Estimate of Recharge of a Rising Water Table in Semiarid Niger from ^{3}H and ^{14}C Modeling

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Abstract

A hydrodynamic survey carried out in semiarid southwest Niger revealed an increase in the unconfined ground water reserves of approximately 10% over the last 50 years due to the clearing of native vegetation. Isotopic samplings (^{3}H, ^{18}O, ^{2}H for water and ^{14}C, ^{13}C for the dissolved inorganic carbon) were performed on about 3500 km^2 of this silty aquifer to characterize recharge. Stable isotope analyses confirmed the indirect recharge process that had already been shown by hydrodynamic surveys and suggested the tracers are exclusively of atmospheric origin. An analytical model that takes into account the long-term rise in the water table was used to interpret ^{3}H and ^{14}C contents in ground water. The natural, preclearing median annual renewal rate (i.e., recharge as a fraction of the saturated aquifer volume) lies between 0.04% and 0.06%. For representative characteristics of the aquifer (30 m of saturated thickness, porosity between 10% and 25%), this implies a recharge of between 1 and 5 mm/year, which is much lower than the estimates of 20 to 50 mm/year for recent years, obtained using hydrological and hydrodynamic methods and the same aquifer parameters. Our study, therefore, reveals that land clearing in semiarid Niger increased ground water recharge by about one order of magnitude.

Introduction

In the last decade, hydrodynamic characteristics of the unconfined aquifer in southwest Niger were investigated in detail over an area of 8000 km^2 (Leduc et al. 1997; Leduc et al. 2001). More than 15,000 level measurements at 250 sites showed an unexpected but regular median rise in the water table of 0.20 m/year since 1991. Further into the past, reanalysis of 100 older measurements revealed that present levels are the highest ever recorded: Since the 1950s, the mean increase is 3 m, i.e., approximately 10% of the ground water reserves. This phenomenon is explained by the intense clearance of native vegetation over recent decades, which increased surface runoff. As runoff concentrates in temporary ponds and then infiltrates the water table, higher runoff implies higher ground water recharge. Hydrodynamic data resulted in an estimate for ground water recharge that presently exceeds 20 mm/year. However, the long-term rise in the water table following the change in land use indicates that recharge must have been lower in the past. The objective of this paper is to use isotopic methods to estimate the long-term, preclearing recharge in a representative region of 3500 km^2. Stable (^{18}O, ^{2}H, ^{13}C) and radio (^{3}H, ^{14}C) isotopes of ground water were interpreted by taking into account the long-term rise in the water table. The recharge estimate was then compared with present calculations.

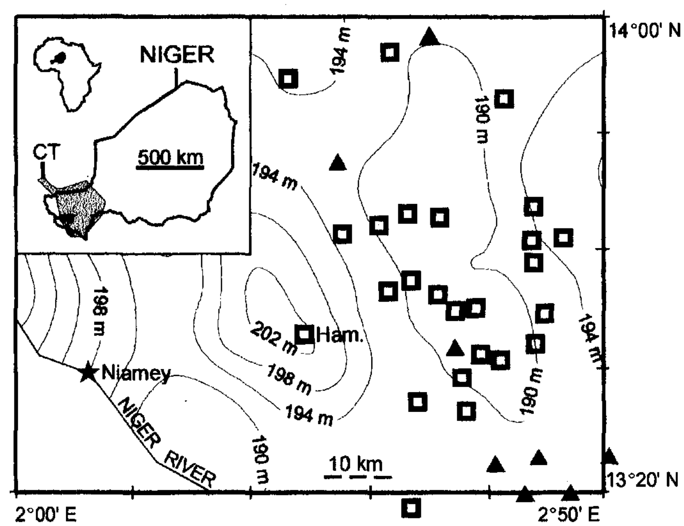

Figure 1. General map of Niger showing the Continental Terminal (CT) outcrops (small inset), and potentiometric map of the water table in 1998 (in meters above mean sea level) with the sample locations (□: well, ▲: borehole). ★Niamey; Ham.: Hamdallay.

Geochemical methods are widely used in arid and semiarid areas to provide recharge estimates. Among others, Allison et al. (1990) in southeast Australia, Cook et al. (1992) in north Senegal, and Selaolo (1998) in the Botswana Kalahari used tracers in the

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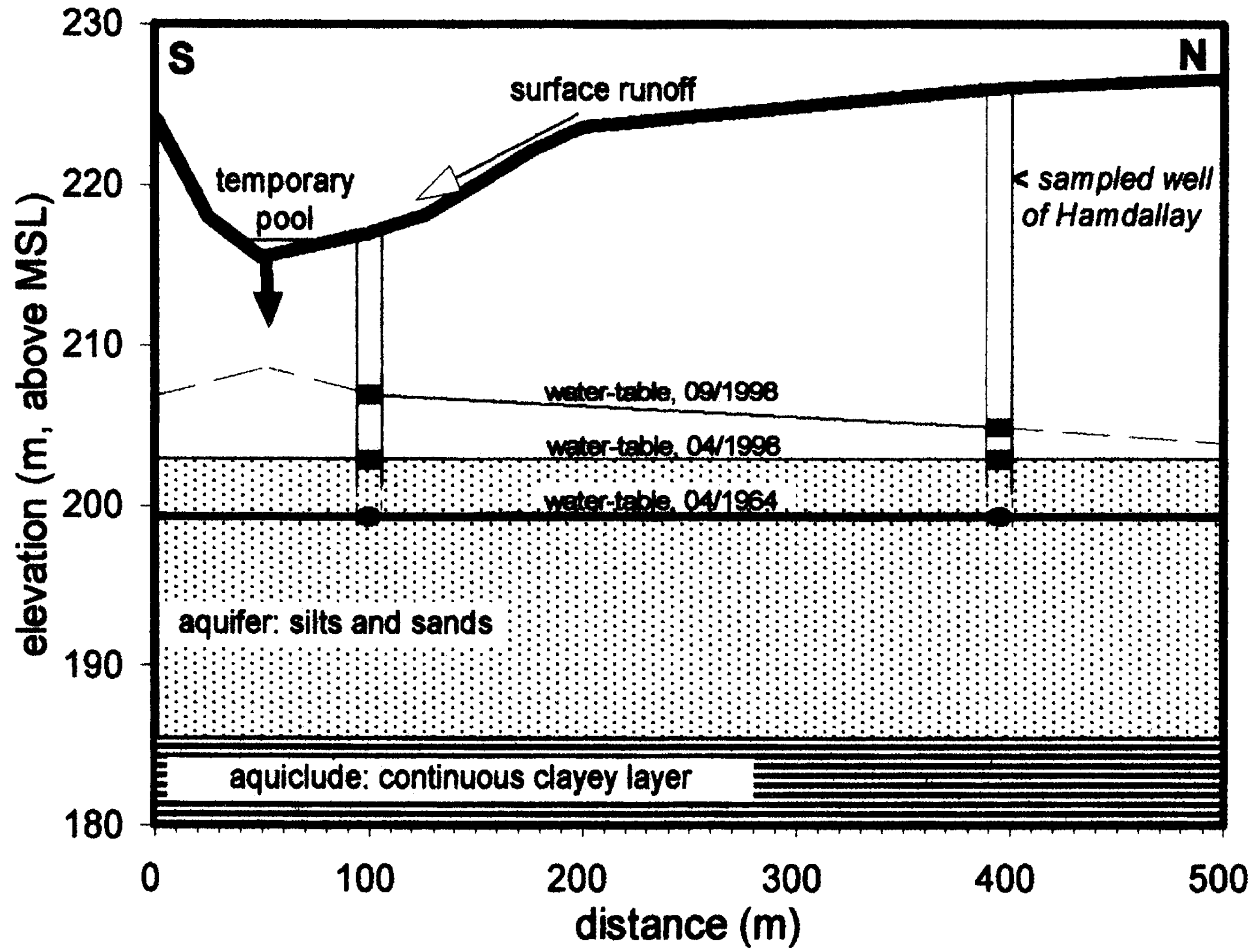

Figure 2. Cross section (south to north) at Hamdallay (located in Figure 1) showing the transient potentiometric mound of the water table below the temporary pond before (April 1998) and after (September 1998) the rainy season. The water table elevation measured in 1964 is also displayed.

unsaturated zone to characterize diffuse recharge. In other places, geochemical data of ground water were used to estimate a more integrated recharge in time and space (Fontes et al. 1991; Wood and Sanford 1995; Sultan et al. 2000). In regions where recharge processes are indirect, this approach is particularly relevant, as correct interpretation of unsaturated zone profiles often requires multiple techniques (Scanlon et al. 1999). In semiarid areas, studies dedicated to the quantification of the increase in recharge following a change in land use are rare, and most of them concern Australia (Allison et al. 1990). Our study provides one of the first estimates in semiarid Africa.

Hydrogeologic Setting

The study area covers approximately 3500 km^2 of the Continental Terminal, a formation that extends over a surface area of 150,000 km^2 in southwest Niger (Figure 1). The sediments are made up of continental clays (kaolinite), silts, and sands, and date from Eocene to Pliocene. A continuous and impervious clayey layer about 10 m thick constitutes the substratum of the unconfined aquifer. The climate is semiarid with average rainfall of 567 mm (from 1905 to 1998, Niamey, Niger), an average annual temperature of 29°C, and a potential evapotranspiration rate of 2500 mm/year. During the rainy season from June to September, surface runoff concentrates in temporary ponds, natural outlets of closed basins on the order of 1 km^2. In this environment, all hydrological data indicate that most of the aquifer recharge is indirect and occurs under the ponds (Desconnets et al. 1997). Volumes infiltrated from a pond to the aquifer are frequently 10,000 m^3/year, but can reach 100,000 m^3/year in some places or during the rainiest years. In most cases, infiltration takes place over a few hours or days. Consequently, transient water table mounds up to meters of amplitude are created below infiltrating ponds (Leduc et al. 1997; Figure 2). Elsewhere in the landscape, there seems to be no infiltration below 4 m (Peugeot et al. 1997).

The depth to the water table varies between 75 m below the lateritic plateaux and less than 10 m below the dry valleys. The saturated thickness of the aquifer ranges from 10 to 60 m (mean value of 30 m). At the study scale, the water table displays low gradients (< 1 ‰). A closed potentiometric depression exists in the study area (shown by the 190 m water table elevation contour in Figure 1); this natural anomaly, which is often observed in the Sahel south of the Sahara, can be explained in terms of small fluxes with evapotranspiration exceeding infiltration in the center (Aranyossy and Ndiaye 1993). However, this natural discharge is probably weak, considering the relatively important depth to the water table in the study area. In the absence of irrigation, anthropic withdrawals are limited: Current

Table 1
Sample Location and Isotopic Results

Long./Lat. 2° E/13° N	Local Name	Water Table/Aquiclude (Screen Depths) in m	Sampling Date	$\delta^{18}O$ ‰ VSMOW	$\delta^{2}H$	$\delta^{13}C$ DIC ‰ VPDB	$\delta^{13}C$g. eq.	^{14}C DIC pmC	^{3}H TU
				Wells					
35'00 / 27'74	Balal Sagui	–69/–90 (–69/–71)	04/98	–4.6	–30	–16.0	–16.0	85.5 ± 0.9	< 2.2
37'49 / 35'17	Banikane	–27/–50 (–28/–31)	08/97	–4.6	–28	–17.6	–16.6	91.3 ± 0.8	3.5 ± 0.7
44'68 / 33'72	Bani Koubey	–56/–95 (–56/–58)	04/98	–5.2	–33	–17.0	–16.9	84.1 ± 0.8	< 2.3
39'58 / 31'72	Banizoumbou	–20/–45 (–21/–24)	02/95	–4.0	–26	–14.2	–13.7	90.4 ± 0.7	ND
27'73 / 41'81	Birni Kol.	–45/–65 (–47/–49)	11/93	–5.1	–31	–15.6	–18.5	74.1 ± 0.5	< 2.0
32'10 / 42'85	Birnin L.	–60/–80 (–60/–62)	07/97	–5.3	–33	–16.9	–18.0	81.3 ± 0.5	< 1.3
44'11 / 31'75	Ciminti K.	–58/–100 (–59/–61)	04/98	–5.0	–32	–17.5	–17.5	73.2 ± 0.6	< 1.6
36'04 / 36'70	Dantiandou T.	–39/–60 (–39/–42)	08/97	–4.7	–30	–15.4	–17.0	86.6 ± 1.2	< 1.3
43'83 / 40'49	Datché	–45/–90 (–45/–47)	04/98	–5.7	–36	–17.5	–19.1	83.4 ± 0.8	< 1.8
31'64 / 36'95	Goguieizé K.	–47/–70 (–48/–50)	11/93	–5.0	ND	–13.8	–14.8	85.1 ± 1.1	< 2.0
24'47 / 33'47	Hamdallay	–23/–45 (–25/–28)	02/95	–4.6	–28	ND	ND	89.5 ± 1.0	1.4 ± 0.5
43'98 / 42'85	Kafina	–49/–100 (–51/–54)	09/97	–4.4	–31	–17.8	–17.3	82.6 ± 0.7	< 1.2
38'80 / 26'74	Kampa Zarma	–65/–95 (–66/–68)	02/95	–3.9	ND	–14.7	–14.8	88.5 ± 0.7	1.2 ± 0.5
40'87 / 53'10	Kirib Béri	–49/–75 (–51/–53)	02/95	–5.5	ND	–15.9	–16.7	78.6 ± 0.7	1.6 ± 0.4
36'10 / 43'22	Kolo Diogono	–27/–60 (–29/–31)	07/97	–4.1	–29	–13.4	–15.5	94.4 ± 1.4	1.4 ± 0.8
37'89 / 29'81	Koma Koukou	–54/–80 (–55/–57)	04/98	–4.6	–29	–16.2	–16.2	60.4 ± 0.5	<1.0
34'59 / 43'49	Kouabiri K.	–38/–65 (–38/–39)	07/97	–4.8	–32	–16.4	–16.9	ND	< 1.9
33'19 / 37'31	Kountché T.	–69/–90 (–69/–71)	04/98	–5.2	–34	–17.1	–18.5	76.0 ± 0.6	< 1.6
34'59 / 18'73	Kouré	–34/–60 (–35/–37)	11/93	–4.6	ND	–13.1	–14.1	71.4 ± 0.6	< 2.0
39'04 / 35'44	Maourey K. Z.	–20/–45 (–21/–24)	08/97	–5.2	–31	–15.9	–17.5	85.2 ± 0.9	2.0 ± 1.0
43'88 / 41'30	Niné Founo	–34/–85 (–37/–40)	11/93	–5.1	–33	–9.1	–10.6	97.1 ± 0.9	4.0 ± 1.0
31'74 / 57'20	Téko B. K.	–52/–85 (–52/–55)	04/98	–4.8	–29	–21.7	–21.6	89.0 ± 0.6	< 1.4
46'36 / 41'61	Tidrika Sud	–64/–115 (–64/–65)	04/98	–5.8	–38	–15.8	–14.7	78.5 ± 0.7	< 2.1
41'20 / 31'23	Tondi Kiboro	–29/–60 (–31/–33)	08/97	–4.0	–28	–14.1	–17.2	94.0 ± 1.2	2.0 ± 1.0
23'20 / 54'92	Tongom	–63/–85 (–64/–67)	02/95	–2.9	ND	–16.3	–16.4	94.8 ± 0.7	0.8 ± 0.4
				Boreholes					
47'61 / 20'12	Boula K. T.	–27/–80 (–51/–62)	04/98	–4.4	–26	–12.2	–11.9	88.9 ± 0.9	1.8 ± 0.6
50'63 / 22'71	Ko Dey	–38/–85 (–67/–78)	04/98	–4.1	–24	–11.1	–11.1	88.5 ± 1.1	< 1.6
27'34 / 47'85	Oura Tondi	–50/–65 (–52/–63)	04/98	–4.7	–30	–15.7	–15.5	84.4 ± 0.8	< 2.2
41'12 / 22'51	Sourgourou	–39/–75 (–54/–65)	04/98	–4.0	–25	–13.4	–13.0	80.2 ± 0.7	< 1.1
43'49 / 20'15	Tollo	–62/–100 (–81/–92)	04/98	–3.7	–21	–12.7	–12.8	77.9 ± 1.2	ND
44'51 / 22'83	Touliel	–28/–70 (–41/–52)	04/98	–4.6	–27	–14.5	–14.1	83.1 ± 0.7	1.8 ± 0.8
37'30 / 32'25	Youloua	–21/–40 (–25/–36)	04/98	–4.8	–28	–18.0	–17.6	84.1 ± 0.6	2.0 ± 1.0
34'08 / 58'50	Zébani Fiti	–30/–60 (–48/–59)	11/98	–5.6	–39	–10.2	–10.2	65.2 ± 0.6	< 1.5

ND: not determined. "δ" is defined as $\delta = [(Rx/Rs - 1] * 1000$, where Rx is the isotopic ratio of the sample and Rs is the isotopic ratio of the standard (VSMOW for $\delta^{18}O$ and $\delta^{2}H$, and VPDB for $\delta^{13}C$). Analytical accuracies are ± 0.2‰ for $\delta^{18}O$, ± 2‰ for $\delta^{2}H$, and ± 0.1‰ for $\delta^{13}C$. Depth values (third column) are expressed in meters below the soil surface.

anthropic pumping is estimated at approximately 0.3 mm/year (Favreau 2000). Ground water chemistry was studied by Leduc and Taupin (1997) and Favreau (2000). The ground water temperature is consistent with climatic conditions, i.e., approximately 30°C. The water is acid (pH values between 5.0 and 6.0) and circulates in an oxidizing medium (Eh between +300 and +500 mV). Total dissolved solid values are low, consistent with the quartzic nature of the aquifer (20 to 200 mg/L, with a median of 45 mg/L for the samples of Table 1). The dominant cations and anions are Na^+, Ca^{2+} and NO_3^-, HCO_3^-, respectively. No difference appears between samples at different depths below the water table. As the aquifer is carbonate-free, the saturation indices with respect to carbonate minerals are very low, between –3 and –5.

Various methods were used to estimate the current recharge of the unconfined aquifer. Preliminary results obtained from water balance calculations of endoreic ponds and from seasonal variations in the water table were given by Desconnets et al. (1997) and Leduc et al. (1997). Both methods produced similar results, which ranged from 10 to 80 mm/year for a given year, with an average value near 50 mm/year. A minimal threshold for the current recharge was obtained by considering the mean rise in the water table of 0.20 m/year over the last decade (Leduc et al. 2001). For a porosity ranging between 10% and 25%, the increase in ground water reserves is between 20 and 50 mm/year; the total recharge, shared between this resource increase and the ordinary flux, is necessarily higher.

Materials and Methods

Sampling and Analyses

Samples of ground water were taken for isotopic measurements from 25 domestic wells, which penetrated only the upper meters of the aquifer, and from eight boreholes screened over a length of 11 m at different depths below the water table (Table 1). In wells open to the atmosphere, renewal of the water column was obtained by pumping before sampling when necessary to obtain water representative of the aquifer (Favreau et al. 2000).

Stable isotopes of the water molecule and of the dissolved inorganic carbon (DIC) were measured using the usual protocols of Epstein and Mayeda (1953), Coleman et al. (1982), and McCrea (1950). For ^{14}C of DIC, the conventional method of Fontes (1971) and the CO_2 to C conversion method for AMS (Kut et al. 1986) were used at the University of Paris-Sud and at the Centre National de la Recherche Scientifique at Gif sur Yvette, France. Tritium was analyzed by the electrolytic enrichment method (Kaufman and Libby 1954) at Thonon (Centre de Recherches Géodynamiques, Université de Paris-6, France).

Conceptual Model of Recharge

Hydrodynamic observations have proved that deep infiltration is localized in time and space, but significant in volume at the location where it occurs (Figure 2). The ground water chemistry of wells and boreholes displays neither zonation nor any relationship with hydrogeologic parameters as, for instance, the depth to the water table or its elevation (Leduc and Taupin 1997; Favreau 2000). Both these characteristics suggest the absence of any steady stratification in ground water within the relatively small saturated thickness of the aquifer. A simple model of good vertical mixing thus appears to be the most suitable method of estimating recharge rate:

$$\frac{dC}{dt} = \frac{R}{V}(Cr - C) - \lambda C \qquad (1)$$

where V is the saturated aquifer volume (L^3), C the radioisotope (^3H or ^{14}C) aquifer concentration, R the aquifer recharge (L^3/T), Cr the atmospheric radioisotope concentration input (^3H or ^{14}C), and λ the radioisotope decay constant (5.58 10^{-2}/year for ^3H, 1.21 10^{-4}/year for ^{14}C). When R is summed over one year, the R/V ratio is the annual renewal rate. Equation 1 is for a stable water table balance ($R = R_0$ and $V = V_0$, unchanged throughout the years). As the climate has not significantly changed for the last 4000 years in southwest Niger (Durand and Lang 1986), this steady-state model was run for 4000 years, until 1950. Then, as stated by Leduc et al. (2001), the saturated aquifer volume has steadily increased for the last five decades in response to land clearance, a change in land use that increased surface runoff and ground water recharge below temporary ponds (Figure 2). Considering this regional increase in ground water reserves, after 1950 Equation 1 becomes

$$\frac{d}{dt}(CV) = RCr - QC - \lambda CV \qquad (2)$$

where Q is the outflow (discharge) from the aquifer (L^3/T) and $R > Q$. Because of the low level of pumping (< 0.3 mm/year) and of the essentially endoreic nature of the region, Q is assumed to be constant over the period from 1950 to 1998 and equal to the value of R in 1950. Three changes in ground water reserves for 1950 to 1998, representing changes in recharge, were considered: (1) a steady water table, $V(t) = V_0$, and no change in recharge with time; (2) an exponential rise in the water table, $V(t) = V_0 e^{kt}$, where t is the time since 1950; and (3) a linear rise in the water table, $V(t) = V_0 (1 + kt)$. Considering Equations 1 and 2, the initial aquifer renewal rate in 1950 (R_0/V_0) is inferred from the present radioisotope concentration (C) measured in the aquifer. The atmospheric radioisotope chronicles (Cr) used in the model are described later.

Input Function for ^3H and ^{14}C

Nuclear tests carried out during the 1950s and 1960s brought huge quantities of ^{14}C and ^3H into the atmosphere. In southwest Niger, there was no regular sampling of ^3H in rain or of tropospheric

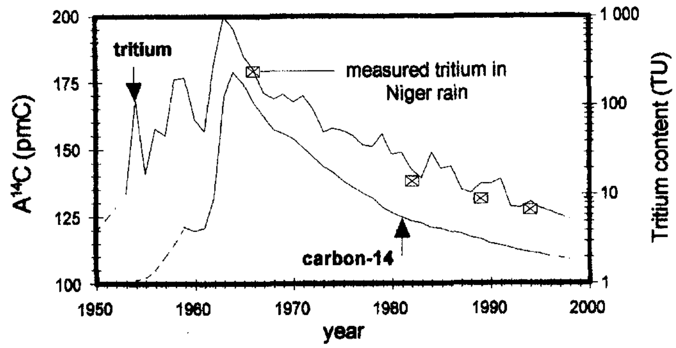

Figure 3. Reconstructed mean annual values (lines) and measured (squares) contents of ^3H in rain and of ^{14}C in troposphere in Niger since 1950.

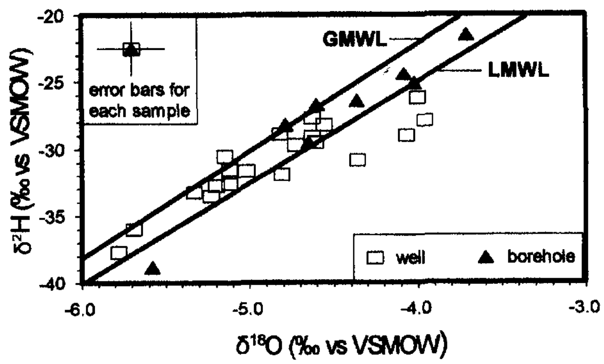

Figure 4. Graph of δ^2H as a function of $\delta^{18}O$ for ground water samples. Global meteoric water line (GMWL): $\delta^2H = 8\ \delta^{18}O + 10$. Local meteoric water line (LMWL): $\delta^2H = 7.6\ \delta^{18}O + 6.0$ (Taupin et al. 1997).

$^{14}CO_2$. The ^3H input function was reconstructed using annual means weighted with rainfall figures from African stations (Bamako, 1963 to 1979; Kano 1971 to 1973; Khartoum, 1960 to 1978; Ndjamena 1963 to 1978; Tunis 1968 to 1997) and with the long Ottawa record (1953 to 1997), all these data came from the International Atomic Energy Agency/World Meteorological Organization network (IAEA/WMO 1998). The background level at the beginning of the 1950s is assumed to be 5 TU (Kaufman and Libby 1954). The few data concerning ^3H contents in rainfall in Niger mentioned in the literature (Leduc et al. 1996) are consistent with this reconstruction (Figure 3). For ^{14}C, the annual values from African stations (Dakar, 1963 to 1968; N'Djamena, 1966 to 1976; Tenerife, 1963 to 1990) reported by Nydal and Lövseth (1996) were used. Longer chronicles from central Europe between 1959 and 1997 (Levin et al. 1985; Levin and Kromer 1997) were used to extrapolate missing data. Small differences in ^{14}C appear between stations, in agreement with the hemispheric homogeneity of CO_2 in the atmosphere.

Results and Discussion

Stable Isotopes ($^{18}O/^2H$, ^{13}C)

Stable isotope contents of ground water vary between –5.8 ‰ and –2.9 ‰ for $\delta^{18}O$, and between –39 ‰ and –21 ‰ for δ^2H (Table 1). Mean values are –4.7 ‰ and –30 ‰, respectively, in good agreement with the mean composition of local precipitations (Leduc and Taupin 1997). Most of the samples are close to or

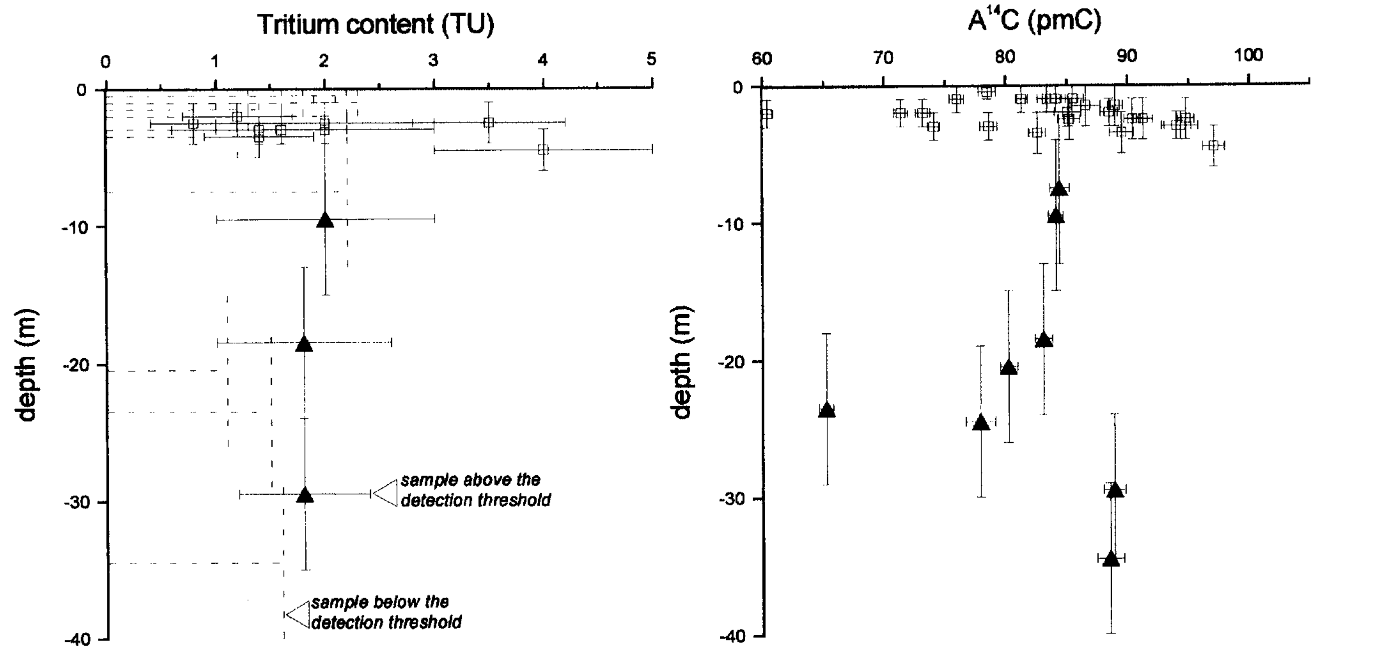

Figure 5. Measured ^3H content and ^{14}C activity versus depth below the water table. Horizontal bars indicate the analytical accuracy for each sample. Vertical bars indicate the vertical extent of the well or borehole screens (□: well, ▲: borehole).

between the global meteoric water line (GMWL) and the local meteoric water line (LMWL) in a δ^2H versus $\delta^{18}O$ diagram (Figure 4). In this semiarid environment where the potential evapotranspiration is 2500 mm/year, this confirms that recharge occurs rapidly through the unsaturated zone with no significant isotope fractionation due to evaporation. The similarity of isotopic composition of samples from boreholes (screened at various depths below the water table) and wells (the upper meters of the saturated zone) is an indication of the absence of ground water stratification, already suggested by the similarity of their chemical composition.

The measured $\delta^{13}C$ contents of DIC range from –21.7 ‰ to –9.1 ‰ (Table 1). The independence of $\delta^{13}C$, ^{14}C activities, and Ca^{2+} contents confirms the absence of carbonate dissolution within the aquifer; thus, DIC is likely to have a single soil/biogenic origin. The calculated $\delta^{13}C$ contents of the theoretical CO_2 in equilibrium with the DIC at 30°C range between –21.6 ‰ and –10.2 ‰ ($\delta^{13}Cg$. eq., in Table 1). This wide range of contents is consistent with the diversity of plants in the Sahel, distributed among C-3 plants ($\delta^{13}C$ near –27 ‰ such as *Acacia sp.*) and C-4 plants ($\delta^{13}C$ near –12.5 ‰ such as *Pennisetum sp.*, millet). Calculated pCO_2 in equilibrium with DIC lies between $10^{-2.1}$ and $10^{-1.2}$ atm., which is compatible with soil pressures. As for $\delta^{18}O$ and δ^2H, there is no noticeable difference in $\delta^{13}C$ contents sampled at different depths within the saturated thickness of the aquifer (Table 1). Because the dissolved carbon in ground water comes from the atmosphere only via the soil, ^{14}C activity measured can be considered representative of the residence time of water in the aquifer.

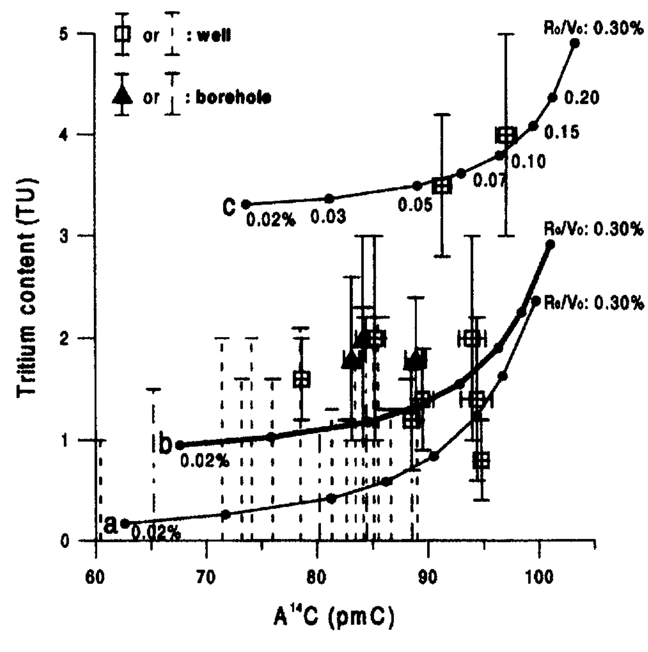

Figure 6. Graph of ^3H content as a function of ^{14}C activity for ground water samples. The three curves represent the $^3H/^{14}C$ couples calculated for various annual renewal rates (R_0/V_0, in %) and different increases in ground water reserves for the 1950 to 1998 period: (a) stable water table; (b) exponential rise in the water table, resulting in an increase of reserves of 10% in 48 years; (c) linear rise in the water table, resulting in an increase of reserves of 20% in 48 years. Dotted lines refer to samples below the detection threshold in ^3H.

Radioisotopes (^3H and ^{14}C)

In the study area, the ^3H content of ground water is low (maximum of 4.0 ± 1.0 TU) and is below the detection threshold for 60% of the samples (Table 1). Taking into account all ^3H analyses, the mean and median values are below 2.1 TU. Tritium is found in both wells and boreholes; the presence of measurable ^3H at great depth below the water table once again suggests the absence of ground

Table 2
Expected vs. Measured ^3H in Ground Water for a Stable or Rising Water Table Model

Local Name	Radioisotopic Data		Stable Water Table Model			Rising Water Table Model[a]		
	Measured ^{14}C	Measured ^3H	^{14}C R_0/V_0[b]	Expected ^3H	Difference[c]	^{14}C R_0/V_0[b]	Expected ^3H	Difference[c]
Wells								
Banikane	91.3 ± 0.8 pmC	3.5 ± 0.7 TU	0.11%	1.0 TU	+1.8/+3.2 TU	0.09%	1.5 TU	+1.3/+2.7 TU
Hamdallay	89.5 ± 1.0 pmC	1.4 ± 0.5 TU	0.09%	0.9 TU	* *	0.07%	1.5 TU	* *
Kampa Zarma	88.5 ± 0.7 pmC	1.2 ± 0.5 TU	0.08%	0.8 TU	* *	0.07%	1.5 TU	* *
Kirib Béri	78.6 ± 0.7 pmC	1.6 ± 0.4 TU	0.04%	0.4 TU	+0.8/+1.6 TU	0.04%	1.2 TU	* *
Kolo Diogono	94.4 ± 1.4 pmC	1.4 ± 0.8 TU	0.15%	1.3 TU	* *	0.12%	1.8 TU	* *
Maourey K. Z.	85.2 ± 0.9 pmC	2.0 ± 1.0 TU	0.07%	0.6 TU	+0.4/+2.4 TU	0.05%	1.2 TU	* *
Niné Founo	97.1 ± 0.9 pmC	4.0 ± 1.0 TU	0.21%	2.2 TU	+0.8/+2.8 TU	0.17%	2.5 TU	+0.5/+2.5 TU
Tondi Kiboro	94.0 ± 1.2 pmC	2.0 ± 1.0 TU	0.14%	1.2 TU	* *	0.11%	1.7 TU	* *
Tongom	94.8 ± 0.7 pmC	0.8 ± 0.4 TU	0.16%	1.5 TU	−1.1/−0.3 TU	0.13%	2.0 TU	−1.6–−0.8 TU
Boreholes								
Boula K. T.	88.9 ± 0.9 pmC	1.8 ± 0.6 TU	0.09%	0.7 TU	+0.5/+1.7 TU	0.07%	1.3 TU	* *
Touliel	83.1 ± 0.7 pmC	1.8 ± 0.8 TU	0.06%	0.5 TU	+0.5/+2.1 TU	0.05%	1.2 TU	* *
Youloua	84.1 ± 0.6 pmC	2.0 ± 1.0 TU	0.06%	0.5 TU	+0.5/+2.5 TU	0.05%	1.2 TU	* *

[a]The model of rising water table is for an exponential rise for the 1950 to 1998 period, resulting in an increase in ground water reserves of 10% in 48 years (Figures 6 and 7).
[b]The "^{14}C R_0/V_0" is the aquifer renewal rate in 1950 (in % year^{-1}) inferred from the measured ^{14}C activity; the model takes into account the exact sampling date (in Table 1).
[c]Difference in ^3H (measured ^3H—expected ^3H); for both models, the calculated ^3H content corresponding to the "^{14}C R_0/V_0" is compared with the measured value: a positive difference suggests a higher increase in reserves, a negative difference suggests a lower rise or a decrease in reserves; the symbol * * indicates that there is no difference between the measured and expected ^3H. Samples below the detection threshold in ^3H are not displayed here, because they are not discriminant between both scenarios.

water stratification (e.g., 1.8 ± 0.6 TU for Boula K.T., where the screen depths are between 24 and 35 m below the water table; Figure 5).

The ^{14}C activities of DIC range from 60.4 ± 0.5 to 97.1 ± 0.9 pmC, for a mean value of 83.4 pmC (quartiles are 78.5, 84.3, and 88.9 pmC). As for ^3H contents, there is no difference in ^{14}C activity of samples from wells or boreholes, even though these represent various hydrodynamic situations; the mean and median values are 84.0 and 85.1 pmC, respectively, for wells, and 81.5 and 83.6 pmC for boreholes. Moreover, the lowest activity was measured in a well (60.4 ± 0.5 pmC in Koma Koukou) and the deepest sampling below the water table displays activity above the mean values (88.5 ± 1.1 pmC at Ko Dey, with screen depths between 29 and 40 m below the water table; Figure 5). In accordance with these results and taking into account the localized recharge process in the area (Figure 2), the vertical movement of ^{14}C in the aquifer probably occurs mainly by convection; the ^{14}C diffusion within the saturated thickness of the aquifer (Sanford 1997) is thus assumed negligible and not to affect the representativeness of the radioactive decay in ground water.

Estimates of the Renewal Rate

The first approach used to estimate the ground water renewal rate was to consider each sampled point separately. The ^3H and ^{14}C couples for different models of increase in ground water reserves and for a range of annual renewal rates are given in Figure 6. Theoretical contents for each model are calculated with Equations 1 and 2, considering the reconstructed ^3H and ^{14}C input functions (Figure 3). The corresponding data and results for samples above the ^3H detection threshold are presented in Table 2. The stable water table model is not satisfactory because only 30% of the expected ^3H values fit with the measured contents. For a rising water table model, the expected ^3H contents fit with the measured values in 90% of the samples. Samples with background concentrations do not discriminate between these two scenarios (Figure 6). These results are in good agreement with the mean long-term increase in ground water reserves of about 10% over the last five decades (Leduc et al. 2001). For the rising water table model, the $^3H/^{14}C$ couples suggest a range of preclearing annual renewal rates (under natural vegetation, in 1950) of between 0.02% and 0.20% (Figure 6).

The second approach used the aquifer mean characteristics to calculate a representative, regional renewal rate. A model of exponential rise in the water table, resulting in an increase in the aquifer reserves of 10% for the period 1950 to 1998 was retained. The representative annual renewal rate in 1950 is close to 0.05% for the median (84.3 pmC) of the ^{14}C activities, and is below 0.20% with respect to the median (less than 2.1 TU) of the ^3H contents (Figure 7). Such an agreement obtained with two independent tracers would appear to validate the results; however, a sound examination of the sensitivity of the model assumptions is still required.

Sensitivity Testing

The impact of the model assumptions on the results were tested for both tracers. Hereafter the estimated errors are given for the model of exponential increase in ground water reserves (10% during the 1950 to 1998 period), for tracer contents in ground water in the 1990s, and for a range of annual renewal rates in 1950 of between 0.02% and 0.20% (Figure 7).

The primary area of uncertainty involves the reconstructed input chronicles. Tritium in rainfall is highly variable in time and space; considering the upper and lower annual contents of north African chronicles results in an uncertainty factor lower than 1 TU. For the ^{14}C, fluctuations in the atmospheric levels between 98 and 110 pmC have been reported for the last several thousand years in response to changes in cosmic ray intensity, climate, or anthropic release of fossil fuel (Suess 1971; Tans et al. 1979). The corresponding uncertainty concerning the ^{14}C content in ground water is of 3.5 pmC. A second source of uncertainty is the relative mean increase in ground water reserves. Differences of about 0.3 TU for ^3H and near 1

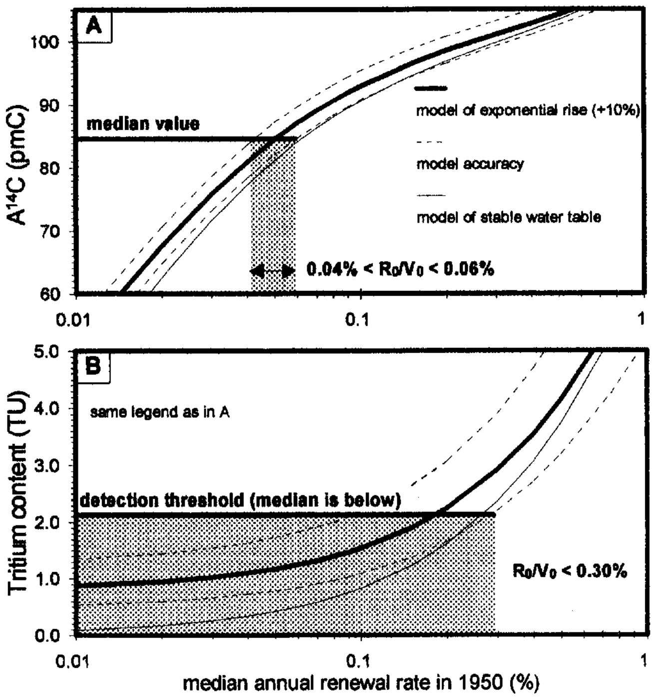

Figure 7. Median annual renewal rate in 1950 (R_0/V_0, in %) inferred from the ^{14}C and ^{3}H median values in ground water: (a) the ^{14}C median value (84.3 pmC) indicates a R_0/V_0 between 0.04% and 0.06% when using the model of exponential rise (+10% in 48 years); (b) same diagram, for ^{3}H. Because more than half of the samples are below the detection threshold, the ^{3}H median value is below 2.1 TU: R_0/V_0 is lower than 0.30%.

pmC for ^{14}C are obtained for relative increases in ground water reserves of ± 2% around the representative value of 10% considered for the 1950 to 1998 period. A third source of uncertainty may be the variability of recharge as a function of annual rainfall. The weighting of the renewal rates by annual precipitation does not modify ground water contents by more than 0.2 TU for ^{3}H, and 0.2 pmC for ^{14}C. Another source of uncertainty tested is a possible time-lag in the ^{14}C inputs in ground water via the soil, where carbon is naturally stored. A maximum delay of three years (Dörr and Münnich 1980) for the transit of the $^{14}CO_2$ from the atmosphere to the ground water via the soil can influences recent levels by less than 0.3 pmC. For ^{3}H inputs, as infiltrated water reaches the water table in a few hours or days (Desconnets et al. 1997), there is no need to take time-lag into consideration (Figure 2).

Cumulated uncertainties on the results are shown in Figure 7 (model accuracy). Calculations by the ^{3}H method appear to be very sensitive to the uncertainties and provide only an upper limit (0.30%) for the median annual renewal rate in 1950. Because of its longer half-life, ^{14}C provides more accurate results, with a range of possible values between 0.04% and 0.06% for the median value of ^{14}C (84.3 pmC) in ground water. Considering the mean ^{14}C activity instead of the median, or the mean and median values of wells and boreholes, does not modify this range. A reverse approach would be to independently calculate the renewal rates corresponding to each ^{14}C activity; the median and mean values for all the annual renewal rates being 0.05% and 0.06%, respectively. The annual renewal rates inferred from the lower (78.5 pmC) and upper (88.9 pmC) quartiles of the ^{14}C data are of 0.03% and 0.08%, respectively; for the mean aquifer characteristics (30 m of saturated thickness, estimated porosity of 10% to 25%), this is equivalent to preclearing recharge rates of between 0.9 and 6.0 mm/year. These figures do not significantly differ from the representative range of 1.2 to 4.5 mm/year obtained for the median value of ^{14}C in ground water (i.e., annual renewal rates between 0.04% and 0.06%; Figure 7). For the same aquifer parameters, the rise in the water table over the last decade (0.20 m/year) results in a recharge rate of between 20 and 50 mm/year. This discrepancy can be interpreted as a quantification of the increase in recharge following the change in land use.

Conclusion

In semiarid southwest Niger, the isotopic contents of ground water characterize the long-term, natural recharge. Stable isotopes of water (^{18}O, ^{2}H) confirm an indirect recharge process and, together with the ^{13}C content of DIC, indicate the purely atmospheric origin of tracers. The absence of stratification of ground water within the aquifer is suggested by stable and radioisotopes sampled at different depths below the water table. Combining the independent ^{14}C and ^{3}H methods appears to be a satisfactory way to estimate recharge. Modeled annual renewal rates are consistent in both methods. However, accuracy is lower with ^{3}H due to its short half-life and low atmospheric contents in present rainfall; this provides only an upper limit of 0.30% for the annual renewal rate. For the opposite reasons, the long-term annual renewal rate can be calculated more precisely using the ^{14}C method, and ranges between 0.04% and 0.06%.

A good understanding of the hydrodynamic context is essential to interpret tracers in ground water. Considering stable ground water reserves would have led to an embarrassing discrepancy between the $^{14}C/^{3}H$ recharge estimates (a few mm/year) and previous figures obtained by hydrodynamic methods (of few tens of mm/year). The long-term rise in the water table allows this discrepancy to be interpreted as an increase in recharge of about one order of magnitude.

Few estimates of increase in recharge following deforestation have been reported for semiarid environments and, to the best of our knowledge, these concern mainly semiarid Australia. Allison et al. (1990) calculated an increase of about two orders of magnitude in the western Murray basin (from 0.1 to about 5 to 30 mm/year) by matric suction and unsaturated zone chloride profiles. In eastern Australia, using nonsteady chloride profiles, Thorburn et al. (1991) calculated recharge rates of 7 mm/year for uncleared areas, and of 29 to 70 mm/year under deforested catchments. These results are consistent with our estimates in semiarid Niger, where recharge increased from between 1 and 5 mm/year to about 20 to 50 mm/year as a result of clearance of the native vegetation. Because our study area in southwest Niger is representative of semiarid African environments (endoreic runoff and localized recharge; land clearance for several decades) our results suggest that an increase in recharge following deforestation may be a common process in Africa, even if it is not yet well documented.

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