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ESTIMATION OF NATURAL GROUNDWATER RECHARGE IN THE KAROO AQUIFERS OF SOUTH AFRICA

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ABSTRACT

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A 3-year project, to study the natural groundwater recharge of aquifers in the semi-arid Karoo formations of South Africa, was undertaken. Two typical Karoo aquifers, at Dewetsdorp and De Aar, were selected for study purposes. Data were collected from both the saturated and unsaturated zone. Neutron probe measurements showed that there was no increase in the water content beneath a depth of ~ 1 m below the surface. Even with the exceptionally high rainfall in February 1988, neutron measurements indicated that very little soil matrix flow occurred, which implied that most of the recharge occurred along preferred pathways. A triangular finite element network was used to determine the relative saturated volume fluctuations (SVF) of the aquifers from the observed water-level changes over a period of time, which allowed determination of the storativity S and the recharge. The main advantage of the method lies in determining recharge and storativity which are both unknowns in the water balance equation and which both contribute to the water-level response. The SVF method showed that the recharge in the Karoo formations of South Africa varies between 2 and ~ 5% of the annual rainfall. In areas which are overlain by a thick soil cover, the recharge is less than 3%, while recharge in hilly areas with a thin soil cover may be of the order of 5%.

INTRODUCTION

A project to study the natural groundwater recharge of the Karoo aquifers which lie in a semi-arid region, with annual rainfall between 200 and 600 mm, was born out of the need to supply communities in the central part of South Africa with water. In this part of the country, with its relatively low rainfall, the only perennial stream is the Orange River which originates in the Drakensberg of Lesotho. Yet, a considerable number of villages and small towns (> 100), have developed in the Karoo over the past 100–200 years, which draw their water supplies from local aquifers. As these aquifers are more extensively exploited and additional ones have to be developed, the queston of their potential becomes important.

Because the groundwater bearing properties of the subsoil depend mainly on

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the geology and as ~50% of the surface of South Africa (in essence the whole central part) is underlain by rocks of the Karoo Supergroup, the investigation of the groundwater potential of Karoo aquifers was included in the Master Plan of the Water Research Commission of South Africa. The findings presented here are some of the results obtained during a three-and-a-half-year study which was carried out on contract for the Commission (Kirchner and Van Tonder, 1990).

The exploitation potential of an aquifer was defined as the long-term yield which depends on recharge from rainfall, underground inflow and outflow and groundwater storage.

The main objective of the study 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. Two typical Karoo aquifers (the Dewetsdorp and De Aar aquifers) were selected for the purpose of the study (Fig. 1). Dewetsdorp, with its 20 km^2 study area covering most of the townlands and parts of adjoining farms, has a mean annual rainfall of 584 mm, whereas the De Aar area, 120 km^2 in size, has a mean annual rainfall of 287 mm.

Rainfall varied during the study period between less than normal and normal. Towards the end of the project a major flood occurred; after an extended dry spell with soaring temperatures, soaking rains started on February 17, 1988. Within 48 hours, rain exceeding 250 mm fell within a 40 km-wide band, especially in the Dewetsdorp area. A probability analysis showed that this flood, with a total precipitation of 417 mm within a 30-day period, has

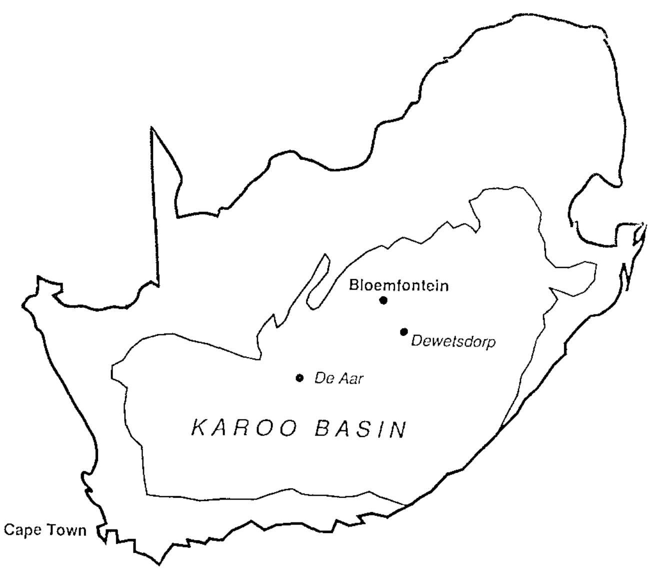

Fig. 1. Locality map of the Dewetsdorp and De Aar aquifer in the Karoo Basin of South Africa.

a return period of 500 years. In order to monitor the effects of this event on groundwater recharge the original 3-year project was extended by six months.

Because the results for the two aquifers were more or less the same, only the Dewetsdorp aquifer will be discussed in this paper.

RECHARGE ESTIMATION TECHNIQUES

Estimating the rate of aquifer replenishment is probably the most difficult of all measures in the evaluation of groundwater resources. Estimates are normally and almost inevitably subject to large errors. In the preface of a workshop recently held by the NATO Advanced Research Group, Professor Ian Simmers (Simmers, 1987) reported that no single comprehensive estimation technique can yet be identified from the spectrum of those available, which does not give suspect results.

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:

(1) the soil water balance method (soil moisture budget);

(2) the zero flux plane method;

(3) the one-dimensional soil water flow model;

(4) the saturated volume fluctuation method (groundwater balance);

(5) inverse modelling for estimation of recharge (two-dimensional groundwater flow model);

(6) istope techniques and solute profile techniques.

The saturated volume fluctuation method and the two-dimensional groundwater flow model are regarded as indirect methods, because groundwater levels are used to determine the recharge. Although the present state of knowledge in the field of groundwater 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 groundwater recharge estimations should be based on the groundwater level fluctuation method (Sinha and Sharma, 1987). Contradictory to this statement, Rushton (1987) and Johansson (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 (Johansson, 1987; Rushton, 1987; Thiery, 1987). Sharma and Hughes (1985) studied the groundwater 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 were the case for a sandy coastal region, one can imagine that this value would be higher for the hard-rock formations of South Africa and results obtained by use of the unsaturated soil water model must be treated with caution when this model is applied to the Karoo formations of South Africa.

Another example of preferred flow paths in Australia is given by Johnston et al. (1983).

Soil water balance method

Water balance models were developed in the 1940s by Thornthwaite (1948) and revised by Thornthwaite and Mather (1955). The method is essentially a book-keeping procedure which estimates the balance between the inflow and outflow of water.

In a standard soil water balance calculation, the volume of water required to saturate the soil is expressed as an equivalent depth of water and is called the soil water deficit. The soil water balance can be represented by:

$$G_r = P - E_a + \Delta S - R_o$$

where: G_r = recharge; 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 soil water balance 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 Selaola (1987) stated that the soil water balance method is of limited practical value, because E_a is not directly measurable.Moreover, storage of moisture 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 deserves attention is the depth of the root zone which may vary in semi-arid regions between 1 and 30m.

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). Johannson reported that for a sandy till aquifer in Sweden, recharge had to be allowed, even when a moisture deficit existed, in order to correctly reproduce the dynamics revealed as groundwater level fluctuations. Steenhuis et al. (1985) and Steenhuis and Van der Molen (1986) found that this 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.

Zero flux plane method

The zero flux plane 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 zero flux plane is usually determined by installation of tensiometers. Unfortunately, the method fails to work during 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 the unit area at any depth, is given by Darcy's law:

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

where: $K(\theta)$ = unsaturated hydraulic conductivity; θ = total water potential = $z + \Psi(\theta)$; z = depth beneath the surface (negative); Ψ = matric potential (negative); θ = water content. Thus, knowing the unsaturated hydraulic conductivity and the potential gradient, the flux may be determined. Water 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 approach 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 $z+\mathrm{d}z$:

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

where q_z is the vertical component of the Darcian water flux (positive in the upward direction). At the zero flux plane depth, 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 (c.f. Cooper, 1980):

$$F(Z) = \int_{t_1}^{t_2} q(z)\cdot\mathrm{d}t = \int_{z_0}^{Z} |\,\theta(t_1) - \theta(t_2)|\cdot\mathrm{d}z$$

where $Z = z + \mathrm{d}z$ and $z_0 = z$.

Wellings (1984) obtained good estimates of recharge in the U.K. for periods of the year when a zero flux plane exists. Sophocleous and Perry (1985) used the unsaturated zone hydraulic conductivity, $K(\theta)$, and hydraulic potential data, $\Psi(\theta)$ to solve Darcy's equation in the unsaturated zone below the zero flux plane. Application of the Sophocleous and Perry version of the zero flux plane depends on the success of accurately determining $K(\theta)$, which is one of the most difficult parameters to determine directly in the field. A technique to predict $K(\theta)$ is given in the following section.

The work of Stephens and Knowlton (1986) highlights the problem of using the flux technique for estimation of groundwater recharge, especially where the fluxes are low. They found that the annual recharge flux was either 7 or 37 mm/year^{-1}, depending on how they calculate their mean hydraulic conductivity.

Soil water flow model

For recharge to occur, 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 equation of a moisture retention curve is a non-linear relation of the water content. In more physical terms, it is said to show a hysteresis effect (Sophocleous and Perry, 1985; Rushton, 1987). Since the moisture 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 (1980) function. The Van Genuchten equation deserves special attention. In this equation, the moisture 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 moisture 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 and θ_r = the residual water content.

The three parameters, namely, (1) the water content, (2) the matric potential (fluid pressure) and (3) 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 of two or more orders of magnitude. The one-dimensional equation for vertical flow in the unsaturated zone can be expressed as (Richards, 1931):

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left(K(\Psi)\cdot\frac{\partial \Psi}{\partial z}\right) + \frac{\partial}{\partial z}\Big(K(\Psi)\Big) \tag{1}$$

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 (1986) indicate that the solutions of eqn. (1) are more sensitive to $\Psi(\theta)$ variations than $K(\theta)$ variations. No evidence in the literature exists that the $K(\theta)$ relationship exhibits a significant hysteresis therefore it is safe to assume that K is a unique function of θ.

Following Richards (1931), Darcy's law for unsaturated flow can be expressed as:

$$q = -K(\theta)\cdot\frac{\partial \phi}{\mathrm{d}z}$$

where $\phi = z + \Psi(\theta)$ and $K(\theta)$ are 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 (1) can be solved by either a finite difference or a finite element model.

Inverse modelling technique

The inverse modelling technique is a two-dimensional finite element (or finite difference) groundwater model of the saturated zone. Current methods of calibrating groundwater 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 modelling technique represent imprecise measurements and processed information that give a distorted picture of the system's true state (Neuman, 1973).

The calculation of recharge to an aquifer by the inverse modelling technique must be regarded with caution if the true S-values of the aquifer are not known. If, however, the calibrated S-values can be regarded as being 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.

Saturated volume fluctuation (SVF) method

Theoretical development

Inputs and outputs for conventional hydrological models are generally water volumes per unit time, such as recharge, discharge and surface inflows 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, for example:

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

In the same manner, the geohydrological balance equation for a groundwater reservoir is given as:

$$I - O + G_r - Q = \frac{\Delta W}{\Delta t} \qquad (2)$$

where:

$I = \dfrac{I_1 + I_2}{2}$ = mean lateral inflow (m^3 day^{-1}) during time $t_2 - t_1 = \Delta t$;

$O = \dfrac{O_1 + O_2}{2}$ = mean lateral outflow (m^3 day^{-1}) during Δt;

G_r = groundwater recharge into the reservoir in m^3 day^{-1} (also named percolation or deep infiltration);

Q = discharge out of (or into) the reservoir (boreholes, rivers, etc.) in m^3 day^{-1} during Δt;

ΔW = change in groundwater volume (m^3) = $S \cdot \Delta V$;

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 eqn. (2), it is assumed that there is no vertical movement through the base of the water-table aquifer. In the case where the roots of plants extract water from below the water table, the evapotranspiration must be added to the Q-term in eqn. (2). The accuracy of estimating this evapotranspiration term is not very reliable, and, for this reason, it is assumed that the G_r-term already includes the evapotranspiration, i.e. effective G_r = actual G_r − evapotranspiration from the saturated zone. In the application of the SVF-method in this paper, the estimated groundwater recharge is the effective groundwater recharge.

The term I in eqn. (2) can be expanded with the aid of Darcy's law:

$$I = T_i \cdot L_i \frac{i_i^1 + i_i^2}{2} = A_1 \cdot T_i \qquad (3)$$

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_i^1 = groundwater gradient at inflow boundary at time t_1; i_i^2 = groundwater gradient at inflow boundary at time t_2.
The same reasoning can be followed at the outflow boundary to yield:

$$O = T_o \cdot L_o \frac{i_o^1 + i_o^2}{2} = A_2 \cdot T_o \tag{4}$$

Substitution into eqn. (2) yields:

$$G_r + A_1 \cdot T_i - A_2 \cdot T_o - A_3 \cdot S = Q \tag{5}$$

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

Equation (5) is the general groundwater balance equation for an unconfined aquifer. The boundaries of an area usually studied, do not represent streamlines, i.e. they are not perpendicular to the equipotential lines. Hence, the lateral inflow and outflow of groundwater 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 of the water table changes (Kessler and De Ridder, 1974), especially for water tables < 3 m deep. Furthermore, it should be noted that if the water table drops, part of the water is retained by the soil particles; if it rises, air can be trapped in the interstices that are filling with water. Hence S for rising water is, in general, less than for a falling water table (Kessler and De Ridder, 1974).

To apply eqn. (5) correctly, it is essential that both the area and the period for which the balance is assessed, be carefully chosen. By comparison of groundwater 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 will be described at the end of this section. Equation (5) can be applied for a number of specified assumptions.

(a) Where the inflow terms are balanced by the outflow terms, the change in groundwater storage is zero (i.e. $\Delta V = 0$). This provides the necessary conditions to derive safe yield estimates and to predict recharge from precipitation:

$$G_r = A_2 \cdot T_o - A_1 \cdot T_i + Q \tag{6}$$

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 groundwater storage or the inflow that takes place following the recession.

(b) By incorporating the "no recharge" recession (i.e. G_r = O) for ΔV = maximum decrease during Δt, eqn. (5) reduces to:

$$A_1 \cdot T_i \; - \; A_2 \cdot T_o \; - \; A_3 \cdot S \; = \; Q$$

from which S can be calculated as:

$$S = \frac{A_1 \cdot T_i \; - \; A_2 \cdot T_o \; - \; Q}{A_3}$$

The S calculated with the above equation is a minimum, because G_r may not be zero as assumed.

(c) If the aquifer is bounded by impervious dykes or by groundwater divides, A_1 and A_2 in eqn. (5) are zero. For this case:

$$G_r \; - \; A_3 \cdot S \; = \; Q$$

from which the groundwater recharge can be calculated, if S is known.
The following procedures for the application of eqn. (5) are recommended:

(1) Groundwater levels (m a.m.s.l.) in observation boreholes, which are well distributed over the whole of the aquifer within certain well-defined groundwater boundaries (such as groundwater divides or no-flow boundaries), are required on a regular, preferably monthly, basis.

(2) The region must be divided into a number of small triangles (constructing a mesh).

(3) 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 in ΔV over a Δt is needed.

(4) Repeat (3) for all months.

(5) Construct a graph of V_i against time (months). For times where $V_i = V_j$ (i.e. $\Delta V = 0$), eqn. (6) can be used to estimate the groundwater recharge G_r, during the time $t_j - t_i$.

It is very important to realise that eqn. (5) is subject to a number of possible errors. 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 groundwater 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 eqn. (5) is dependent on the accuracy of the transmissivity values at these boundaries.

A microcomputer program which performs the calculation of saturated volumes and groundwater recharges, was developed to deal with the above-mentioned equations and is named SVF.

Isotope and solute profile techniques

^{3}H, ^{2}H, ^{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, 1978; Foster and Smith-Carrington, 1980). Although a proven tool for qualitative recharge estimation, environmental tritium has several disadvantages (Edmunds, 1987), e.g. (1) tritium is not conservative and is lost from the system by evapotranspiration; (2) contamination during sampling and processing is a factor which is enhanced in remote areas and at low total moisture levels; (3) analysis is highly specialized and costly; (4) 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 suitable for determining 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 precipitated.

If the assumption of chloride as a conservative ion is accepted, the groundwater recharge is given by:

$$G_r = \frac{D}{C} \text{ (mm/year}^{-1}\text{)}$$

where D = wet and dry chloride deposition (mg m^{-2} /year^{-1}) and 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 groundwater, if there are no evaporite sources, is from the atmosphere. In this case, the recharge can be expressed as (c.f. Houston, 1987):

$$G_r = \text{rainfall} \times \text{Cl of rainfall/Cl of groundwater}$$

The chloride method must be treated with caution, as accession of chloride near the soil surface may violate the assumption of a steady state chloride flux density throughout the unsaturated zone, because of evapotranspiraton (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 groundwater to a high degree, resulting in an overestimate of the mean annual recharge.

DATA COLLECTION

Groundwater data, generated during the early stages of the project, were entered into the National Groundwater Data Base. A complete soil survey was carried out in both research areas by the Department of Soil Science at the University of the Orange Free State. After a first reconnaissance of the area, inspection holes were dug and described and, where necessary, samples taken. Criteria for the identification of the different soil types were established and boundaries between them determined with the aid of auger drills.

Soil water and soil water movement belong to those hydrological variables which are most difficult to comprehend. This is not only because of space–time variations, but also the measuring methods. For the current project, two Troxler 3331 Series Depth Moisture Gauges were acquired. The instruments measure the moisture content of the soil and have the ability to store the data for downloading to a computer.

Dry and wet bulk densities have been determined in the soil laboratory of the Department of Soil Sciences at the University for between four and six different depths in 31 sample pits, dug 5m away from one of the respective moisture tubes. The samples were wrapped in plastic foil to prevent desiccation. At the same time, neutron gauge readings were taken in the access tubes and, for control purposes, gamma–gamma readings were recorded with a Nucletronix probe in the sample pits at the depths the samples were taken. For texture analysis, a total of 116 samples, 44 in De Aar and 72 in Dewetsdorp, were taken at the same sites.

Drilling of all 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. Water samples were taken of all waters struck and analysed for the usual macro elements by the Department's Hydrological Research Institute. Water samples from selected holes were analysed for their isotopic composition.

A total of five rain gauges were installed in each of the two study areas.

DEWETSDORP

General features

Dewetsdorp is situated in the eastern part of the Karoo Basin and has an annual precipitation of $\sim$600 mm. The research area stretches $\sim$8.5km in a SSW–NNE direction, is about 3 km wide and has a surface area of 21.1 km^2, with a well-defined ground-watershed in the west, south and southeast, which follows a chain of rather steeply sloping hills, partly capped by dolerite sills. The eastern boundary is formed by a shallow topographical ridge between the Kareefontein Creek and the Modder River Valley. In the north, where the Kareefontein Creek joins the Modder River, there is no boundary and surface water as well as groundwater outflow, does occur.

In the uppermost parts of the area, the topographical gradient is $\sim$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 $\sim$1:40.

Climate

Rainfall is mainly in the form of showers and thunderstorms, falling between October and March, with a peak in the months January–March. Temperatures show large diurnal and seasonal variations of between − 11 and 41°C.

Much effort was expended in obtaining meteorological data. A network consisting of a weather station, two automatic rainfall recorders and a rain meter were set up. Precipitation within the research area was also measured at a Weather Bureau Station in operation since 1905. With hindsight, it seems doubtful whether all the additional meteorological data generated, helped to obtain a better estimate of the aquifer recharge.

In groundwater potential calculations, it is common practice to establish a relationship between the annual rainfall and the annual groundwater recharge. The yearly exploitable groundwater 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 groundwater supply can be depleted seriously during a low rainfall period of several consecutive years. Secondly, the annual recharge is not merely dependent upon the annual rainfall depth, but upon the annual distribution of storm events as well.

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 rainfall. The mean annual rainfall for Dewetsdorp, as measured from 1905 to 1987, amounts to 587 mm (median = 564, 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 in the hydrological year, October 1985 to September 1986, equalled 564 mm and for the following year it was 612 mm. Except for the flood event in 1987/1988, the rainfall could be considered as quite normal during the study period. Precipitation in that season amounted to 1075 mm, and during the period November 1, 1987 to October 31, 1988, it was even higher at nearly 1200mm.

Geology

The area is underlain by sediments of the Beaufort Group which consist of fine-grained cross-bedded sandstone and coarse arkose, alternating with greenish-, bluish-grey and greyish-red mudstone. The sediments reflect a cyclical succession of channel and flood plain deposits.

Dolerite sills cap on the hills west and northwest of the village. A number of prominent dolerite dykes are also 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. Here they had to be detected and followed by means of a magnetometer survey. The dykes strike in various directions and give rise to fountains in the upper parts of the Townlands and to a number of fairly good yielding boreholes which have been sited on them.

Geohydrology

Unsaturated zone

During the first year of the project, neutron measurements were taken eight times between October 17 and December 24, 1985, seven times during 1986, twelve times during 1987 and five times during 1988. Based on these measurements, the soil water content and the soil water matric potentials were calculated. Because no noticeable change in the potential gradient was detected in 18 of the 20 neutron access tubes, situated in the flat area below the village, it was concluded that in this area with its deeper and fine-grained soil cover, no significant recharge took place *by means of matrix flow* in the unsaturated zone.

During long periods of drought, the soil water profile showed no variation and even after the good rains of February 1988, no increase in the soil water content was observed below a depth of ~ 0.8 m. Recharge through the soil matrix does, however, occur in the higher lying areas near the foot of the hills, where the remaining two neutron tubes were situated, and possibly along the major watercourses, especially during extended high rainfall periods.

Using a constant head permeameter, a number of saturated K-values (K_s) were determined for undisturbed soil samples. Measured permeabilities varied between 2×10^{-1} for top layers and 4×10^{-5} m day^{-1} for deeper layers. It is interesting to note that, for two of the samples, no K_s-value could be calculated, due to 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 the soil, there was still no flow that could be measured.

Saturated zone

Borehole yields were generally very weak ($< 1\,\mathrm{l\,s^{-1}}$), if water was struck in the weathered zone only. Higher yields were associated with dolerite intrusions or prominent jointed zones. Of the forty exploration holes drilled in Dewetsdorp in the beginning of the project, only three had yields in excess of $1\,\mathrm{l\,s^{-1}}$. Original yields of the municipal production holes varied between ~ 3.5 and $6.5\,\mathrm{l\,s^{-1}}$. The water-level observation network consisted of 72 boreholes (Fig. 2).

Evaluations of pumping-test data and double-packer testing were used to determine hydraulic parameter values. Transmissivity values range between 5 and 50 m^2 day^{-1} with S values lying in the range 10^{-3}–10^{-4}.

Estimation of recharge

The saturated volume fluctuation (SVF) method

The saturated volume fluctuation (SVF) method transforms groundwater

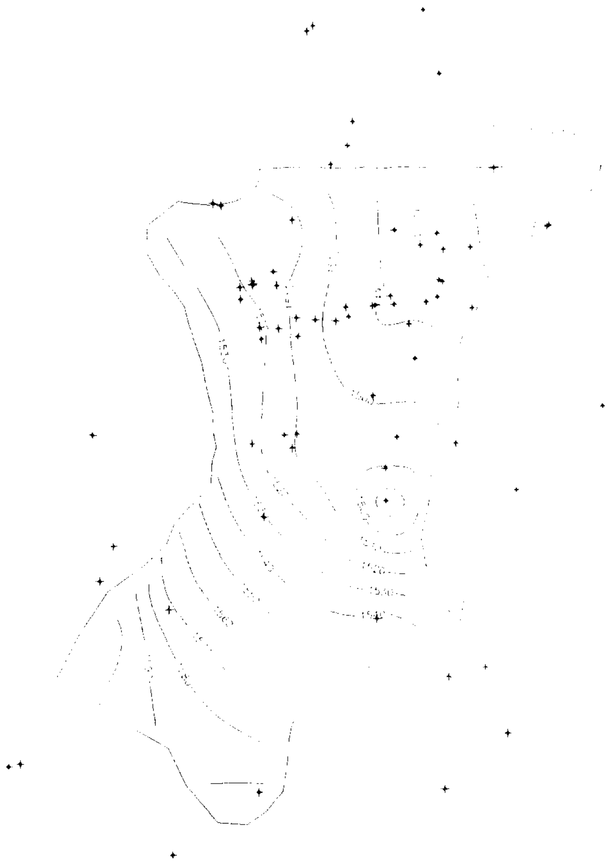

Fig. 2. Position of the seventy-two observaton boreholes in Dewetsdorp together with the groundwater level contours.

level fluctuations 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 eqn. (5).

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

(1) Firstly, the position of the topographic groundwater divide was deduced from reports (Dziembowski, 1977). Figure 2 shows the position of these boundaries for the southern, western and eastern parts of the aquifer. Groundwater outflow occurs in the northern part of the aquifer.

(2) The part of the aquifer outlined on Fig. 2, was discretized into 934 triangles with 513 nodes (Fig. 3).

(3) Monthly groundwater levels at 72 boreholes were used by the SVF program to calculate the saturated volume for each triangle and consequently for the whole aquifer. Figure 4 shows the saturated volume for the period October 1985 to July 1988 (a period of 34 months). Figure 4 was obtained by

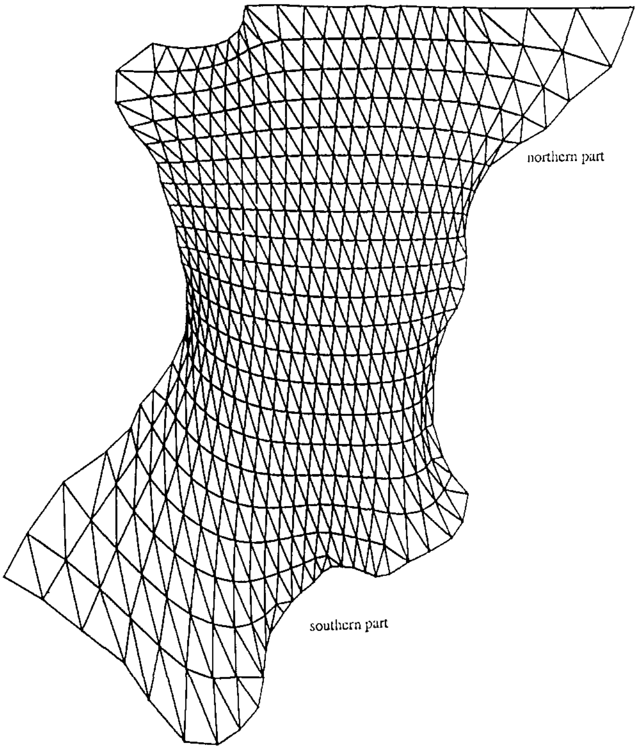

Fig. 3. Triangular finite element mesh used for the SVF method and two-dimensional modelling of the aquifer.

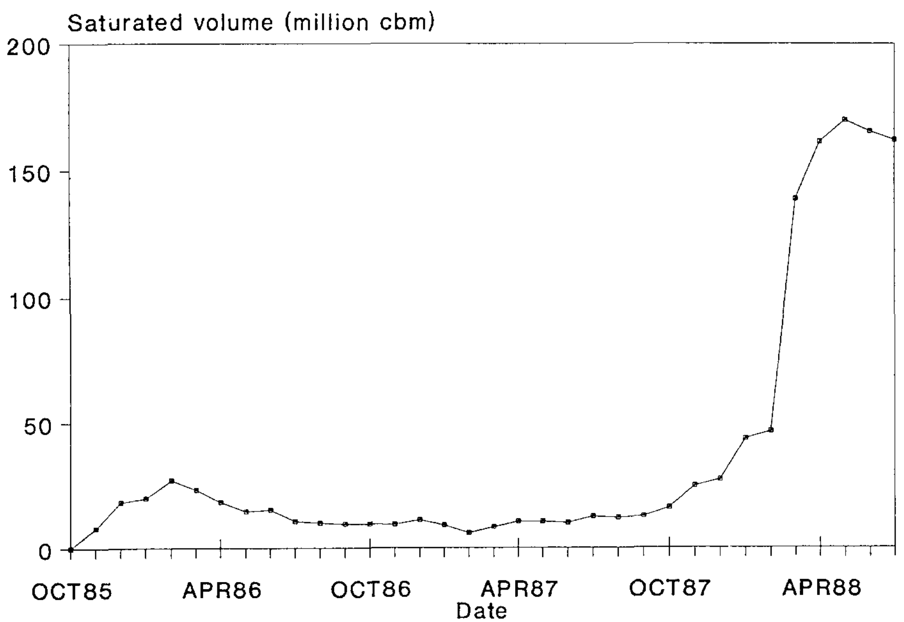

Fig. 4. The relative saturated volume of the Dewetsdorp aquifer for the study period.

subtracting the saturated volume of each month from that of October 1985. The SVF method will subsequently be applied to a number of special cases for Dewetsdorp.

Case 1

Equation (6) describes the case where the change in groundwater storage is zero (i.e. $\Delta V = O$):

$$G_r = A_2 T_o - A_1 T_i + Q$$

which simplifies to:

$$G_r = Q + A_2 T_o \tag{7}$$

for Dewetsdorp, because $A_1 T_o = 0$ for the region.

The Q-term in eqn. (7) can easily be quantified, because the abstraction rate of the municipality boreholes is monitored ($\sim$50 000 $m^3 year^{-1}$). Groundwater abstraction by other consumers in the township amounts to $\sim$13 500 $m^3 year^{-1}$ (Dziembowski, 1977). The groundwater abstraction from the aquifer by farmers 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_o in the second term of eqn. (7) is very important. The mean transmissivity of 10 $m^2 day^{-1}$ 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 groundwater flow model (which will be described in a following section), it was necessary to allow for a mean outflow at this boundary of 470 $m^3 day^{-1}$, which is in agreement with a transmissivity, T, of 10 $m^2 day^{-1}$ at the outflow boundary. The A_2-term in eqn. (7) represents the mean groundwater gradient times the width of the outflow boundary and is an output of the SVF program.

From Fig. 4, 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 eqn. (7) is applied for all such periods, the graph illustrated in Fig. 5 is obtained, which shows the relationship between rainfall and the groundwater recharge (infiltration) for different periods of time.

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

$$G_r = 0.024\,(P - 51)\ \text{(mm)} \tag{8}$$

with a correlation coefficient $r = 0.99$. With the exception of one recharge value, eqn. (8) was determined using periods of at least three months. The coefficient of variation (C.V.) equals 26% for precipitation in the Dewetsdorp region. For a mean annual rainfall of 587 mm, a mean annual groundwater 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 also 26%, which implies that the mean annual groundwater recharge of the Dewetsdorp region usually varies between 9.5 and 16.1 mm.

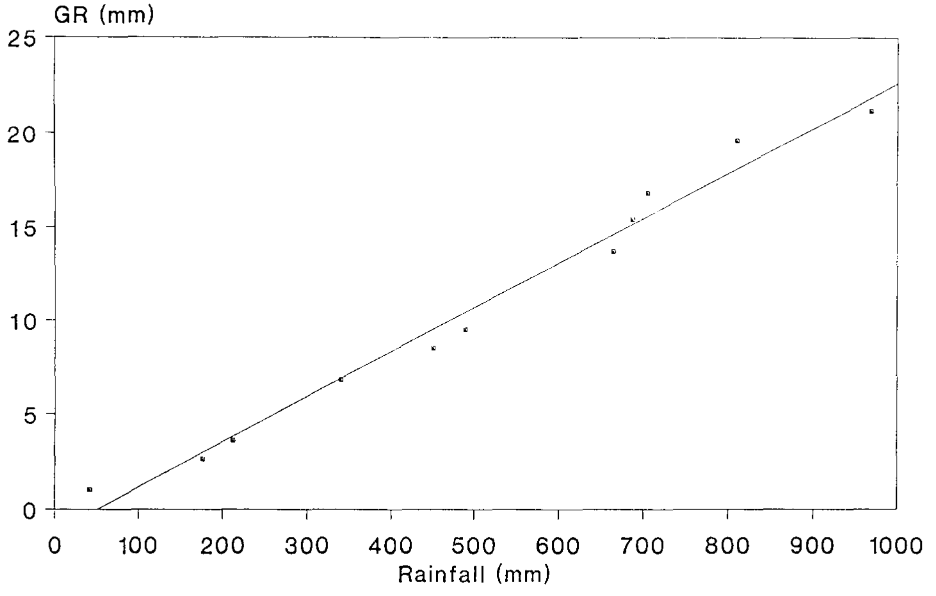

Fig. 5. Relationship between rainfall and groundwater recharge as obtained with the SVF method.

Case 2

Equation (5) is the general groundwater balance equation for an unconfined aquifer and for the Dewetsdorp aquifer it reduces to:

$$G_r = Q + A_2 T_o + A_3 S \tag{9}$$

where:

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

Whereas eqn. (7) is independent of the specific yield (S), it is not the case with eqn. (9). 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 April 1 to June 1, 1986, the maximum decrease in saturated volume occurred and equalled $-4.938 \times 10^6 \mathrm{m}^3$, so that:

$$S = \frac{-A_2 T_o - Q}{A_3}$$

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

$$= 4.1 \times 10^{-3}$$

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

$$S = \frac{-6\ 020}{-4.938 \times 10^{-6}} = 1.2 \times 10^{-3}$$

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

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 groundwater table. This time-lag depends on a number of factors that cannot usually be measured for each rainfall event (e.g. soil moisture 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 groundwater recharge. Table 1 shows that a lag of one month yields the highest cross-correlation coefficient. By lagging the rainfall with one month, annual percentage recharge values, shown in Fig. 6, were obtained (for Δt = 12

TABLE 1

Cross-correlation of the lag between rainfall and groundwater recharge

Lag (months)	Cross-correlation
0	0.12
1	0.85
2	0.20

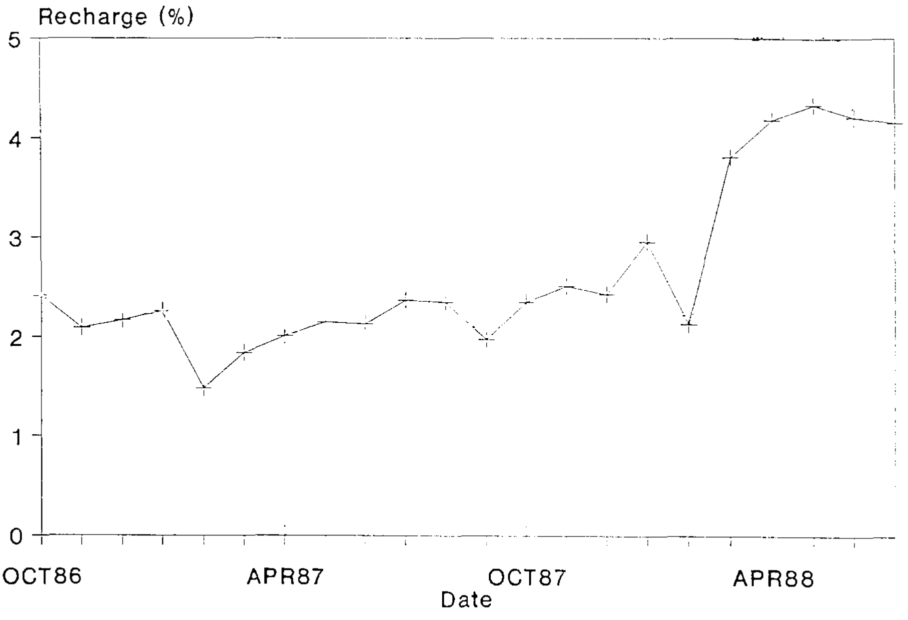

Fig. 6. The percentage annual recharge during the study period for the Dewetsdorp aquifer as estimated with the SVF method.

months). From this figure, it can be seen that the mean percentage recharge for the whole aquifer is of the order of 2.5%, except for the flood period when the recharge amounts to 5% of the rainfall.

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 southwest. The aim was to 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 (Fig. 2).

A T-value of $10\,m^2\,day^{-1}$ was used for the northern outflow boundary. For the southern part, an outflow T of $3\,m^2\,day^{-1}$ was used, a value that was obtained from the two-dimensional finite element model, which will be discussed in the next secton. This value of T accounts for the steeper water-table gradients in the vicinity of the northern boundary in the southern part. 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, respectively. By excluding the flood values, recharge values of 22 mm (1.7%) and 65 mm (5%) were obtained. Figures 7 and 8 show the recharge for the northern and southern parts of the aquifer during the study period, from which it can be concluded that the recharge is twice as high in the southern aquifer (with the thin soil cover) as it is in the northern aquifer (where the soil cover is thick).

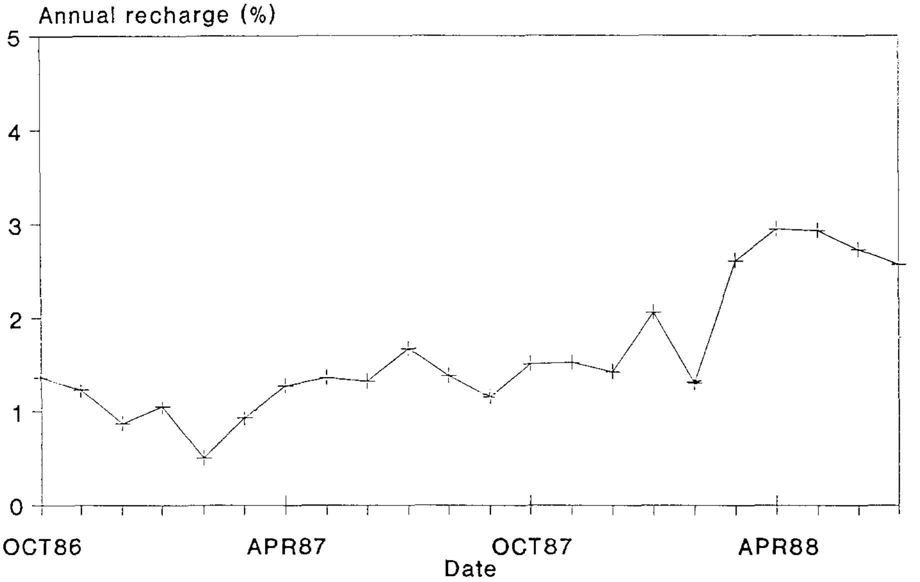

Fig. 7. Percentage annual recharge for the northern part (thick soil cover) of the Dewetsdorp aquifer as estimated with the SVF method.

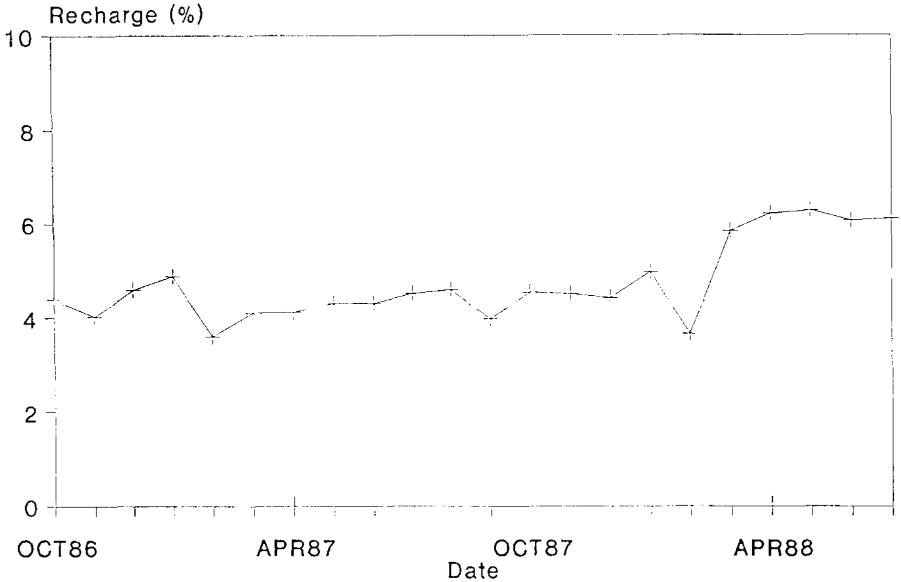

Fig. 8. Percentage annual recharge for the southern part (thin soil cover) of the Dewetsdorp aquifer as estimated with the SVF method.

Case 4

In the previous cases, 72 boreholes were used for the SVF method. To determine the sensitivity of the SVF method regarding the number of boreholes used in the analysis, initially 15 and then six boreholes were randomly selected in the Dewetsdorp area and the same analysis as in Case 1 was performed. If the results obtained with the 72 boreholes are supposed to be 100% correct, then the results with 15 boreholes are 68% and those with six boreholes are 67% correct. It is thus clear that the SVF method loses accuracy with a decrease in the number of boreholes used.

Two-dimensional finite element model (inverse modelling technique)

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 (Fig. 3) was constructed. The boundaries for the Dewetsdorp aquifer can be treated as groundwater 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.

The next step was to calibrate the model. The method proposed by Van Tonder (1989) was used to find the transmissivity distribution in the aquifer. T-values of between 1 and 20 $m^2 day^{-1}$ were found to give a good description of the steady state solution of the Laplace equation. S-values ranging between 0.003 and 0.008 simulated the physical behaviour of the aquifer well. An S-value of 0.001 was also used to calibrate the model, but without success.

An attempt was made to simulate the February 1988 flood between the period February 23 and the end of March 1988. For a first run, an S-value of 0.004 was used with a groundwater recharge value of 18 mm (4.5% of the rainfall) over the whole aquifer.

The results, in terms of the difference in metres between the actual and the simulated water levels, were not very encouraging. A constant recharge value for the whole aquifer thus tends to overestimate the water table in the northern part, and underestimate it in the southern part of the aquifer.

In a following step, varying recharge rates and S-values through the aquifer were applied on a trial and error basis. A good fit was obtained (with S-values lying in the range 0.003–0.008 and recharge values of between 1 and 8%), except at one or two boreholes, which displayed recharge behaviours which could not be simulated by the model. The calibrated model showed that the recharge in the hilly southern part of the aquifer was as much as four times higher than in the northern part of the aquifer.

Natural isotopes

For determination of natural isotopes, groundwater samples were collected from boreholes after completion and rain samples at regular intervals. Groundwater samples show a certain small depletion in deuterium which lie not far from the weighted mean of the precipitation that has fallen during the period. Overall, the stable isotopes display as postulated in the literature and the results support the assumption that local rainfall is responsible for recharge. No indication of groundwater of a different origin was found.

Recharge via the unsaturated zone

It has been stated above that no measurable matrix flow occurred in the flat-lying areas. 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 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 relatively heavy clay layer is reached, from where the water movement is predominantly horizontal. At specific locations, this clayey layer may have cracks, which form a pathway for water to travel along. Dieleman and De Ridder (1963) found similar results when they studied the recharge near Lake Chad in central Africa.

Recharge estimates in which the unsaturated zone model is used, are not applicable in the Dewetsdorp aquifer, because the recharge mechanism does not favour models that use soil water matrix flow.

CONCLUSIONS

This study has shown that the groundwater balance method is the only method which yielded reliable estimates of the groundwater recharge. Methods which use the unsaturated zone data are not suitable for calculating the natural recharge of groundwater in the Karoo formations of South Africa. The main recharge mechanism of groundwater in the Karoo formations of South Africa is flow along preferred pathways.

The saturated volume fluctuation method showed that the recharge in the Karoo formations of South Africa varies between 2 and $\sim 5\%$ of the annual rainfall. In areas which are overlain by a thick soil cover, the recharge is less than 3%, while recharge in hilly areas with a thin soil cover is of the order of 5%.

This study has shown that the saturated volume fluctuation method is the only practical method to obtain a reliable recharge estimate, as well as a mean storativity value of an aquifer in the Karoo formations of South Africa.

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