

DEPARTMENT OF WATER AFFAIRS AND FORESTRY

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THE GEOHYDROLOGY OF THE KURUMAN EYE AND QUANTITATIVE
ESTIMATION OF RECHARGE AND STORATIVITY
OF THE AQUIFER SYSTEM

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THE GEOHYDROLOGY OF THE KURUMAN EYE AND QUANTITATIVE ESTIMATION OF RECHARGE AND STORATIVITY OF THE AQUIFER SYSTEM

1. Introduction

The Kuruman eye is a major spring issuing from the Ghaap Plato dolomite and has been the sole water supply of the town of Kuruman since the first settlement in 1885. The flow of the spring was maintained during the most severe droughts, though it fluctuates according to the prevailing rainfall conditions and is affected by the pumpage of the municipality.

It is important to ascertain the longterm average flow as well as the annual variability of its flow in relation to the recharge on the catchment that feeds the system. An in depth study of the Kuruman eye and other eyes in the drainage area, has been carried out by Smit (1978) who related the flow of the springs to the rainfall and estimated losses that occur from the system. He also successfully applied a lumped hydrological model to simulate the recharge from rainfall. Bredenkamp (1978) analyzed the C^{14} and tritium concentrations of the water emerging from the spring and was able to determine the minimum ratio of recharge to the total volume of the water stored in the aquifer, based on the mixing ratios of the C^{14} concentration and tritium concentrations in the atmosphere with that of the water in the system.

The present study of the spring flow was aimed at improving the simulation of the observed values of the C^{14} and tritium, and incorporates essentially four methods to estimate the system's behaviour and the quantification of the recharge.

These were:

- 1) An analysis of the flow in relation to the rainfall variations;
- 2) Determining the flow in relation to the Wondergat water levels, which show a linear correspondence for springs in the Bo Molopo region (Bredenkamp et al 1991);
- 3) By incorporating the isotopes C^{14} and tritium as natural tracers to reveal the mixing of the recent recharge with water contained in the aquifer. In this way the ratio of recharge to the aquifer storage was obtained;
- 4) By applying a finite element approach of the mass balance of the aquifer system, which allowed the estimation of the aquifer storativity using an integrated mass balance approach;
- 5) Determining the aquifer storativity also from the inferred depth of the water circulation using the geothermal characteristics of the aquifer.

2. GEOLOGY (refer Figure 2.1)

A discussion of the geology of the area is important as it is closely related to the geohydrology. The lithological sequence of the geological formations is given in the following schematic layout which is furthermore depicted in profile (refer Figure 2.2).

Table 2.1 Lithology of the area

LITHOLOGY	FORMATION	MEMBER	AGE
Wind-blown sand, calcrete, talus	Gordonia		Quaternary
Dolerite dykes (intrusive)			Quaternary
Jaspilite and crocokolite with shale	Asbestos Hills	Danielskuil	Vaalian
Banded ironstone with amphibolite and crocokolite		Kuruman	
Dolomite with lenses of limestone, chert and shale	Ghaaplato	Lime Acres	
Chert with interbedded dolomite	Grootfontein		

In general the formations dip at 4° to the west. The Grootfontein member outcrops only in the far eastern part of the study region.

The largest part of the area is underlain by dolomite of the Lime

FLOW OF MANYEDING B EYE

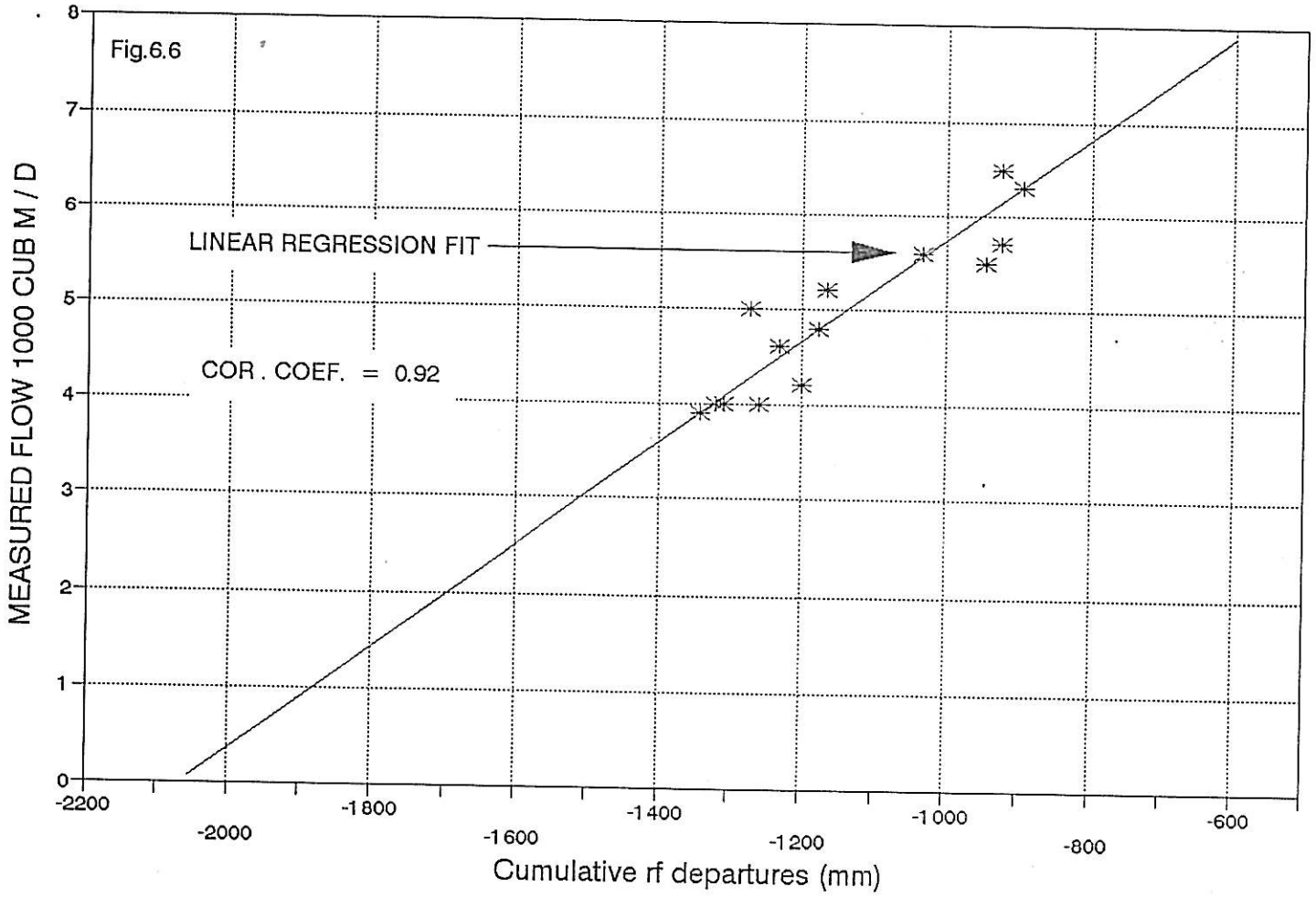
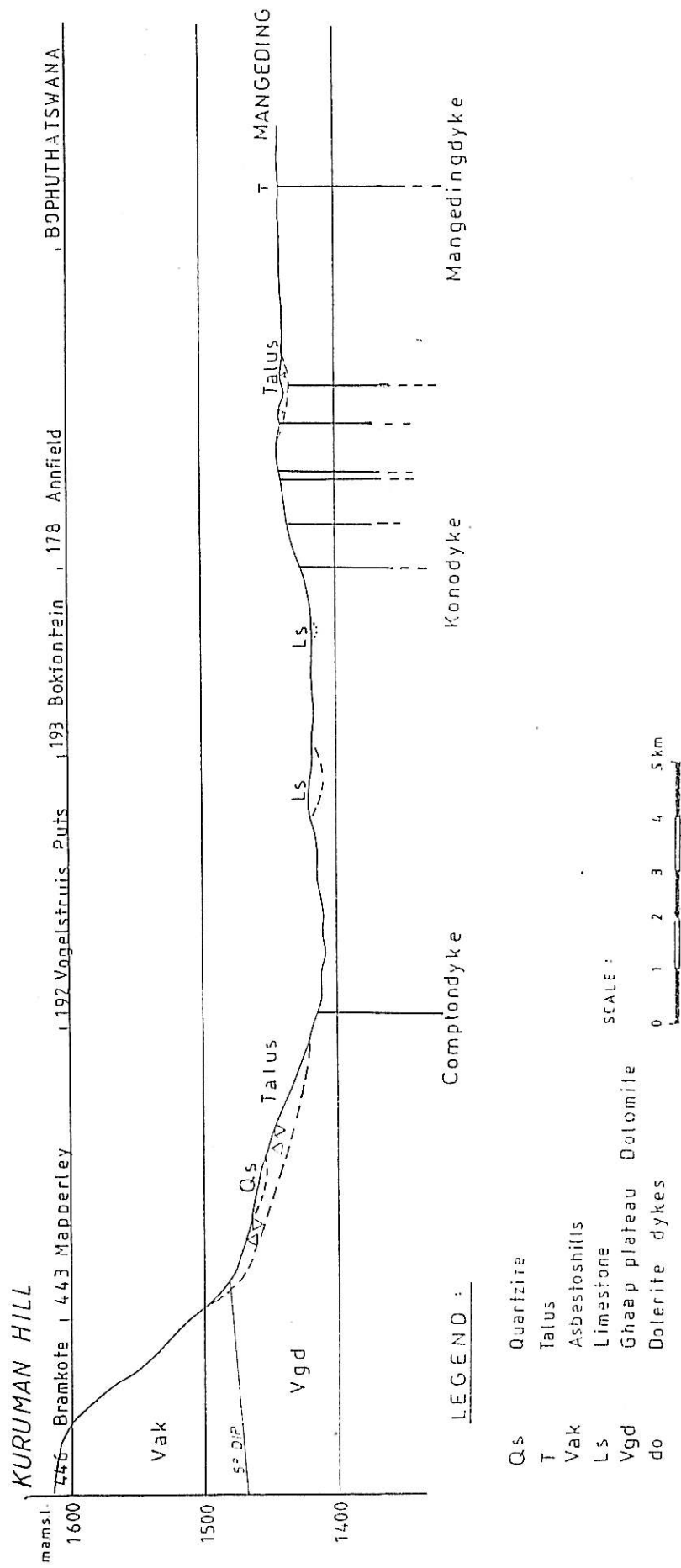


Figure 6.5 Flow of Manyeding B eye in relation to the cumulative rainfall departures

PROFILE OVER THE KURUMAN HILL EYE CATCHMENT

S.W. N.E.



- LEGEND :**
- Qs Quartzite
 - T Talus
 - Vak Asbestoshills
 - Ls Limestone
 - Vgd Ghaap plateau Dolomite
 - do Dolerite dykes
- SCALE :**
- 0 1 2 3 4 5 km

Figure 2.2 Section of the lithology

Acres member, which outcrops rather extensively. Several chert lenses outcrop in the region - extending from the Grootfontein member in the east, to right next to the Kuruman hills in the west. Although shale outcrops are rather scarce, black carbon-rich shale bands have been extensively intersected in boreholes. The upper end of the Lime Acres member is characterized by a limestone layer about 15 m thick followed by a prominent black layer of shale.

The Asbestos Hills formation follows concordantly on the Ghaap plateau and forms the north-south striking hills to the west of Kuruman. At the base of the Asbestos Hills, the banded ironstone of the Kuruman member is found. Ferruginous brecciated banded ironstone, generally referred to as the Blindklip breccia, occurs. Amphibolite bands and lenses of crocokolite, tuff and a flattish alluvial stone conglomerate appears quite generally.

Brownish jaspilite, crocokolite and shale of the Danielskuil member lies concordantly on the Kuruman member. It contains three prominent markers, namely:

- the bottom "speckled" marker
- the upper "speckled" marker
- the potsherd marker

Talus of quaternary age occurs widespread in a band extending north-south to the east of the Kuruman hills. This band is about 10 km wide and thicknesses up to 80 m have been intersected in

boreholes. Aeolian sand and surface limestone also occur fairly generally, but these deposits are nowhere as extensive as the talus depositions.

The dolerite dykes criss-crossing the area characteristically have a dominant north-east/south-west strike. Major exceptions however are:

- the Kuruman dyke, with an east-west strike, and
- the Manyeding dykes, which has a southwest-northeast strike.

Outcrops of the dykes occur very seldom, but they can be traced by the typical limestone ridges that are associated with the dykes. Some of the dykes e.g. the one causing the small Kono eye appear to be weathered to depths of up to 50 m.

Many of the dykes have intruded into fault zones as is signified by different geological formations on its sides, for example on the farm Happy Valley where the western side of the dyke constitutes black shale, whilst on the eastern side calcified dolomite is found to depths up to 200 m.

3. GEOHYDROLOGY

The groundwater occurring in the area is typically of the secondary aquifer type. The groundwater is found in fractures and dissolution channels in the dolomite and occasionally also

Known Kuruman waterlevels @ topography

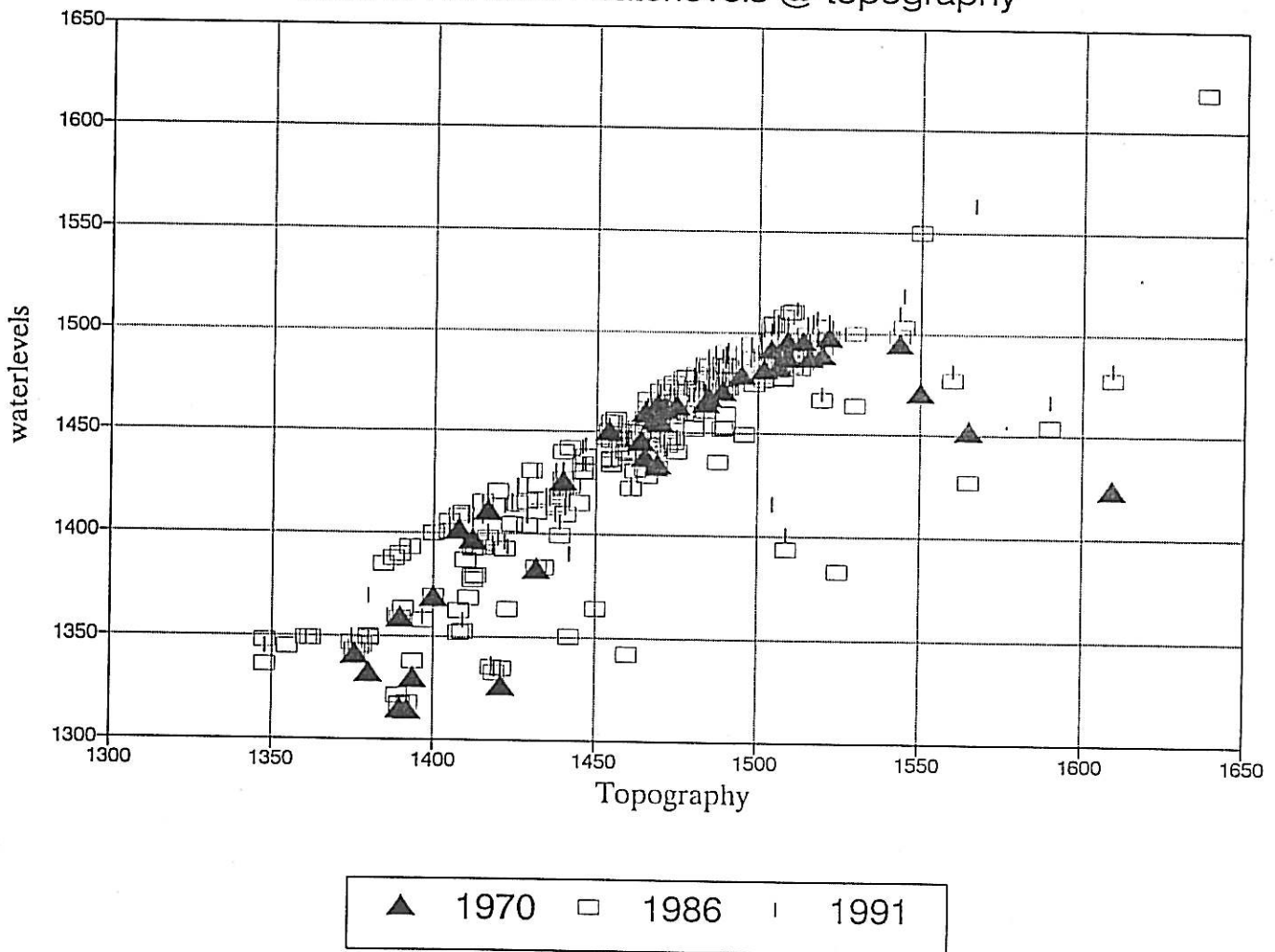
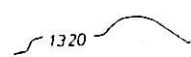



Figure 3.1 Water levels in relation to the topographical heights

d	m	y	Vogel C-14	WITS	Vogel Tritium	TU	Verhagen Tritium TOL	0-18	Deut	flow Kur A corr
		1967	66.3		0					7.138
		1968								6.401
26	11	1969	69.6		0.6			-6.6		6.516
		1970								5.871
22	3	1971	68		1.4					6.131
		1972								6.969
		1973								6.032
		1974								10.26
9	7	1975	75.3			1	0.3			11.62
11	8	1976	80.7	84.6		2.5	0.4	-6.6	-42.3	14.89
11	1	1977	77.8	79.4		1.9	0.3	-6.5	-41.2	15.8
		1978		79.5		5	1.6	-6.5	-42.4	16.32
		1979								16.03
		1980								15.04
		1981	69.9							15.7
		1982								15.35
		1983								13.7
		1984								13.78
		1985								12.32
		1986								11.88
		1987								11.37
		1988								14.26
		1989								16.59
		1990	68.8							16.3
		1991	69.8							11.3
		1992								

Table 3.1 Measured values of C¹⁴ and Tritium for Kuruman A eye



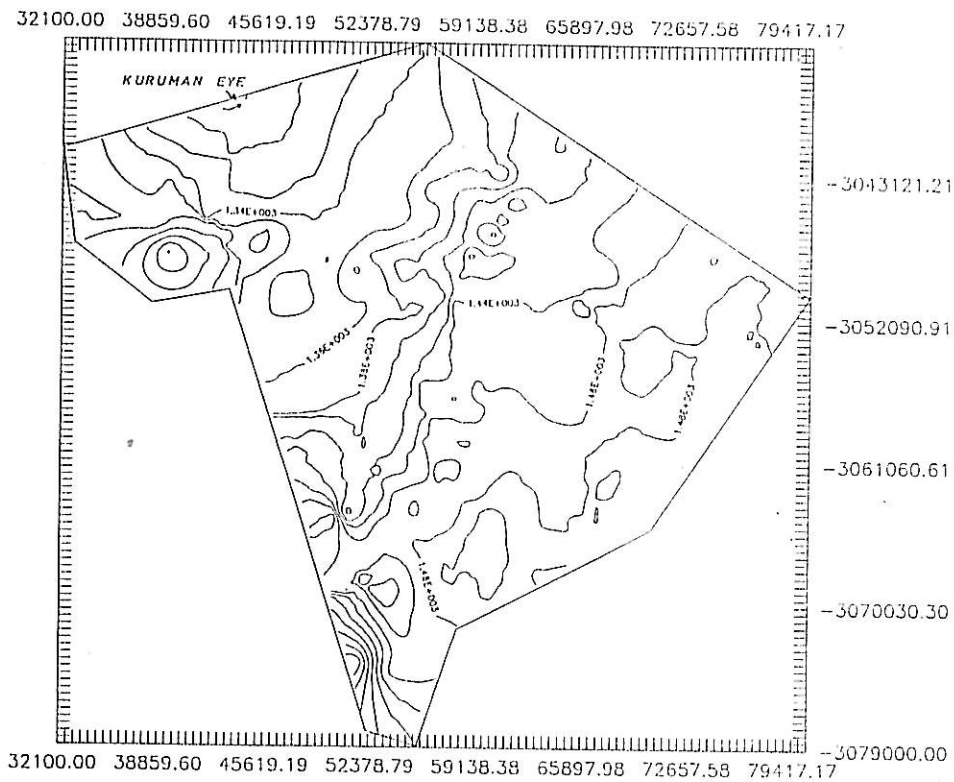
 Water level contour (a.m.s.l.)
 Boundary of KURUMAN RECHARGE AREA

Scale
 0 2,5 5 10 km



Figure 3.2 Hand contoured map of the piezometric surface

1986 Waterlevels of Kuruman catchment



1991 Waterlevels of Kuruman catchment

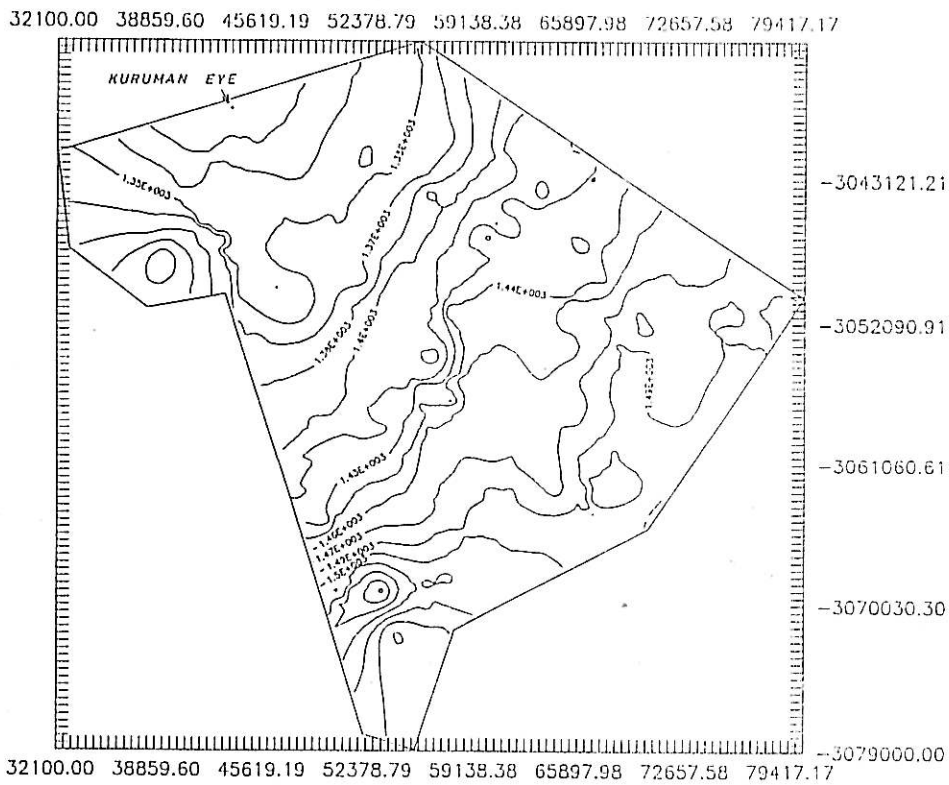


Figure 3.4a Reconstructed groundwater contour maps for 1986 and 1991

Topography of Kuruman catchment

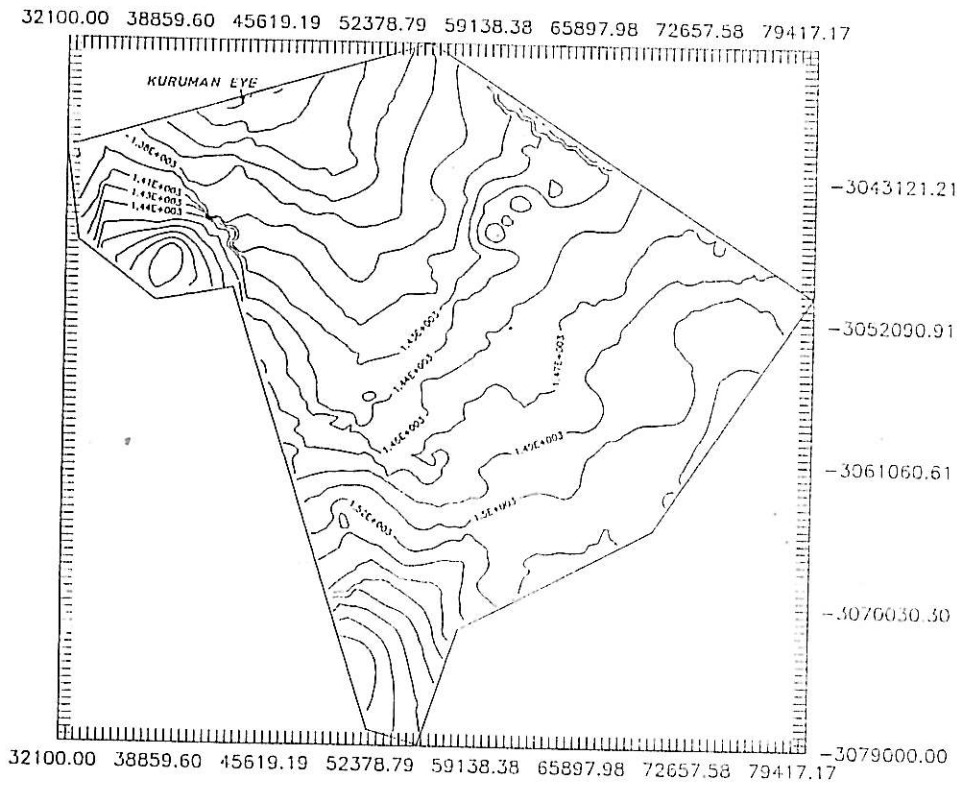


Figure 3.3 Topographical contours by computer

1970 Waterlevels of Kuruman catchment

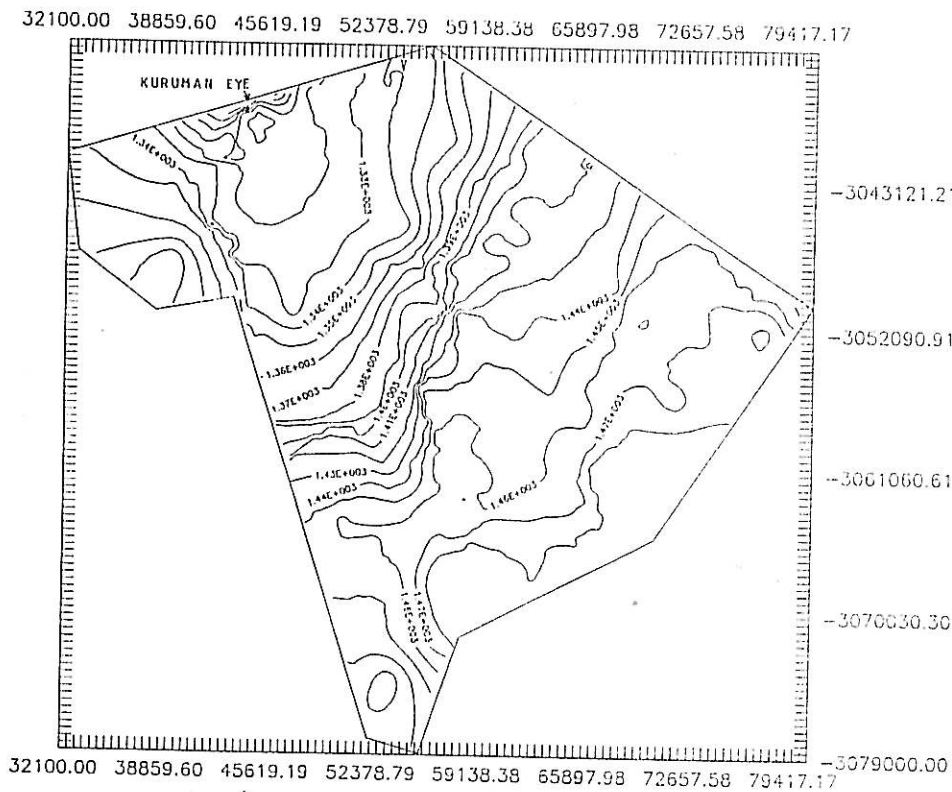


Figure 3.4b Topographical- and reconstructed groundwater contour map for 1970

in the talus, which is actually a primary aquifer.

The depth to the water level varies quite significantly based on a superficial scrutiny, however if the levels are examined more closely a striking linear relationship with the topographical heights is revealed. This is clearly demonstrated by Figure 3.1 which shows a remarkable uniform relationship for each of the water level sets. The almost 1:1 relationship between the data sets implies that the points which apparently do not conform to the straight line relationship with the topography, are indeed outliers. The reason for the outliers could be due to errors in the collar elevations of the boreholes, but these were found to be reasonably reliable, although the values had only been inferred from the topo maps. Some points that deviates could be explained as a zone of higher transmissivity as is depicted by the outlier points in Figure 3.1. According to the hand contoured map of the piezometry it represents a zone of preferential flow (Figure 3.2). The respective gradients towards the eye are related to the transmissivity of the dolomite aquifer, but overall indicates a remarkable uniform pattern, even considering that steps in the water level occur over most of the dykes. The computer drawn piezometric map (Figure 3.4) reveal contours that conform very closely to the topographical contour map (Figure 3.3). The irregularities could only be due to errors in the collar elevations of boreholes, which in the case of a dolomite aquifer must preferably be surveyed accurately instead of deriving the heights from a topographical map.

At some places perched water tables were found, the most prominent occurrence is on the farms Ruimte and Bokfontein about 20 km south-east of Kuruman. This perched water table is apparently caused by a thick shale layer, but after the good rains of 1988 the perched water table appeared to have merged with the main water levels.

The dykes traversing the dolomite cause several springs to issue from the drainage area, similar to other dolomite aquifers. The dykes act as boundaries forcing lateral groundwater flow to emerge as springs. The drainage areas of the individual springs have not been delineated precisely, but according to Smit (1978) the composite drainage area of the Kuruman eyes and the Manyeding springs comprises about 1140 sq.km. This area shown in Figure 2.1 was treated as a hydrological unit for which the recharge was derived from the spring flows relating to virgin conditions.

The south-eastern boundary of the system was delineated according to the groundwater divide, which slopes away from the spring. The boundary to the west was assumed to be the Kuruman hills, which comprise of banded ironstone that appear to be much less permeable than the dolomite.

The Manyeding No.2 dyke (Man 2) was taken as the north-eastern boundary, mainly on account of the water level step of 10 meters that occurs over this dyke. This step however does not necessarily signify that the dyke is fully impermeable.

In view of the uncertainty about the leakage of dykes, the separate compartments that feed the Kuruman and the Manyeding eyes cannot be delineated with the present available data and for this reason a lumped system approach was adopted.

The northern boundary is the Kuruman dyke, which is a major feature causing the groundwater from great depth to issue at the main Kuruman spring referred to as Kuruman A eye. Two more outflow points for the Kuruman compartment exist along the Kuruman dyke, namely Kuruman B and Kuruman C eyes. Of the latter two the B eye is the larger spring, although both flow only during periods when the water levels in the compartment has risen substantially after periods of good recharge. On the basis of the C^{14} concentration of the springs the main spring (Kuruman A) with a lower C^{14} concentration (69% modern) and a temperature of $22,4^{\circ}\text{C}$, represents water emanating from a depth of about 360 meter. The B eye has a C^{14} concentration of 83% modern and a temperature of $21,6^{\circ}\text{C}$, indicating that the depth of circulation is shallower (about 160 metres). The C eye emerges at a topographical point that is lower than the Kuruman B eye, and flows only when the water levels are high and seemingly is related to times when the Compton eye, some 20 km south-east, is flowing strongly. The Compton eye which is an ephemeral spring only flows after periods of exceptional good recharge, as had been the case around 1974-78 and again from 1985 to the present time. Some of its flow re-enters the lower part of the compartment and probably leaves via the Kuruman B and C eyes.

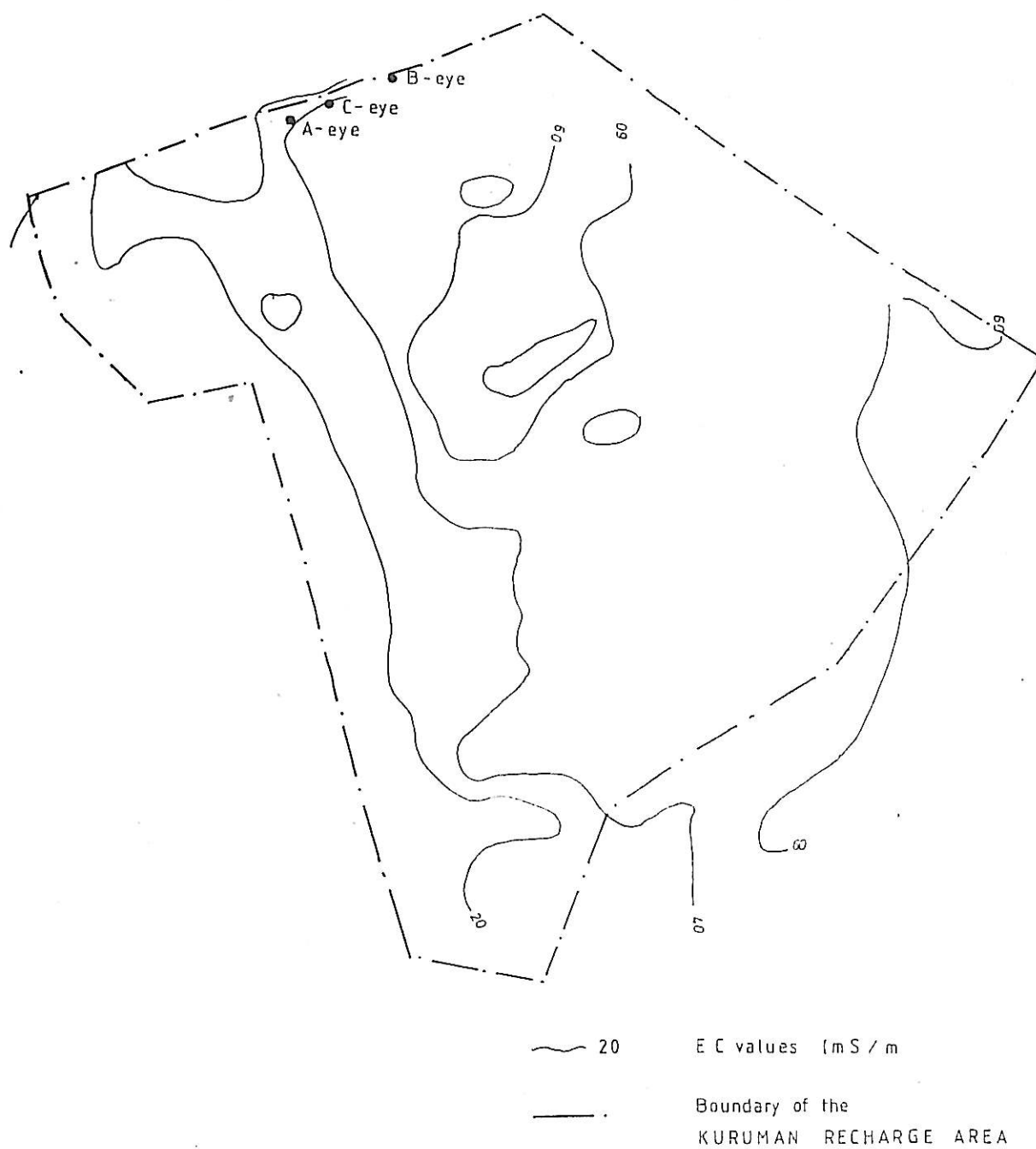


Figure 4.1 Iso-lines of electrical conductivity

The three Manyeding eyes, (A, B, and C) of which the middle eye (B eye) is the largest and the only one with perennial flow, all emanate along the contact of the prominent Manyeding dyke (Man 1 in Figure 2.1). These springs represent additional outflow points from the greater Kuruman eye Compartment, but their recharge areas can not be delineated. An attempt will however be made to infer the extent of their drainage areas, based on their average discharge in relation to the total recharge - assuming the same recharge characteristics.

4. WATER CHEMISTRY

The ground water quality is generally good and the best water occurs in the region of the Asbestos Hills (< 20 mS/m). From there the conductivity increases to the east as is indicated by the iso-lines in Figure 4.1. A plume of fresher water extends along the high transmissive zone from the Kuruman Hills to the outlet of the Kuruman A eye. This plume appears to coincide with a doline structure that is characterized by a lineation of banded ironstone.

Superficially it would appear as if most of the water from the Kuruman eye originates from the plume of fresh water as the water quality of the spring is consistently about 35 mS/m, and differs significantly from the waters to the east which is about 60 mS/m. A degree of nitrate contamination is found in the catchment area but very little in the spring water. To the west of the Kuruman A eye in the Bothidistad area, there is significant nitrate

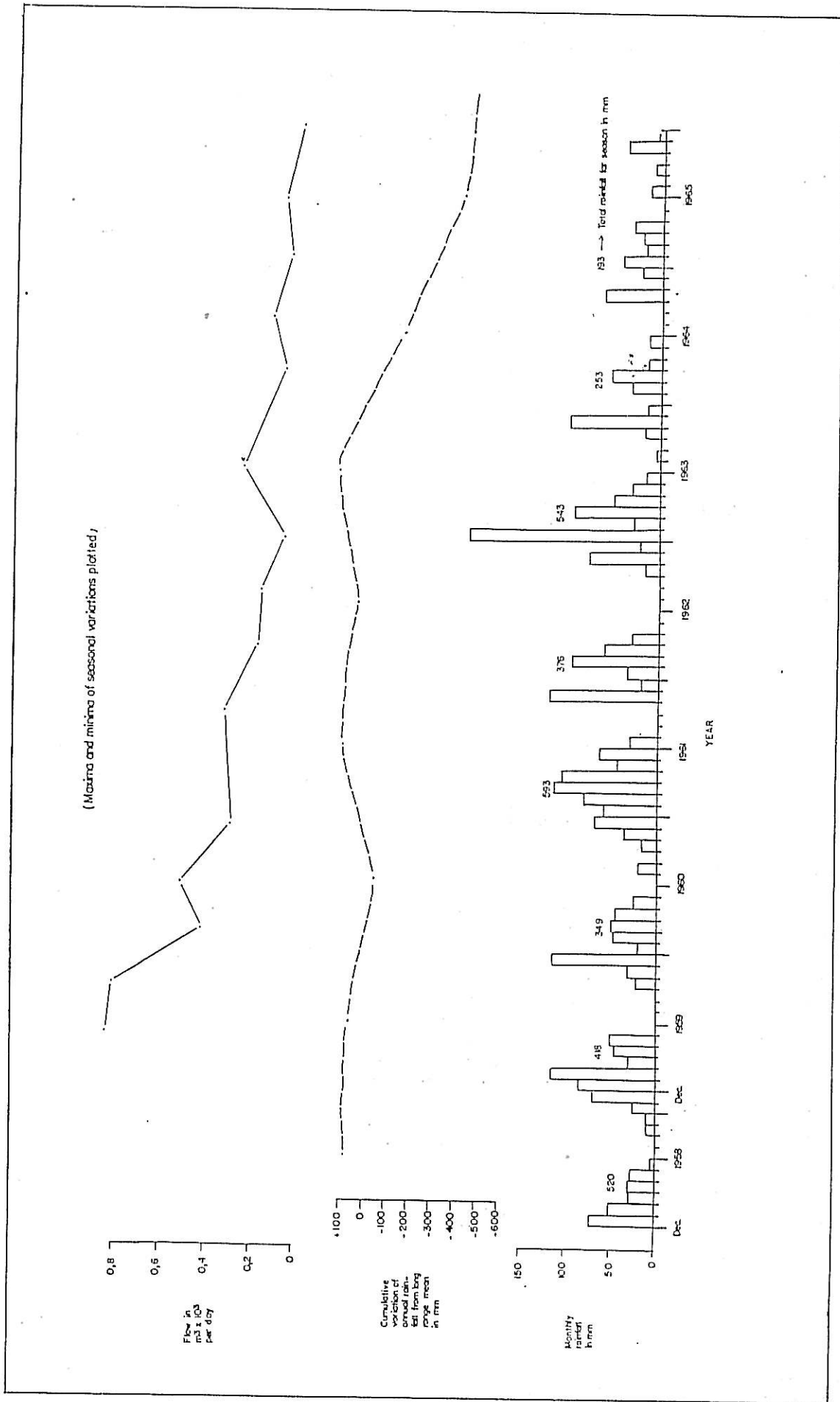


Figure 5.1 Temporal variation of the Kuruman B eye and the cumulative rainfall departures (according to Smit, 1978).

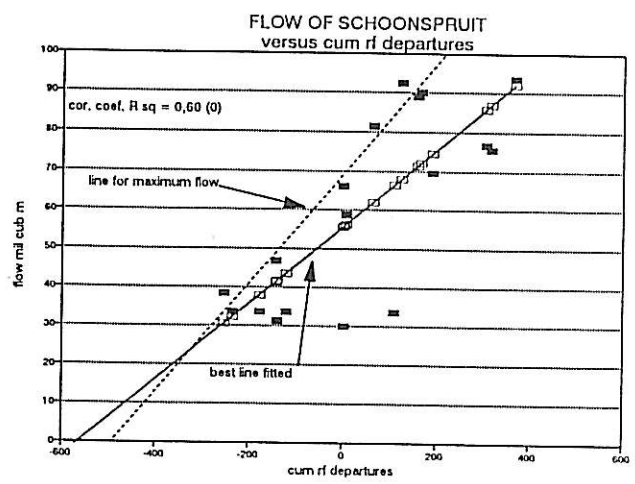
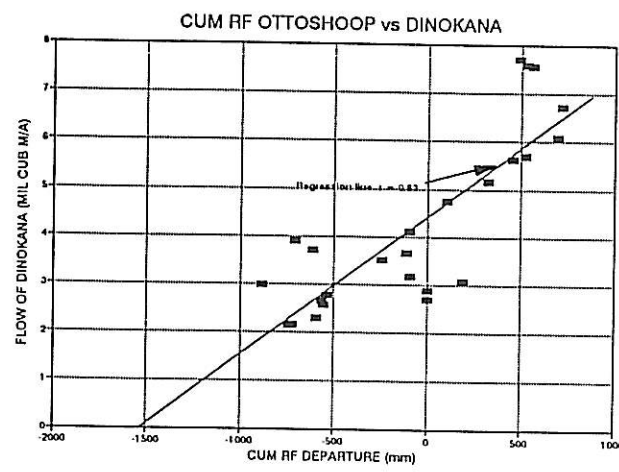
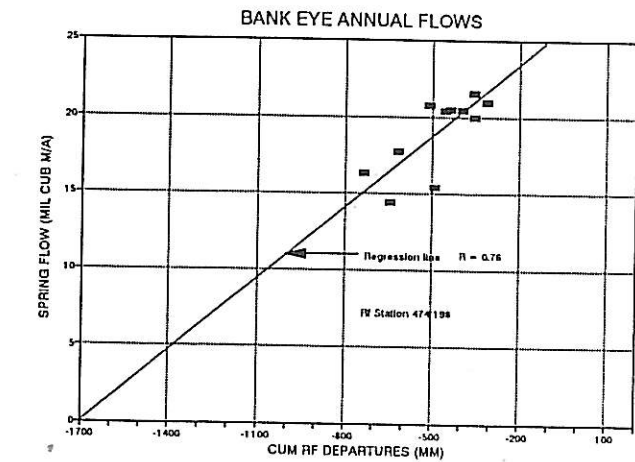


Figure 5.2 Flow of the Transvaal dolomite springs in relation to the cumulative rainfall departures

contamination, due to improper sanitation.

5. RECHARGE

Flow of Kuruman eye

The flow of the main spring (Kuruman A) has been gauged from 1959 to 1972 by means of a partial flume, but pumping by the municipality had affected the flow measurements. Smit however checked the flow gaugings and retained only readings during periods when the pumpage had been stopped for a reasonable long time. In this way an acceptable record of the flow for the period 1959-1972, was obtained. For the period 1974-1986 the flow gaugings had been abandoned because the spring flow had reached an all-time high rate in 1976 and again in 1988, which caused a drainage problem to the municipality and the partial flume was destroyed.

Smit (1978) had already indicated, by means of graphs, that there is an apparent relationship between the flow of the spring and the monthly cumulative rainfall differences (Figure 5.1).

Bredenkamp (1990) had proven that the relationship is a linear one, similar to that found for springs of the Transvaal area. This allows one to infer the annual flows over a long period by means of the cumulative rainfall departures from the mean. (refer Figure 5.2).

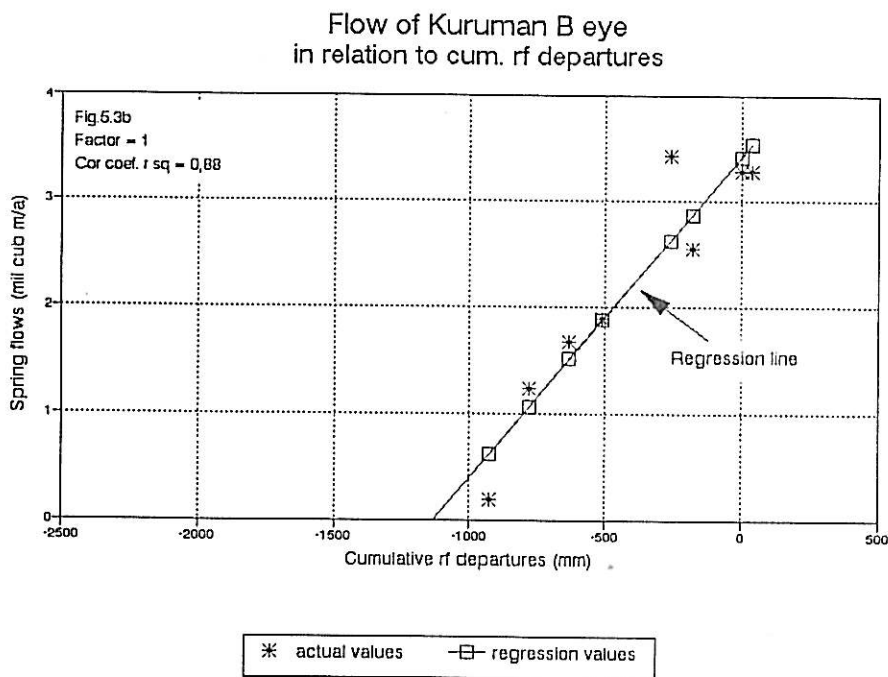
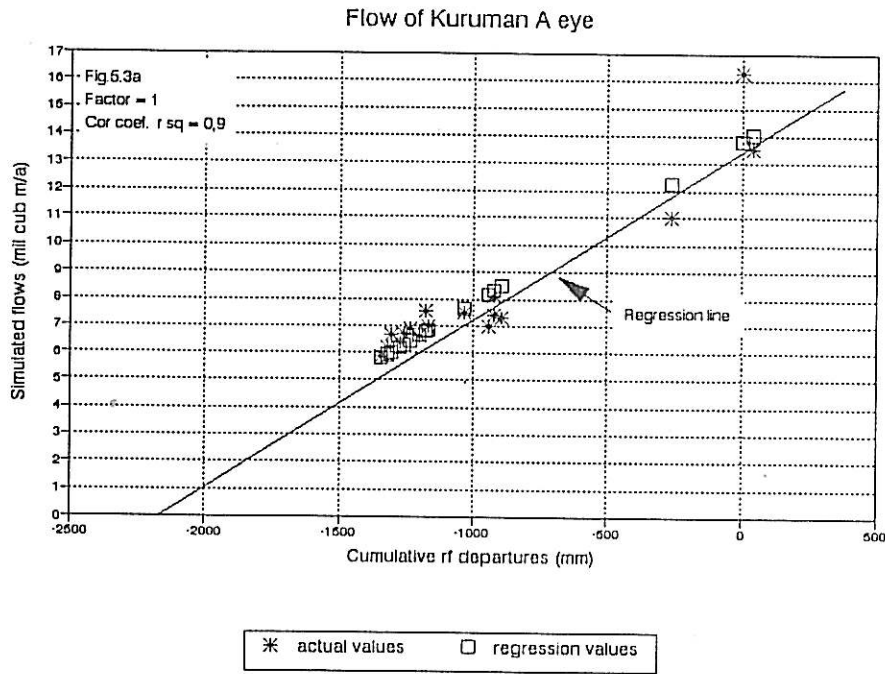


Figure 5.3 Flow of the Kuruman A and Kuruman B eyes in relation to the cumulative rainfall departures

Graph to show the rainfall factor shows the best correlation

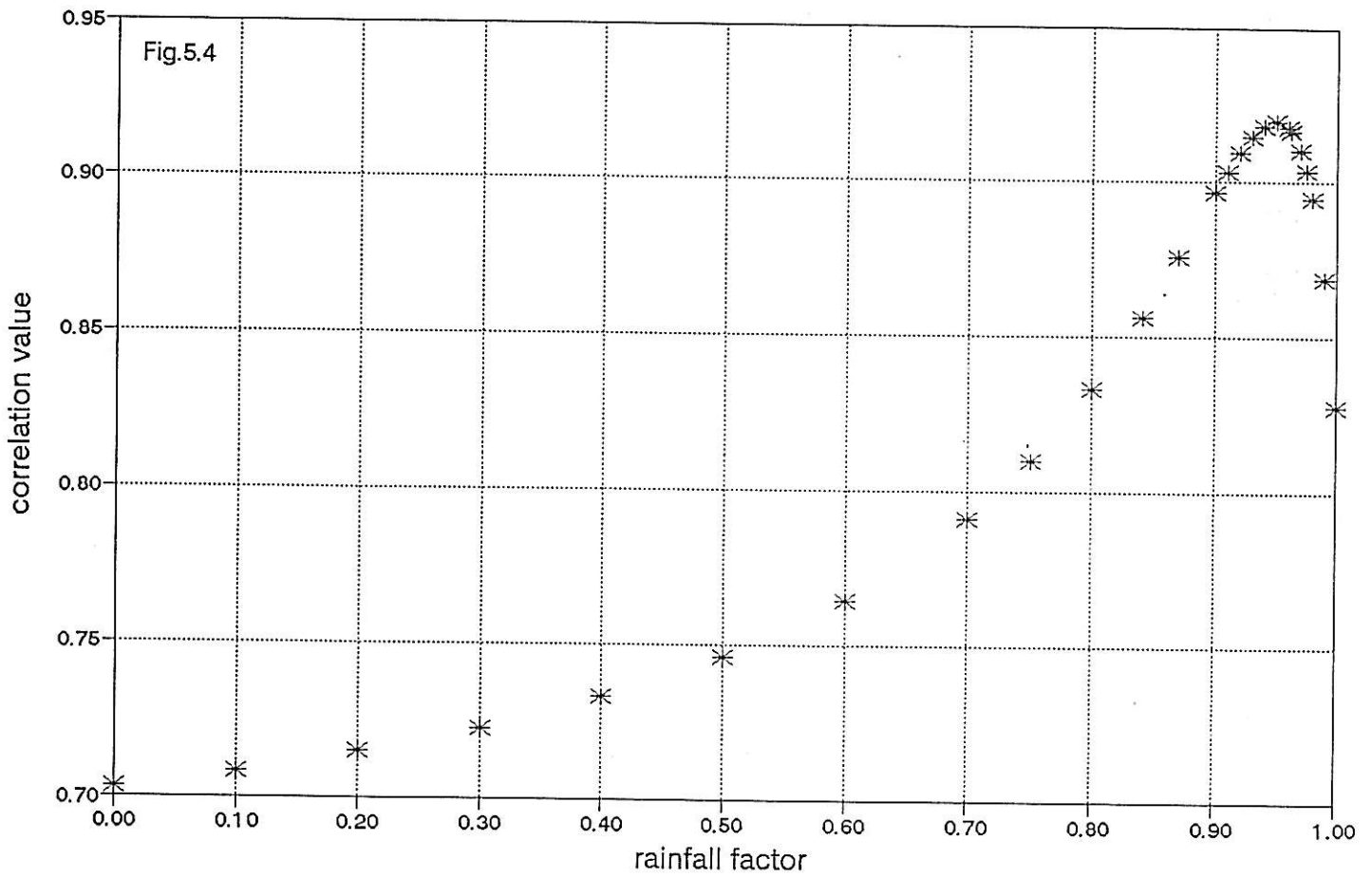


Figure 5.4 Relationship between the factor K and the best fit of cumulative rainfall departures with the flow of the spring

For both the of Kuruman and Manyeding B eyes the linear relationship with the cumulative rainfall departures is depicted in Figure 5.3, based on the data from Smit as well as a few records that were obtained by Van Wyk for the period 1985 - 1988. This linear relationship was used to estimate the annual discharge over an extended period for which a reliable rainfall record could be obtained.

The cumulative rainfall departures were derived using a reconstructed rainfall record that incorporated data from stations in the vicinity of which there are only a few. The following stations were used: 358/049, 358/216, 358/268, 358/316, 358/403, 358/577, 358/893, 358/808 as well as 393/778, 393/864. The data of the different stations were compared for the period 1925-1990. In total 15 outliers were eliminated. For the period March 1927 to October 1931 only the values of station 358/808 were used because records for the other stations were missing. For the rest of the period there were always at least three monthly values from which an average was determined for the final data set. The annual rainfall values for the period 1925 to 1990 were determined from the monthly data to find a representative rainfall record over the recharge area.

The correlation coefficient for the linear relationship, incorporating the cumulative rainfall departures, was checked using different factors by which to multiply the rainfall departures relative to the mean value. This revealed that a best linear regression is obtained if $K = 0.95$ was used in the

formula:

$$CRFD(I) = (RF(I) - K \times ARF) + CRFD(I - 1)$$

where CRFD = cumulative rainfall departure

RF(I) = rainfall

ARF = average rainfall

K = constant

Values for $K = 0.90$ and 1.0 produced much the same correlations (Figure 5.4), however to ensure a consistent application of the cumulative departures, a factor of 1 was used in all the cases, which implies that the rainfall departures relative to the average value of the precipitation was used.

6. ESTIMATION OF THE TRUE FLOW OF THE SPRINGS

The measurements of the flow of a spring are often not reliable because of inaccurate gaugings and improper structures that could allow leakage to take place underground. The reliable estimation of the spring flows, as had been indicated by Smit (1978) is the only way by which to estimate the recharge of the area in a reliable way, obviously making provision for irrigation development in the catchment and other losses. In the case of the Kuruman catchment the recharge is equivalent to the flow of all the springs emanating from the total drainage area. For the Kuruman eye drainage area there is some irrigation development, and some additional losses to the system occur in the form of

Simulated flow of Kuruman A and B eyes

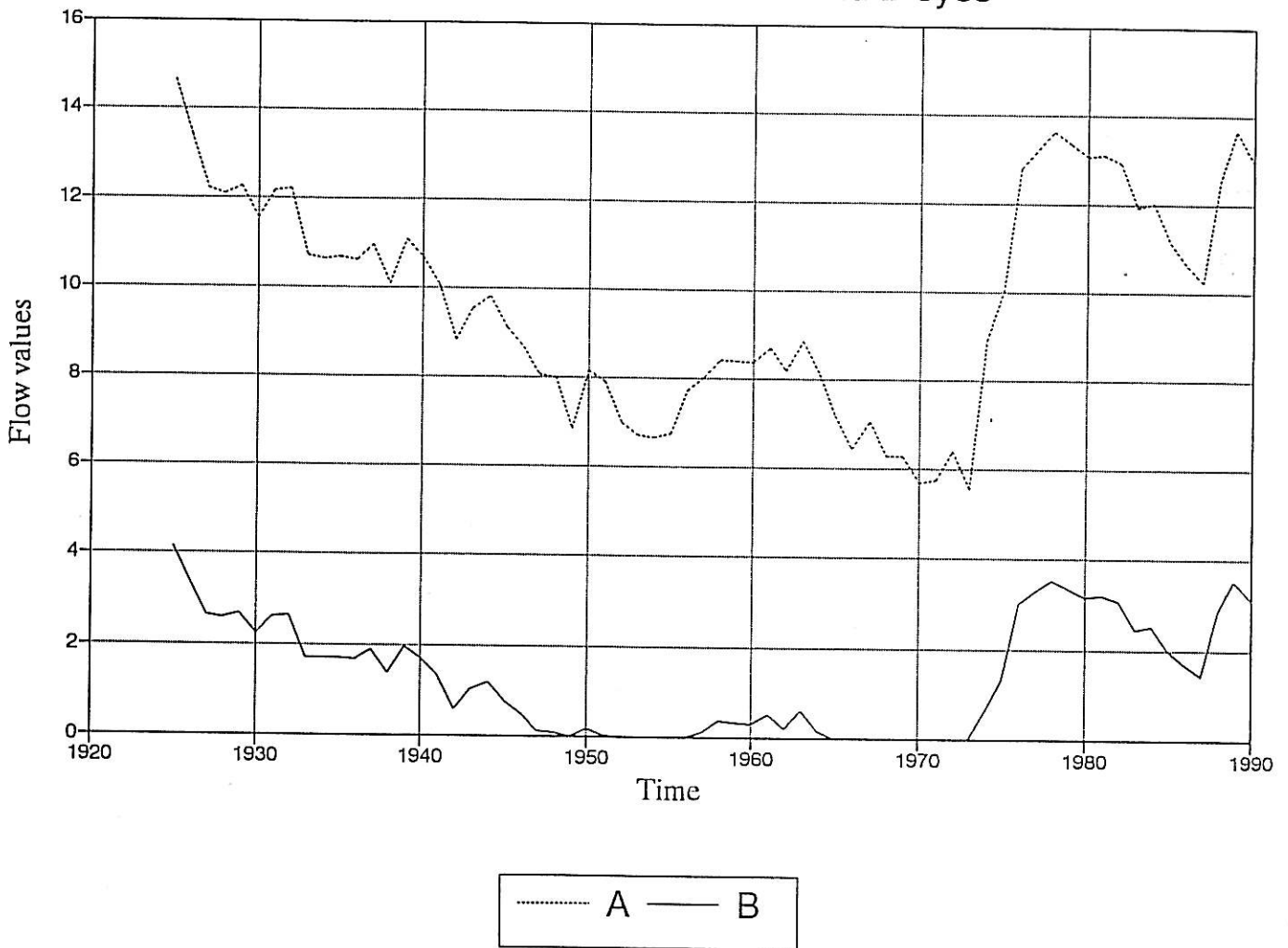


Figure 6.1 Flow of the Kuruman A spring for the period 1926 to 1991 partially inferred by means of the cumulative rainfall departures

evaporation from open waters at the springs, and water that is being used for domestic purposes and cattle farming.

Kuruman A, B and C eyes

Cumulative rainfall departure method

The annual variation of the flow of spring A, as was inferred from the cumulative rainfall relationship, is depicted in Figure 6.1. The high flows of 1974-1976 and those experienced from 1985-90 onwards are clearly shown. The flow of spring B, calculated in the same way, (Figure 6.1) indicates that the spring flow had been zero for the period 1948-1949, 1952-1956 and also from 1964 to 1974. The longterm average flow of Kuruman B eye is estimated at $1.07 \times 10^6 \text{m}^3/\text{a}$. The flow of the C eye can only be inferred by assuming that the ratio of flow at any stage allows the unknown spring in any year to be obtained. At the same time the variability of flow of the C eye will follow a line parallel to the linear regression for either Kuruman A or B eyes. There are unfortunately no reliable flow measurements of the C spring that could be used to reconstruct its flow reliably. For this reason a very rough estimate was made by setting the average flow at one third that of the B eye.

Utilizing a relationship with the Wondergat levels.

A study of the flow of several dolomite springs in the Transvaal region (Bredenkamp et al 1991) has indicated that a linear

Groot Kono Oog Spring
 Relationship with Wondergat levels

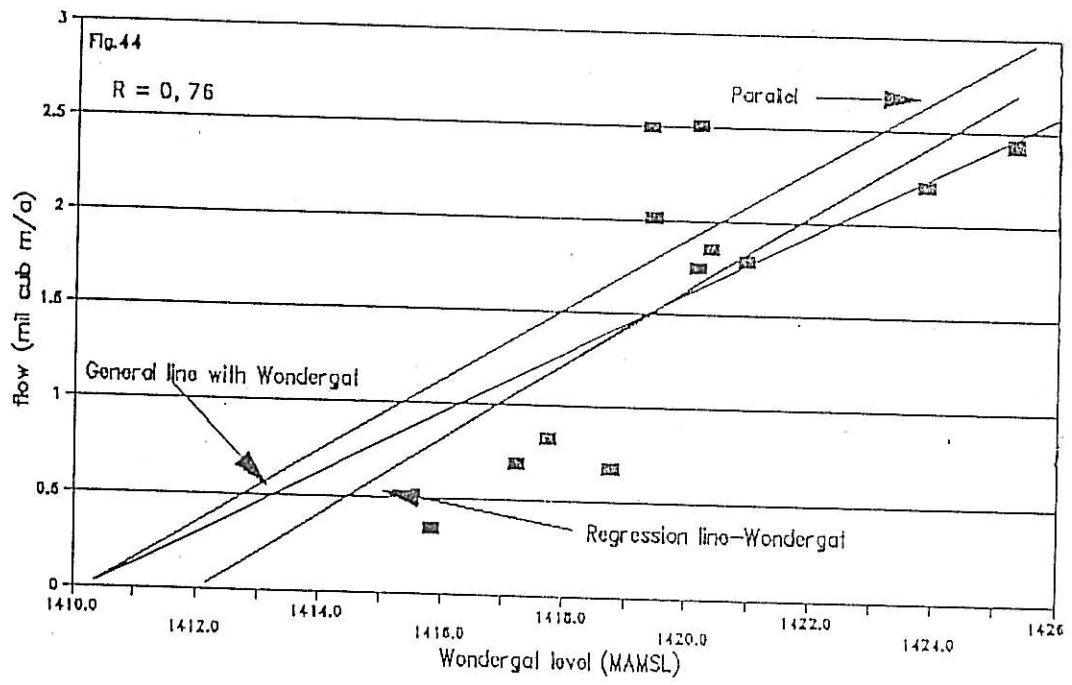


Figure 6.2 Flow of Groot Kono eye in relation to the Wondergat water levels

Kuruman A + B eye flow

A measured - B regression cum rf dif

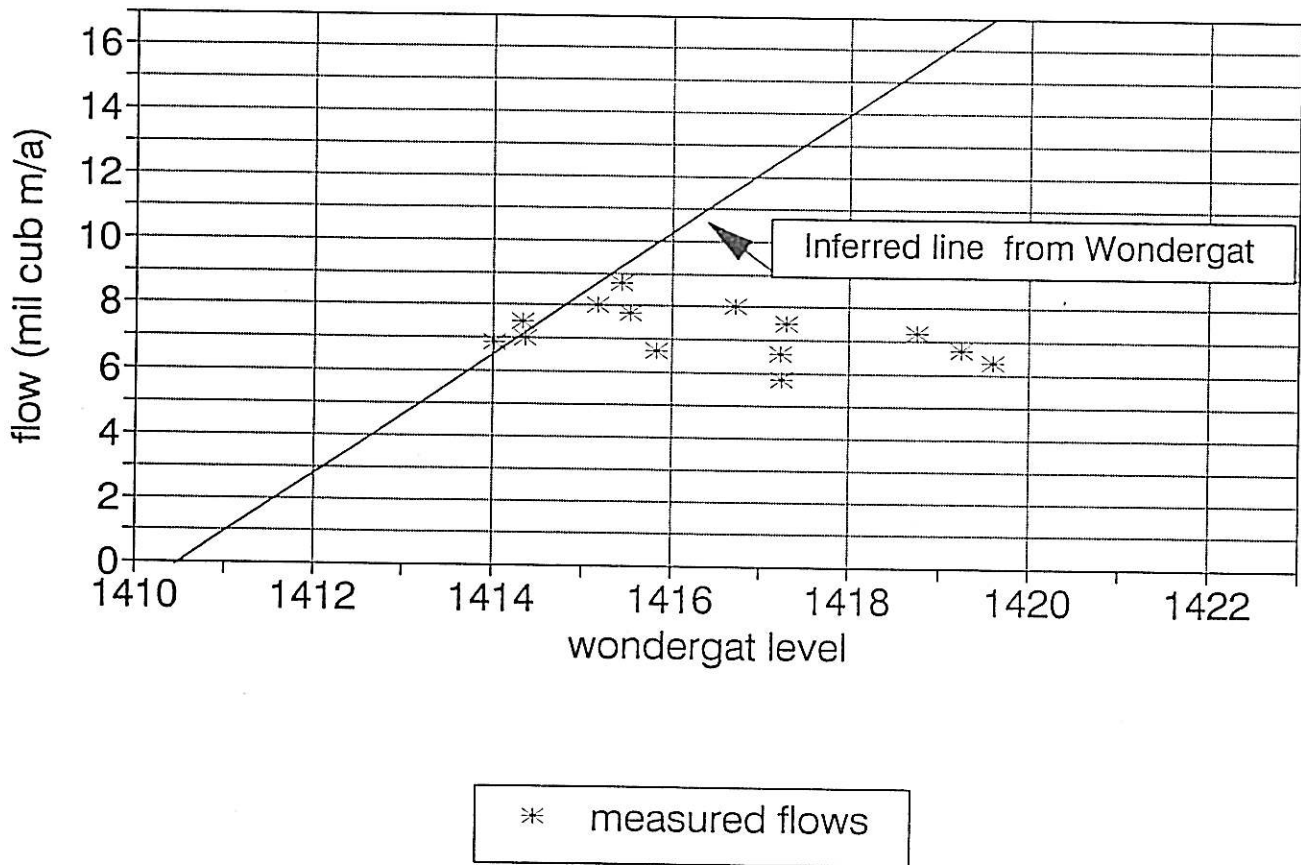
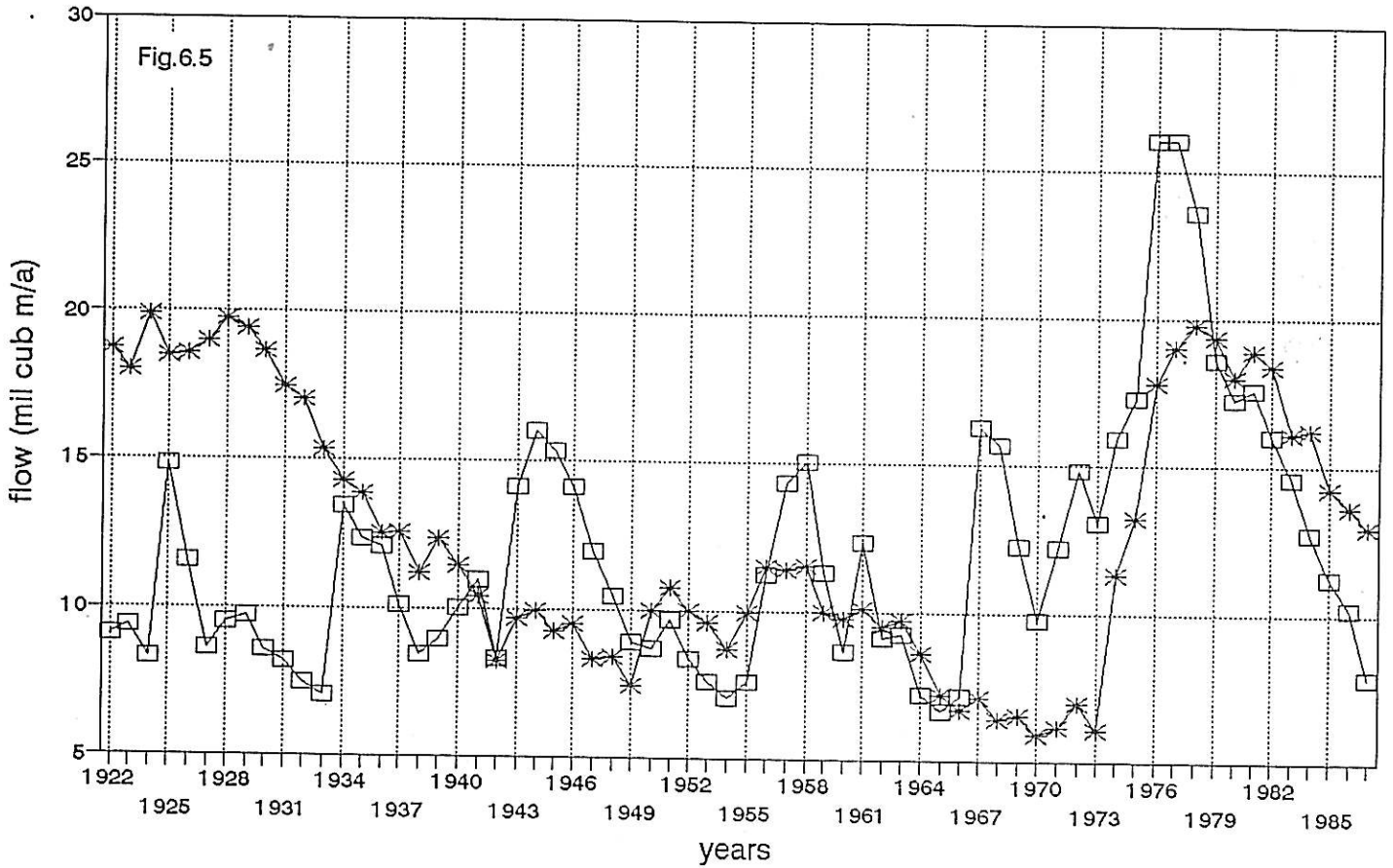


Figure 6.3 Flow of the Kuruman A and B eyes plotted against the Wondergat levels

Kuruman A + B eye flow vs time



*— inferred cum f dep □— wongat inferred

Figure 6.4 Comparison of the flow of Kuruman eyes derived from the cumulative rainfall departures and from the Wondergat levels

relationship between the Wondergat water levels and the flow of the springs exists. This relationship appeared to also hold for the Groot Kono eye (Figure 6.2) which also complied in respect of the magical cut-off level in the Wondergat level, indicating that the flow of the spring would stop if the corresponding level of 1410,5 mamsl in the Wondergat is reached.

This general relationship can be tested in the case of the Kuruman A eye, but it appears for the relationship to hold, that the high flows of the spring are underestimated. Even if the flow of the Kuruman B spring is added to the flow of the A spring, the general relationship with the Wondergat does not fit the data (see Figure 6.3). This could be due to the fact that the rainfall characteristics of Kuruman is different from the Bo Molopo (Wondergat), but it is more likely that the discrepancy is due to a portion of the flow leaking across the dyke that is not being measured, over and above the excess flow that appears at Kuruman B and C eyes. According to the piezometric map and high yielding boreholes along the Kuruman dyke, there could be outflow of subsurface leakage of an unknown quantity over the dyke.

By assuming that the general relationship between the spring outflow and the water levels of the Wondergat also holds for Kuruman A, a second estimate of the longterm average flow of the spring was obtained. This gave a flow of 11,9 mil cub m/a, which closely corresponds to the value 10,7 mil cub m/a, that was obtained from the cumulative rainfall departure relationship.

A plot of the two sets of annual values shows that the correspondence is surprisingly good for the full record from 1934 to 1991 (Figure 6.4). Incidentally the rainfall from 1934 onwards are the best because it was simulated from rainfall stations in close proximity to the Kuruman A spring. The main outliers appearing in Figure 6.4 are

- 1) the 1944 period, when the Kuruman area had apparently not experienced the same degree of excess rainfall as the Molopo area;
- 2) the 1985-1986 period, when the Kuruman region had received exceptionally high rainfall, which had not occurred to the same degree in the Molopo region.

Manyeding springs

There are only flow records available for the Manyeding B eye, which also exhibit a linear relationship with the cumulative rainfall departures (Figure 6.5). This allows the full flow record over the period 1934 to 1991 to be simulated and yields an average flow of 2,41 mil cub m. The flow of Manyeding A and C could only be determined approximately on a proportional basis according to the ratio of the available measured flow of Manyeding B in relation to its longterm average flow. The flow of Manyeding C had been measured at $0,12 \times 10^6$ cub metres per annum, yielding a ratio of 0,0265 to that of Manyeding B for the same year. On this basis the average flow of Manyeding C was

calculated at 0,19 mil. cub m/a. In the case of Manyeding A, no reliable records exists, but it appears that its flow is much less than even that of Manyeding C. If the flow is assumed to have been about 0,05 mil cub metres per annum in 1970, the longterm average flow of this spring is estimated at 0,06 mil cub metres per year.

7. RECHARGE OF THE KURUMAN DRAINAGE AREA

The recharge to the drainage area includes the combined flow of the springs discharging from the outer boundaries of the area, but it does not include the flow of the two Kono eyes or the Compton eyes which re-enter the drainage area to reappear eventually at one of the main eyes on the Kuruman dyke.

In addition to the spring flows, the losses that could be accounted for, also form part of the outflow that balance the recharge. These losses, comprising of domestic consumption, stock watering, irrigation, evaporation from the open pools of water at the springs as well as the evapotranspiration from the reed beds, have been estimated at 1,5 mil. cub metres per annum by Smit (1978). A reassessment of the irrigation has indicated that more irrigation is presently practised and that the water abstracted amounts to $2 \times 10^6 \text{m}^3$ per annum.

The sum of the reassessed water balance elements, measured in million cub m/annum, are the following:

Flow of Kuruman A = 10,7	Flow of Manyeding A = 0,06
Flow of Kuruman B = 1,07	Flow of Manyeding B = 2,41
Flow of Kuruman C = 0,357	Flow of Manyeding C = 0,19
Total of Kuruman eyes = 12,13	Total for Manyeding eyes = 2,66
Irrigation = 2	

Additional water balance elements (mil. cub. m/a) according to Smit (1978):

Stock watering = 0,277

Domestic consumption = 0,102

Evapotranspiration from wetland and open waters = 0,08

Total recharge to the area = 17,24

Since the area of recharge according to Smit is 1140 sq km the recharge expressed as an equivalent depth of precipitation is 15,1 mm. This means that the average recharge is 3,3% of the average precipitation and this was the average value used in the analysis of the C¹⁴, tritium and mass balance presentations, to infer the storativity of the aquifer. This estimate of the recharge is higher than that obtained by Smit (12,7 mm/a or 2,9% of the average rainfall of 440 mm/a).

8. NATURAL ISOTOPIC CONTENT OF THE SPRING WATER

8.1 Introduction

The Carbon-14 and tritium in the atmosphere can serve as useful environmental tracers to reveal the circulation of water in the aquifer and the admixture of the recharging water with the water stored and transmitted by the aquifer.

Although tritium and the C^{14} isotopes have been present in the atmosphere at a fairly constant concentration, before 1958, the C^{14} and tritium concentrations have both increased, especially in 1962 when hydrogen bomb fall-out had caused an increase (Figure 8.1). Both the tritium and Carbon-14 spikes can be used as a time reference for the hydrological input that had occurred in a specific period. At the same time the value for the C^{14} as measured in the spring, allows the effective average age of the water resulting from a mixture of the young recharging water with the base flow of the water emerging from depth to be calculated. The age refers to the time taken for the water with C_0^{14} concentration indicating young water to reach a concentration of C_t^{14} by decay of the C^{14} radio-isotope according to the formula:

$$C_t^{14} = C_0^{14} \times e^{-\frac{0,693 t}{t_{\frac{1}{2}}}}$$

where t = time for decay and

$t_{\frac{1}{2}}$ = half life of isotope (5546 yrs).

C-14 AND TRITIUM IN ATMOSPHERE

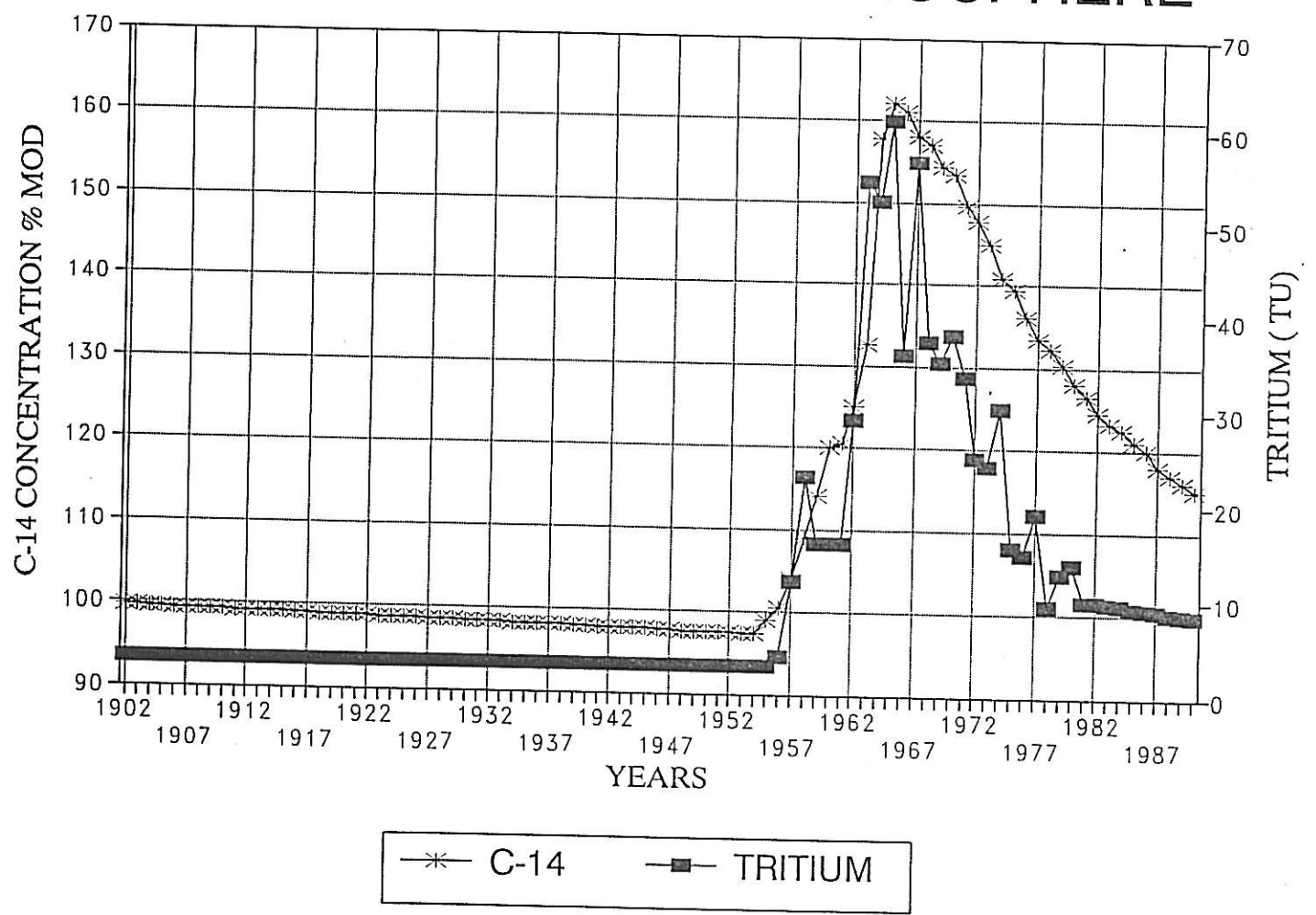


Figure 8.1 C¹⁴ and tritium in the atmosphere indicating the hydrogen bomb input

By means of a water balance, combined with the isotope concentrations of the spring water, the ratio of the recharge in relation to the water stored in the aquifer could be inferred. The recharge was derived by different rainfall recharge relationships which were found to give a good simulation of the Wondergat water levels in the Bo Molopo area. Furthermore the relative contribution of recharge from a particular year to the effective recharge that mixes with the water that is stored in the aquifer, can be estimated. Unfortunately, due to irregular sampling of the spring, only a few measurements of the carbon-14 and tritium in the spring water (refer Figure 8.3) were available to obtain the reliable quantitative interpretation.

8.2 Interpretation of C^{14} and Tritium concentrations

8.2.1 Introduction

The first isotope measurements date from 1967, which were already past the occurrence of the 1962/3 increased concentrations in the rainfall (Figure 8.1). However, most surprisingly the first tritium assay of the spring water showed zero tritium within the uncertainty limits of the measurements. The plot of C^{14} and tritium (Figure 8.2) shows a high degree of similarity in the responses. The measured C^{14} concentration of the spring in 1967 were 69,3% of modern C^{14} which represent water with a corrected age of about 2000 years assuming that recent groundwater has a C^{14} concentration of 85% of modern C^{14} (Vogel, 1966). In the next series of measurements, the tritium levels remained close

KURUMAN EYE ISOTOPE DATA

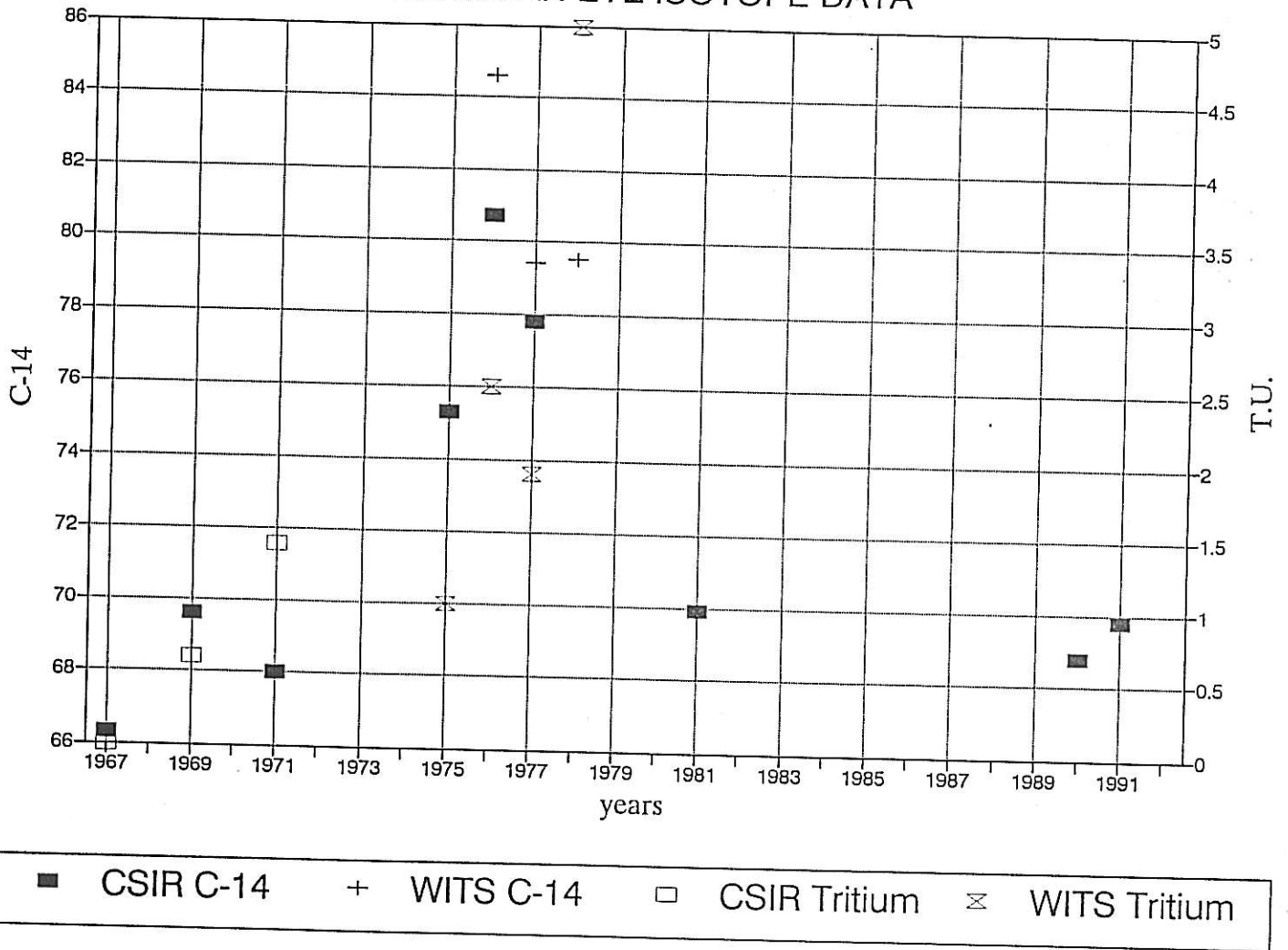


Figure 8.2 Measured C^{14} values versus tritium concentrations

KURUMAN A EYE C-14 DATA AND FLOW DATA

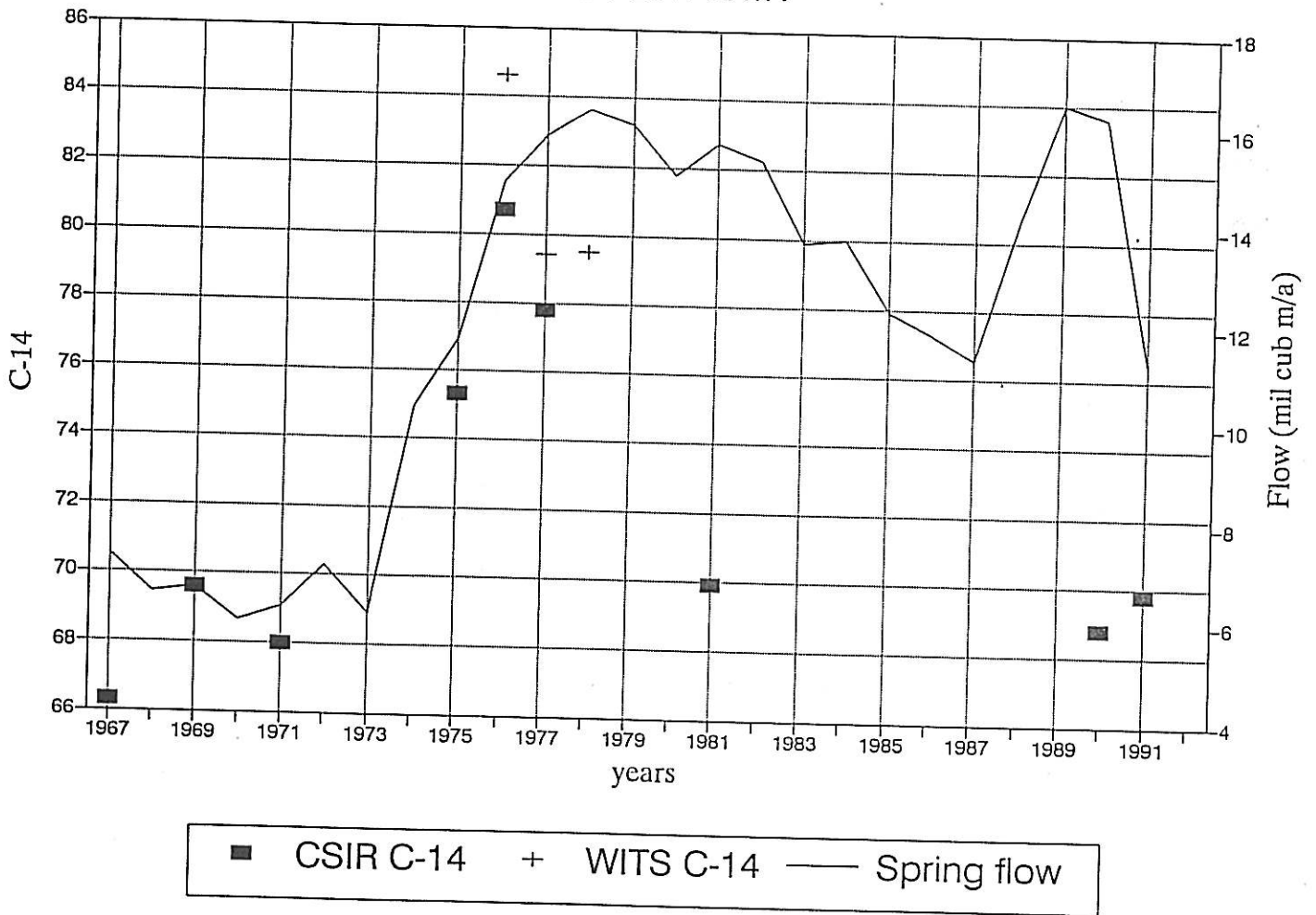


Figure 8.3 Composite graph of the flow of the Kuruman A eye and the measured C^{14} and tritium concentrations of the spring

to zero but the carbon-14 concentration of the spring water showed a definite increase (refer Table 3.1). There appeared to be a slight lag between the response of the C^{14} and tritium in the water flowing from the spring (Figure 8.2 and 8.3) but for the rest the correspondence between the C^{14} and tritium, was reasonably good. The maximum tritium value was reached about one year later than the peak of the C^{14} . This unfortunately had been the last tritium measurement taken of the Kuruman A spring, whence only C^{14} concentrations were obtained and measured irregularly.

8.2.2 Interpretation of natural isotopes occurring in the spring water

The first interpretation of the tritium and carbon-14 content of the spring water, had already produced some interesting results (Bredenkamp 1978). The carbon-14 content of the Kuruman spring was simulated whilst it was simultaneously tried to match the levels of tritium that had been observed.

The first simulation of the spring C^{14} values incorporated the following simple mass-balance equation:

$$V \times C_{i+1}^{14} = V \times C_i^{14} + RE \times C_0^{14} - RE \times C_{i+1}^{14} \dots \text{eq. 8.1}$$

where V = the volume of the water stored in the aquifer,
 RE = the recharge, which was assumed to be the same
 for all years,

C_0^{14} = C^{14} concentration of the recharge water that
 reaches the water level,

C_i^{14} = C^{14} concentration of the spring water in the year
 i and C_{i+1}^{14} the values for the following years.

Considering that $V = RE \times RT$ where RT is the turnover time of the
 water in the aquifer, eq. 8.1 reduces to

$$C_{i+1}^{14} = \frac{RT \times C_i^{14} + C_0^{14}}{RT + 1} \quad \dots, \text{ eq. 8.2}$$

which shows that the concentration of the isotope for year $i+1$
 is not dependent on the recharge RE assuming the recharge to be
 constant for all years. The concentration C_{i+1}^{14} however depends
 on

- C_i^{14} the isotope concentration of the water emanating
 from the spring,
- the isotope concentration (C_0^{14}) which represents
 recent water entering the aquifer,
- and the turn-over time RT of the water that is stored

in the aquifer, where RT is the ratio of the water stored in the aquifer to the annual recharge.

According to this simplified approach, the effect of the annual volume change of water in the aquifer does not come into play, basically, because it does not change in view of the assumption that the recharge is constant, and therefore the water that annually recharge will also flow out at the same rate. If the turnover time RT of the aquifer is large, equation (b) reduces to

$$C_{i+1}^{14} = C_i^{14} + \frac{C_0^{14}}{RT}$$

or by rearrangement gives:

$$C_{i+1}^{14} \times RT = C_i^{14} \times RT + C_0^{14}$$

The carbon-14 concentration of recent recharging waters, is generally assumed to be 85% modern C^{14} (Vogel 1966, Verhagen 1972), and not 50% modern, considering that the C^{14} concentration of carbon being dissolved from the dolomite must be zero. The higher value of 85% modern is due to the input of additional C^{14} by exchange biogenic CO_2 in the root zone. The acceptability of this initial level of C^{14} is still a point of uncertainty and controversy, especially for semi-arid regions where it is possible that recharge water can enter the aquifer whilst only partial admixing with biogenic carbon dioxide, could occur. Thus

the initial value of the groundwater for juvenile waters could lie between 50% and 85% modern C^{14} . This means quite a large variation which, at the present stage, cannot be defined more accurately in view of the many uncertainties about the recharge mechanism and the difficulties of quantifying it, as well as the mixing that takes place within the aquifer.

The tritium and C^{14} concentrations of the Kuruman A spring had been measured by the CSIR, and a few samples were measured by the laboratory of the Schonland Institute. The present re-evaluation of the data was based mainly on the C^{14} isotope values of the CSIR, whilst for the tritium both the Schonland Research Institute and the CSIR measurements were used.

In the first interpretation of the Kuruman eye isotope values, Bredenkamp (1978) has indicated that the initial C^{14} content points towards 75% of modern C^{14} , in order for the measured values of both C^{14} and tritium in the spring water to give a reasonable match. The volume of the water stored in the aquifer, derived from this balance, had to be at least 22 times the volume of water that is recharged annually. For both the tritium and the C^{14} results to match, the concentration of the tritium of the water entering the saturated aquifer, had to be 5 TU. This is much lower than the tritium levels of the rainfall at the time of recharge. These results of the first simulations are depicted in Figure 8.4.

If the latest isotope data set for the spring is analyzed, (refer

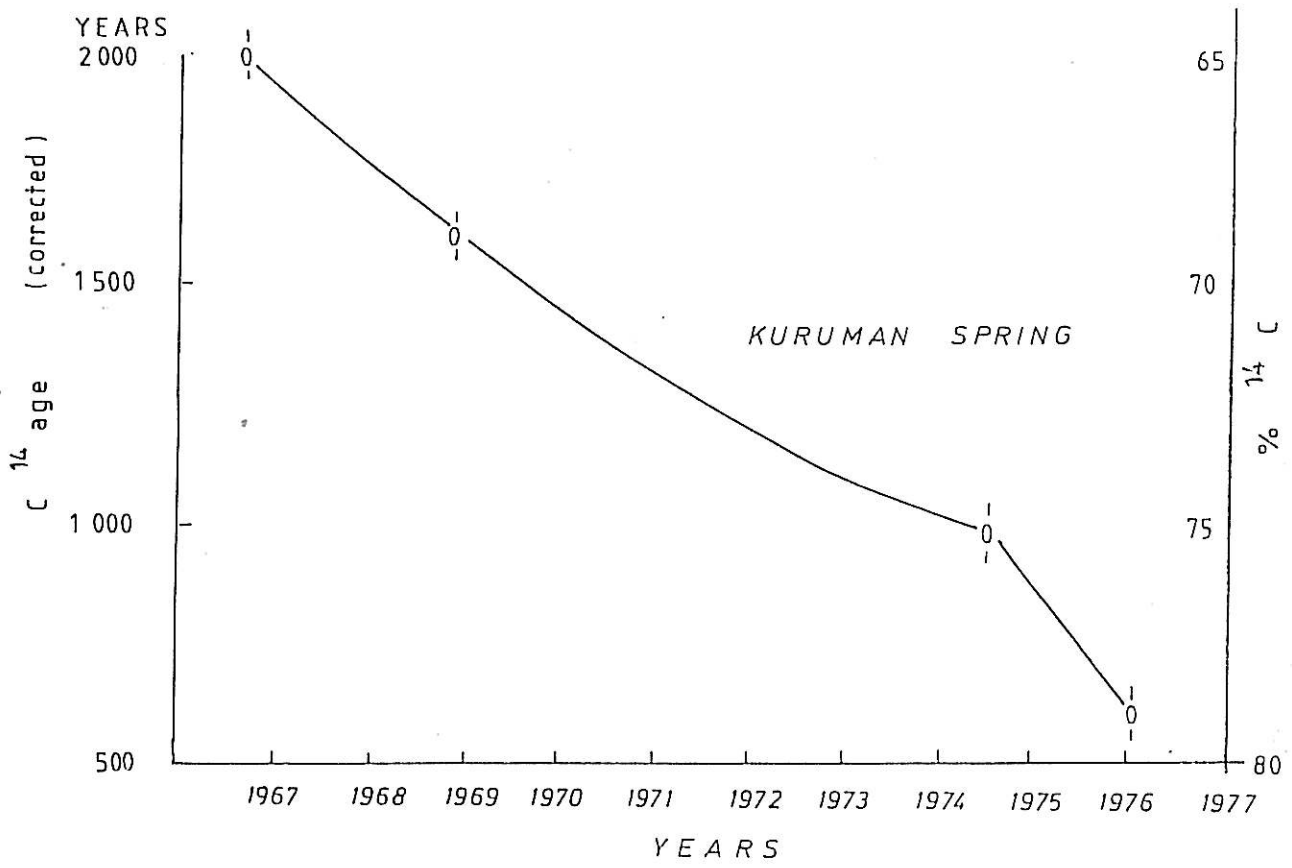


Figure 8.4 Simulated C^{14} concentrations for Kuruman A spring according to Bredenkamp (1978)

Figure 8.3) it appears that the highest value of tritium was also the last measurement. The lag in the appearance of tritium compared to that of the incident rainfall in a specific year, agrees with the premise of piston-like displacement of the water in the soil zone. This could explain why the higher levels of the tritium in the rainfall, had only reached the saturated aquifer some time after the C^{14} pulse.

This bears evidence that the mechanisms of tritium and C^{14} input are indeed different. The C^{14} input is effected at the root zone of the plants which occurs at some depth below the surface while the tritium is displaced through the soil profile by the infiltrating water. This in essence imply that the water originating from a few years back would reach the water table before the water entering as recharge (see schematic presentation in Figure 8.5).

The first part of the tritium and C^{14} response shown in Figure 8.3, was much the same as the reaction of the flow during the rising phase but the decline of flow was much slower than that of the C^{14} concentrations, after the peak had been reached. In view of the absence of tritium values beyond the time when the peak concentration was measured, the response of the tritium could only be inferred by simulation which allowed for a better comparison of the tritium and C^{14} response to recharge.

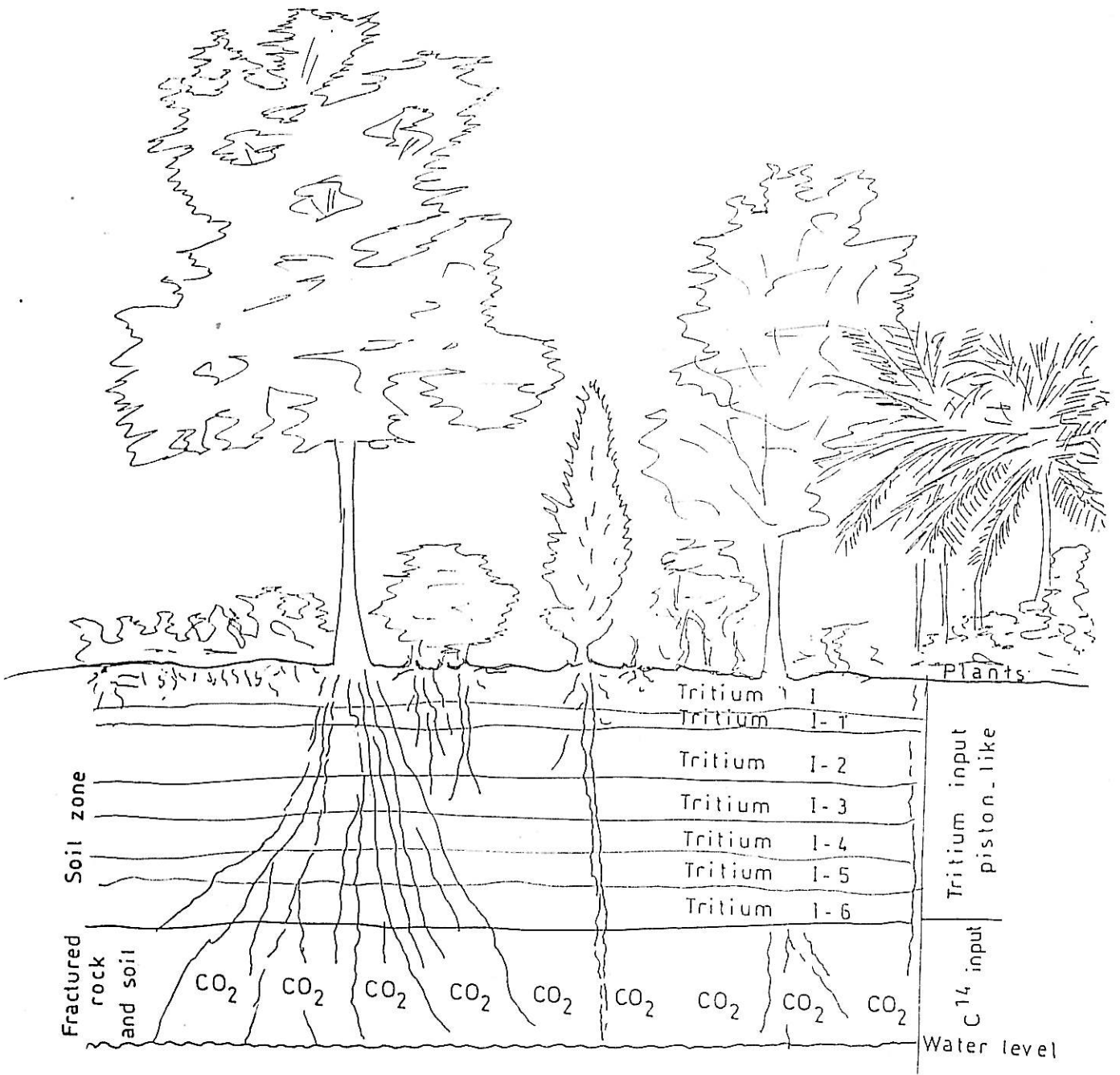


Figure 8.5 Schematic diagram of the input of tritium and C^{14}

9. RE-INTERPRETATION OF THE C¹⁴ AND TRITIUM CONCENTRATIONS OF THE SPRING

9.1 Introduction

A more refined quantitative interpretation of the reappearance of the C¹⁴ isotopes was attempted, using the available data and incorporating a more exact balance of input and output. Provision was made for variable recharge input and outflow. The variable recharge, similar to the simulation of the Wondergat recharge, could be inferred from the rainfall by simple rainfall-recharge equations. The effect of the recharge in a specific year was shown to depend on the recharge over the past 4 years. At the same time the effective flow of the spring, after it had been corrected to account for consumptive losses (refer section 7) was incorporated in the mass balance.

For the analysis of the admixture of the recharge with the water stored in the aquifer, a simplistic approach using a single cell mixing model was adopted.

The C¹⁴ concentrations and tritium content of the spring water were simulated in the following way:

- (i) The variation of Carbon-14 in the atmosphere which is shown in Figure 8.1 depicts the values since the turn of the century. This clearly marks the nuclear C¹⁴ pulse which is the signal that affected the concentration of the

emerging spring water and that could be related to the admixture of the rainfall recharge water, with the bulk of the water contained in the aquifer system having a lower C^{14} concentration. In the case of C^{14} no allowance for radioactive decay had been made due to its long half-life, of about 5546 years.

- (ii) The annual recharge was estimated from the rainfall using different rainfall-recharge equations, which according to Bredenkamp (1989), can produce good simulations of the annual variation of recharge in the Bo Molopo area.
- (iii) For tritium the input is depicted in Figure 8.1 according to measurements of tritium in precipitation for Pretoria and Johannesburg (Verhagen 1972).

9.2 Recharge equations

The first equation implies a constant percentage of rainfall as the recharge:

$$RE(I) = A \times RF(I) \quad \dots \text{eq. 9.1}$$

where $RE(I)$ is the rainfall for year I , and A represents the coefficient (%) of the rainfall that effectively constitutes the recharge to the groundwater.

The second recharge equation is:

$$RE(I) = \frac{A \times RF(I) \times RF(I)}{ARF} \dots \text{eq. 9.2}$$

which is the same as eq. 9.1 except that the coefficient A is weighted by the ratio of the actual rainfall to average rainfall. This conceptually conforms to an increased percentage recharge for rainfall exceeding the average value. This equation has produced reasonable simulations of the Wondergat levels.

The third equation assumes the recharge to be related to the departure of the rainfall relative to the mean rainfall:

$$RE(I) = A(RF(I) - ARF) \dots \text{eq. 9.3}$$

This is the same as eq.(9.4) where B = 1, and implies that recharge will only be effected if the outcome of eq. 9.3 is positive. This equation is essentially the one that was used by Temperley (1978), Smit (1978) and others to show a linear relationship between the cumulative rainfall departures and the water levels or springflow - A was conveniently assumed to be 1. Different values of A would not affect the degree of correlation in the linear relationship, but by implication the average recharge could have different values.

The fourth equation:

$$RE(I) = A(RF(I) - B \times ARF) \dots \text{eq. 9.4}$$

is the general recharge equation that depicts a regional relationship as well as the annual variation of recharge (Bredenkamp 1990). The value of B could vary from about 0,30 to close to 1 and still produce a good fit, and the equation implies that the recharge effectively only occurs, if the rainfall exceeds a threshold value. The factor $B \times ARF$ represents the threshold rainfall value, whilst the variable A denotes the average portion of the aquifer area that effectively produces the recharge. The parameter A and B represent a lumped parameter aquifer model to obtain the recharge. For the Bo Molopo area the value of $B \times ARF = 310$ yielded good results for the simulation of the annual variability of recharge and for regional recharge, $B \times ARF = 313$ proved to be the threshold value.

The fifth recharge equation, stems from the first but implies that the recharge percentage is inversely proportional to the rainfall in relation to the mean. The equation reduces to the following:

$$RE(I) = A \times ARF \left(1 - \frac{B \times ARF}{RF}\right) \dots \text{eq. 9.5}$$

Application of the different recharge equations is discussed in the next section.

9.3 Single cell admixture model for C^{14} and Tritium

9.3.1 General

The admixture of the recharge water, which contains the higher concentration of hydrogen bomb isotopes, was investigated

- 1) determining the annual recharge by means of each of the recharge equations mentioned in the previous section, and
- 2) varying the relative contribution of the annual recharge that effectively causes the water level response in the aquifer.

Allowance was made for four years preceding a specific year to contribute to the recharge of that year e.g. u % of the recharge of year I , v % for year $I-1$, w % for $I-2$, x % for $I-3$ and y % for year $I-4$, as being the relative contributions to the recharge of a specific year I .

The C^{14} concentration of the spring water was estimated by the following mass balance:

$$Q(I) \times C^{14}(I) = (RT \times z \times C^{14}(I-1) + t(u \times RE(I) \times C^{14}(I) + v \times RE(I-1) \times C^{14}(I-1) + w \times RE(I-2) \times C^{14}(I-2) + x \times RE(I-3) \times C^{14}(I-3) + y \times RE(I-4) \times C^{14}(I-4)) - Q(I-1) \times C^{14}(I-1)) \div z$$

where $Q(I)$ to $Q(I-4)$ refer to the spring flow during the

respective years, z is the turnover period (volume of the water in the aquifer relative to the average recharge), while $C^{14}(I)$ is the carbon-14 concentration of the precipitation of the year (I) . The value of $C^{14}(I-i)$ represents the isotope concentration of the main groundwater component that mixes with the recharge water, and was assumed to be the C^{14} concentration i years prior to the year (I) . This value of the C^{14} concentration signifying the C^{14} of the base flow could have been assumed constant for the present interpretations, corresponding to the initial constant C^{14} concentration in the atmosphere, but it must change as the effect of the increased C^{14} concentrations in the atmosphere will eventually admix with the base flow. If the C^{14} measurements were to be continued for a long enough period in the future, it would shed light on the expected change of the base flow value.

9.3.2 Simulation results

For the various recharge equations the goodness of the fit between the actual and the simulated values is shown in Figures 9.1 to 9.5. The degree of correspondence that was obtained between the measured and the simulated values is reflected by the sum of the squared residuals, which is denoted by ERR ie. :

$$ERR = \sum [MI(I) - SI(I)]^2$$

where $MI(I)$ = measured value of C^{14} isotope in the spring for year I ,
 $SI(I)$ = simulated value of C^{14} isotope for spring water.

KURUMAN EYE C-14 VALUES

$RE = a[RF(I)]$

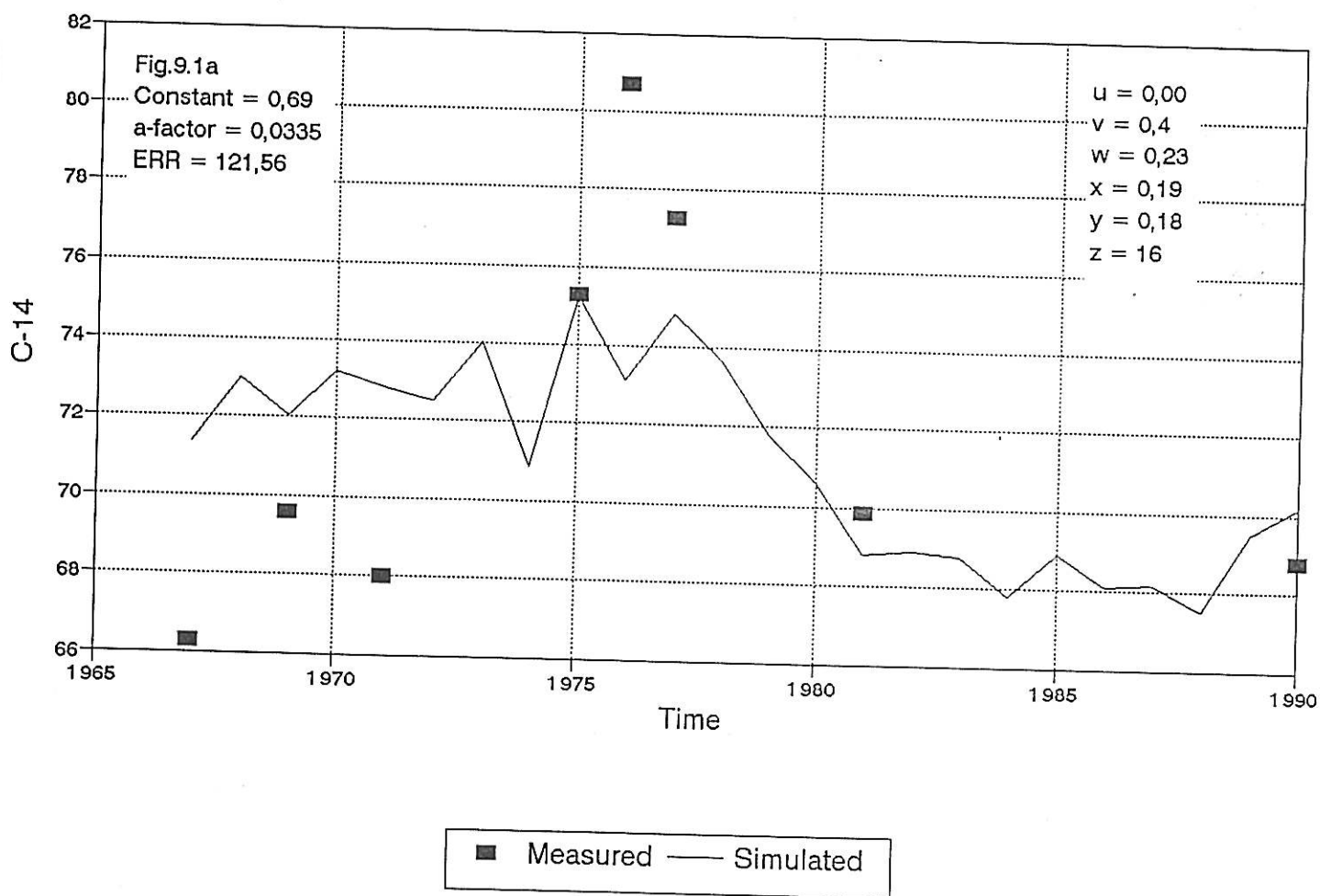


Figure 9.1(a) Simulated C^{14} concentrations based on recharge eq. 1 - constant %. Variable contribution from preceding years

KURUMAN EYE C-14 VALUES

$RE = a[RF(I)]$

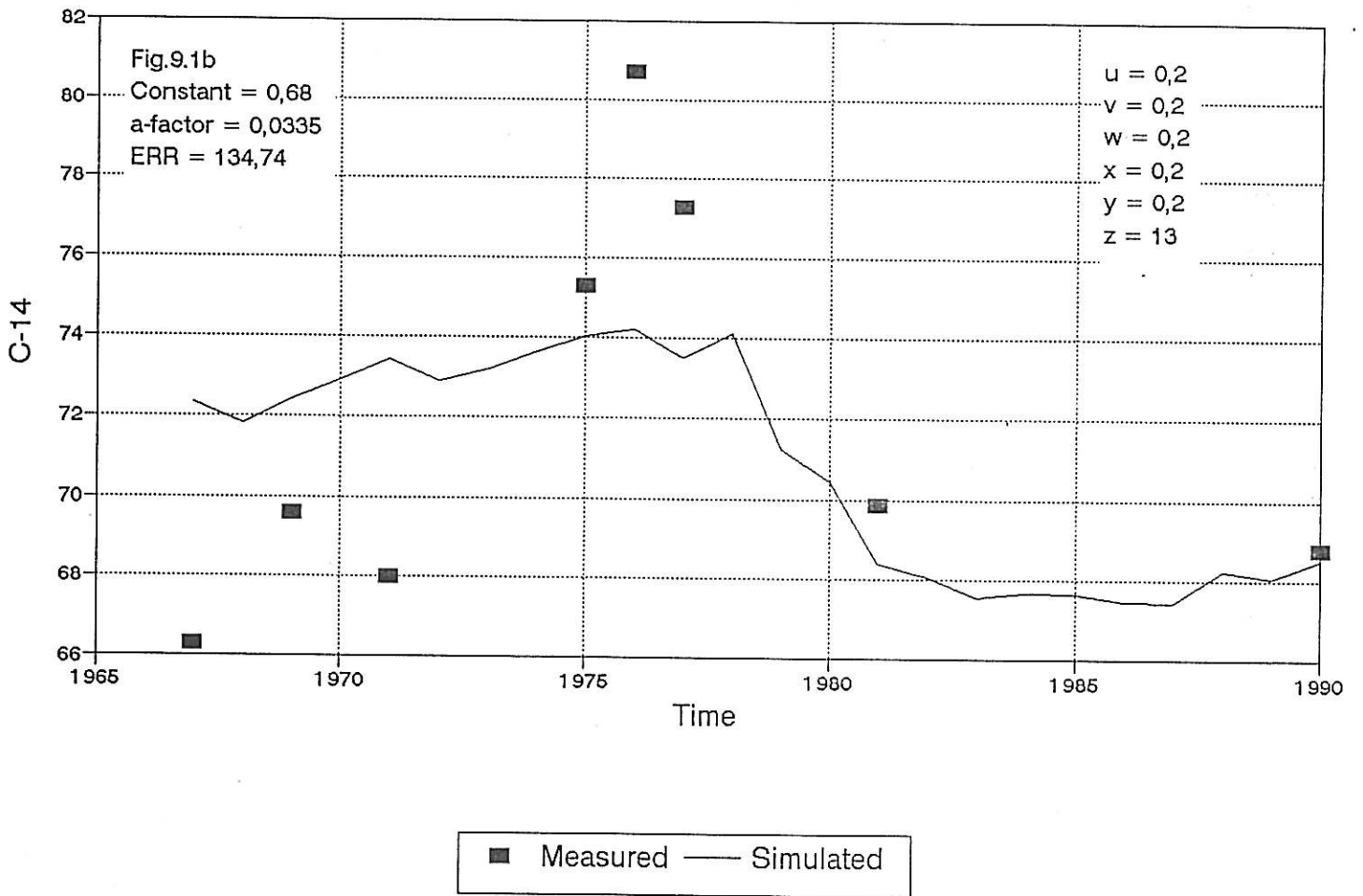


Figure 9.1(b) Simulated C^{14} concentrations based on recharge eq. 1 ; but with equal contribution from preceding years

KURUMAN EYE C-14 VALUES

$RE = a[RF(I)/ARF]RF(I)$

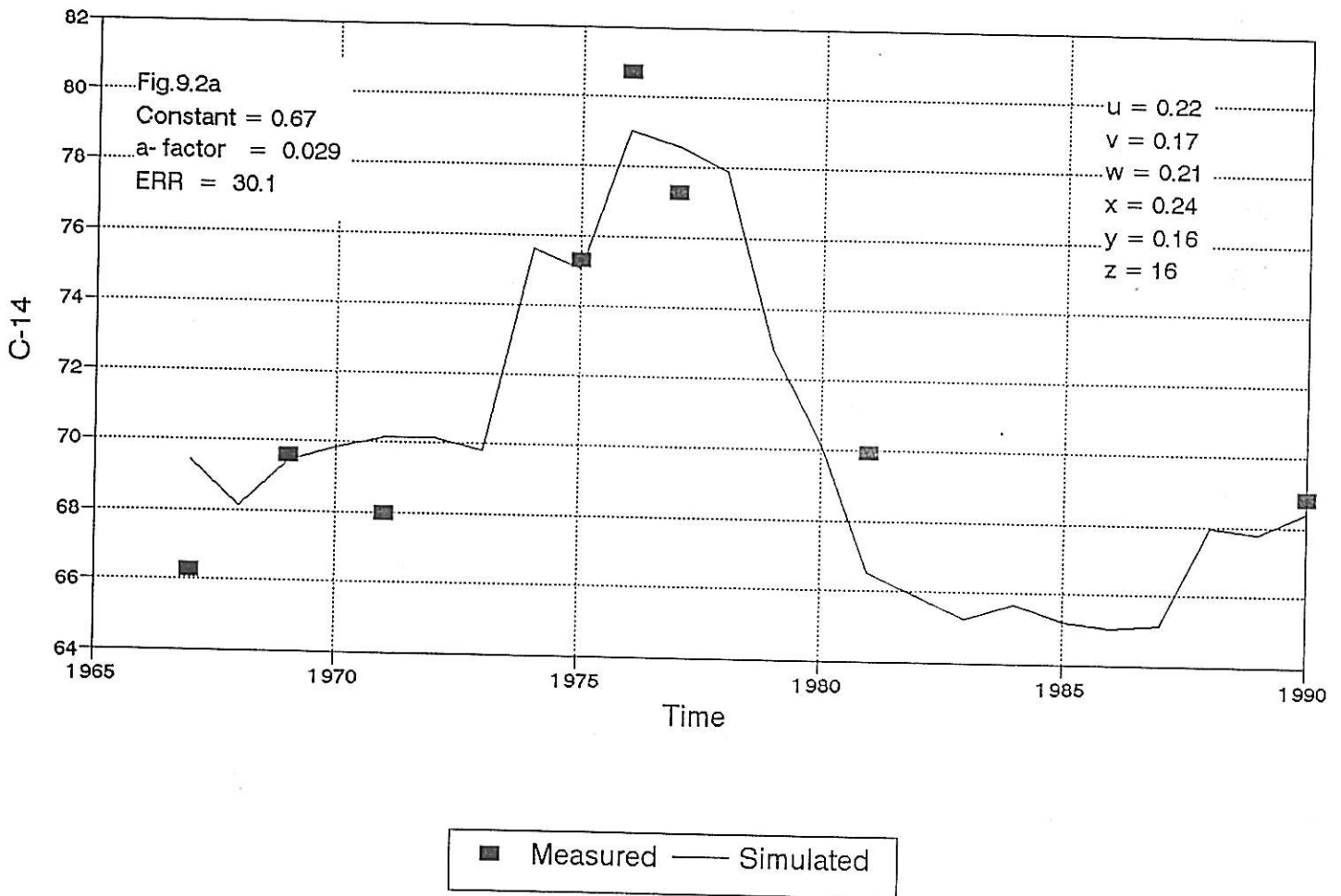


Figure 9.2(a) Simulated C^{14} concentrations for recharge eq. 2 - weighted %. Variable contribution from preceding years

KURUMAN EYE C-14 VALUES
 $RE = a[RF(I)/ARF]RF(I)$

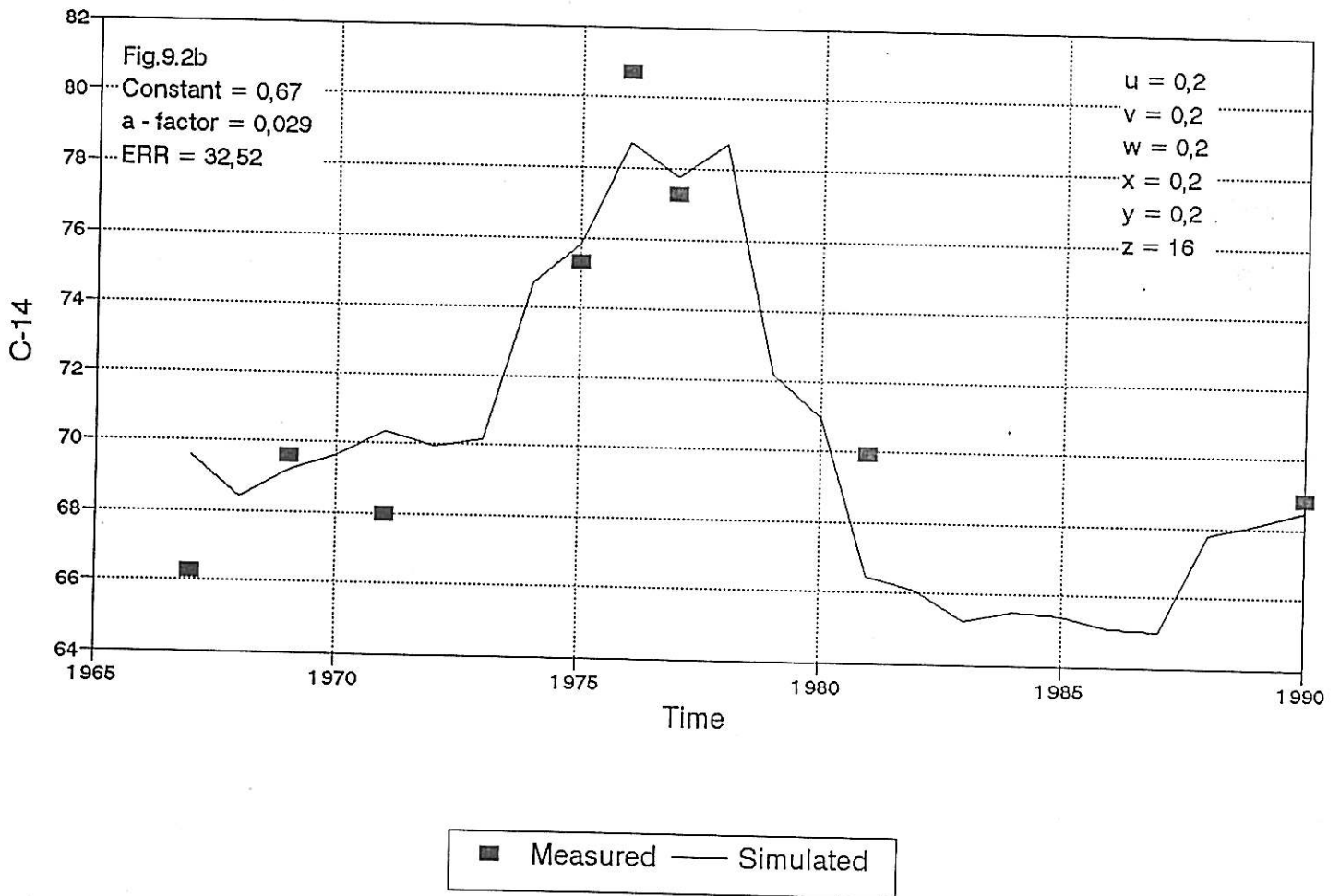


Figure 9.2(b) Simulated C^{14} concentrations for recharge eq. 2 - weighted %. Equal contribution from preceding years

KURUMAN EYE C-14 VALUES

$RE = a[RF(I) - ARF]$

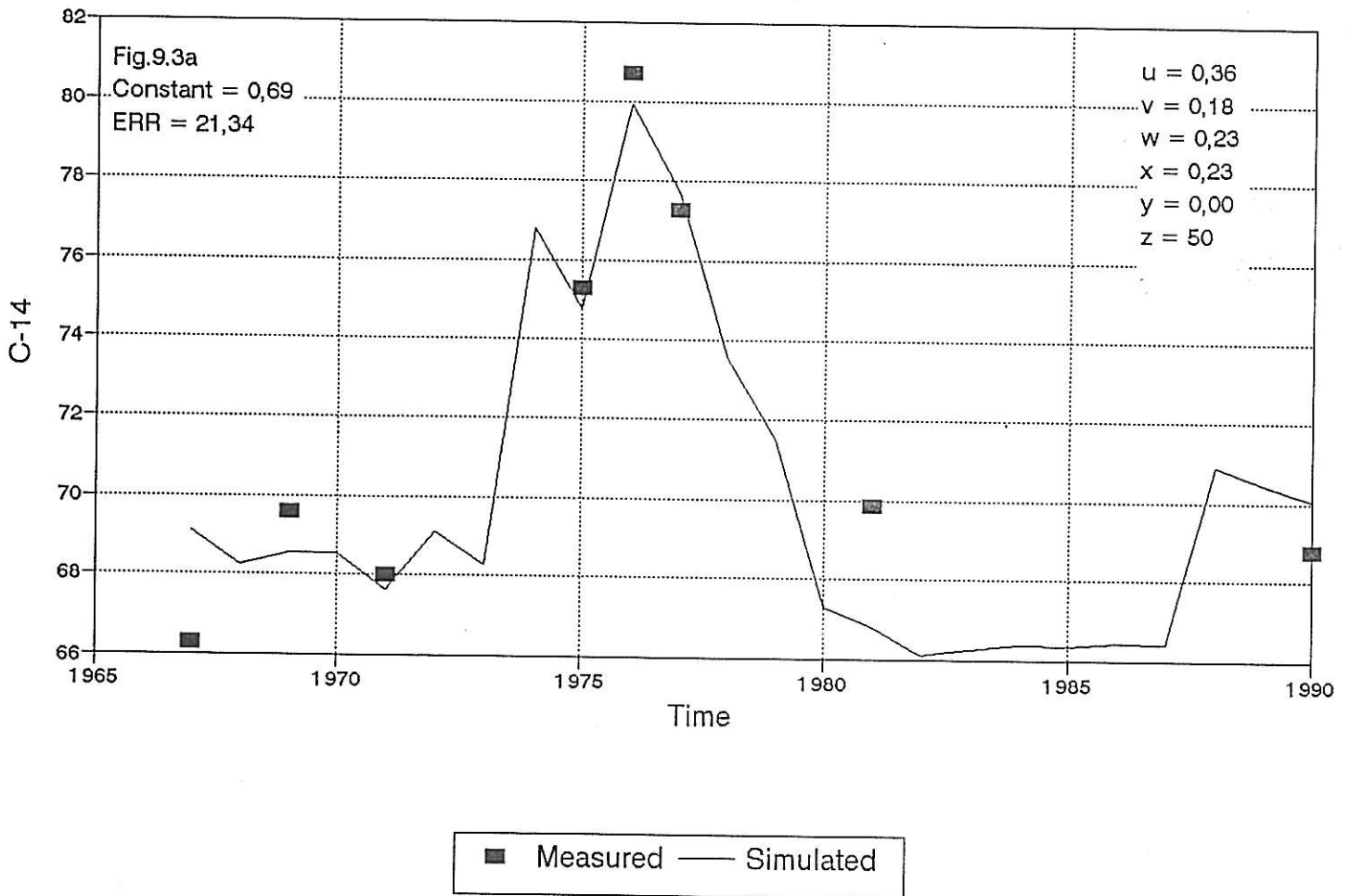


Figure 9.3(a) Simulated C^{14} values based on recharge eq. 4 - excess to average annual rainfall. Variable contribution from preceding years

KURUMAN EYE C-14 VALUES
 $RE = a[RF(I) - ARF]$

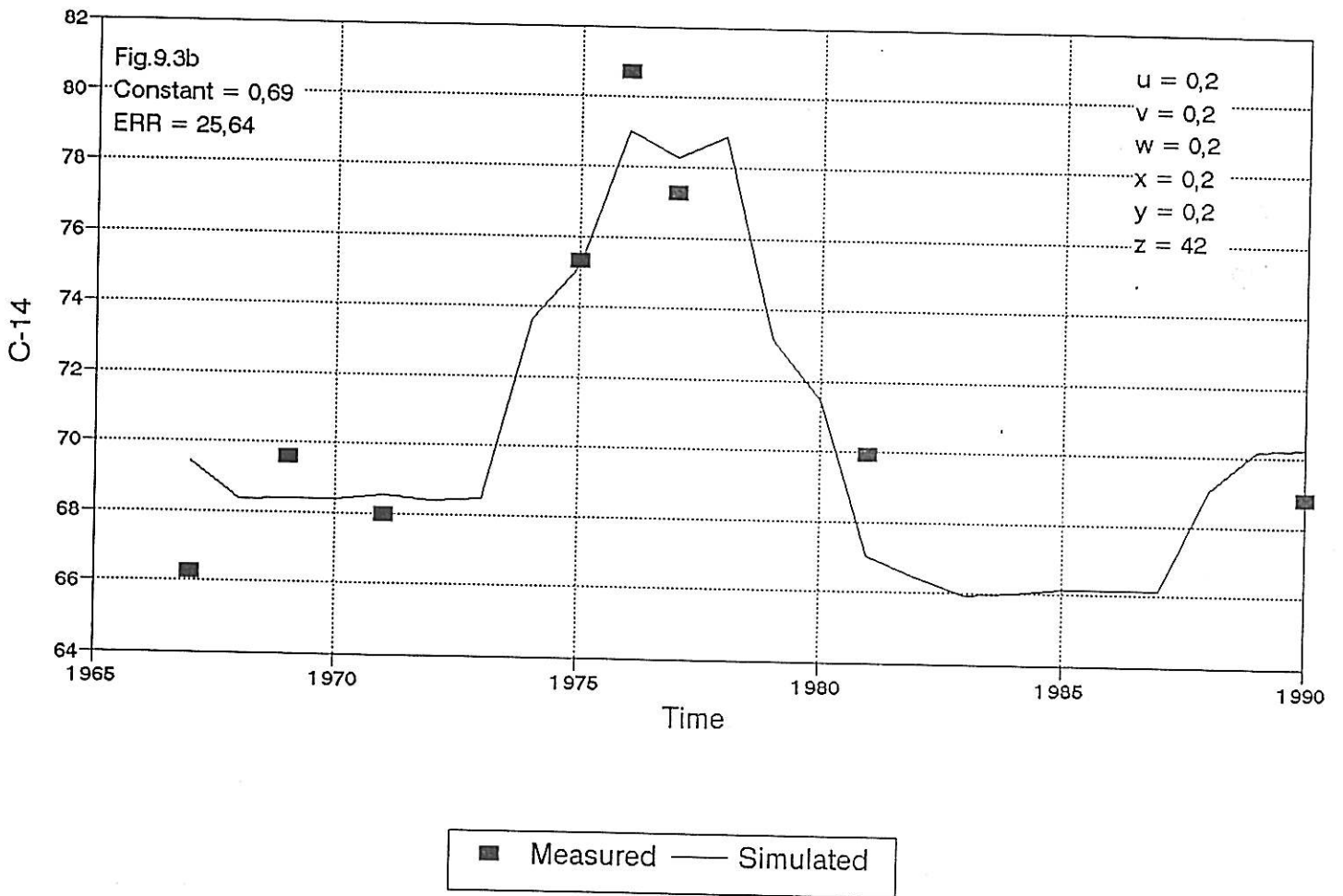


Figure 9.3(b) Simulated C^{14} values based on recharge eq. 4 - excess to average annual rainfall. Equal contribution from preceding years

KURUMAN EYE C-14 VALUES

$RE = a[RF(l) - 310]$

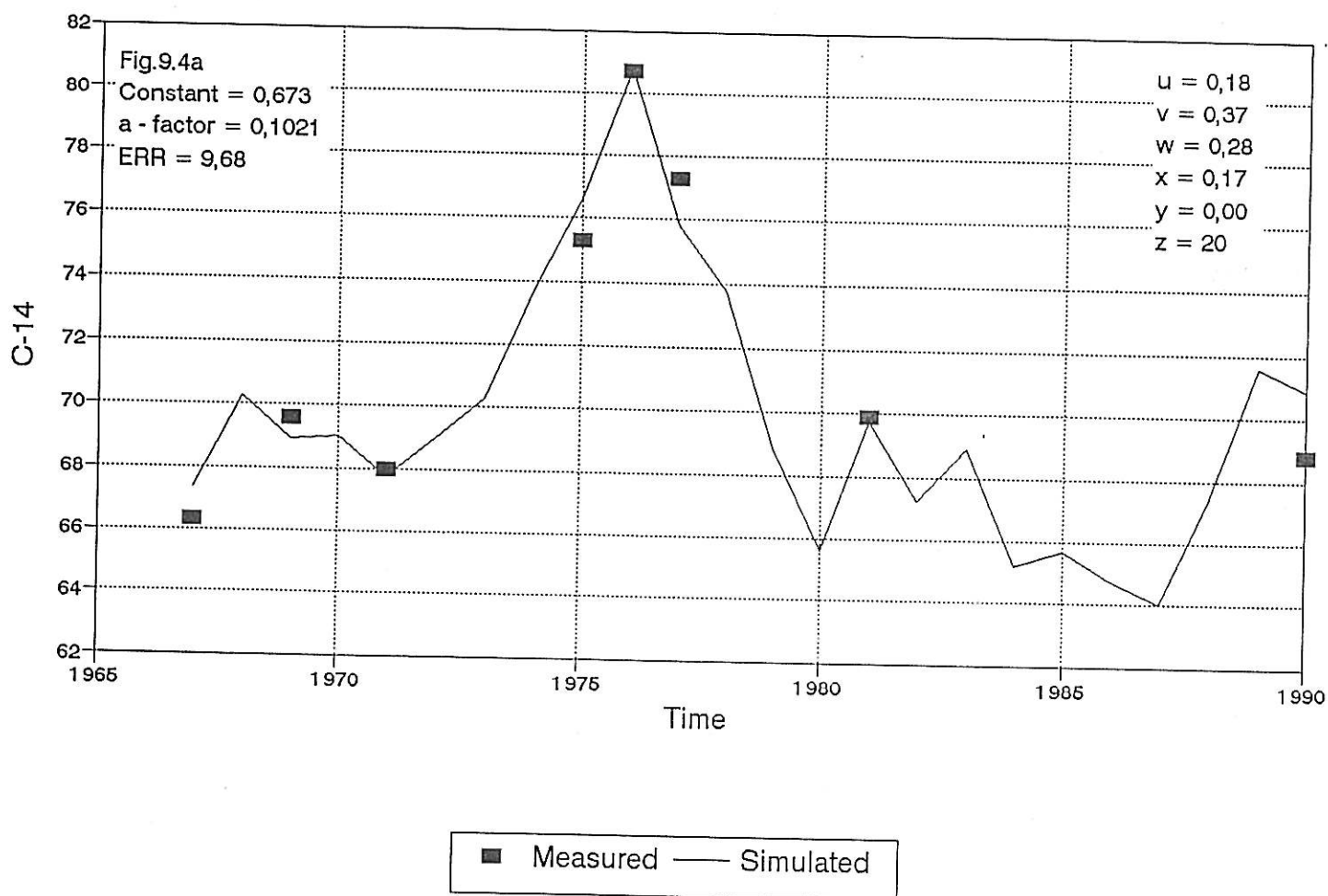


Figure 9.4(a) Simulated C^{14} concentrations for recharge eq. 3 - general equation 310 mm threshold value. Variable contribution from preceding years

KURUMAN EYE C-14 VALUES

$RE = a[RF(I) - 310]$

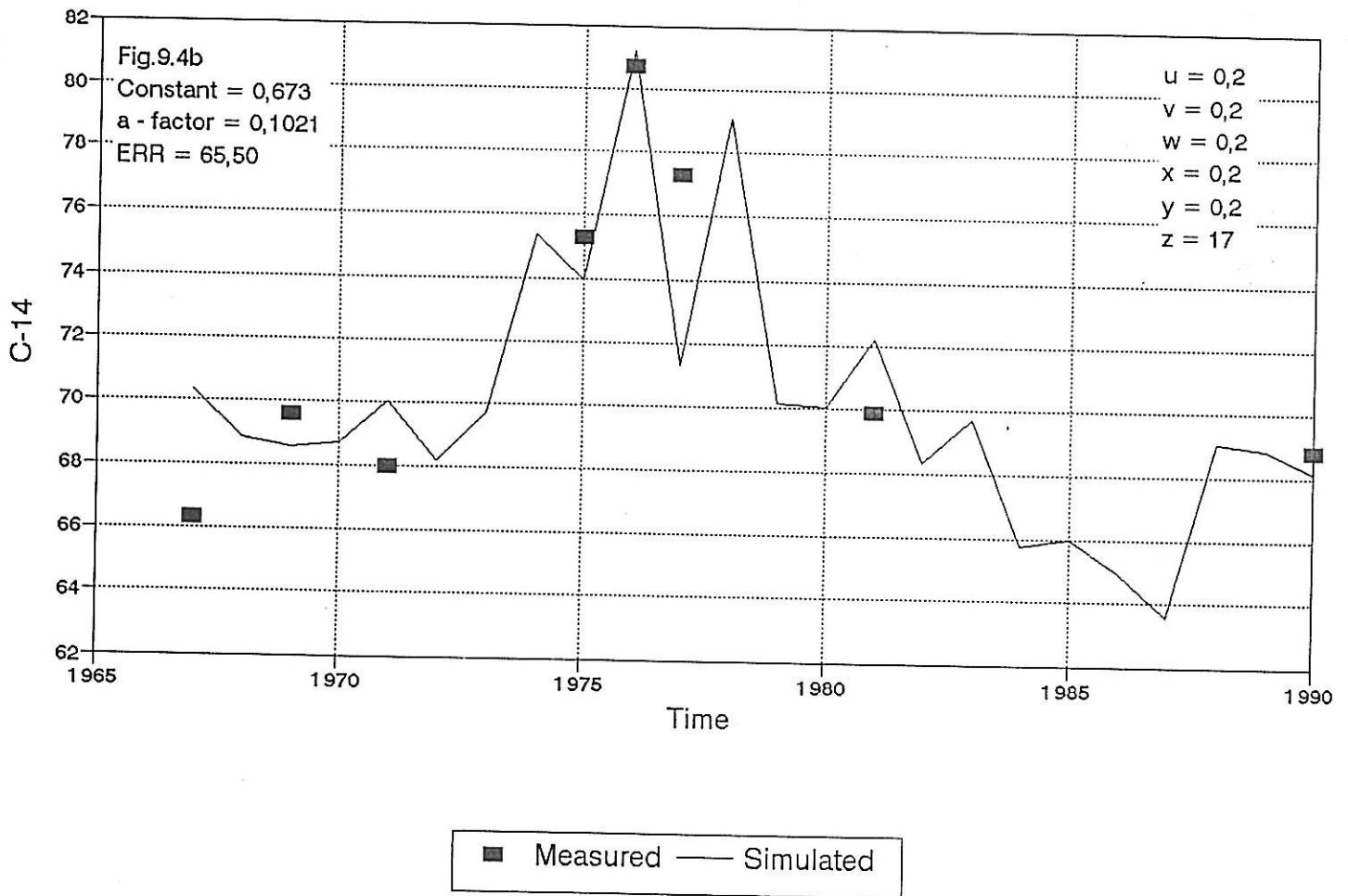


Figure 9.4(b) Simulated C^{14} concentrations for recharge eq. 3 - general equation 310 mm threshold value. Equal contribution from preceding years

KURUMAN EYE C-14 VALUES

$$RE = a \text{ ARF} \{1 - b[\text{ARF}/\text{RF}(t)]\}$$

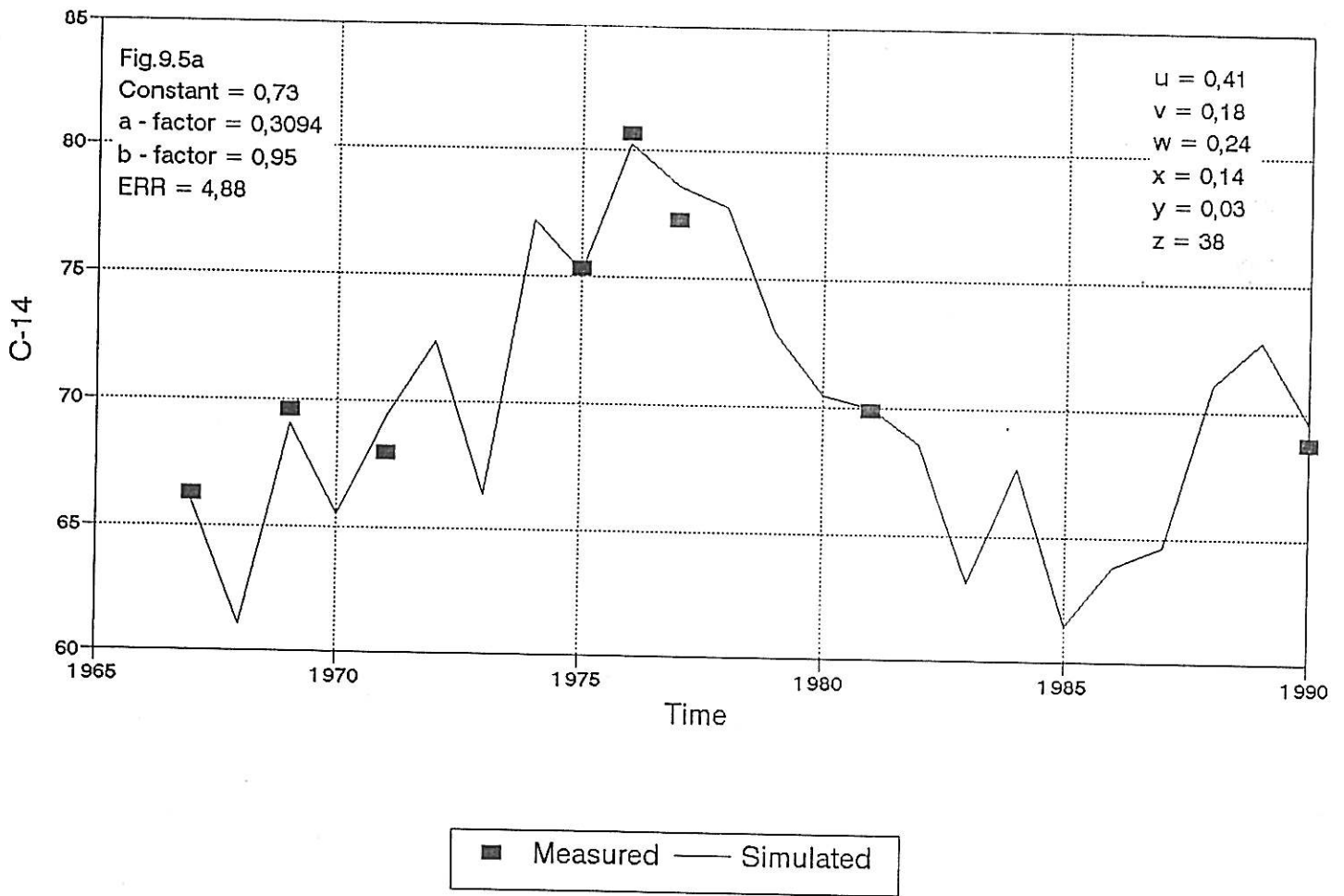


Figure 9.5(a) Simulated C^{14} values based on recharge eq. 5 - decreasing % recharge

KURUMAN EYE C-14 VALUES

$RE = a \text{ ARF} \{1 - b[\text{ARF}/\text{RF}(I)]\}$

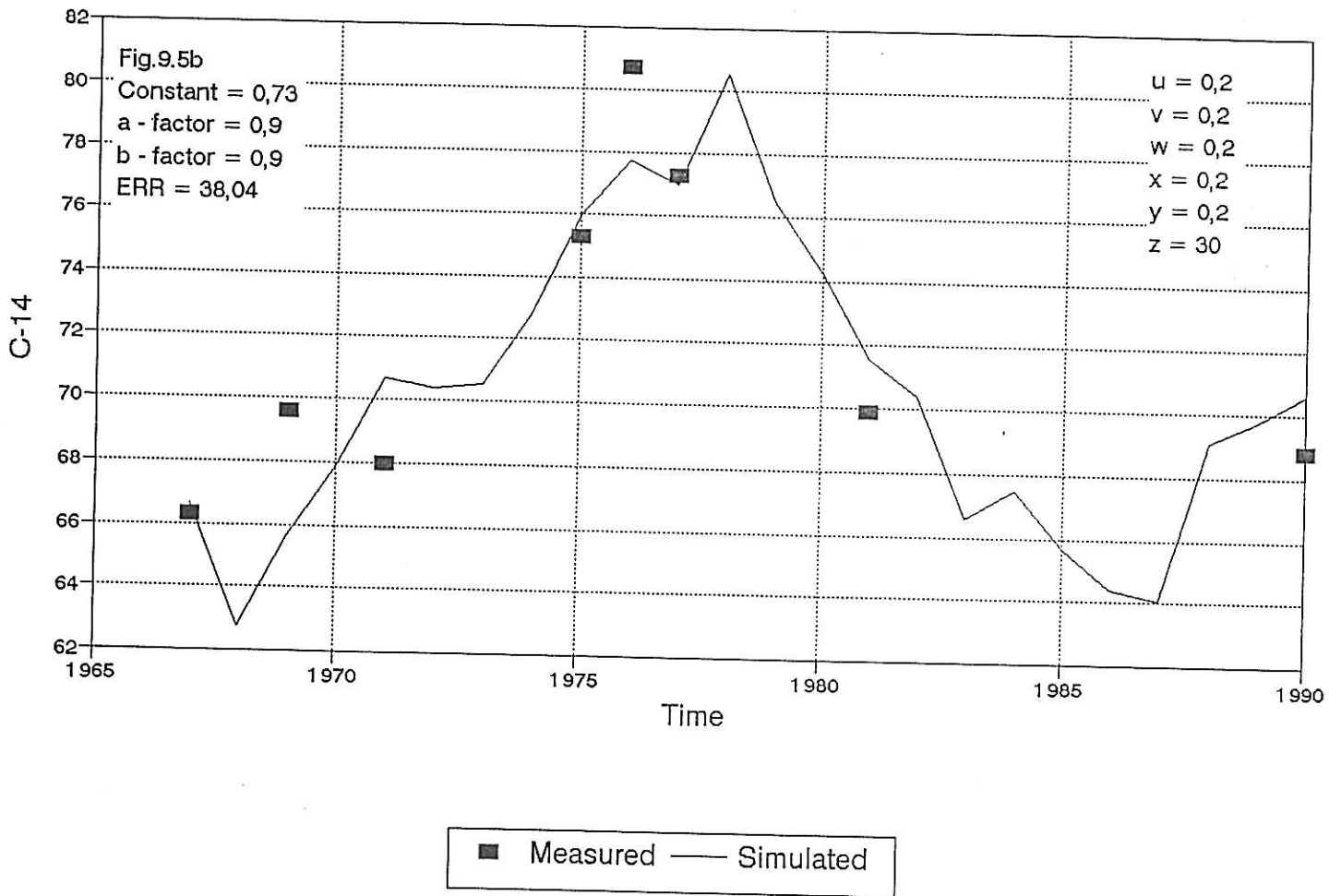


Figure 9.5(b) Simulated C^{14} values based on recharge eq. 5 - decreasing % recharge. Equal contribution from preceding years

Although more than one recharge equation had produced reasonable simulations of the C^{14} concentration of the water emanating from the Kuruman A eye, the following two recharge equations proved to be the best:

(i)

$$RE(I) = a(RF(I) - 310) \dots \text{eq. 9.6}$$

The value of ERR was 9,68 and the relative contribution of recharge from preceding years occurs in the ratio 18%, 37%, 26%, 17% and 0% for years I to I-4 (refer also Figure 9.4a). An aquifer storage of 20 times the average recharge emerged from this simulation.

ii)

$$RE(I) = a \times ARF \left(1 - \frac{(b \times ARF)}{RF(I)} \right) \dots \text{eq. 9.7}$$

This simulation gave the best results (ERR = 4,88) with recharge contributions of 41%, 18%, 24%, 14% and 3% from years I to I-4 (refer Figure 9.5a). Compared to eq. 9.6 the pattern of the relative contributions was much the same for years I-2, I-3 and I-4, but for years I and I-1 the percentages are roughly interchanged. The storage to recharge ratio z proved to be 50 which is much higher than that produced by the previous equations.

The remaining recharge equations, especially eq. 1, which assumes a constant percentage of recharge, were less successful and as is evident from Figure 9.1a the fit was poor (ERR = 121,57).

9.3.3 Tritium concentration simulation

The simulation of the tritium concentrations of the water in the spring was also done according to the different recharge equations, that were used for the C^{14} simulations.

As had been indicated previously the response in the tritium change in the spring water, seemed to lag the C^{14} response (refer Figure 8.3). As is indicated in Figure 9.6, which represents the simulated tritium values if the recharge is assumed equal to the rainfall in excess of the average precipitation, the simulated pulse shows a definite lagged reaction for the first arrival of tritium. This lag in a way corresponds to the first attempt by Bredenkamp to match the C^{14} and tritium results (Bredenkamp 1978), where it was merely indicated that the initial tritium values entering the aquifer had to be about 5 TU to yield a good simulation of both C^{14} and tritium.

In the present attempt different lag periods were tried and surprisingly a lag of about 7 years produced the best simulations. This implies that the tritium values of 7 years back represent the year in which recharge is effected.

This could manifest that the so-called piston-like input of

KURUMAN EYE - SIMULATION OF TRITIUM
 $RE=A(RF-ARF)$ year(i)to(i-4)

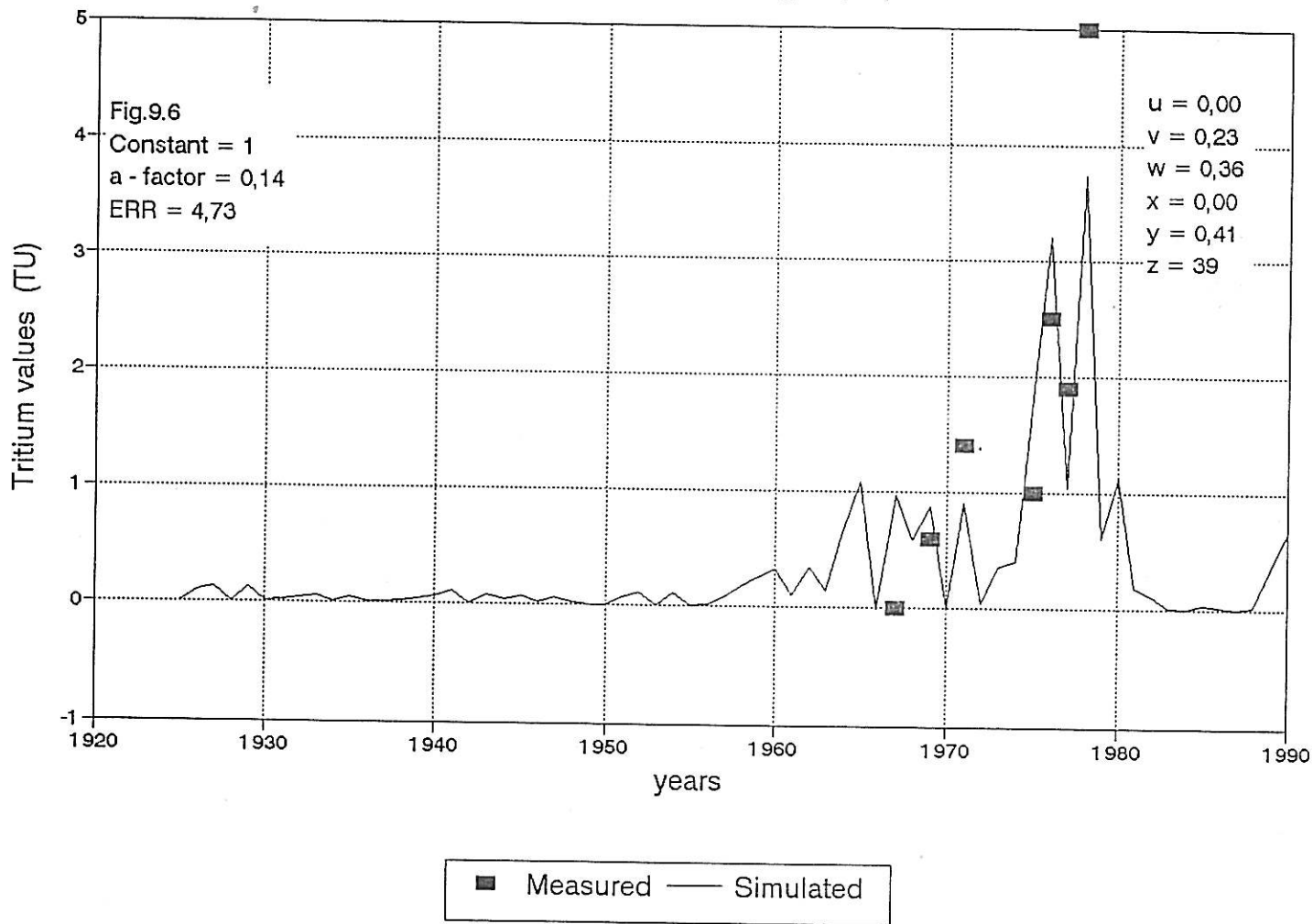


Figure 9.6 Simulated tritium response based on recharge eq. 4 with variable contributions from preceding years

tritium in fact applies (refer Figure 8.5).

Discussion of the results obtained using the different rainfall-recharge relationships.

It must be remembered that the accuracy of the tritium measurements is subject to an uncertainty of between 0,5 to 1 TU because of the low levels of tritium and associated measurement errors.

Equation 9.1 with 7 year lag (refer Figure 9.9.1a).

This recharge equation produced a good simulation (ERR = 2,42) with relative contributions of recharge from the preceding years, as is indicated in Table 9.1a. The storage/recharge ratio of 42 accords with a high volume of aquifer storage in relation to the recharge, much the same as that obtained in the C¹⁴ simulations.

The second equation ie.

$$RE(I) = \frac{a \times RF(I) \times RF(I)}{ARF} \dots \text{eq. 9.2}$$

produced the best results (ERR = 0,90, refer Figure 9.9.2), but with a non-uniform contribution of recharge from the different years preceding a specific year. The recharge contribution was 15%, 0%, 9%, 17% and 59% from the years I to I-4. The storage to recharge ratio of $z = 22$ is in close agreement to the earlier results of Bredenkamp (1978) which had indicated a z value of about 20, but which was about half the value yielded by the good

KURUMAN EYE - SIMULATION OF TRITIUM

$RE = A \cdot RF(l)$ (7 year lag)

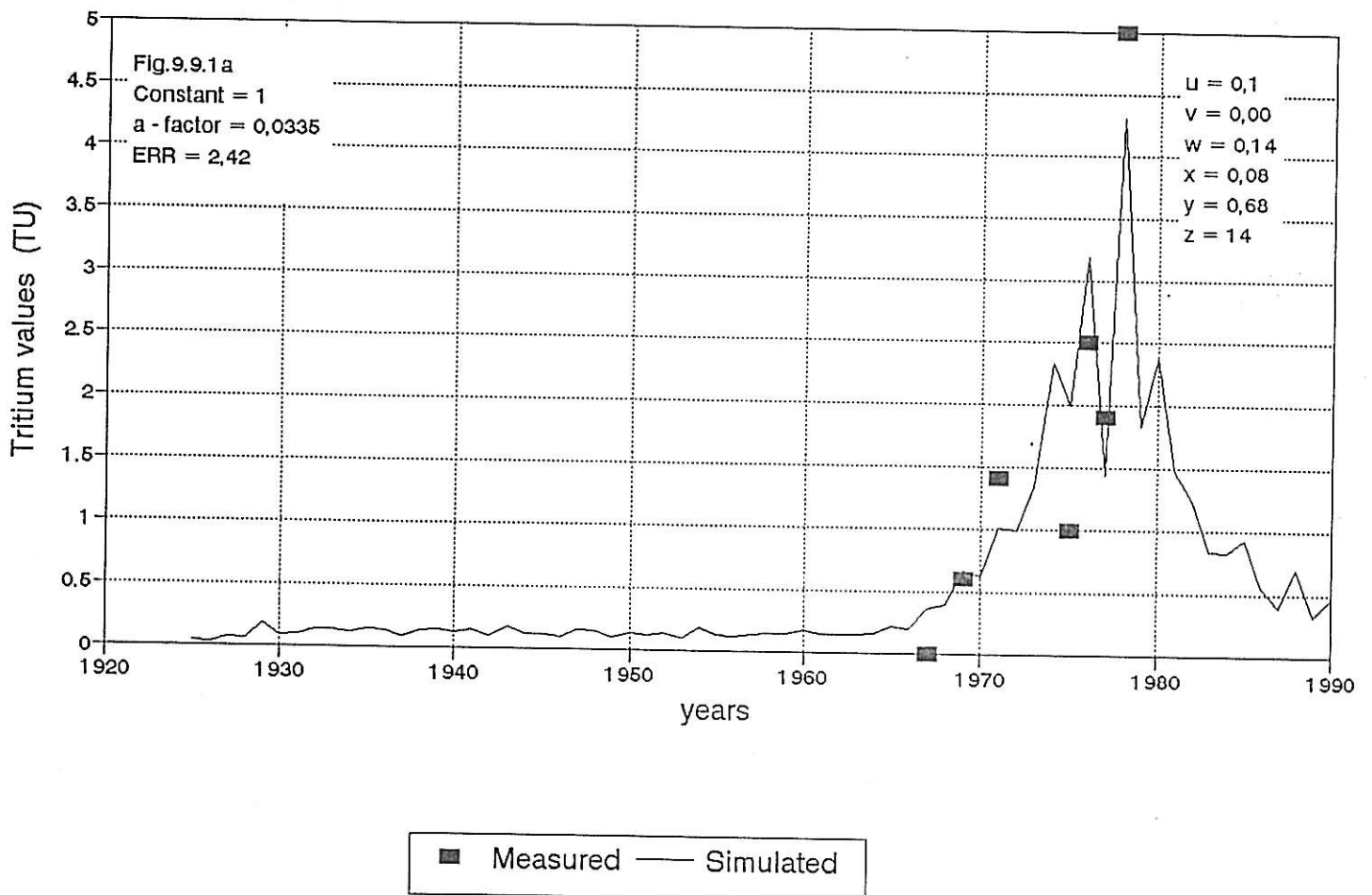


Figure 9.9.1a Simulated tritium values based on recharge eq. 1 (constant % recharge) with variable input from preceding years (7 year lag)

KURUMAN EYE - SIMULATION OF TRITIUM
 $RE = A * RF(l)$ (7 year lag)

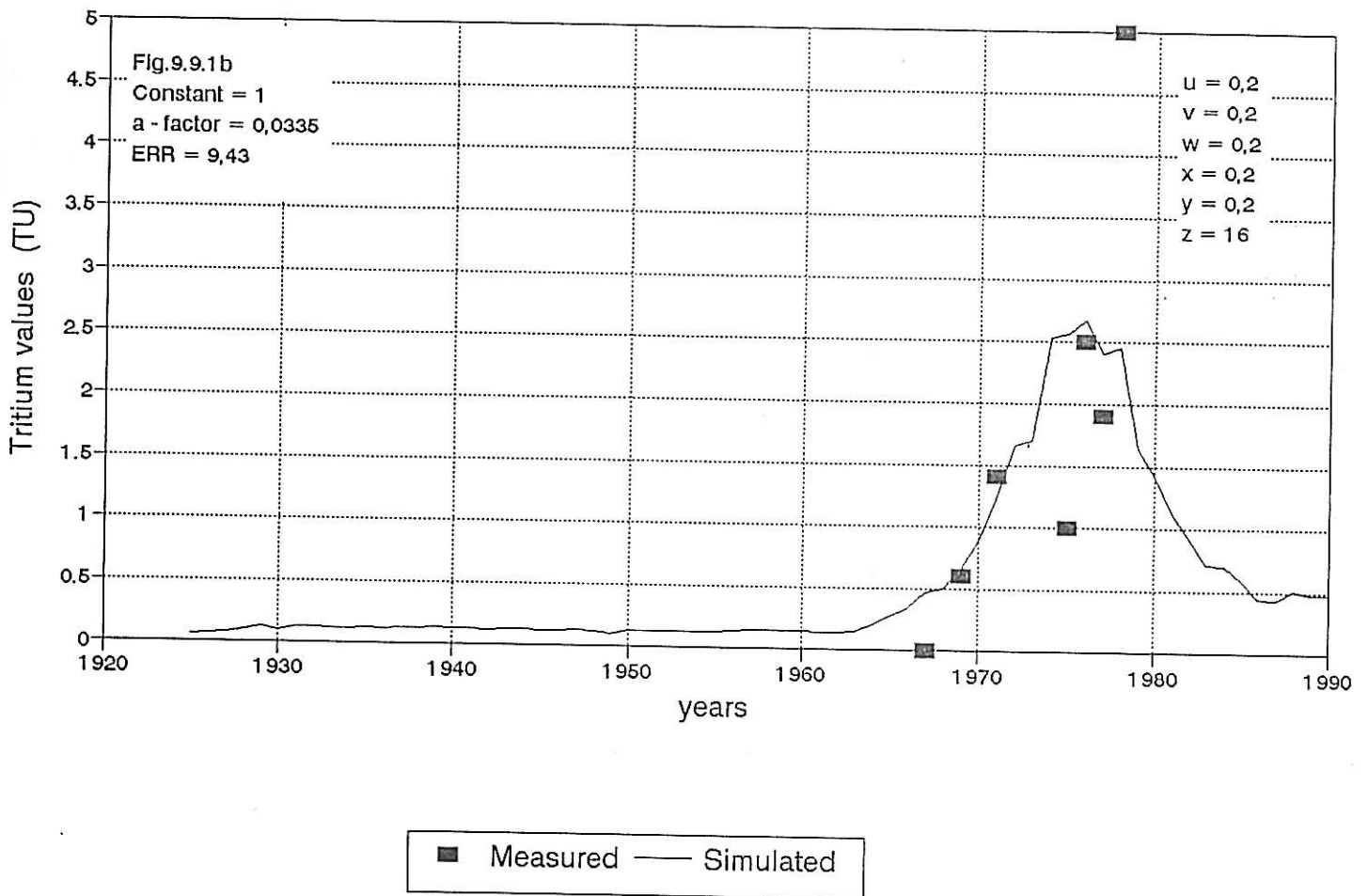


Figure 9.9.1b Simulated tritium values based on recharge eq. 1 but with equal contribution of recharge from preceding years (7 year lag)

KURUMAN EYE - SIMULATION OF TRITIUM

$RE = A \cdot RF(I) / ARF \cdot RF(I)$ (7 year lag)

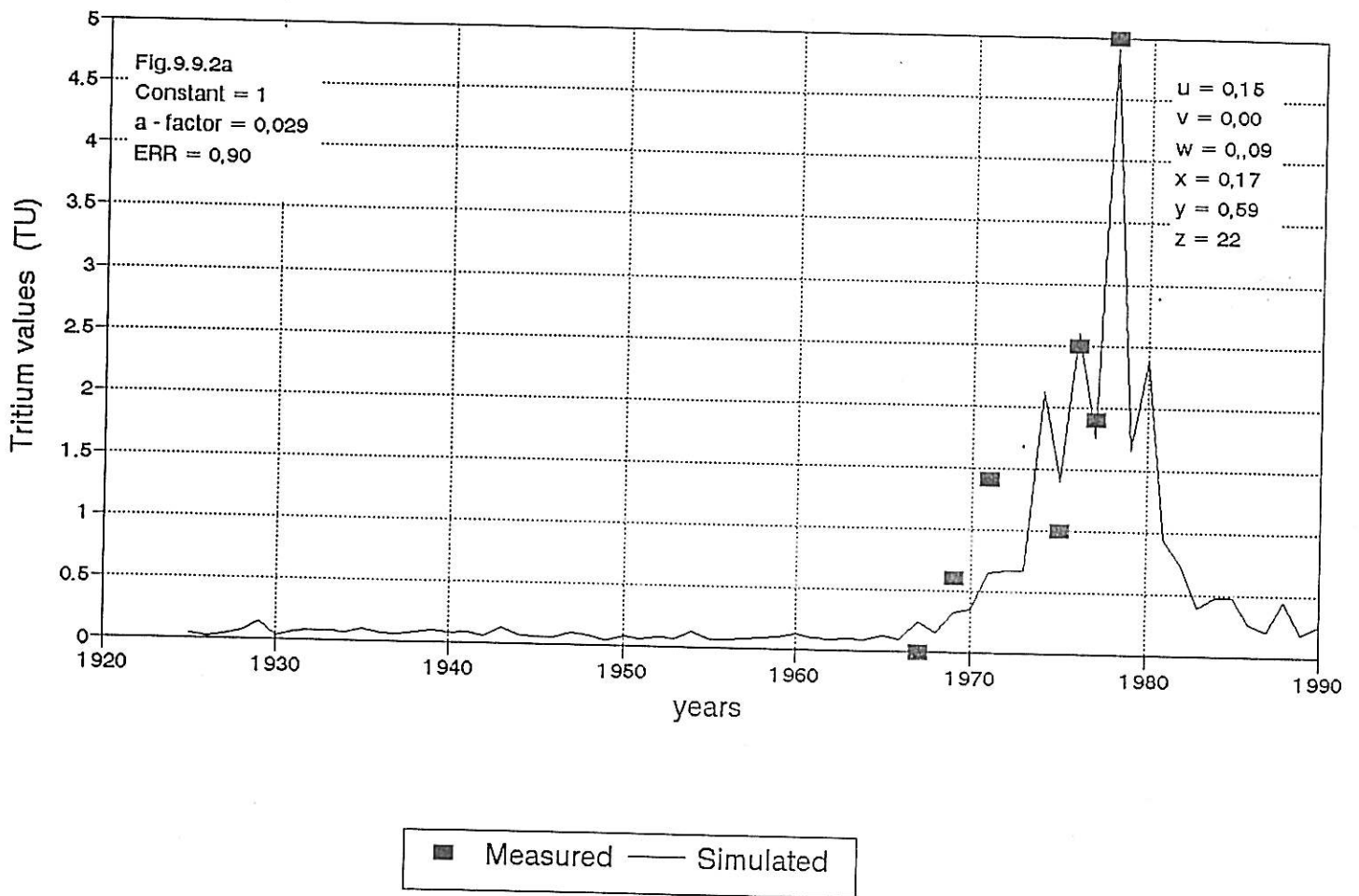


Figure 9.9.2a Simulated tritium values based on recharge eq. 2 but with variable contribution of recharge from preceding years (7 year lag)

KURUMAN EYE - SIMULATION OF TRITIUM
 $RE = A \cdot RF(l) / ARF \cdot RF(l)$ (7 year lag)

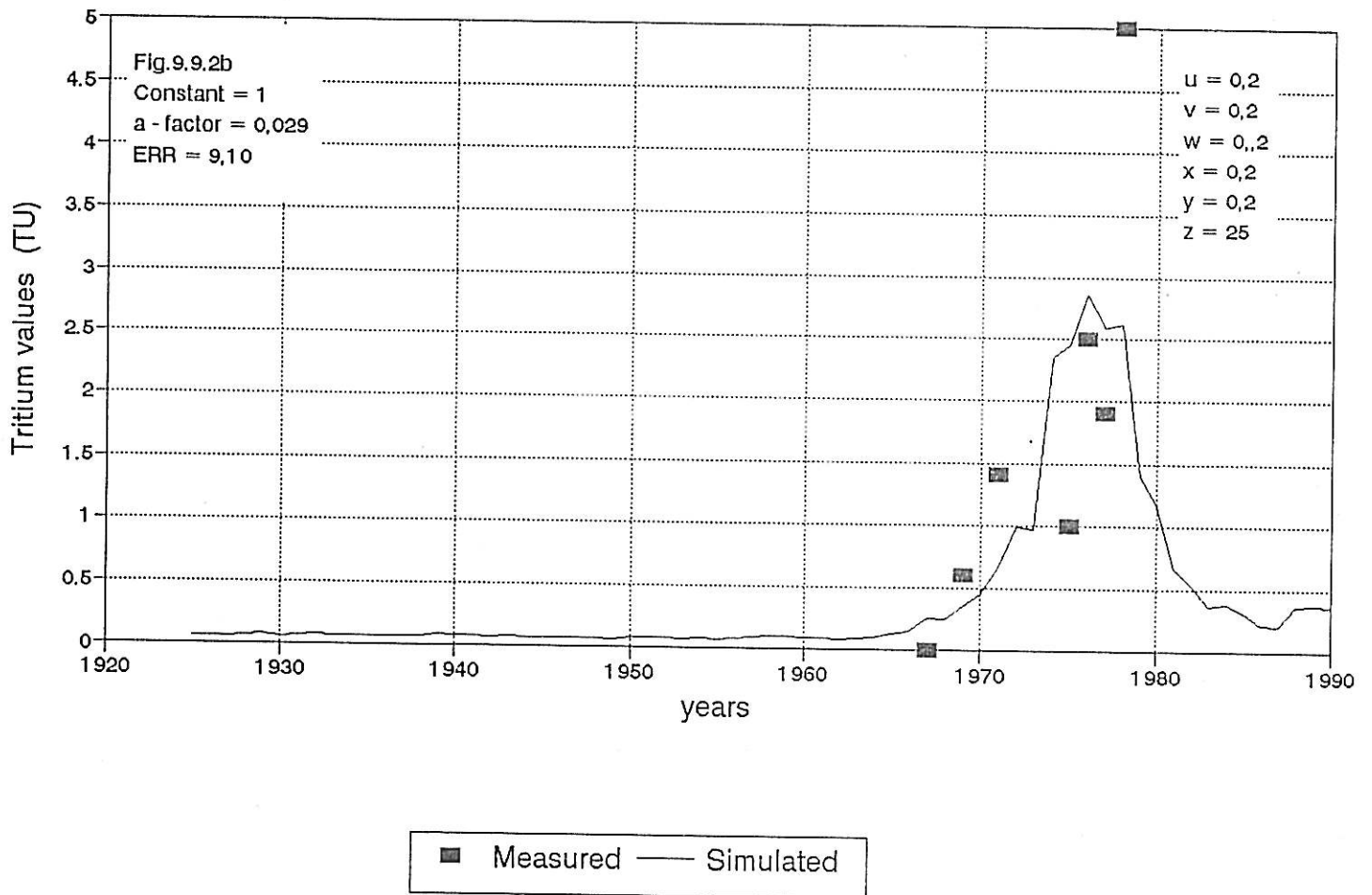


Figure 9.9.2b Simulated tritium values based on recharge eq. 2 but with equal contribution of recharge from preceding years (7 year lag)

C^{14} simulations.

Third recharge equation :

$$RE(I) = A(RF(I) - ARF) \dots \text{eq. 9.3}$$

This recharge equation corresponds to the cumulative rainfall departure from the mean rainfall, the only difference being that A allows for the right quantitative values for the annual recharge to be obtained, which give the required average value for the aquifer recharge.

The simulation that was effected by using this recharge formula produced a good fit (refer Figure 9.9.3), again using the tritium values of rainfall with a 7 year lag (ERR = 2,38). The relative recharge contribution from the preceding years were 15%, 11%, 11%, 5% and 58% from the years I to I-4, which is fairly close to the pattern of recharge which had been obtained with Eq. 9.2.

In the case of equation 9.4 ie.

$$RE(I) = A(RF(I) - B \times ARF) \dots \text{eq. 9.4}$$

Here again $B \times ARF = 310$ was used as the cutoff value below which no recharge would take place.

As was the case with the C^{14} results which had been obtained using this formula, the simulation of the tritium values was very good (ERR = 1,18, refer Figure 9.9.4a). The relative recharge contribution from the preceding years were confined to years I-2,

KURUMAN EYE - SIMULATION OF TRITIUM
 $RE = A(RF[I]-ARF)$ (7 year lag)

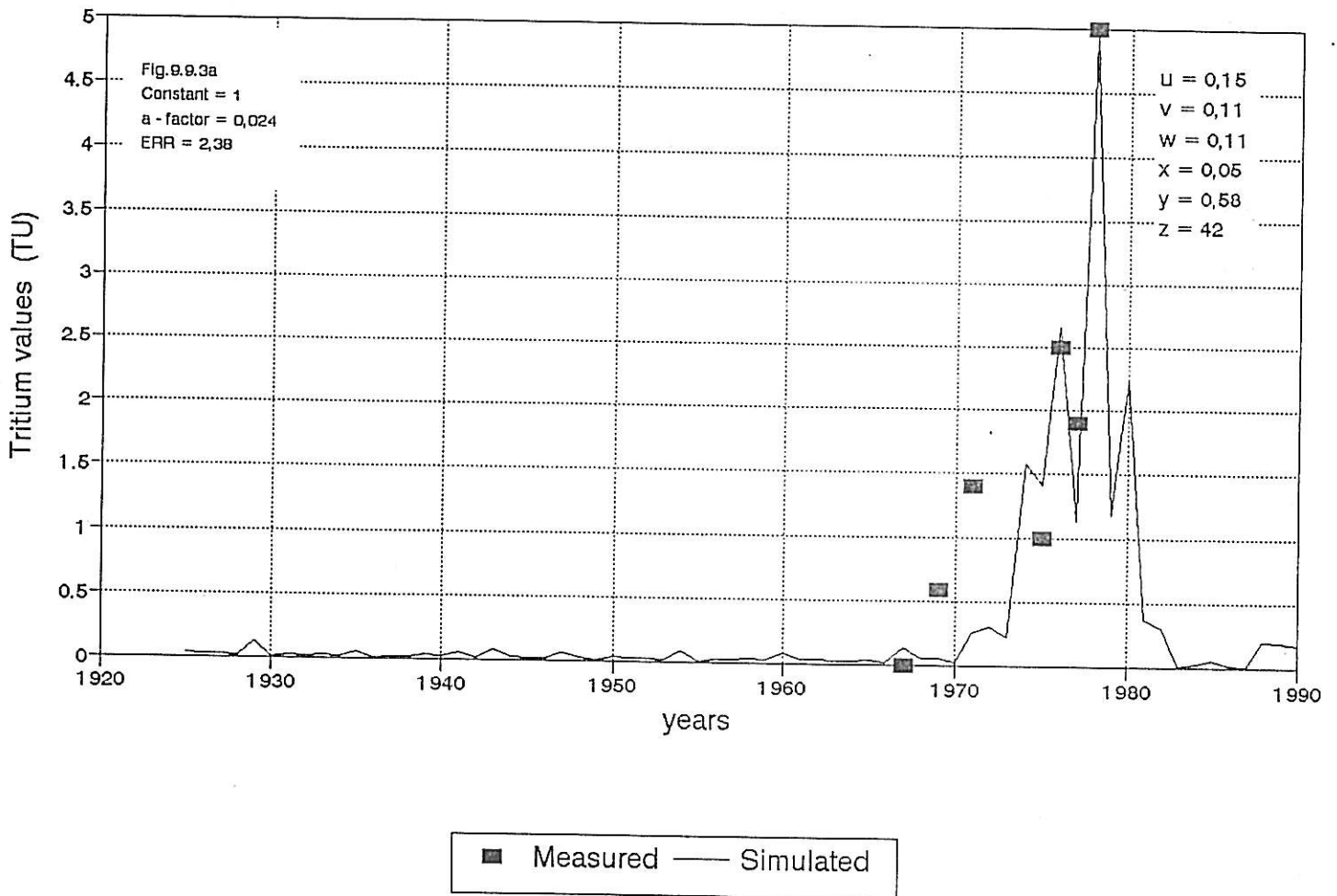


Figure 9.9.3a Simulated tritium values based on recharge eq. 3 but with variable contribution of recharge from preceding years (7 year lag)

KURUMAN EYE - SIMULATION OF TRITIUM
 $RE = A(RF[I-7]-GRF)$

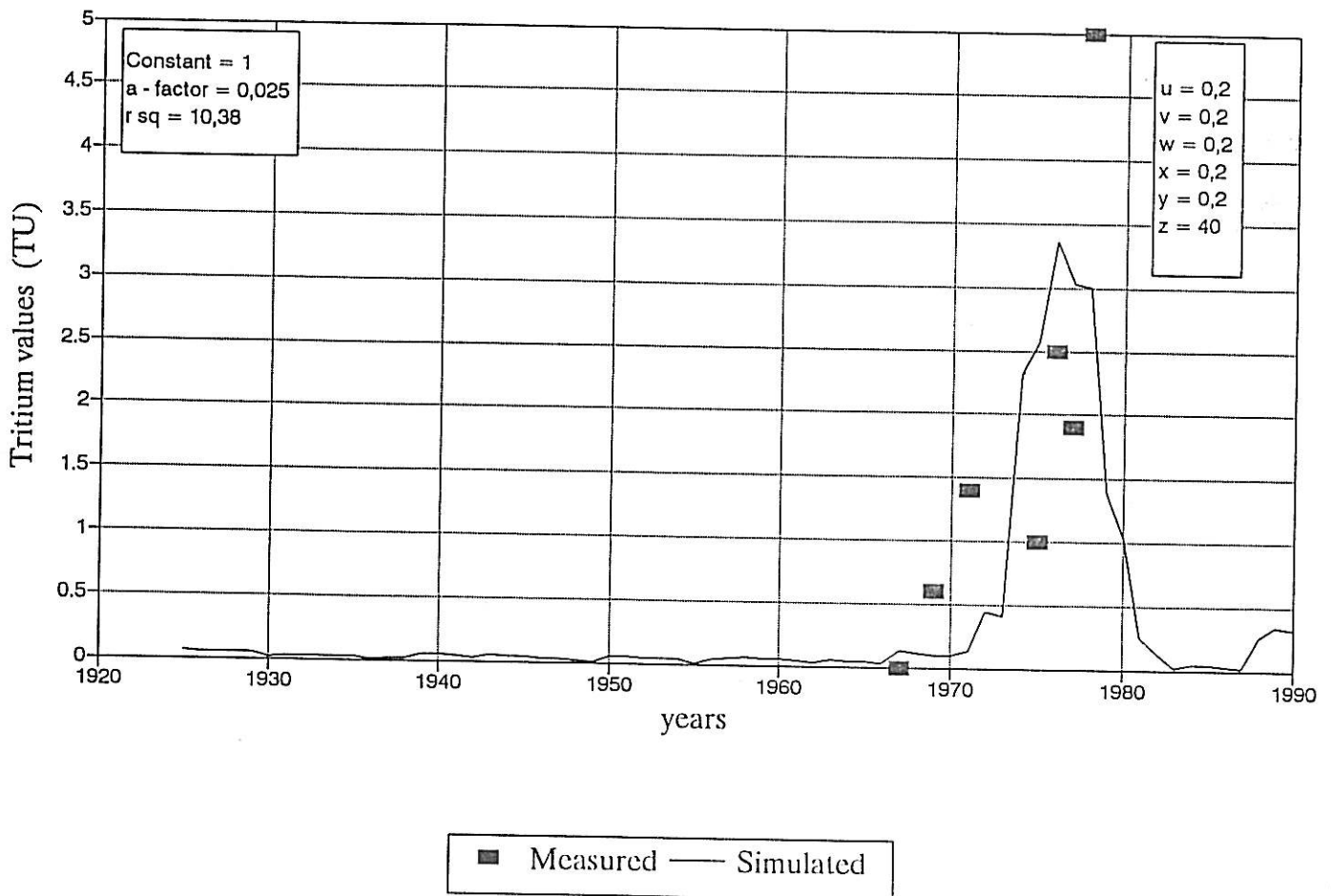


Figure 9.9.3b Simulated tritium values based on recharge eq. 3 but with equal contribution of recharge from preceding years (7 year lag)

KURUMAN EYE -SIMULATION OF TRITIUM
 $RE = A(RF[I]-310)$ (7 year lag)

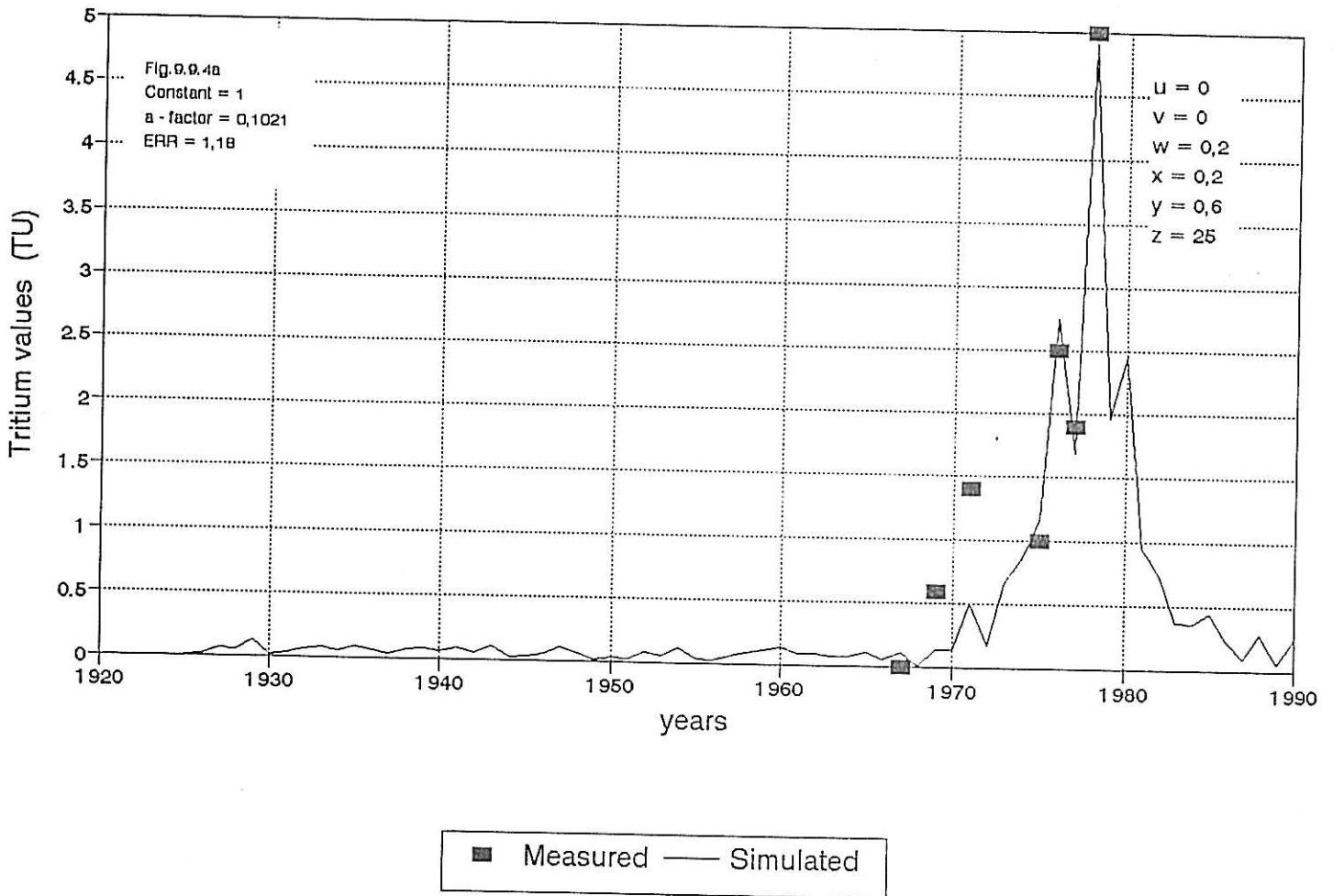


Figure 9.9.4a Simulated tritium values based on recharge eq. 4 but with variable contribution of recharge from preceding years (7 year lag)

KURUMAN EYE -SIMULATION OF TRITIUM

$RE = A(RF[I]-310)$ (7 year lag)

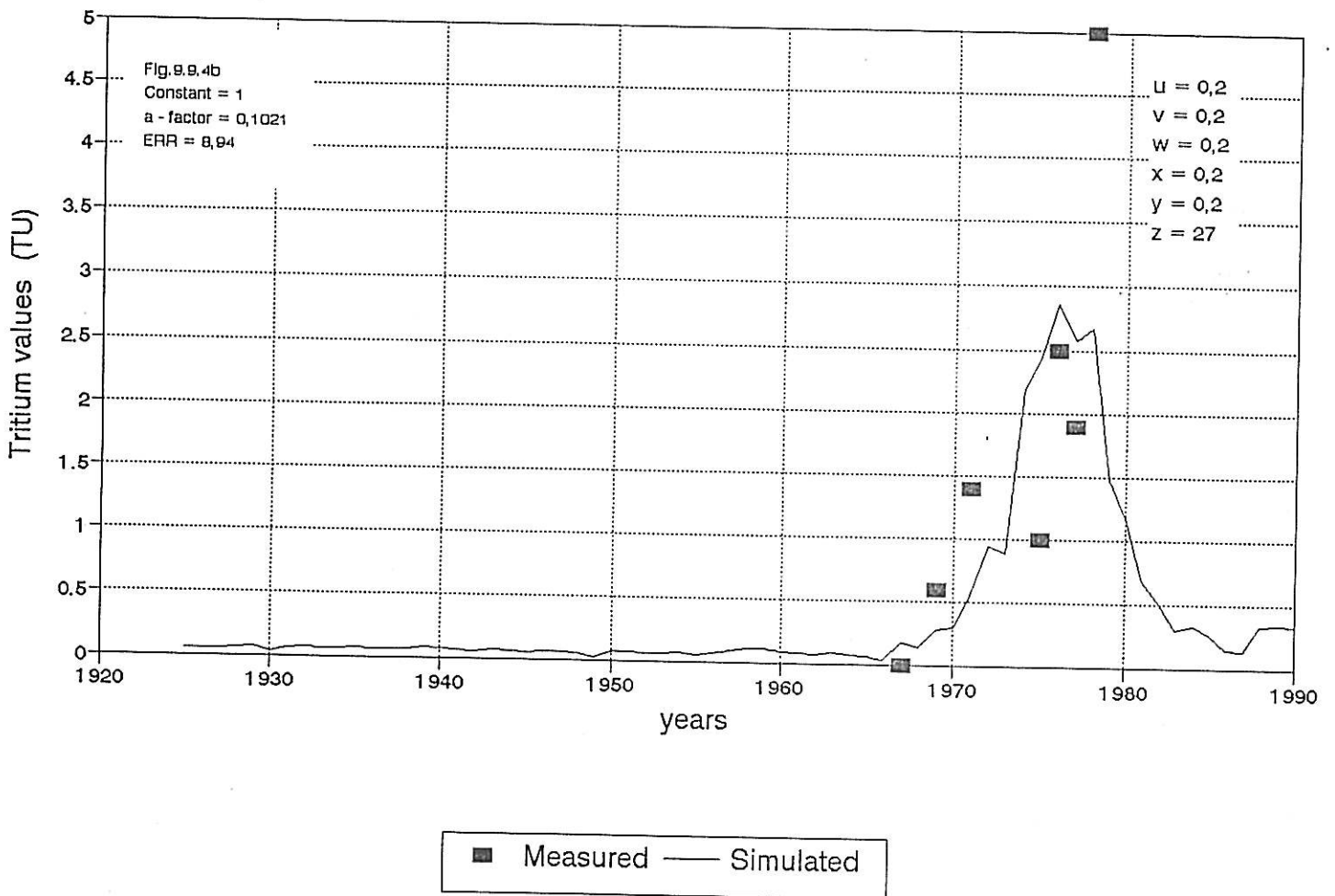


Figure 9.9.4b Simulated tritium values based on recharge eq. 4 but with equal contribution of recharge from preceding years (7 year lag)

I-3 and I-4 in the following proportions of 20%, 20% and 60% which again show the predominance of the year I-4 in the case of the tritium simulations. The storage to recharge ratio was $z = 25$ which is about the range of values that were found for the other simulations.

$$RE(I) = A \times ARF \left(1 - \frac{B \times ARF}{RF(I)} \right) \dots \text{eq. 9.5}$$

This equation differs from eq. 9.1 in that the weighting of factor A was inversely related to the ratio of RF(I) to ARF. The fit was excellent (ERR = 2,10, refer Figure 9.9.5a) and the relative recharge contributions were 0%, 0%, 28%, 19% and 53% for the respective years I to I-4. The storage to recharge ratio in this case was 20 which is close to the predominant value for the tritium simulations.

9.4 Reconciliation of C^{14} and Tritium simulations

The validity of assuming that there is a lag of several years between the recharge and the tritium value of the rainfall, could be queried because no tritium values of the spring water prior to 1967 were available. The correspondence that could be achieved without incorporating the lag in all of the recharge equations, was rather poor.

A possible explanation for the lag of the tritium values could also be that the tritium gets retarded due to absorption on soil and clay particles. Evidence to this effect had been reported

KURUMAN EYE -SIMULATION OF TRITIUM

$RE = A*ARF(1-B*ARF/RF[I])$ (7 year lag)

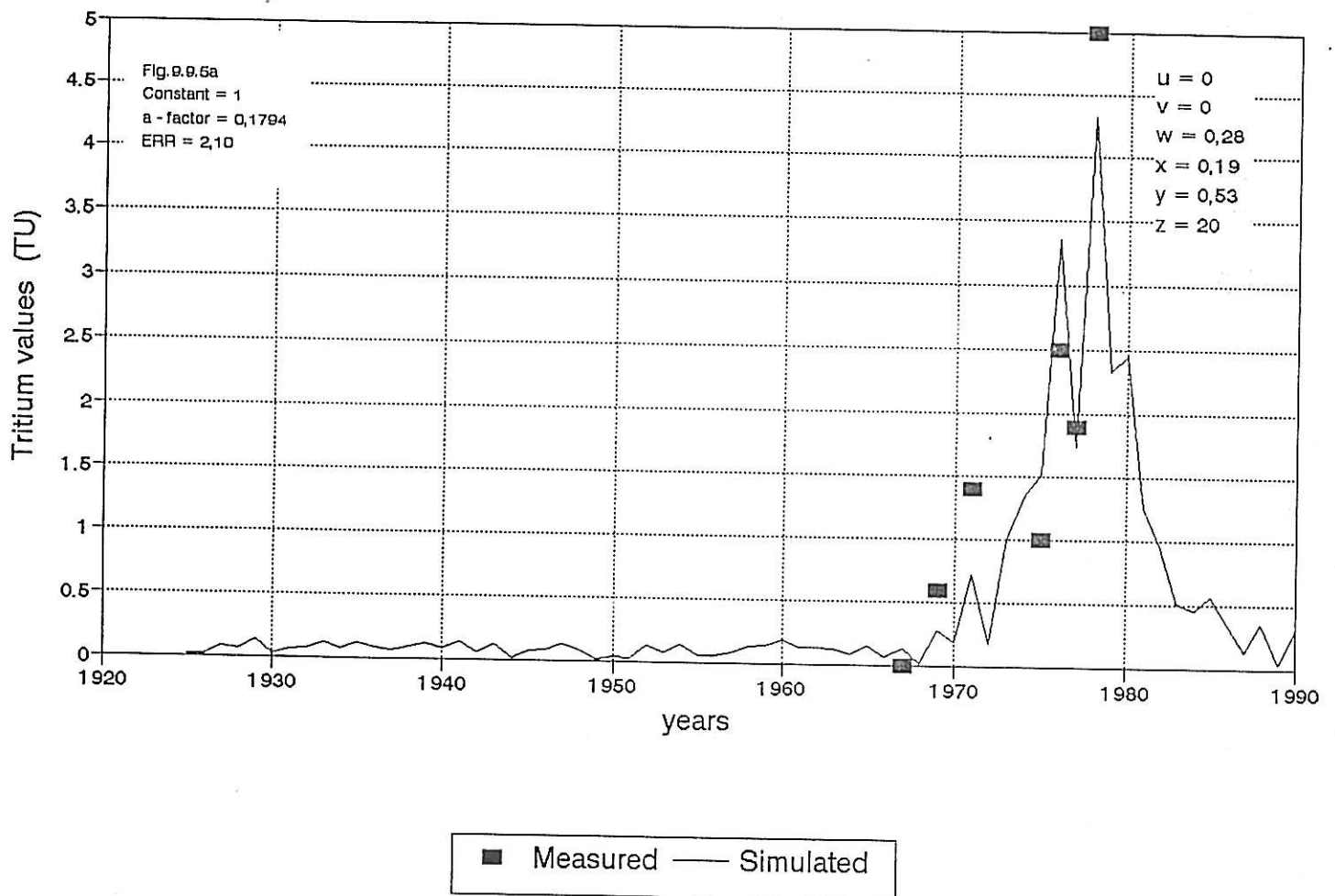


Figure 9.9.5a Simulated tritium values based on recharge eq. 5 but with variable contribution of recharge from preceding years (7 year lag)

KURUMAN EYE - SIMULATION OF TRITIUM

$RE = A \cdot ARF(1 - B \cdot ARF / RF[I])$ (7 year lag)

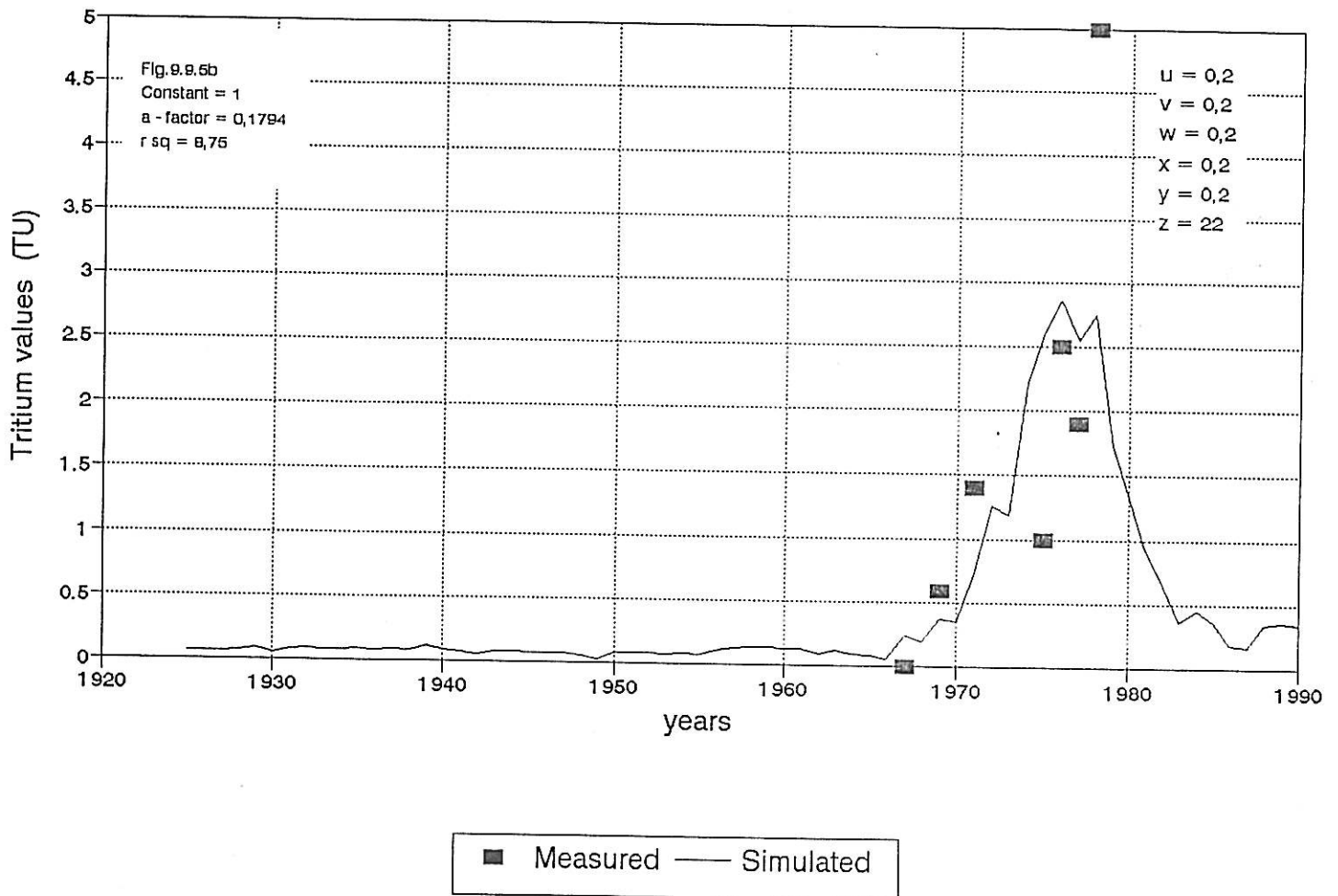


Figure 9.9.5b Simulated tritium values based on recharge eq. 5 but with equal contribution of recharge from preceding years (7 year lag)

in the literature.

The variable recharge contributions from the preceding years (u, v, w, x, y) for the different recharge equations made it difficult to compare the performance of the different equations. For this reason a constant contribution of recharge (u, v, w, x, y all equal to 0.2) was assumed and the rest of the parameters were adjusted until the best fit was obtained. These results are listed in Table 9.1b - 9.2b which allow the following interpretations - (refer also Figures 9.9.1b to 9.9.5b):

Table 9.1a: Comparison of the simulated C^{14} results using different recharge formulae

Recharge model	u	v	w	x	y	z	ERR
Eq. 1	0	0,4	0,23	0,19	0,18	16	121,75
Eq. 2	0,22	0,17	0,21	0,24	0,16	16	30,1
Eq. 3	0,36	0,18	0,23	0,23	0	50	21,34
Eq. 4	0,18	0,37	0,26	0,17	0	20	9,68
Eq. 5	0,41	0,18	0,24	0,14	0,03	38	4,88

Table 9.1b: Results of C^{14} simulations with equal contributions of recharge from the preceding years.

Recharge model	u (I)	v (I-1)	w (I-2)	x (I-3)	y (I-4)	z	ERR
Eq. 1	0,2	0,2	0,2	0,2	0,2	13	134,7
Eq. 2	0,2	0,2	0,2	0,2	0,2	16	32,52
Eq. 3	0,2	0,2	0,2	0,2	0,2	42	25,64
Eq. 4	0,2	0,2	0,2	0,2	0,2	17	65,50
Eq. 5	0,2	0,2	0,2	0,2	0,2	30	38,04
Average						23	

Table 9.2a: Comparison of the simulated tritium results using different recharge formulae

Recharge model	u	v	w	x	y	z	ERR
Eq. 1	0,1	0	0,14	0,08	0,68	14	2,42
Eq. 2	0,15	0	0,09	0,17	0,59	22	0,90
Eq. 3	0,15	0,11	0,11	0,05	0,58	42	2,38
Eq. 4	0	0	0,2	0,2	0,6	25	1,18
Eq. 5	0	0	0,28	0,19	0,53	20	2,10

In the case of the C^{14} values the contributions from preceding years were also more or less equal (refer Table 9.1a).

Table 9.2b: Summary of the best tritium simulations obtained with equal recharge contributions from the different years (7 year lag)

Recharge model	u (I)	v (I-1)	w (I-2)	x (I-3)	y (I-4)	z	ERR
Eq. 1	0,2	0,2	0,2	0,2	0,2	16	9,43
Eq. 2	0,2	0,2	0,2	0,2	0,2	25	9,10
Eq. 3	0,2	0,2	0,2	0,2	0,2	40	10,38
Eq. 4	0,2	0,2	0,2	0,2	0,2	27	8,94
Eq. 5	0,2	0,2	0,2	0,2	0,2	22	8,75
Average						26	

The average values for factor 0,2, the ratio of the aquifer storage to the average recharge is very much the same i.e. being 26 for tritium and 23 for the C^{14} simulations with equal recharge contributions.

In the case of the best simulation for a non uniform recharge contribution the respective values for z are listed in Table 9.3.

Table 9.3: Comparison of C^{14} and tritium simulations with variable contributions of recharge from preceding years

Recharge model	Tritium z	ERR	C^{14} z	ERR
Eq. 1	14	2,42	16	121,5
Eq. 2	22	0,90	16	30,1
Eq. 3	42	2,38	50	21,34
Eq. 4	25	1,18	20	9,68
Eq. 5	20	2,10	38	4,88
Average	24		28	

The general pattern is that the z value seems to vary depending on the recharge equation that is used. On average about the same z value is obtained for tritium and C^{14} . For the best simulations with non uniform recharge from the respective years, the average z value is 26, whilst it is 24,5 if equal recharge contributions were assumed. The average value of 25 will later on be used to obtain the storativity of the aquifer.

10. DETERMINATION OF THE AQUIFER STORATIVITY

10.1 General

The aquifer storativity S is also one of the most difficult parameters to be quantified, especially in the case of a dolomite aquifer, such as the Kuruman groundwater system. Pumping tests

have proved to be unreliable for determining S in a heterogenous aquifer because of the fractured and cavernous nature of the dolomite.

A water balance approach is regarded to be the most suitable method to obtain storativity and allows a lumped estimate to be obtained and not only a local value, as a pumping test would provide.

In the case of the Kuruman eye compartment the saturated volume fluctuation method (SVF) was applied but only after some of the essential data were inferred from the temporal sets of water table measurements. The second method was to determine S by means of the geothermal gradient of the aquifer and the ratio of storage of the aquifer to the average annual recharge.

10.2 Saturated Volume Fluctuation Method

10.2.1 Regional drainage of the groundwater

Three sets of water level measurements in boreholes that are fairly evenly scattered over the compartment and representing the piezometry for different years, were available for interpretation. The same boreholes have not always been remeasured during each of the surveys and the first aim was to reconstruct water levels for the boreholes that were missing from any of the sets.

A simple plot of the water levels relative to mean sea level

against the topographical heights of the collar elevations, showed a good linear relationship for all three sets (Figure 3.1). The outliers of which there are several proved to lie in a zone of higher transmissivity which forms a trough of preferential flow, that seems to link up with the Kuruman A eye. It is interesting that the bounding dykes of which some signify a step in the water table, do not feature as outliers in the linear plot. By means of the linear relationship the missing water levels could be inferred. A better method, however is to use a Bayesian estimation (Van Tonder 1990) which infers the missing values by incorporation of the values of the five nearest points to a missing point as well as the relationship to the topography. The correspondence between the linear and the Bayesian interpolations is good, as is depicted in Figure 10.1.

The contours (Figures 3.4) indicate that the Kuruman B and C eyes are probably fed by groundwater that is not part of the zone of preferential flow that feeds the main eye. This also explains the higher C^{14} concentration of the B eye, which implies that the depth of circulation of its water is not as deep as that of the water emerging from the Kuruman A eye.

10.2.2 Temporal variations of the groundwater levels

The temporal fluctuation of the groundwater levels is confined to the three years of groundwater surveyed for 1970, 1986 and 1991. The water level state for each of the years since 1925 could however be derived by means of the saturated aquifer

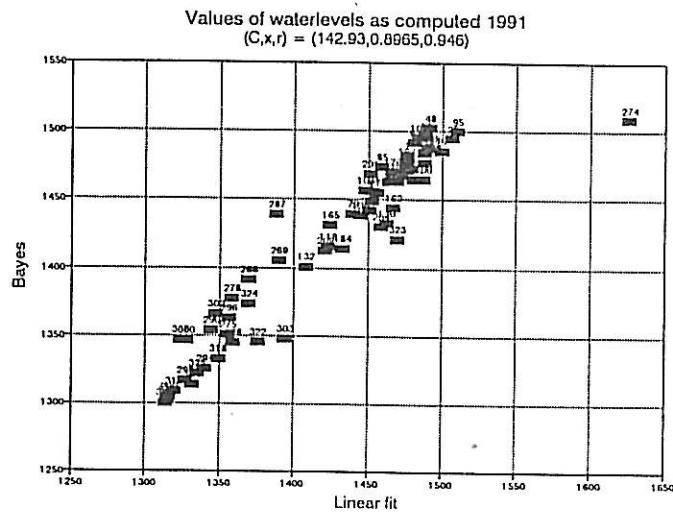
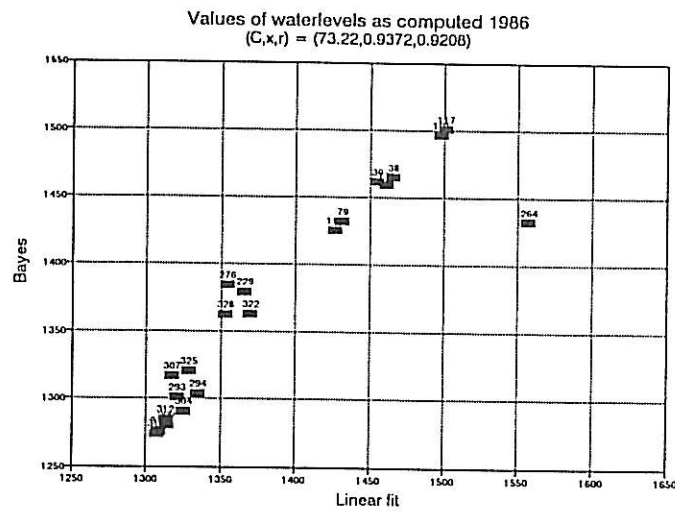
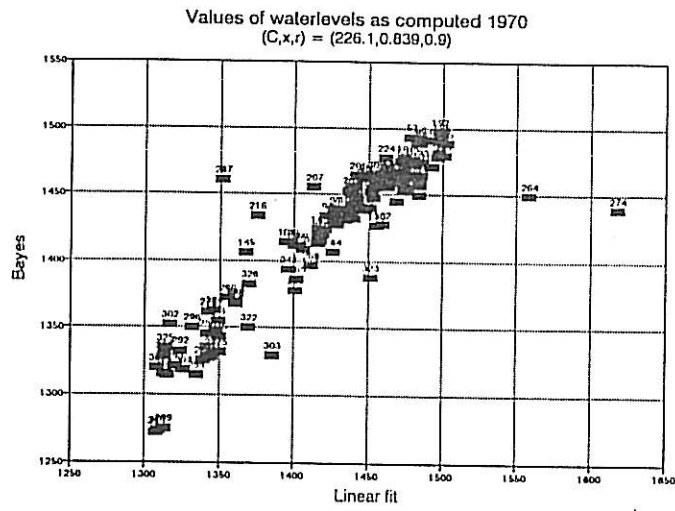


Figure 10.1 Relationship between values obtained by the linear relationship between piezometric levels and topographical height to those inferred from the Bayesian extrapolation

Interpolated SVF values
plotted against catchment yield

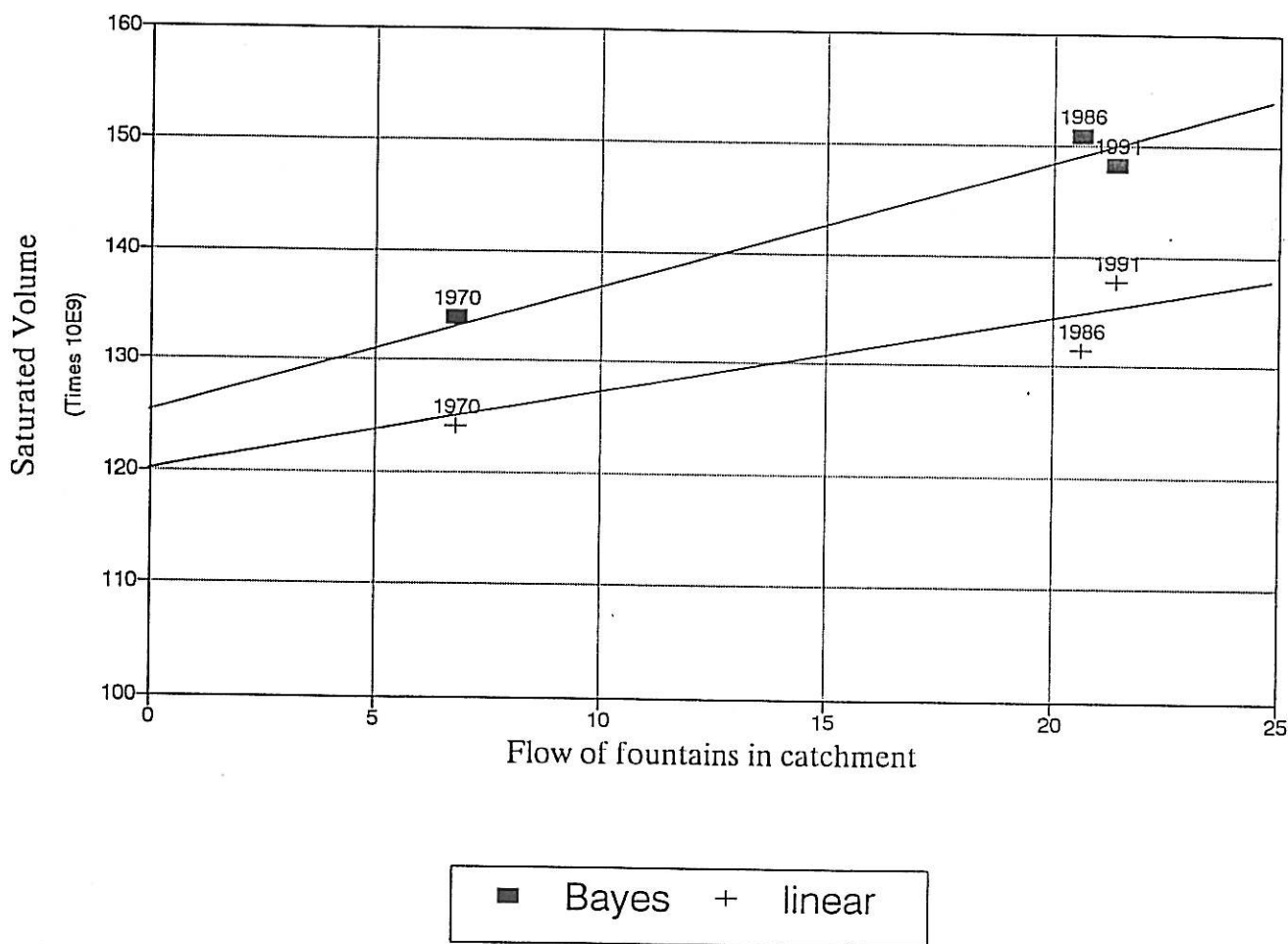


Figure 10.2 Linear relationship between the saturated volumes of water in the aquifer and the combined equivalent flow discharging from the total aquifer

Comparison of the Linear & Bayes method

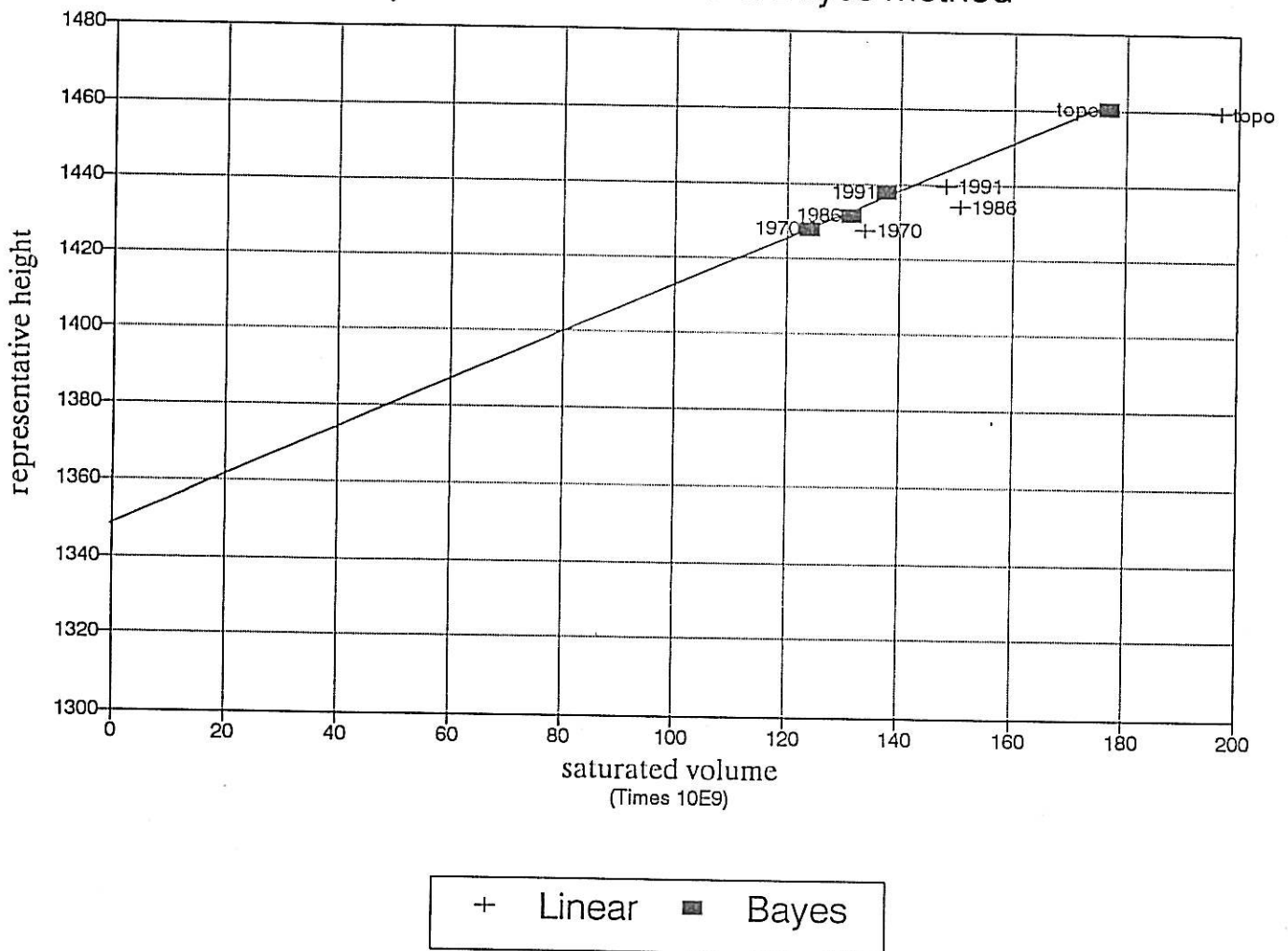


Figure 10.3 Relationship between the saturated aquifer volume (SV) and the representative height of the water in the aquifer relative to mean sea level

volumes in relation to the flow of the springs for the three years mentioned above using some of the relationships that came forth from the studies of springs.

The analysis of the Wondergat levels proved that a linear relationship applies with the flow of several springs in the area. The Wondergat level however merely reflects the temporal variation of the saturated volume fluctuation of the Wondergat aquifer in a simple way, being only a measurement at a single point. Therefore by means of the more accurately determined fluctuations of the aquifer volumes in the respective years 1970, 1986 and 1991, a linear relationship with the flow is also expected and is manifested in Figure 10.2. From this figure it is clear that the spring would stop flowing if the saturated volume reaches a value of $125 \times 10^9 \text{m}^3$, using the Bayes values. Figure 10.3 depicts the relationship between the saturated volume (SV) and the representative height of the water in the aquifer above mean sea level. The representative height is the saturated volume divided by the surface area for each element in the network. The base value for the aquifer was assumed to be 10 metres below the outflow level of 1310 mamsl, although in reality the bottom of the aquifer is about 1000 mamsl. The difference between the 1991 representative height and the zero-flow height represents the total water level decline that would cause the spring to stop flowing. This difference is only about 20 metres which is alarmingly small, considering the small value of the aquifer storativity (refer Figure 10.2).

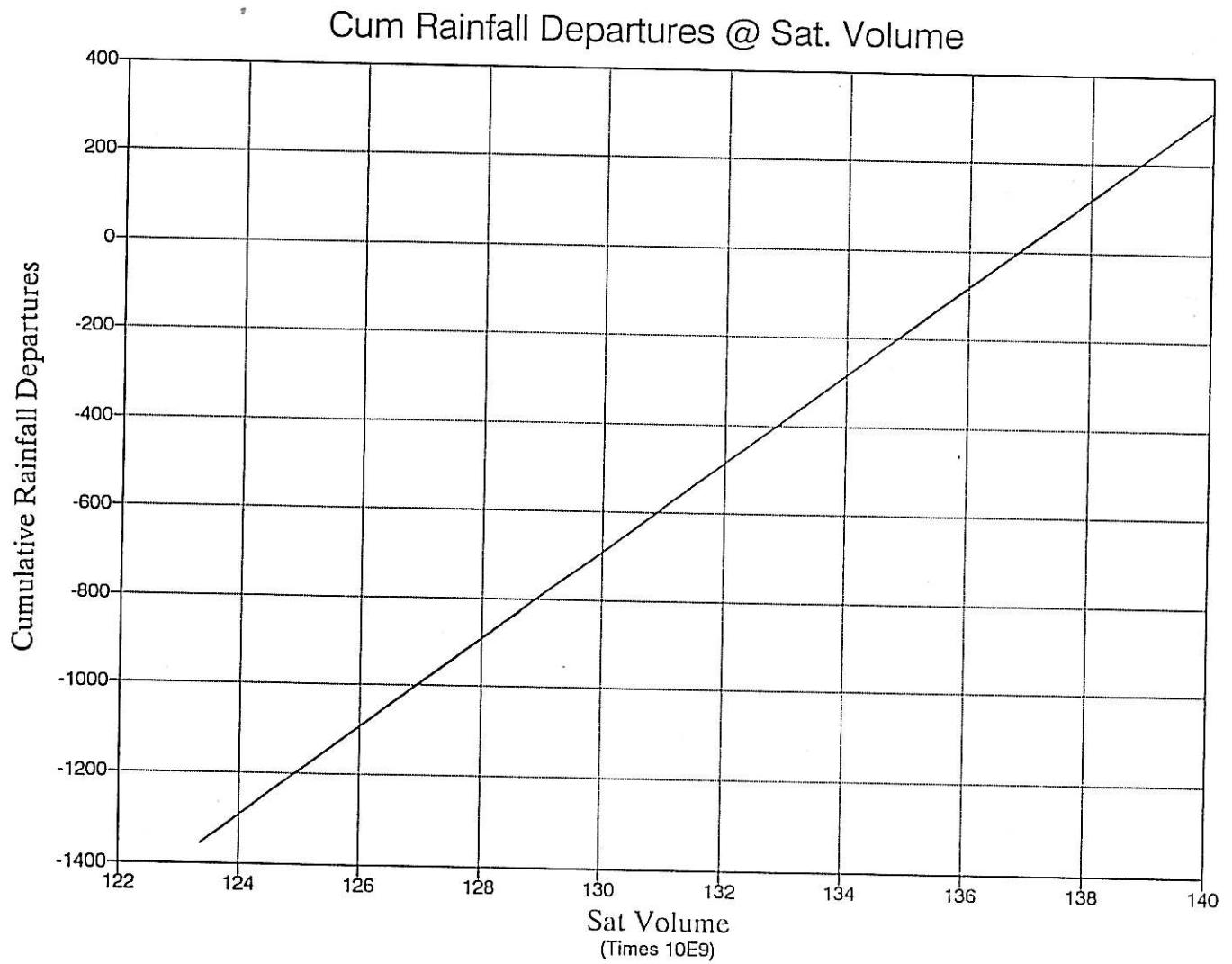


Figure 10.4 Relationship between the cumulative rainfall departures and SV values

The total flow of the Kuruman fountains
plotted against cum rainfall departures

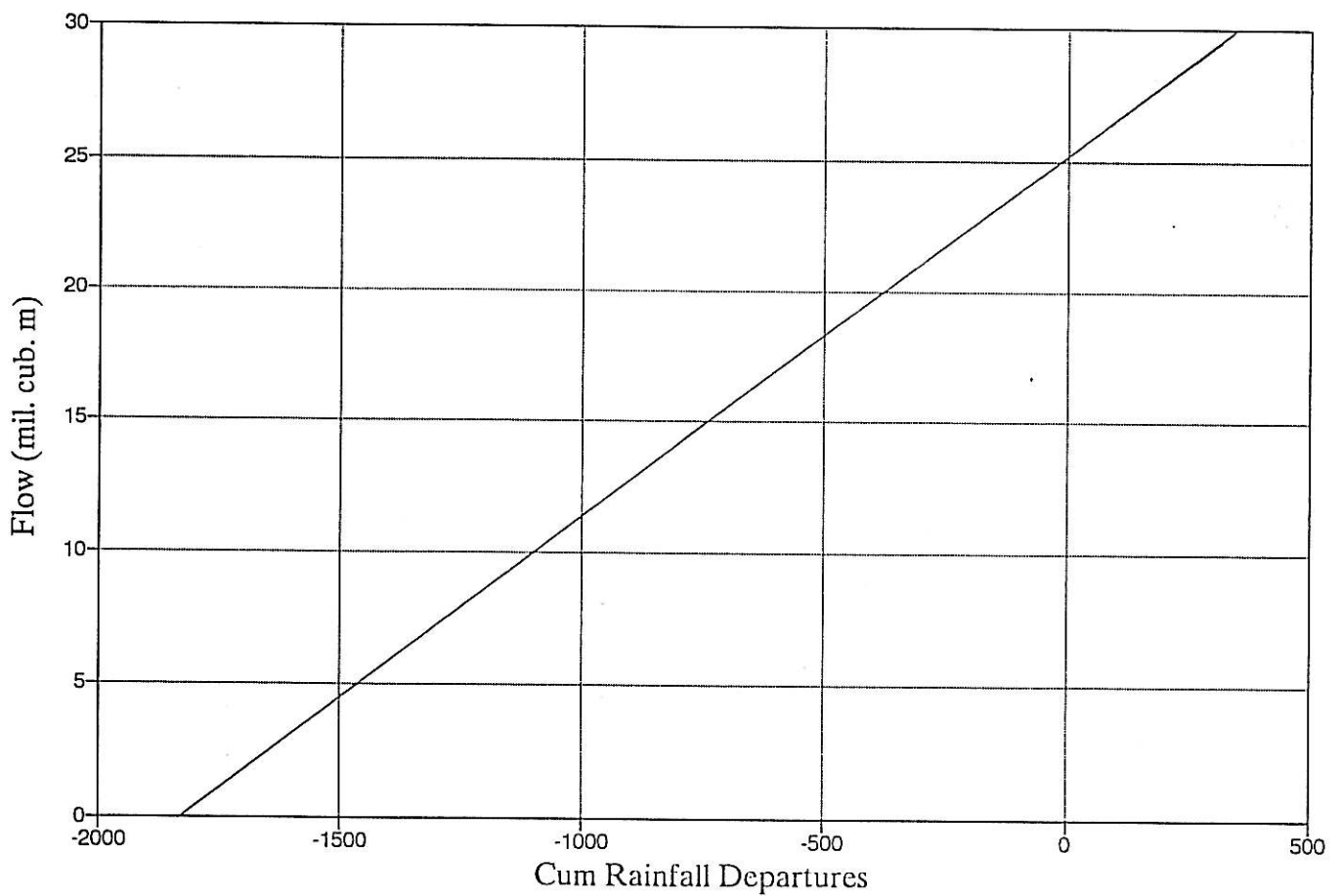


Figure 10.5 Relationship between the cumulative rainfall departures and the equivalent flow from the groundwater system, which was derived by means of the cumulative rainfall departure and measured values of the Kuruman and Manyeding springs. (refer Fig. 5.3 and 6.5) The loss component was also incorporated.

Change in Flow @ change in Sat. Vol.

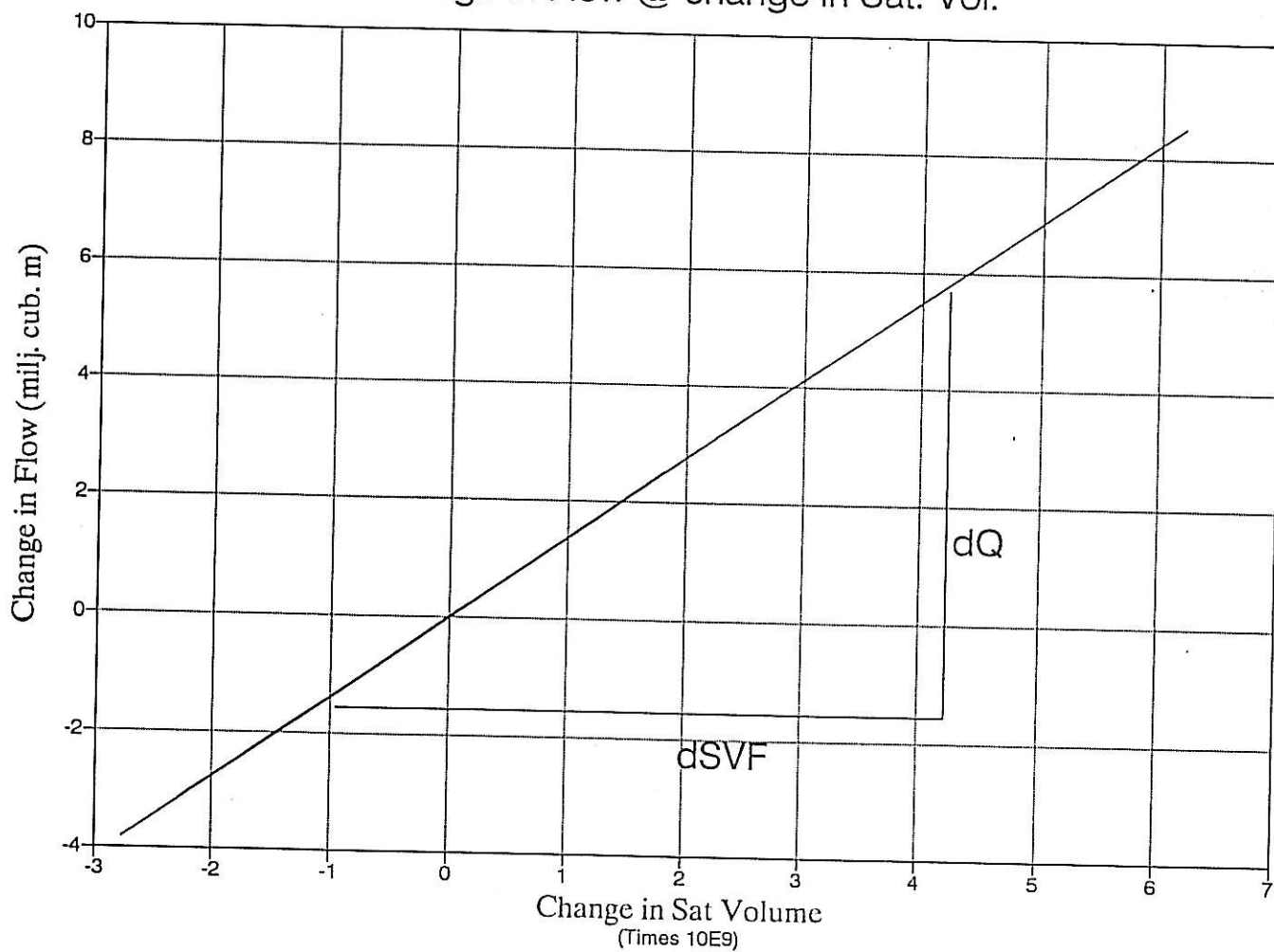


Figure 10.6 Application of the adapted Hill method to infer the aquifer storativity

The saturated volume for the period 1925 to 1991 could either be inferred from the relationship between the yield of the aquifer for 1970, 1986 and 1991, or by means of the cumulative rainfall departures which stands in a linear relation to springflow or recharge (refer Figure 5.3). The relationship between the cumulative rainfall departures (CRFD) and the SV values, which is also a linear one, is shown in Figure 10.4. By means of the linear regression between the CRFD and the spring flows the SV for each of the years could be deduced (refer Figure 10.5).

10.2.3 Estimation of the aquifer storativity

The change in the total aquifer recharge per year is accompanied by a change in aquifer storage. The annual recharge can be represented by the equivalent flow of all the springs for virgin conditions. Figure 10.6 shows the values of the change in SV plotted against the corresponding change in virgin conditions. This yields a linear relationship from which the aquifer storativity could be calculated (refer Figure 10.6) using the adapted Hill approach (Bredenkamp et al, 1987) - this yields a storativity of

$$S = \frac{\Delta Q}{\Delta SVF} = \frac{7 \times 10^6}{5 \times 10^9} = 1,4 \times 10^{-3}$$

10.2.4 Verification of the S value

The value of S, obtained in the previous section, which appears to be rather low, can be verified by an independent estimate

based on the storage/recharge ratio that was obtained from the interpretation of the C^{14} and tritium pulses that appeared in the spring water (refer Section 9).

The water temperature of the Kuruman A eye is $22,4^{\circ}\text{C}$ which is $2,8^{\circ}$ higher than the average ambient temperature of the region which is $19,6^{\circ}\text{C}$. Very few measurements of the geothermal gradient in the Kuruman eye catchment area are available, but the few measurements indicate a gradient at depth of about $0,78^{\circ}\text{C}/100\text{m}$, which is close to the general gradient for dolomite over the country which is about $0,88/100\text{m}$ according to Jones 1992.

This implies that the water from the Kuruman A spring emanates from an average depth of $2,8/0,78 \times 100$ metres ie. 360 metres. The simulation of the C^{14} and tritium pulses of hydrogen bomb tritium that had entered and mixed with the main body of base flow water, revealed that the storage/recharge ratio is about 25 on average, varying between about 20 and 40 depending on the recharge equation that was used (refer Section 9.4). The average recharge of the aquifer during the period 1958 to 1992, was estimated at 15 mm/a which means that the total column of water in the aquifer is equivalent to 375 mm of water. This water is spread over 360 metres of the aquifer which yields a storativity value of $0,375/360 = 1 \times 10^{-4}$. The maximum storativity value could be $40/25 \times 1,10^{-4}$ ie. $1,44 \times 10^{-3}$ and the minimum value 7×10^{-4} depending on whether the storage to recharge ratio is 40 or 20. This is still in close agreement to the water balance

estimate bearing in mind that the water balance estimate reflects the storativity of the aquifer at the levels of the water table fluctuation. In the case of the geothermally derived value, the storativity is the average value over the entire aquifer depth and the storativity is known to decrease with depth (Enslin and Kriel, 1967). This implies that the actual depth of the aquifer is probably higher than 360 m.

11. ANALYSIS OF RESULTS AND CONCLUSIONS

The study of the Kuruman groundwater system has revealed a number of significant results regarding the recharge and the storativity of the aquifer.

The flow of the springs conforms to linear relationships with the cumulative rainfall departures from the average rainfall, both with respect to monthly and annual values. This is in agreement with the behaviour of springs in other dolomite regions and allows the longterm variation in flow to be derived from the rainfall.

The average recharge to the Kuruman system was inferred from the combined yields of the springs issuing from the drainage area and the estimated losses due to irrigation, evaporation losses occurring from ponds and vlei areas as well as the domestic/cattle consumption as drinking water.

The average recharge of the area was estimated at 15mm/a which

represents 3,3% of the average longterm rainfall which is estimated at 466 mm.

The annual variation of recharge appears to conform to the results which had been obtained for the Bo Molopo region (Wondergat) and the same rainfall recharge equations could be used to derive the annual recharge values.

This was concluded from the good simulations of the C^{14} and tritium values that had been measured in the Kuruman A eye. These isotope values provided a unique opportunity to test the different recharge formulae and the input and admixture of the C^{14} and tritium to the water stored in the aquifer.

The simulations proved that a simple mixing model incorporating the recharge and the column of water stored in the aquifer, provided good results. From these it was concluded that the ratio of water stored in the aquifer to the average recharge, is of the order of 25.

In order to get reasonable correspondence between the C^{14} and the tritium simulations a lag of about 7 years had to be incorporated for the tritium input. The initial value of C^{14} (for water of zero age) was assumed to be 85% modern whilst that of the bulk of the system appeared to be about 69% modern. This implies a corrected age of 2000 to about 500 years for the average age of water in the system, which is about 20 to 100 times the ratio given by the simulation of the C^{14} and tritium pulses in the

spring water. This is an aspect still requiring further investigation and clarification.

The aquifer storativity, estimated at about $1 - 1,4 \times 10^{-3}$, was derived by two methods ie.

- 1) based on the equivalent total volume of water stored in the aquifer ($\approx 375\text{mm}$) and the inferred depth of the aquifer which was derived from the geothermal gradient and the temperature of the water of the main spring.
- 2) based on an analysis of the saturated volume fluctuation of the water in the aquifer in relation to the change in spring flow. This method entails the application of an extended version of the Hill method by Bredenkamp et. al.

The study is probably one of the most extensive investigations of the recharge, the flow characteristics and the storativity as well as the total volume of water stored in an aquifer. The results agree very well with that which had been obtained for the Bo Molopo area especially regarding the best rainfall recharge relationships.

The C^{14} measurements of the spring water should be continued for another year at intervals of about 3 months. Some tritium measurements must also be obtained, whence a critical re-examination of results are required. This would hopefully clarify some of the uncertainties about C^{14} age determinations as

a viable method of estimating groundwater recharge quantitatively.

The low value of the aquifer storativity and the small average annual recharge to the system might require a decision as to the need for a control on the abstraction of water for purposes other than primary consumption.

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