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Groundwater recharge in the Kalahari, with reference to paleo-hydrologic conditions

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Abstract

The Kalahari is situated in the semi-arid center of southern Africa and can be characterized as a savannah with a sandy subsurface, deep groundwater tables and annual rainfall ranging from 250 mm in the southwest to 550 mm in the northeast. A high infiltration rate and high retention storage during the wet season and subsequent high transpiration by the dense vegetation during the dry season, make that very little water passes the root zone and contributes to aquifer recharge. A lively debate has continued for almost a century on the question whether the Kalahari aquifers are being replenished at all under present climatic conditions. The present paper reports on results of an extensive recharge research project at the eastern fringe of the Kalahari, which is the most favorable part for groundwater replenishment. Additional observations were made in the central Kalahari. Environmental tracer studies and groundwater flow modeling indicate that present-day recharge is in the order of 5 mm yr^{-1} at the eastern fringe of the Kalahari where annual rainfall exceeds 400 mm. Figures in the order of 1 mm were obtained from the central Kalahari with lower precipitation. A dry valley system refers to more humid paleo-climatic conditions with a higher groundwater recharge. A tentative reconstruction of the groundwater depletion history suggests a time lapse of several thousands of years since the end of the last wet period.

Keywords: Groundwater recharge; Tracer studies; Dry valleys; Kalahari desert; Paleo-hydrology

1. Introduction

The semi-arid Botswana Kalahari is situated on the southern African plateau, which is characterized by a flat, slightly undulating topography and an elevation of around 1000 m above sea level. The Kalahari forms a tectonic as well as a morphological basin, filled with tens of meters of unconsolidated sandy deposits of Tertiary and Quaternary age. It is underlain mainly by Paleozoic to Mesozoic sedimentary rocks and basalt of the Karoo Supergroup. Rainfall is restricted to the summer period from September to April and on average ranges from 250 mm yr^{-1} in the southwest to 550 mm yr^{-1} in the northeast (Fig. 1). Perennial water at the surface is restricted to the far north where the Okavango and Chobe Rivers bring water from catchments in Angola and Zambia.

Groundwater is mainly found in Karoo rocks with water tables ranging from 20 m below surface at the fringe of the Kalahari to more than 100 m in the central part. The main regional groundwater flow is

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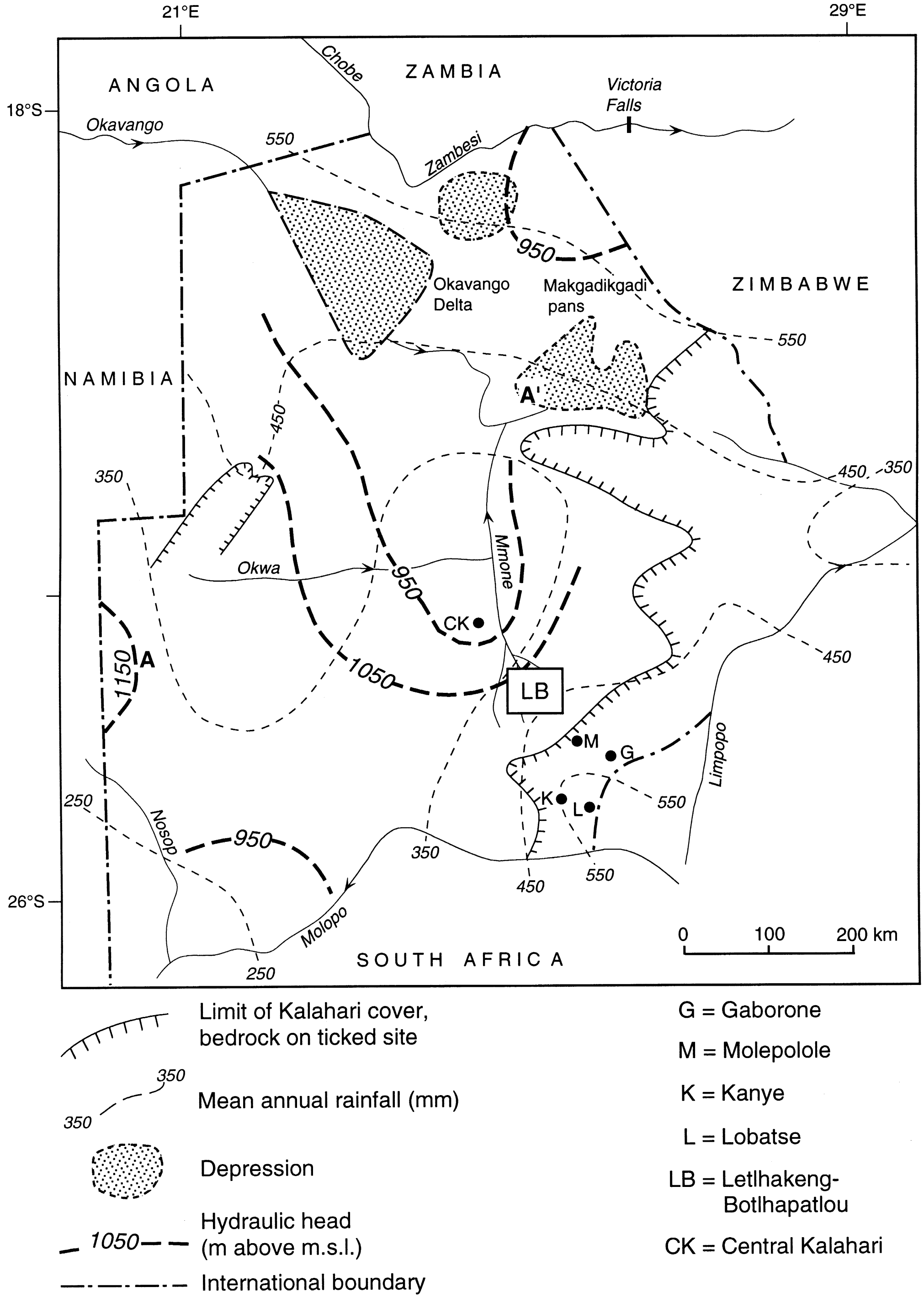

Fig. 1. Botswana: physiographic features and locations referred to in text; hydraulic head contours inferred after Hydrogeological Reconnaissance Map of Botswana, scale 1:500,000.

from the water divide in the southwest at a land surface elevation of 1250 m, to the lowest depression, the Makgadikgadi Pan, 600 km to the northeast at an elevation of 925 m. The groundwater table slopes from 1125 m at the divide to near surface at the Makagadikgadi Pan (Fig. 1). The regional hydraulic gradient is thus 0.03% whereas the surface slope is slightly higher with 0.05%. A small groundwater component flows southward to the depression of the Molopo dry valley. No springs or other groundwater outcrops occur in the area under present climatic conditions. The Kalahari thus forms more or less a closed basin with an internal groundwater drainage system.

The large Okwa-Mmone dry valley system follows the main groundwater system to the northeast. This poorly integrated system, with low tributary bifurcation ratios, consists of broad shallow depressions with more deeply incised headwater branches (Figs. 1 and 3). Other depressions are formed by hundreds of pans which probably reflects poor drainage conditions during wet paleo-climatic periods with high groundwater tables. Extensive calcrete and silcrete duricrust precipitates are found along the depressions, produced by subsurface leaching of the felspathic matrix and removal and precipitation of calcite. Virtually no paleo-climatic information is available for the Kalahari because of a lack of datable organic remnants. Proxy data of cave and pan deposits in the northern part of South Africa, just south of Botswana, gave evidence for a pluvial period during the last part of the Pleistocene, from 30,000–11,000 yr BP, followed by another wet episode during the early Holocene, from about 8000–4500 yr BP. From 4500 BP onwards more variable conditions with second order fluctuations prevailed (see Thomas and Shaw, 1991 and references therein). The late Pleistocene pluvial period is also known from the paleo-lake Makgadikgadi and the Okavango Delta in the northern Kalahari. The paleo-hydrologic conditions in this area however, were dominated by the Okavango and Zambezi river systems, coming from the northern wet tropical area.

The average annual rainfall of about 400 mm supports rather dense vegetation, classifying most of the Kalahari as a bush and tree savannah with alternating grass steppe. Real desert conditions in the sense of poorly vegetated and unstable moving sands only prevail in the extreme southwest where rainfall is below 250 mm. Groundwater has been encountered almost everywhere in Karoo and older Precambrian rocks beneath the Kalahari Beds, but the Kalahari sand itself does not at present form extensive aquifers. Perched water bodies are found locally in the Kalahari Beds in areas with large pans, related to episodic flooding and rapid infiltration through fractured duricrust deposits. For a detailed account of the Kalahari environment, one is referred to Thomas and Shaw (1991).

A lively debate has continued for almost a century as to whether groundwater recharge by infiltrating rainwater through Kalahari sands is taking place under the present climate (De Vries and Von Hoyer, 1988). Early studies (Boocock and Van Straten, 1961, 1962) contested the occurrence of recharge because of the high moisture retention storage in the sandy subsoil during the rainfall season and subsequent high evapotranspiration by the dense savannah vegetation during the dry season. Similar conclusions were drawn by Foster et al. (1982) from a few tracer profiles in the unsaturated zone at the eastern fringe of the Kalahari.

De Vries (1984) considered regional flow through the Kalahari from southwest to northeast, on the basis of hydraulic gradient and transmissivity and concluded a flux of less than 1 mm yr^{-1}, which for steady-state conditions would mean an overall recharge of the same order. He argued however, that the present gradient could also be a residual from a previous wet period, which would mean an even lower recharge. Starting from a wet period with the groundwater table close to the surface, it would, according to his calculations, take more than 10,000 yr of depletion by discharge to reach the present groundwater table. The present hydraulic head could thus equally be explained as a residual of a previous wet pluvial period.

To the contrary, several environmental tracer studies since the early 1970s demonstrated evidence of widespread and active replenishment by the occurrence of high ^{14}C and tritium contents in groundwater far below the root zone (Jennings, 1974; Mazor et al., 1974, 1977). Subsequent exploitation and exploration study at the fringe of the Kalahari also revealed evidence for recharge. The question then arose, if recharge is substantial, what then happens to this water, taking into consideration the low regional groundwater discharge flow and the fact that the groundwater table is everywhere below the root zone so that evapotranspiration from the saturated zone is apparently not possible. Some researchers however, have speculated in (oral) discussions about the possibility of the loss of water from greater depths by vapor transport.

To solve this groundwater recharge question, an extensive research project has been carried out by the present authors since 1992, via the Botswana Department of Geological Survey in cooperation with the Vrije Universiteit Amsterdam and the University of Botswana. The main aim of this project was to assess the recharge in the Kalahari and to investigate the reason for the discrepancy between recharge and discharge flux evidences. This Groundwater Recharge and Evaluation Study (GRES II project) focused on the fringe of the Kalahari where the most favorable conditions for replenishment were expected from evidence of earlier exploration. Additional observations were carried out in the central and dryer part of the Kalahari (Fig. 1).

An extensive account on the results of this GRES II project (as well as a former GRES I project that focused on the Precambrian hardrock area in the east) is reported by the present authors and their collaborators in Beekman et al. (1996, 1997) and Selaolo (1998). The present paper summarizes and evaluates part of the results of GRES II, viz., groundwater recharge assessment by application of environmental tracers and groundwater modeling. It furthermore presents a tentative evaluation of paleo-hydrologic conditions to distinguish a possible influence of residual flow components. Finally a discussion is added on moisture and vapor transport from depths below the root zone, because such processes could cause an imbalance between the total amount of percolation to the groundwater table and the total regional discharge flux through the saturated zone. The overall result of the present study is that substantial recharge occurs at the eastern fringe of the Kalahari, whereas very little groundwater replenishment prevails in the central en western part of the area, covering 80% of the Botswana Kalahari.

2. Results of the groundwater recharge investigations

2.1. The Letlhakeng–Botlhapatlou (L–B area, Figs. 1 and 3)

The study area covers 65×75 km^2 and is situated at the eastern fringe of the Kalahari, close to the boundary of Precambrian and Karoo outcrops. It forms the headwater catchment of the Mmone dry valley system, which is characterized by several tributaries with a width up to several kilometers and a maximum depth of tens of meters. This paleo-stream system forms a marked topography in this otherwise flat area. The Kalahari sand cover reaches a thickness of 10–40 m, with extensive horizons of calcrete and silcrete precipitates along (former) drainage lines and shallow pan depressions. The topographic slope is in a northwesterly direction with a gradient of about 0.2%. These huge valleys with small catchments and gorge type valley heads are most likely initiated by subsurface leaching of the felspathic matrix along fractures, producing pseudo-karstic features, followed by groundwater sapping erosion during wetter periods with higher groundwater tables (Shaw and de Vries, 1988). The relatively deep incision is caused by gradual uplift along the Zimbabwe–Kalahari Axis during the Tertiary. The present groundwater table in the L–B area is at a depth of about 50 m and more or less reflects the topography. Fig. 3 shows the dry valley system and the piezometric head contours.

The main aquifer is formed by the Mmabula Sandstone which subcrops the Kalahari sand in the southeast. Toward the northwest the Mmabula aquifer dips underneath less pervious mudstone and basalt from the Dibete Formation and the Stormberg Lava Group. The unconfined part forms the main recharge area, as indicated by the groundwater head culmination in the southeast, and by an increasing ^{14}C and ^{4}He groundwater age in a northwesterly direction; from hundreds of years in the unconfined area to more than 10,000 of years at the western fringe of the study area (Beekman et al., 1997; Selaolo, 1998).

The recharge process was studied by sampling soil moisture in vertical profiles for the environmental tracers, chloride, tritium, ^{18}O and deuterium. (A full account on the applied methods and the results are

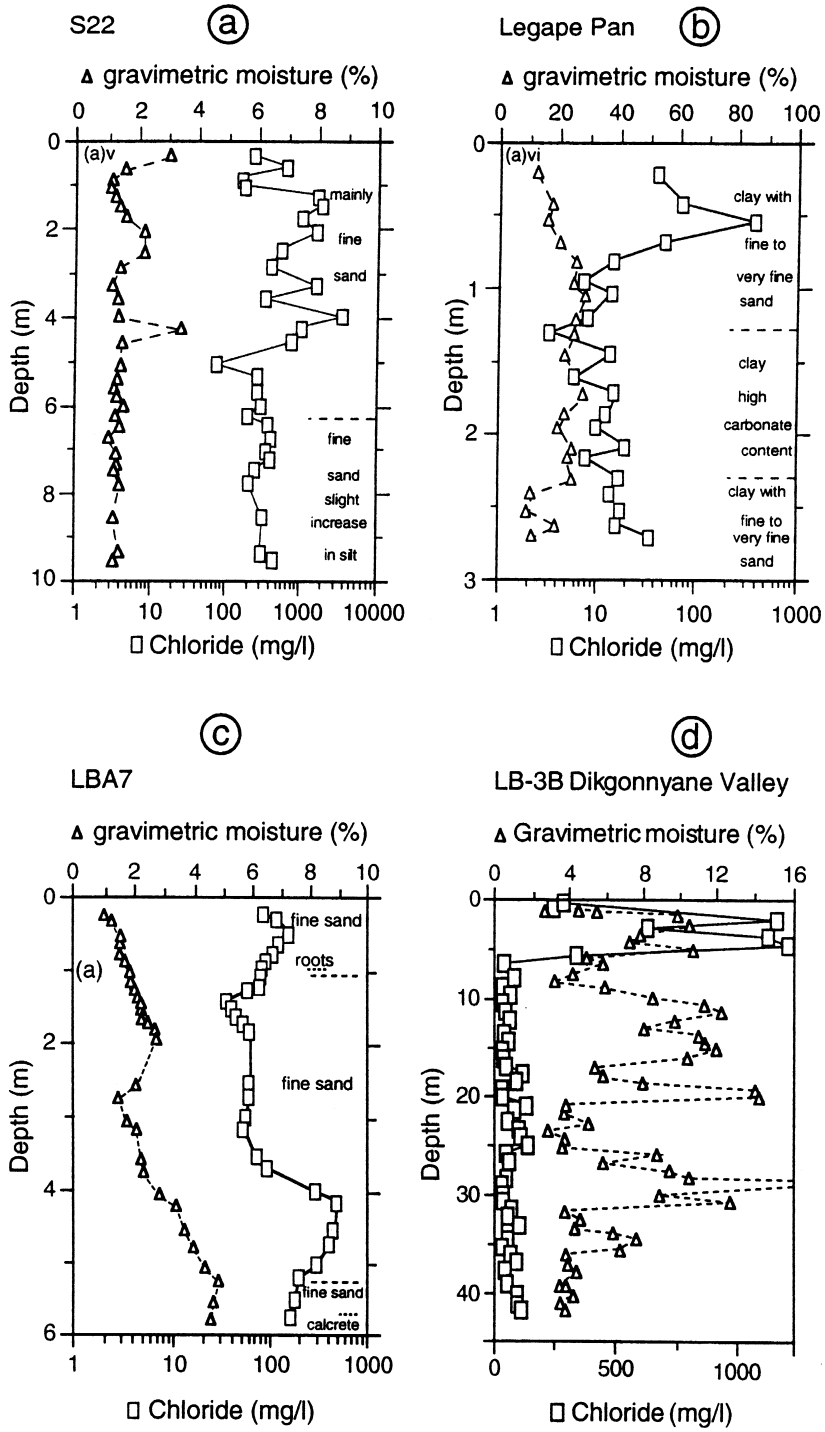

Fig. 2. Examples of vertical soil moisture and chloride profiles for the Letlhakeng-Botlhapatlou area (for locations see Fig. 3), for two level areas (a, c), a pan (b) and a dry valley (d); average chloride content at Dikgonnyane is 65 ppm. (Redrawn from Beekman et al., 1997; Selaolo, 1998.)

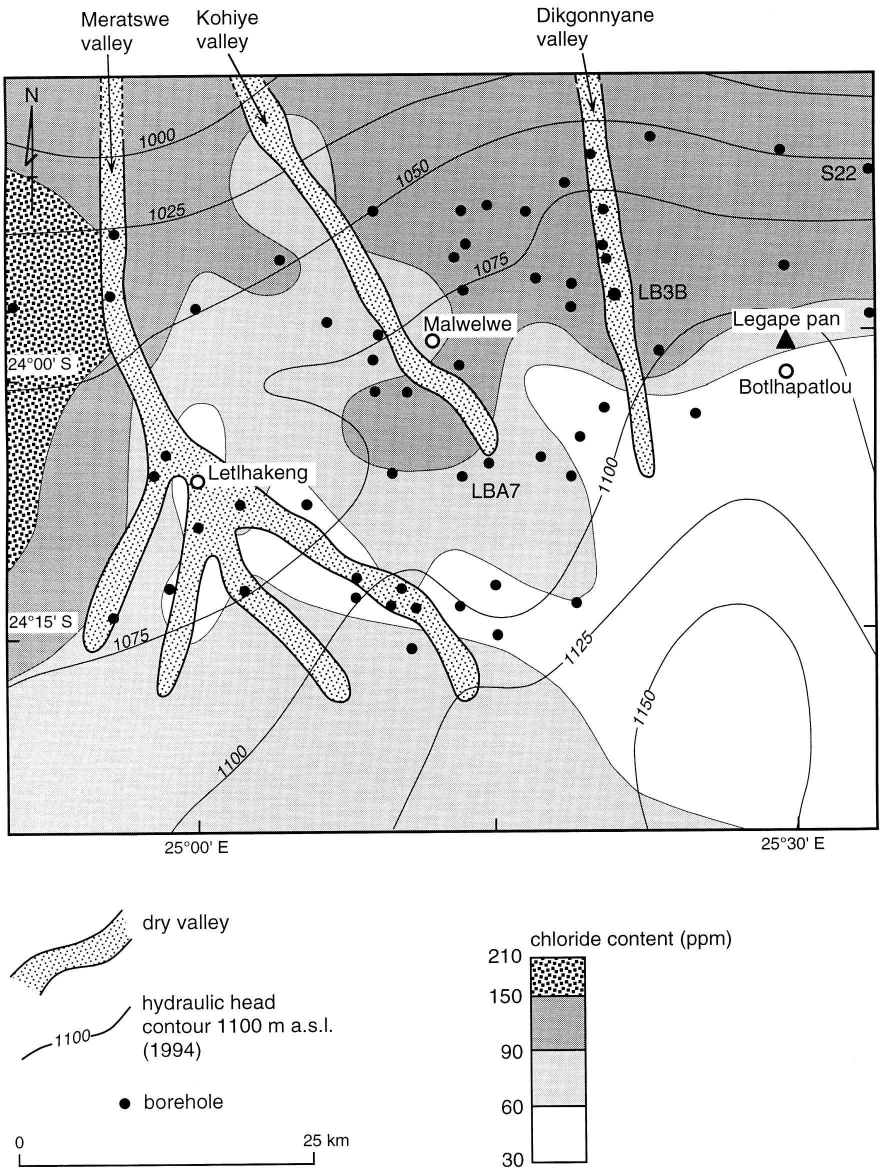

Fig. 3. Chloride and hydraulic head distribution in the Letlhakeng–Botlhapatlou area; for location see Fig. 1. (Compiled from Nijsten and Beekman, 1997; Beekman et al., 1997.)

given by Selaolo, 1998.) A compilation of results is given in Fig. 2, showing representative chloride profiles for two flat areas, a pan and a dry valley. A chloride mass balance for steady state conditions in the soil moisture zone means that the total input of chloride by wet and dry atmospheric deposition should equal chloride output by transport through the unsaturated zone, assuming that the Kalahari beds itself do not produce any chloride. For diffuse infiltration the following equation applies (see, e.g.

Lerner et al., 1990):

$$\mathrm{TD} = V_{\mathrm{sm}} \times \mathrm{Cl}_{\mathrm{sm}} \qquad (1)$$

where TD is the total chloride deposition at the surface (mg m^{-2} yr^{-1}), V_{sm} the downward soil moisture flux through the unsaturated zone (mm yr^{-1} or l m^{-2} yr^{-1}) and $\mathrm{Cl}_{\mathrm{sm}}$ is the chloride concentration of soil moisture (mg l^{-1}). From a chloride deposition observation network, with long-term regional monitoring and intensive local observations during the GRES II period, a TD value between 400 and 600 mg m^{-2} yr^{-1} was determined, with 500 mg m^{-2} yr^{-1} applied henceforth as the most probable average for the study area (Beekman et al., 1996). $\mathrm{Cl}_{\mathrm{sm}}$ varies for level areas from 50 to 500 mg l^{-1}, so that diffuse transport through the unsaturated zone ranges between 1 and 10 mm yr^{-1}, with an areal average of about 3 mm yr^{-1} (Fig. 2a and c). Higher values are found in depressions (pans and dry valleys) where water accumulates at the surface during high rainfall events (Fig. 2b and d). The chloride content often decreases and stabilizes below the root zone, indicating that part of the infiltration bypasses this zone of strong evapotranspiration via preferential flow paths, formed by cracks, root channels and funneling through unstable wetting fronts. Therefore the chloride balance calculations were based on the chloride concentrations below the root zone where diffuse percolation prevails.

Average chloride content in the saturated zone in the recharge area is about 75 ppm (Fig. 3). Applying Eq. (1) for the saturated zone, with V and Cl representing the groundwater flux and groundwater chloride content, leads to a total areal discharge (including diffuse and concentrated components) in the order of 7 mm yr^{-1}. For long-term steady state conditions, discharge equals recharge. This thus means that total recharge is about twice the average diffuse percolation through the unsaturated zone. A preferential and concentrated flow component that moves directly to the water table is therefore likely. This rapid concentrated percolation most probably originates from surface water accumulation in pans and dry valleys and fast percolation through fractured calcrete surfaces. The Legape Pan, for example, showed a chloride concentration of 10 ppm (Fig. 2b), suggesting a locally enhanced annual recharge of 50 mm. The 42 m deep borehole in the Dikgonnyane dry valley exhibits an intermediate position with an average chloride concentration of 65 ppm; this means an annual recharge of 8 mm (Fig. 2d). A tritium content of 4–5 TU in the samples from depths of 39, 40 and 42 m in this borehole, supports the concept of a rapid preferential flow component, so that the total recharge through this valley is more than 8 mm yr^{-1} (Selaolo, 1998). The total chloride content of this profile is 225 g m^{-2}; with an annual chloride deposition of 500 mg m^{-2}, this would indicate an accumulation period of 450 yr in case of a piston like flow. The tritium content however, indicates younger water, which can only be explained by a fast percolation component.

Hydraulic head contours (Fig. 3) indicate a hydraulic gradient of 0.2%, which combined with an average transmissivity of 500 m^2 day^{-1} (Nijsten and Beekman, 1997) accounts for a groundwater flux in the order of 1 m^2 day^{-1}. Averaging this flow rate over the recharge area, which shows a maximum length of about 50 km perpendicular to the head contours, gives an overall minimum recharge of 7 mm yr^{-1}. A more sophisticated and distributed finite element groundwater flow model produced an average figure of 6 mm yr^{-1} (Nijsten and Beekman, 1997). These results are thus comparable with those from the chloride mass balance approach. Calculation of residence time by combined geochemical and ^{13}C, ^{14}C and ^{3}He/^{4}He — isotope modeling however, produced a much higher maximum age of groundwater in downstream direction than the groundwater flow model: up to 10,000 yr at a distance of 50 km from the intake area, resulting in average long-term annual recharge figures of between 1 and 3 mm (Beekman et al., 1997). These high ages might be caused by an inappropriate assessment of the interaction between water and the mineral phase and/or the presence of stagnating water pockets. Taking into account the rather consistent results of the chloride balance and the groundwater flow calculation, an average recharge estimate of 6 mm yr^{-1} is applied henceforth as the most likely value for the time being.

According to the finite element model (Nijsten and Beekman, 1997), part of the groundwater from the study area discharges downstream in wide and deeply incised fossil dry valley depressions, where the groundwater table comes, according to borehole observations, within a depth of 15–30 m below the floor. A concentration of deeply rooting *Acacia* tree

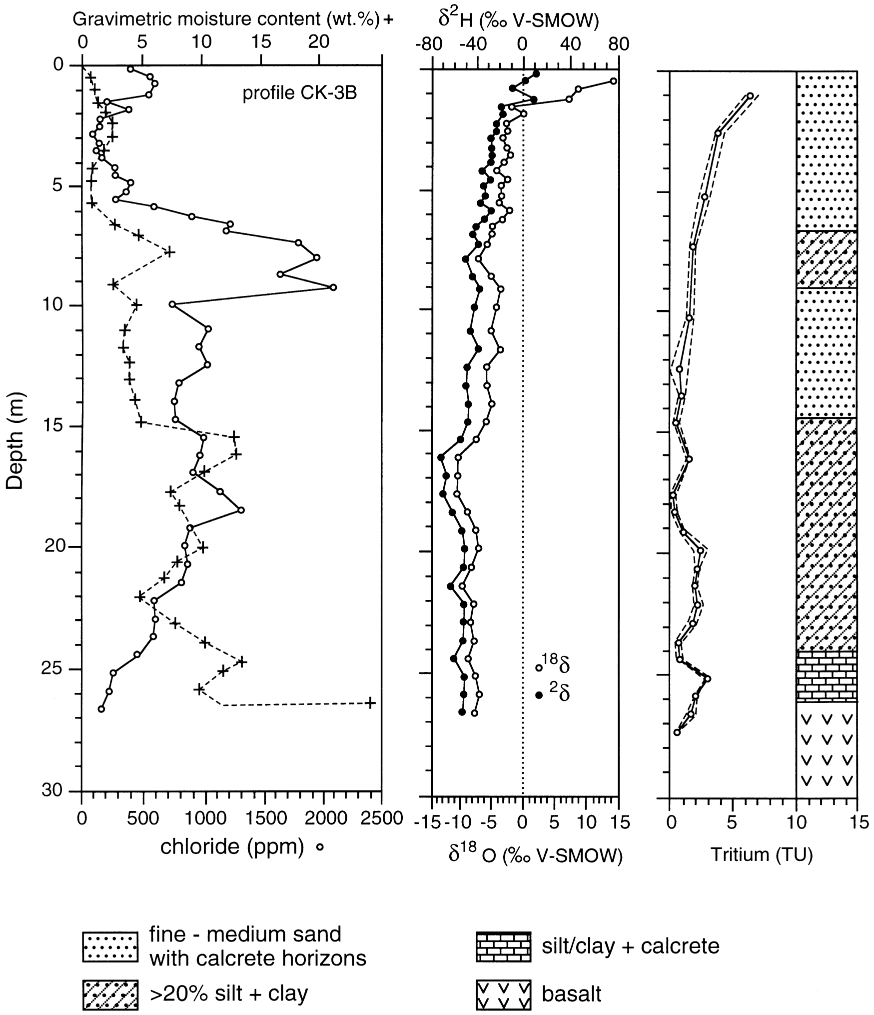

Fig. 4. Central Kalahari deep profile; ^{18}O and deuterium are expressed in mil deviation (δ) from the V-MSOW standard; for location see Fig. 1. (Redrawn from Beekman et al., 1997; Selaolo, 1998.)

species, known for possible root depths of more than 50 m (Ringrose et al., 1997; Selaolo, 1998), indicates that abstraction by root uptake probably occurs. This hypothesis is supported by the preservation of green leaves, as a testimony of continuous transpiration, on several of the larger trees at the end of the dry season when almost all trees and shrubs in the surrounding area have a grey–brown appearance. An increase in salt concentration and precipitates of calcrete give additional evidence of a local loss of groundwater by evapotranspiration, which according to the groundwater model is in the order of 20–30 mm yr^{-1} in these depressions. It is very likely that these areas of discharge did form real groundwater outcrops (springs) during wet climatic periods with higher groundwater tables. The valleys evidently form a surface expression of deeply weathered fractures with concentrated groundwater flow and seepage (Shaw and De Vries, 1988).

2.2. The Central Kalahari (CK-area, Figs. 1 and 4)

The observed general increase in chloride concentration in a downstream direction (Fig. 3) indicates a general decrease in recharge due to a decrease in rainfall in a westerly direction as well as the occurrence of mudstone and basalt confining layers overlying the sandstone aquifer. The latter will result in perched water horizons, facilitating evapotranspiration by deeply rooting trees. Further to the northwest, in the Central Kalahari at 120 km from the L–B-study area (Fig. 1), a 28 m deep profile was core-sampled for moisture, chloride, oxygen-18, deuterium and tritium (Fig. 4). The groundwater table at this site is at 70 m, and groundwater samples were taken from two boreholes with a depth of 150 m. The upper 23 m consists of Kalahari sand and calcrete with macropores, underlain by 24 m of partly weathered Stormberg basalt, followed by fine-grained massive Ntane Sandstone with traces of dissolution features.

The chloride peak between 6 and 9 m is in loamy sediments and probably reflects an accumulation of salt in the small pores, whereas the decrease in chloride below this zone indicates preferential percolation through the larger pores. A chloride mass balance calculation for this lower zone with more or less constant chloride content of 900 ppm, gives an average annual flux of 0.6 mm. Below 20 m the chloride content further reduces, probably due to preferential percolation through larger cracks. The tritium peaks of more than 3 TU at this depth also suggests a rapid flow component. The downward decrease of ^{18}O and deuterium isotopes in the upper 15 m is explained by evaporation enrichment near the surface and subsequent dilution by downward percolation of lighter rainwater (Allison et al., 1984).

Groundwater samples from the saturated zone gave a chloride content of 360 ppm, suggesting an overall (diffuse and concentrated) recharge of 1.2 mm yr^{-1}. The chloride content in the saturated zone, however, might have been affected by groundwater from the upstream area with a lower Cl-concentration, so that the average overall recharge is probably below 1.2 mm.

3. Stable isotopes and tritium

The groundwater recharge figures presented in this paper are based mainly on the chloride balance method and groundwater flow modeling. Another

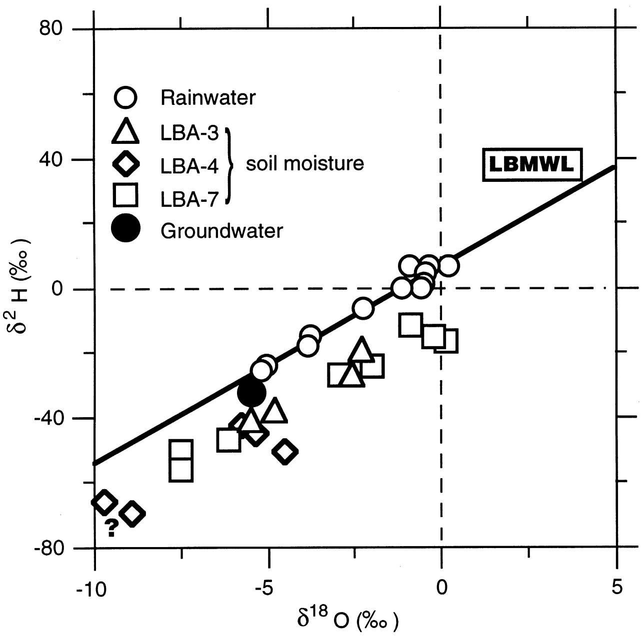

Fig. 5. Relation between δ^2H and δ^{18}O for rainwater (59 samples), soil moisture (108 samples) and groundwater (59 samples) from the Letlhakeng-Botlhapatlou area (symbols represent data clusters); rainwater in monthly weighted values at Malwelwe (period 1993–1995); LBMWL = L–B area Meteoric Water Line. (Redrawn from Selaolo, 1998; original data in Beekman and Selaolo, 1997.)

aspect of the GRES project was the use of environmental isotopes for dating and tracing. Apart from the earlier mentioned dating and hydrochemical process studies in the saturated zone, seven ^{18}O, deuterium and tritium profiles were established in the unsaturated zone; see for example CK-3B in Fig. 4. All ^{18}O and deuterim profiles show the same trend as CK-3B, with an evaporation enrichment near the surface, decreasing to lower values at greater depths. (A full account of the other profiles is given in Selaolo, 1998.)

Stable isotope figures for rainfall, soil moisture and groundwater for the L–B area are summarized in Fig. 5. The average monthly weighted rain values show $\delta^{18}O$ figures between 0 and −5‰ and δ^2H figures between 0 and −35‰ (59 samples from the period 1993–1995). The most depleted values originate from months with the highest rainfall amounts; individual high-intensity storms hardly exceeded these highest monthly depletion values. Soil moisture samples show $\delta^{18}O$ values ranging from 0 to −10‰, and δ^2H values between −20 and −65‰. It is remarkable that the highest depletion values of soil moisture exceed the maximum values of individual high-intensity storm events during the period 1993–1995. An explanation could be the occurrence of more depleted rainstorms in the past.

Groundwater in the saturated zone shows a remarkable uniform stable isotope content: $\delta^{18}O = -5.22‰ \pm 0.21$ and $\delta^2H = -32.8‰ \pm 2.0$. This means that the isotopic character of groundwater closely reflects the average isotopic condition of soil moisture. This also suggests that the 1993–1995 sampling period is not representative for a long period isotopic situation.

Recharge estimations for the Kalahari using the semi-empirical heavy-isotope dilution method proposed for Australia by Allison et al. (1984), gave figures of the same order of magnitude as those obtained from the chloride mass balance calculation (Selaolo, 1998). Less successful were recharge determinations with tritium decay modeling. In accordance with findings of others in areas with low recharge, the calculated recharge figures were much too high (Beekman et al., 1996). This failure is attributed to the relatively high influence of vapor transport on the distribution of tritium (Cook and Walker, 1995).

4. Water and vapor transport from below the root zone

From the foregoing it is evident that almost all rainwater disappears from the root zone in the upper meters by evapotranspiration. Less than 1% of the rainfall on average percolates to greater depths, and will either reach the saturated zone or can locally be extracted by deep rooting *Acacias*. Recent analysis of stable isotope profiles suggests that ascending capillary transport in bare soils under desert conditions, is detectible from water table depths up to 20 m. The flux from such depths is in the order of 1 mm yr^{-1} (Coudrain-Ribstein et al., 1998). For Kalahari conditions, this could form an additional mechanism for the loss of groundwater from local high (perched) water tables in the vicinity of pans and dry valleys.

In earlier discussions on Kalahari groundwater recharge (see Section 1), researchers have speculated about the possibility of a substantial upward transport of water from below the root zone by vapor transport, due to a temperature gradient. This then could possibly explain the discrepancy between the indications for substantial recharge or percolation through the unsaturated zone below the root zone and the low regional groundwater discharge flux. To test this hypothesis, seasonal temperature measurements to a depth of 7 m were carried out during the GRES project (Fig. 6). The maximum temperature difference below the root zone, between 3 m and 7 m, is 4°C, with respective average temperatures of 26.5 and 23.5°C during the summer, and 21 and 24°C during

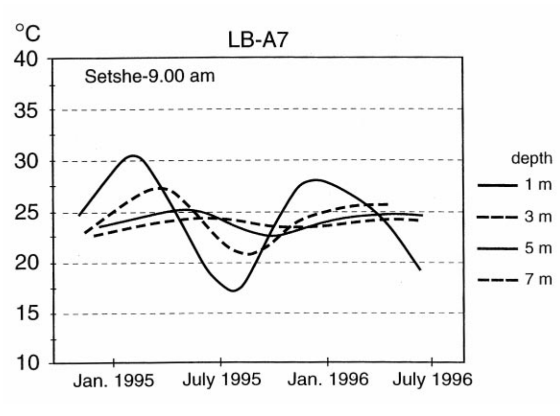

Fig. 6. Seasonal temperature fluctuations to a depth of 7 m for the period 1995–1996; location near LBA-7 (see Fig. 3).

the winter period. This indicates that the average difference in saturated vapor pressure in summer is about 3.8×10^{-6} g cm^{-3} and during winter 2.3×10^{-6} g cm^{-3} (see, e.g. Hanks and Ashcroft, 1980). The downward vapor transport component during the summer and the upward flow component during the winter thus seem to be in the same order of magnitude. Steady-state vapor movement is given by:

$$J = -D\, \mathrm{d}\rho/\mathrm{d}s \tag{2}$$

where J is the vapor flux (g cm^{-2} s^{-1}), D the diffusion coefficient (cm^2 s^{-1}), ρ the vapor density (g cm^{-3}) and s is distance in cm; D is in the order of 0.2 cm^2 s^{-1} (Hanks and Ashcroft, 1980). According to Eq. (2) upward flux during the winter period (150 days) is in the order of 0.2 mm, whereas downward flux during the summer (150 days) is about 0.3 mm. Net transport in the vapor phase is therefore probably low and seems not to play an important role in soil moisture loss from below the root zone. Moreover, if it did, it would have increased the chloride concentration and would thus have been taken into account in the net recharge calculation from the chloride balance. The same applies for the previous mentioned loss of water from greater depths by capillary suction.

Transport in the liquid as well as the vapor phase thus cannot explain a possible discrepancy between a substantial recharge, as determined for the L–B area, and the low regional groundwater discharge flux of less than 1 mm yr^{-1} as calculated by de Vries (1984) for the central and eastern Kalahari. It is evident that the low recharge rate that was determined during the present study for the central Kalahari (CK area), exemplifies the condition of most of the Kalahari, and thus explains the low regional groundwater flux.

5. Drying out of the valley system

The lapse of time since the dry valley system in this area lost its groundwater drainage function, is estimated to evaluate a possible residual nature of the hydraulic gradient in the L–B area. At the beginning of the depletion, the groundwater table must have been above the valley floor. This means a hydraulic head at the divide which was about 50 m higher than at present, or an elevation of 1200 m above m.s.l. (Fig. 3). It is plausible that during this wet period the hydraulic head further to the west, at say 100 km from the L–B study area towards the center of the Kalahari, was not much higher than at present, because the present head at that distance is about 950 m (Selaolo, 1987) which is close to the drainage base of the Makgadikgadi Pan (Fig. 1). To estimate the time needed to produce a decay of the groundwater table at the divide from 1200 to 1150 m due to a decrease in recharge from the wet period to the present-day annual recharge of 6 mm at this fringing area, the following linear-reservoir depletion approach can be applied.

For continuity:

$$S\, \mathrm{d}h = (N - q)\, \mathrm{d}t \tag{3}$$

where S is the specific yield; h the average head above the discharge base (L); N the recharge (LT^{-1}); q the discharge (LT^{-1}).

Schematizing the groundwater discharge as horizontal flow to a parallel drainage system (Dupuit assumptions) lead to a parabolic groundwater table, thus:

$$h = 2/3(h_\mathrm{m}) \tag{4}$$

where h_m is hydraulic head at the divide. The specific drainage resistance R (T) is:

$$R = h_\mathrm{m}/q, \text{ thus } \mathrm{d}h_\mathrm{m} = R\mathrm{d}q \tag{5}$$

The parameter R is a function of the length of the considered flow path and the transmissivity. Substituting Eqs. (4) and (5) in Eq. (3) gives:

$$2/3(SR)\, \mathrm{d}q = (N - q)\, \mathrm{d}t \tag{6}$$

or:

$$\mathrm{d}q = j(N - q)\mathrm{d}t \tag{7}$$

where

$$j = 1.5(S.R)^{-1} \tag{8}$$

j is the reaction factor or reservoir outflow recession constant (T^{-1}).

Integrating Eq. (6) with boundary conditions $q = q_0 : t = 0$ and $q = q_t : t = t$ with $N =$ constant, gives:

$$q_t = q_0 \mathrm{e}^{-jt} + (1 - \mathrm{e}^{-jt})N \tag{9}$$

And (recall Eq. (5)):

$$h_t = h_0 e^{-jt} + (1 - e^{-jt})N.R \quad (10)$$

where h_t is present head above the drainage base at the divide: $1150 - 950 = 200$ m; h_0 is height at the divide above drainage base at the beginning of depletion $(t = 0) : 1200 - 950 = 250$ m.

Since the discharge calculated from the groundwater model accords reasonably with recharge determined by the chloride mass balance (see Section 2), it is concluded that the present conditions are close to steady-state. Substitution of the present recharge, $N = 6$ mm/yr, and the present hydraulic head, $h_t = 200$ m, in Eq. (5) gives $R = 33,000$ yr. Assuming a specific yield of 0.15 for the median coarse Kalahari sand and substitution of this value with R in Eq. (8), gives $j = 3 \times 10^{-4}$ yr^{-1}. Substituting these values in Eq. (10) gives a decay in groundwater head of 50% every 2500 yr. This means that a lowering of the groundwater table with 50 m required a period of at least thousands of years in the case of a recharge of 6 mm/yr during the depletion period. The calculated time lapse is reduced if the recharge during the depletion period would have been lower than at present, and is less than 1000 yr for zero-recharge. However, such a low recharge figure for the depletion phase is not likely because in that case the present groundwater-chloride concentration would have been higher.

The reconstruction is in accordance with the paleo-climatic information from other sources which suggests the end of a major wet period at about 4500 BP (see Section 1). In other words: this analysis makes plausible that the present hydraulic gradient is in steady state if indeed the last major wet period ended some thousands of years ago. A similar approach by De Vries (1984) for the western and central Kalahari resulted in a required time lapse of more than 10,000 yr under conditions of zero-recharge, to reach the present groundwater level. This suggests that only the eastern fringe experienced wet conditions with a high groundwater table during the Holocene. The Kalahari proper seems to have had its last major wet-phase during the Pleistocene pluvial period, when also the Makgadikgadi Pan was full of water.

6. Summary and conclusions

In conclusion, the groundwater recharge studies resulted in an average figure in the order of 5 mm yr^{-1} for the fringe of the Kalahari, decreasing to 1 mm or less for the central part. Since no clear changes in morphological conditions occur, it is likely that the decrease in a recharge reflects the rainfall pattern, which decreases from about 450 mm at the edge of the Kalahari to less than 350 mm in the center. It is remarkable that also in South Africa under different geological and morphological conditions, groundwater recharge seems to become very low or negligible below an average annual rainfall threshold of 400 mm (Bredenkamp et al., 1995).

Part of the relatively high recharge in the fringe area will be depleted into dry valley depressions through transpiration by deep rooting *Acacias* from depths of tens of meters. In addition, water losses from greater depths (up to 20 m) by capillary transport may occur from perched water tables in areas of enhanced recharge, such as depressions and fractured duricrust surfaces, in the fringe as well as in the central parts of the Kalahari. An average regional groundwater discharge flow of less than 1 mm yr^{-1} for the whole of the Kalahari, as suggested by the regional hydraulic gradient of Fig. 1 and earlier argued by De Vries (1984), seems therefore quite likely, and in accordance with the results of the present tracer studies in the central Kalahari. This means that the present regional hydraulic gradient is more or less in steady state and that the depth of the groundwater table is the result of a head decay over a period of more than 10,000 yr. This suggests that the late-Pleistocene pluvial was the last major wet period that affected the whole of the Kalahari (first order climatic effect).

The fossil dry valley system at the Kalahari fringe refers to more humid paleo-climatic conditions with higher recharge and higher groundwater tables in this area during the Holocene. A tentative reconstruction of the water table decline suggests a time lapse of several thousands of years since the end of these wet conditions (second order climatic effect). Restoration of the previous high groundwater table up to at least the depth of the dry valley floor, would require a rise in the groundwater head of about 25%, which will result in an increase in saturated thickness of the aquifer by about 50 m. With an average hydraulic conductivity of about 10 m day^{-1} for Kalahari sand, this means a doubling of the total transmissivity. To establish and maintain this groundwater table would thus require an increase in annual recharge to about 10–15 mm at the fringe of the Kalahari for a period of thousands of years. Recharge figures of 10–15 mm in sandy areas are found in eastern Botswana in areas with a rainfall of around 500 mm (Gieske, 1992; Beekman et al., 1996). To restore the wet conditions would therefore probably require an annual rainfall increase in the Kalahari in the order of 100 mm.

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