the triangular mesh as generated by TFEM
- monthly ground-water levels at a number of boreholes which are well-distributed over the area for at least a period of 18 months
- monthly abstraction from the aquifer
- monthly rainfall values (these values are not used in the recharge estimation procedure, but can be used later to obtain a relationship between recharge and rainfall)
- in- and outflow boundaries with transmissivity values

The above-mentioned data must be supplied to SVF in the appropriate format, as described in the listing of the program in Appendix.

The output of SVF is as follows:

- for the first run, the saturated volume of the aquifer is calculated on a monthly basis
- the ground-water table gradient at the in- and outflow points
- an estimate of the most probable mean S-value of the aquifer
- the recharge (in mm/time period) for a selected time period (e.g. 1, 6 or 12 months).
- an option in SVF allows the user to enter a different S-value at each node; in this case, the volume of ground water is calculated and not the saturated aquifer volume
- an estimate of the lateral in- and outflow
- a file which contains information on recharge against rainfall

Program GLF (which requires much simpler input data) can be used in the place of SVF, if the user is satisfied that the arithmetic mean of the ground-water level is assumed to give a reliable estimate of the saturated volume of the aquifer (i.e. area · mean WL = saturated volume).

6. The ground-water recharge estimates by the SVF-method can greatly be improved by the application of a two-dimensional numerical ground-water flow model (AQUAMOD), which enables the user to analyse the sensitivity of the aquifer in terms of aquifer response to abstraction and errors in recharge estimates. Foster (1987) recommended that such a model must be used as an integral method in conjunction with any other recharge estimation technique. The application of such a numerical model is far-reaching. If sufficient and adequate ground-water level data for a number of boreholes in the aquifer are available, as well as the abstraction during the same time period, the water levels were measured and, together with a sound knowledge of the T- and S-values, such a model can be used to estimate historical ground-water recharge.

For the application of the numerical model, the person familiar with the numerical modelling technique, is required. The triangular mesh, as constructed by TFEM, is used by the program AQUAMOD (triangular finite element program for the simulation of two-dimensional ground-water flow) as input. The data requirements of AQUAMOD are as follows:

- the properties of the triangular mesh (e.g. nodal coordinates and incidences)
- T- and S-values at each node of the mesh
- initial ground-water levels
- abstraction rates of sinks in the aquifer
- information regarding the boundaries of the aquifer

While the numerical prediction model has long been used to aid development and management of ground-water resources in arid and semi-arid regions, the critical use of models in the proposed mode is less common, although it is a logical extension of conventional ground-water balance studies. The main purpose of this technique is to achieve calibration with historical ground-water level data. The limitation of the method is, however, the fact that the calibration achieved, may not be unique and it may prove to be impossible to distinguish between variations in recharge rates and S-values. The use of the SVF, in conjunction with the numerical model, may improve the reliability of the calibrated S-values of the numerical model. The calibration of a numerical model must be viewed as an on-going process, which is refined on a regular basis with the consequence that the accuracy of predictions will improve. The main goal at the end of any model is usually its application for management purposes. The application of such a calibrated model is far-reaching, e.g. the rate at which the ground-water levels will be declining during years of minimum recharge, can be predicted and the necessary operational steps to prevent a possible undesirable situation, could be decided on beforehand.

As already stated earlier, the results must always be viewed with some degree of suspicion, because all exploitation potential estimation methods are almost and inevitably subject to some unknown error. We are, however, quite confident that a reliable estimate of the exploitation potential of a Karoo aquifer can be obtained, if the proposed methodology is applied with care.

Chapter 8 contains a short case study on the recharge phenomena of the dolomitic Grootfontein Compartment in the Western Transvaal. The reason of incorporating this chapter, is to illustrate that the recharge techniques applied to a Karoo aquifer, can also be used for other types of aquifers.

8 RECHARGE ESTIMATIONS OF THE GROOTFONTEIN COMPARTMENT IN THE WESTERN TRANSVAAL

Although studying the replenishment of the dolomitic Grootfontein Compartment near Mafikeng, goes beyond the scope of the present investigation. It was, nevertheless, decided to apply the SVF-method to such a non-Karoo aquifer. Because recharge estimations of this aquifer were thoroughly investigated in the past (e.g. Bredenkamp 1978 and Janse van Rensburg, 1988) and assumed to be a well-known parameter, a short application of the SVF-method may show whether similar results can be obtained with this method.

The present application of the method to the Grootfontein Compartment was performed during a short visit of three days by Hugo Janse van Rensburg of the Directorate of Geohydrology of the Department of Water Affairs to the Institute for Ground-water Studies. At that stage, only information on the ground-water levels, abstraction rates and rainfall amounts, was available on floppy disk for the period January, 1985, to December, 1986. Janse van Rensburg (1988) estimated the recharge for this period to be 33 mm/a. His estimate was based on a two-dimensional finite element model and a ground-water balance.

The aquifer depicted in Figure 8.1 was discretized into 732 elements with 397 nodes (see Figure 8.2) and comprises an area of 169 km^2. The abstraction of ground water from this aquifer amounted to 16,9 x 10^6 m^3/a for 1985 and 1986. The graph of the volume of ground water in the compartment (Figure 8.3(a)) shows an obvious depletion in the total amount of ground water during the study period. Figure 8.3(a) was obtained with the option in program SVF, which allows for the entry of a different S-value at each node of the mesh. The S-values which were used, are those obtained by Janse van Rensburg when calibrating the numerical model. Figures 8.3(b) and (c) show the recharge estimates obtained by SVF. The recharge amounts to 35 mm/a (11,5%) for 1985 and 38 mm/a (8%) for 1986, which is in agreement with the value of 33 mm/a (11,2% and 7,4% respectively for the two years), as estimated by Janse van Rensburg.

Janse van Rensburg (1988) estimated the mean S-value to be 2,4% for the period 1982 to 1984. The option in SVF, which allows for the estimation of a mean S-value for the whole compartment, was applied and resulted in a value of between 1% and 1,9%, with a value of 1,5% to be the most probable mean value. This estimated lower S-

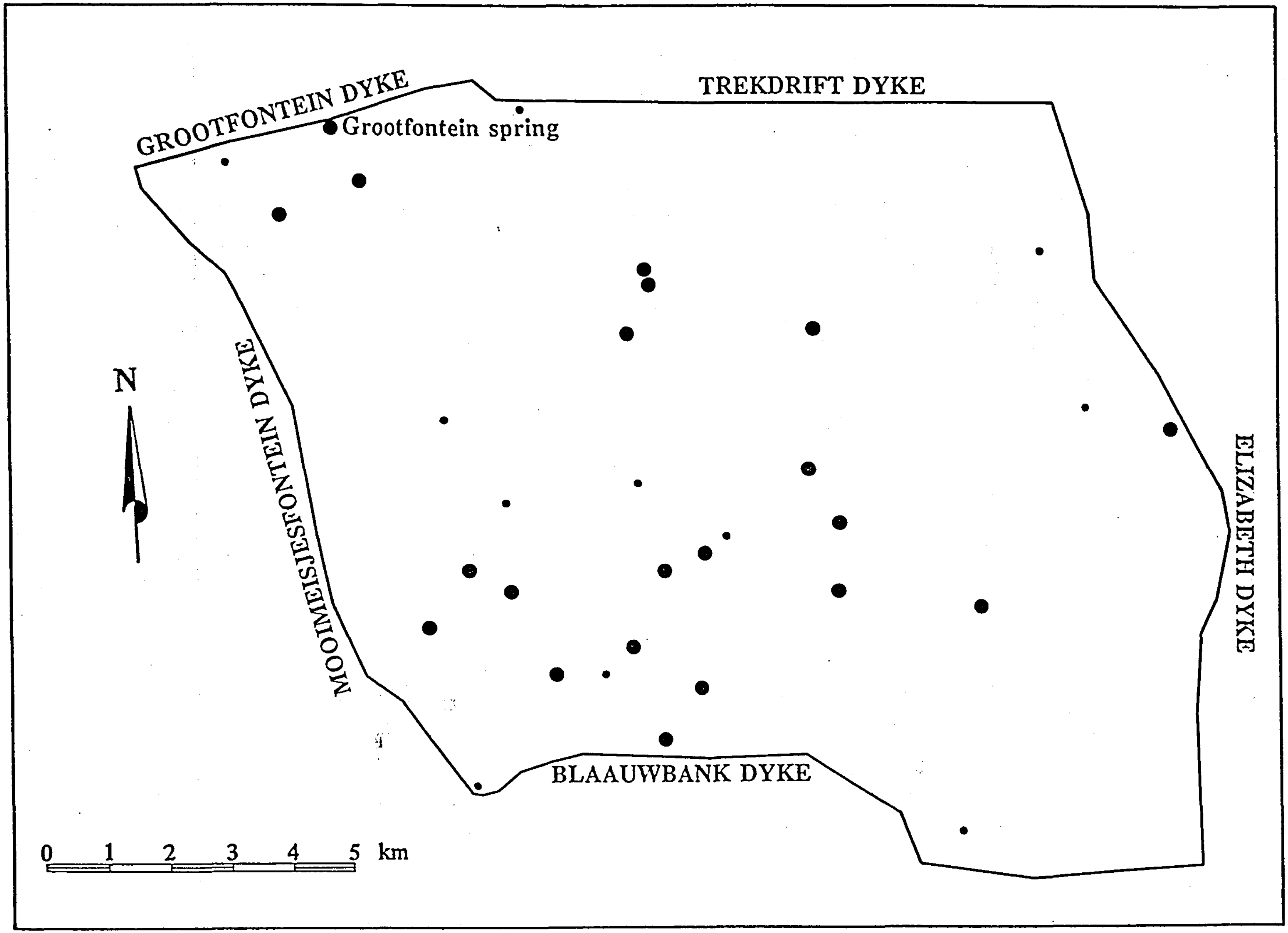


Figure 8.1 Boundaries of the Grootfontein Compartment in the Western Transvaal. The darker points represent abstraction boreholes in the aquifer and the smaller points represent the observation boreholes used in this study.

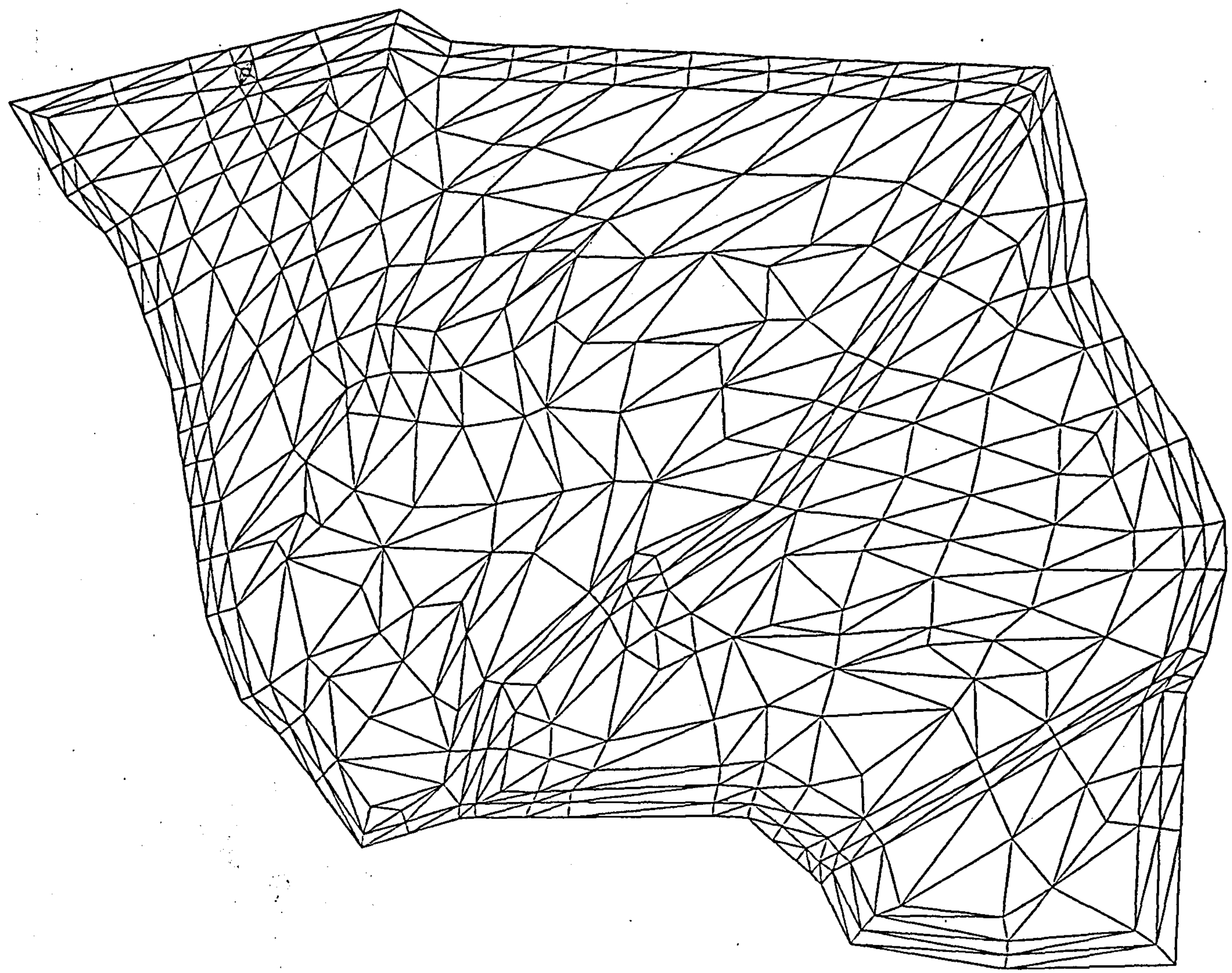


Figure 8.2 Triangular mesh for the Grootfontein aquifer used by the SVF-method for recharge calculations

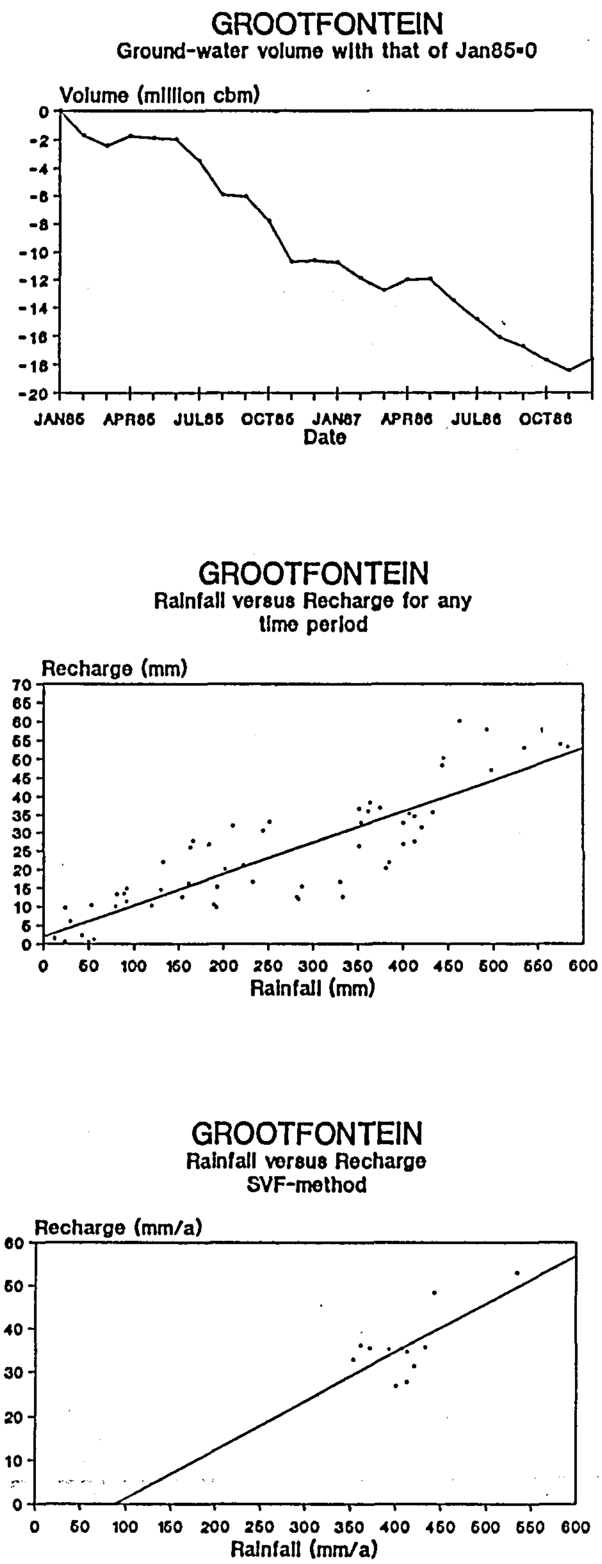


Figure 8.3 Graph showing:

(a) Ground-water volume in the Grootfontein aquifer from January 1985 up to December 1986,

(b) Rainfall-recharge relationship for any time period and

(c) Annual recharge as estimated with the SVF-method for the two-year period.

value for the period 1985/1986 is in agreement with the generally accepted idea that the S-value decreases with depth in the Grootfontein Compartment (Janse van Rensburg, personal communication). When SVF was again run for this period, with a mean S of 1,5%, a recharge value of 12,7% was obtained for the two years.

What can be said at the end of this short trial run regarding the recharge of the Grootfontein Compartment, is that the SVF-method indeed shows promising possibilities for future application to provide ground-water potential estimates of dolomitic aquifers, such as the Grootfontein aquifer.

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