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Renewal rate estimation of groundwater based on radioactive tracers (^{3}H, ^{14}C) in an unconfined aquifer in a semi-arid area, Iullemeden Basin, Niger

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Abstract

Estimation of groundwater recharge in arid and semi-arid areas is difficult due to the low amount and variability of recharge. A combination of radiotracers investigation based on simple mixing models allows direct investigation of relatively long-term renewal rates of an aquifer. The recharge process of the shallow Continental Terminal aquifer in the Iullemeden basin (Niger) was investigated using a geochemical and isotopic approach. This study investigates the area in the one degree square of Nianey (13–14°N, 2–3°E). In this area, recharge is highly heterogeneous and mainly occurs through a drainage system of temporary streams and pools during the rainy season. Heterogeneity of the recharge is reflected through the wide variation in electrical conductivity and oxygen-18 content of the groundwater. The carbon-14 activity range for most of the groundwater falls between 69 and 126 pmc showing pre and post-aerial thermonuclear test recharge. Two renewal rate models have been investigated: the first one models a well-mixed reservoir and the second one is derived from a piston flow model, in which mixing is in equal proportions. Major ions in tritium data analyses allow exclusion of non-representative samples and confirm the carbon-14 renewal rate estimations. Both models give similar results for the relatively low renewal rate investigated in the area. Using carbon-14, the mean annual rates of groundwater renewal range from 3 to 0.03% of the aquifer volume with a median of 0.1%. Assuming the median is representative of the overall renewal rate of the area, the recharge rate is in the order of 5 mm a^{-1}. The shallow aquifer recharge extends from the last small humid period (around 4000 a) up to now. High recharge rates are found in depressions whereas low recharge occurs below the plateaux.

Keywords: Groundwater; Recharge; Isotopes; Semi-arid environment; Niger

1. Introduction

In the Sahelian area of Niger, there is no permanent surface runoff, with the exception of the Niger River. Thus, domestic and pastoral water supply depends on groundwater. Currently, the sustainability of these resources is threatened by both increasing population (3% per year) and drought periods occurring since the 1970s. Hydrogeological information such as total resource and renewal rate of groundwater (i.e. the ratio between the annual recharge and the total reserve) is essential for improving water resources management. Estimation of recharge is difficult, however, due to the very low amounts of recharge, and due to temporal and spatial variability of rainfall

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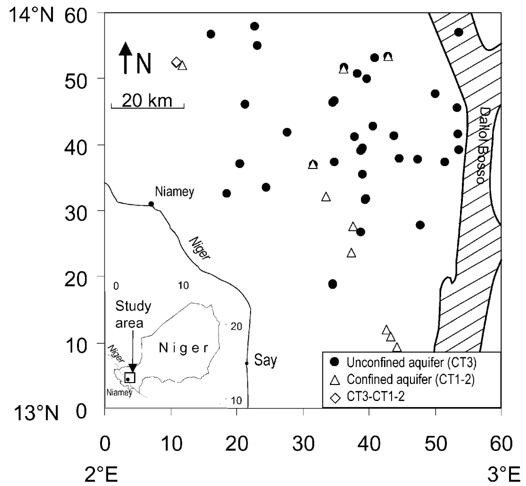

Fig. 1. Study area and the sampling site locations in both the unconfined and confined aquifers.

and recharge processes. This investigation aims to estimate the recharge rate of the unconfined aquifer in the Iullemeden basin (western Niger). This project was carried out under a general hydrogeological study reported by Desconnets et al. (1997) and Leduc et al. (1996,1997).

The Iullemeden basin system is characterised by high aquifer heterogeneity and complex discontinuity of the recharge mechanism. As a result, estimations of the local recharge rate vary widely. Early studies estimated the recent recharge of the unconfined aquifer based on pool water budgets in 1991 and 1993 to be in the order of 50–60 mm a^{-1} (Desconnets et al., 1997; Leduc et al., 1996,1997). In a subsequent investigation, the renewal rate, expressed in percentage of the aquifer volume that is renewed annually, was estimated on the basis of the tritium contents (Leduc et al., 1996). The annual renewal rates calculated in this study, using results above the analytical threshold, ranged from 0.05 to 5% of the aquifer volume. Considering all samples, the overall average rate appeared to be not more than 0.5%. The radioisotope (tritium and carbon-14) approach provides natural time tracers of groundwater and should allow an estimation of local renewal rates integrated over a much longer period of time than the two years of observation. The tritium method is limited here by the short half-life of this isotope, 12.43 a. With a longer half-life, 5730 a, carbon-14 should allow us to confirm and extend the previous calculation of the recharge rate to smaller values. In order to estimate the mean renewal rate of the shallow aquifer, the models considered in this paper take into account the variations of the radioactive tracer input and the mixing processes occurring within the aquifer. The radiotracer input to the groundwater varies according to both (i) variations in atmospheric ^{3}H contents and ^{14}C activities with time and (ii) variations in the amount of annual precipitation.

2. Hydrogeological setting

The study area (2–3°E, 13–14°N) is situated in the Iullemeden basin and covers a region limited by a large dry fossil valley in the east, the Dallol Bosso, and by the Niger River in the south-west (Fig. 1).

The Iullemeden basin is covered by late Tertiary continental deposits, the Continental Terminal (CT). Derived from the erosion of the crystalline formations surrounding the basin, the CT formations mainly consist of sands, silts, clays and sandstones including indurated lateritic layers. The formation can be up to 450 m thick in the central area of the Iullemeden basin. The deposits lie either in concordance with the Continental Intercalaire, a sandstone and clayey sand deposit from the late Cretaceous to early Tertiary, or directly on the bedrock on the western border of the basin, close to the Niger River. Three aquifer formations, a water table aquifer (CT3) and two confined aquifers (CT1–2), have been identified in the studied area. The present study focuses on the CT3 that supplies most of the current water demand.

The average depth to the water table of CT3 is 35 m. However, the water table can get close to the ground surface under topographic depressions or reach up to 75 m deep under the lateritic plateaux, which cover approximately 24% of the area. The aquifer matrix consists mainly of silts and sandy silts with a saturated thickness of 20–50 m. The confining layer between the CT3 and the CT2 consists of clays and grey silts with a thickness of a few meters to 20 m.

Rainfall recorded in Niamey since 1905 shows an annual mean of 562 mm a^{-1} (National Meteorology, Niger). Most rainfall occurs during the rainy season from June to September when the ‘Inter-Tropical Convergence Zone’ migrates northwards and generates

Table 1
Chemical, isotopic composition of groundwater sampled from the shallow (CT3) and confined (CT1-2) aquifers and mean annual renewal rate of the shallow aquifer

	Aquifere	Long. E	Lat. N	Depth to water (m)	T (°C)	E.C. (μS cm^{-1})	pH	Alk. (mequiv. l^{-1})	TDIC (mmol l^{-1})	$\delta^{18}O$ (‰ vs SMOW)		$\delta^{13}C$ (‰ vs PDB)	$A^{14}C$ (pcm)	^3H (UT)	Calculated renewal rate (%)		Selected renewal rate (%)
										1995 samples	Prior 1995				From $A^{14}C$	From ^3H	
NGO-95-01 Kouré village	CT3	2 34.59	13 18.73	33.3	30.0	101	5.8	0.36	1.5	−4.36	–	−13.1	71.4 ± 0.6	< 2,3	0.03	–	0.03
NGO-95-02 Kouré gros puits	CT3	2 34.52	13 18.87	31.4	–	675	6.2	1.24	2.9	−5.01	–	−17.9	109.6 ± 0.7	7.0 ± 1.0	0.87	–	0.87
NGO-95-04 Barkiavel puits no. 3	CT3	2 18.59	13 32.52	27.0	–	113	5.3	0.16	2.0	−4.39	–	−12.3	101.6 ± 0.8	–	0.33	–	–
NGO-95-06 Handallaye Mosquée	CT3	2 24.47	13 33.47	23.0	–	85	6.1	0.16	0.5	−3.44	−4.34	−20.6	89.5 ± 1.0	1.4 ± 0.5	0.09	–	0.09
NGO-95-07 Kida Bazagaizé	CT3	2 34.78	13 37.33	–	–	50.0	5.9	0.28	1.1	–	−4.98	–	–	–	–	–	–
NGO-95-08 Nine Founo Bella	CT3	2 43.90	13 41.30	23.0	–	298	6.0	0.32	1.1	–	−4.82	−9.1	96.9 ± 0.8	4.0 ± 1.0	0.20	–	0.20
NGO-95-09 Ouinditane	CT3	2 53.39	13 45.47	10.0	–	612	6.5	1.20	2.0	–	−3.08	−10.9	113.2 ± 3.0	16.8 ± 0.5	1.20	2.6	2.60
NGO-95-10 Agharous	CT3	2 50.02	13 47.71	34.0	–	110	6.3	0.20	0.4	–	–	−10.5	104.9 ± 1.0	7.9 ± 0.5	0.53	–	0.53
NGO-95-12 Goguize Kouara pastoral	CT3	2 31.64	13 36.95	48.0	–	85	5.8	0.12	0.5	–	−4.88	−13.8	85.1 ± 1.1	< 2,3	0.06	–	0.06
NGO-95-14 Banizoumbou mosquée	CT3	2 39.56	13 31.73	20.5	29.7	203	5.2	0.14	2.3	−3.83	–	−14.2	90.4 ± 0.7	–	0.10	–	–
NGO-95-16 Kampa Zarma	CT3	2 38.80	13 26.74	66.0	31.1	51	5.5	0.20	1.5	−3.86	–	−14.7	88.5 ± 0.7	1.2 ± 0.5	0.08	–	0.08
NGO-95-21 Wankama 30m P1	CT3	2 38.85	13 38.99	19.3	–	145	5.8	1.00	4.7	−4.30	–	−18.6	61.2 ± 0.7	6.8 ± 0.4	0.02	–	–
NGO-95-22 Wankama 80 P2	CT3	2 38.85	13 38.97	20.3	29.9	106	5.6	–	–	−5.23	–	−15.8	70.0 ± 0.6	–	0.03	–	–
NGO-95-23 Wankama 100 P3	CT3	2 38.80	13 38.97	21.9	30.5	76	5.3	0.88	–	−5.47	–	−17.5	44.6 ± 0.4	–	0.01	–	–
NGO-95-24 Wankama puits sud	CT3	2 38.99	13 39.22	16.8	27.0	519	6.0	–	–	−3.94	–	–	–	–	–	–	–
NGO-95-25 Wankama puits nord	CT3	2 39.10	13 39.39	20.4	30.0	345	5.7	–	–	−5.23	−1.89	−16.1	100.5 ± 0.7	–	–	–	–
NGO-95-26 Garbey Tongo	CT3	2 37.92	13 41.20	19.9	28.2	93	5.6	0.40	2.6	−5.21	−3.12	–	–	–	–	–	–
NGO-95-28 Habaka	CT3	2 40.64	13 42.69	75.2	31.1	148	6.2	1.48	3.6	−5.21	–	−8.7	110.2 ± 0.7	1.1 ± 0.4	0.92	0.09	0.09
NGO-95-30 Tongom	CT3	2 23.20	13 54.92	63.1	31.6	39	5.5	0.16	1.2	−2.86	–	−16.3	94.8 ± 0.7	0.8 ± 0.4	0.15	–	0.15
NGO-95-31 Loga	CT3	2 34.83	13 46.56	29.4	30.3	1930	5.3	0.32	3.7	−5.42	−3.85	−14.0	81.5 ± 0.7	–	0.05	–	–
NGO-95-32 Loga Kolo Tassi	CT3	2 34.55	13 46.28	28.2	29.3	319	5.9	0.56	2.2	−5.12	–	−8.9	63.5 ± 0.4	–	0.02	–	–
NGO-95-49 Kokorbé Foundou	CT3	2 36.25	13 51.70	26.7	29.5	588	5.8	0.16	0.8	−3.27	–	–	–	–	–	–	–
NGO-95-50 Kokorbé Foundou peul	CT3	2 38.33	13 50.66	26.1	30.4	36	5.7	0.24	1.3	−5.62	–	–	–	–	–	–	–
NGO-95-51 Dar es Salam	CT3	2 39.73	13 49.96	–	29.3	42	6.5	–	–	−4.09	–	–	–	–	–	–	–
NGO-95-52 Kirib beri	CT3	2 40.88	13 53.12	50.1	31.8	27	5.8	0.20	0.9	−5.50	−5.36	−15.8	78.6 ± 0.7	1.6 ± 0.4	0.04	–	0.04
NGO-95-53 Kirib kaina	CT3	2 42.90	13 53.30	53.4	32.5	54	5.8	0.16	0.7	−5.06	–	–	–	–	–	–	–
NGO-95-55 Tonkobinkani peul	CT3	2 20.49	13 37.09	37.4	28.4	108	5.8	0.34	1.4	−5.70	–	−14.4	102.6 ± 0.7	–	0.40	–	–
NGO-95-56 Sandire	CT3	2 53.70	13 39.10	3.7	28.8	261	6.7	1.28	1.8	–	–	–	104.8 ± 1.0	–	0.52	–	–
NGO-95-58 Boundou Smiti	CT3	2 51.57	13 37.27	32.2	30.1	112	5.8	–	–	−4.30	−4.30	–	–	–	–	–	–
NGO-95-59 Boula Darey	CT3	2 47.45	13 37.68	68.2	31.0	30	5.3	–	–	–	−4.30	–	–	–	–	–	–
NGO-95-60 Karabanga	CT3	2 44.66	13 37.79	51.1	29.7	92	5.5	–	–	–	−4.53	–	–	–	–	–	–
NGO-95-61 Samari mosquée	CT3	2 16.14	13 56.68	18.0	29.8	129	6.5	0.48	0.8	–	–	−8.26	126.1 ± 1.1	14.0 ± 0.5	3.40	1.8	1.80
NGO-95-62 Maourey	CT3	2 53.61	13 56.95	15.3	27.8	257	5.8	0.32	1.4	–	–	−12.47	98.2 ± 1.4	–	0.23	–	–
NGO-95-64 Banizoumbou école	CT3	2 39.53	13 31.60	23.0	–	74	5.9	–	–	–	−3.27	−13.13	85	10.0 ± 1.0	0.06	–	–
NGO-95-65 Kogori Tondi	CT3	2 53.56	13 41.53	5.5	–	440	7.2	–	–	–	−3.00	−14.79	108	20.0 ± 1.0	0.74	5.0	–
NGO-95-66 Birni Kolodia	CT3	2 27.73	13 41.81	46.0	–	75	6.0	–	–	–	−4.40	−14.76	74	< 2,3	0.03	–	0.03
NGO-95-67 Saket Kouara	CT3	2 47.76	13 27.73	63.0	32.2	35	5.8	–	–	–	−4.40	–	–	< 1.5	–	–	–
NGO-95-68 Koïné Kaïna	CT3	2 5.92	13 42	15.0	–	195	5.2	–	–	–	–	–	–	15.0 ± 1.0	–	2.0	–
NGO-95-69 Samari Kaina	CT3	2 22.78	13 57.91	–	–	300	5.2	–	–	–	−4.59	−18.41	106 ± 0.9	7.0 ± 1.0	0.06	–	0.06
NGO-95-Gadabo Fetokadie	CT3	2 21.30	13 46.00	28.0	27?	–	–	–	–	–	–	−14.50	110 ± 0.7	–	–	–	–
NGO-95-Maourey Kouara Zeno	CT3	2 39.04	13 35.44	20.0	–	–	–	–	–	–	–	−17.00	87 ± 0.7	–	–	–	–
NGO-95-29 Karey Bangou	CT3-2	2 10.86	13 52.27	–	29.2	1118	6.8	5.84	7.9	–	–	–	12.1 ± 0.3	–	–	–	–
NGO-95-24b Wankama marre	CT2/1	2	13	–	–	1388	–	–	–	−0.19	–	–	–	–	–	–	–
NGO-95-13 Goguizé	CT2/1	2 31.55	13 37.07	–	31.0	1050	6.6	4.96	7.6	−7.72	–	–	1.3 ± 0.2	–	–	–	–
NGO-95-15 Fétokadié	CT2/1	2 37.65	13 27.60	–	29.4	908	6.5	4.48	7.4	−6.69	–	–	0.6 ± 0.1	–	–	–	–
NGO-95-17 Kanaré	CT2/1	2	13	–	–	1024	7.2	–	–	−7.45	–	–	–	–	–	–	–
NGO-95-18 Tombo dar es Salam	CT2/1	2 42.61	13 11.95	–	30.0	1030	6.8	5.40	7.4	−7.49	–	–	1.2 ± 0.1	–	–	–	–
NGO-95-19 Madina	CT2/1	2 43.30	13 10.86	–	–	1038	6.8	–	–	−7.19	–	–	–	–	–	–	–
NGO-95-20 Kofo	CT2/1	2 44.20	13 9.28	–	–	965	6.5	4.60	8.1	−7.31	–	–	0.8 ± 0.2	–	v	–	–
NGO-95-48 Kokorba foundou	CT2/1	2 36.13	13 51.52	–	29.7	1552	7.3	5.00	5.5	−7.48	–	–	0.7 ± 0.4	–	–	–	–
NGO-95-54 Kirib Kaîna	CT2/1	2 42.94	13 53.31	–	34.5	1043	7.3	5.36	6.0	−7.66	–	–	1.6 ± 1.0	–	v	–	–
NGO-95-71 Zimba	CT2/1	2 11.84	13 52.08	–	30.7	–		5.00	–	–	−7.30	−10.22	2.3 ± 0.2	–	–	–	–
NGO-95-72 Fandouberi	CT2/1	2 33.50	13 32.06	–	32.1	1480	7.7	4.00	–	–	−7.00	−11.46	1.5 ± 0.2	–	–	–	–
NGO-95-73 Baboussey	CT2/1	2 37.39	13 23.65	–	31.7	–	7.8	5.00	–	–	−7.40	−10.34	1.4 ± 0.2	–	–	–	–
NGO-95-70 Diawando	CT2/1	2	13	–	–	–	–	–	–	–	–	–	–	–	–	–	–

high convective thunderstorms. The resulting rainfall shows large temporal and spatial variability (Taupin et al., 1993). Under such a precipitation regime, infiltration is more likely to be related to the amount of precipitation of each event and its spatial and temporal distribution than to the mean annual rainfall. Moreover, most of the infiltration to the shallow aquifer occurs through a drainage system of ephemeral streams and pools filled during the rainy season (Desconnets et al., 1997). As a result, the overall recharge process is quite heterogeneous. From 1991 to 1993, the mean annual recharge rate estimated from pool infiltration and water level was 10% of the annual rainfall, i.e. 50 mm a^{-1} (Desconnets et al., 1997).

3. Sampling and measurements

During 1994 and 1995, groundwater samples were collected from both the shallow and the confined aquifers. Most shallow groundwater samples were collected from open village wells while the samples from the confined aquifer were collected from boreholes equipped with manual pumps.

Physical and chemical parameters, i.e. depth to water table, temperature, pH, electrical conductivity (EC) and alkalinity, were measured in the field. The total dissolved inorganic carbon (TDIC) is calculated from temperature, pH and alkalinity using the equilibrium equations between the different carbonate species in solution (Stumm and Morgan, 1981). (Table 1)

Thirty-eight shallow groundwater samples were selected to be analysed for oxygen-18 content of water and 30 to be analysed for carbon-13/carbon-14 content of TDIC. Ten additional tritium samples were collected to complement the pre-existing tritium data series (Leduc et al., 1996). To allow comparison of isotopic contents between the CT3 and CT1-2 aquifers, nine samples from the confined aquifer were also analysed for their oxygen-18, carbon-13 and carbon-14 contents.

Isotopic contents were measured using the usual protocols (Mac Crea, 1950; Epstein and Mayeda, 1953; Cameron, 1967; Fontes, 1971 and 1983). Carbon-14 analyses were done either by β counting or by accelerator mass spectrometry (Tandetron, Gif sur Yvette, France) for groundwater with very low total dissolved inorganic carbon contents. The ^{18}O and ^{13}C contents are reported in ‰ versus V-SMOW and ‰ versus PDB, respectively, with an uncertainty of 0.2‰. The tritium content and the carbon-14 activity are reported in tritium units (TU) and in percentage of modern carbon (pmc), respectively. The precision of the latter measurements depends on the amount of carbon recovered for analyses and typically varies between 3 and 0.4 pmc.

4. Results

4.1. Mineralisation, pH and oxygen-18 contents

The groundwater temperature of the shallow aquifer ranges from 28 to 32°C reflecting the mean annual temperature. Most water samples are characterised by low total mineralisation, with EC ranging from 20 to 300 μS cm^{-1}, although locally a few of them show higher values. The shallow groundwater is characterised by low pH, ranging from 5.3 to 6.5, and low TDIC contents, around a mean of 1.6 mmol l^{-1} (ranging from 0.4 to 3.6 mmol l^{-1}). According to a large survey carried out in 1994 and 1995 (Leduc and Taupin, 1997), the oxygen-18 content of this groundwater varies widely around an average value of −3.98‰ ($\sigma = 0.98$, $n = 98$). This average is similar to the oxygen-18 weighted annual mean of modern rainfall (−3.98 ± 1.02‰ over the period from 1989 to 1995; Taupin et al., 1995).

The geochemical signatures of the groundwater from the unconfined and the confined aquifers can be distinguished by both salinity and oxygen-18 content. The shallow groundwater is mainly fresh, whereas the water sampled from the confined aquifer is more saline, with electricial conductivity above 1200 μS cm^{-1}. Due to palaeorecharge, the oxygen-18 content of the deeper aquifer shows values at least 2–4‰ lower than those of the groundwater from the unconfined aquifer (Dray et al., 1983; Le Gal La Salle, 1994). In general, the TDIC content of the confined aquifer is also much higher (7 mmol l^{-1}) than in the shallow aquifer and the pH is generally higher, ranging from 6.5 to 7.8.

Of the following parameters EC, TDIC content and $\delta^{18}O$ (Fig. 2a,b) none show any evidence of mixing

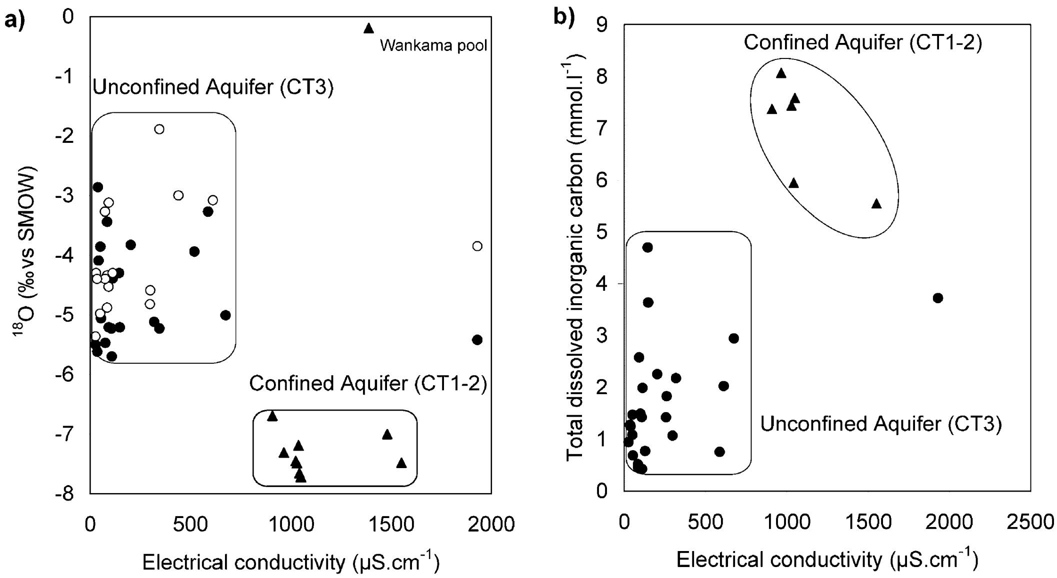

Fig. 2. Oxygen-18 content (a) and total dissolved inorganic carbon (b) versus electrical conductivity. Unconfined aquifer: ● samples collected after 1995, ○ samples collected before 1995; Confined aquifer: ▲.

between groundwater from the two aquifers which could perturb the time tracer signal. However, localised mixing could be hidden by the scattering of the shallow water end-member (Fig. 2a).

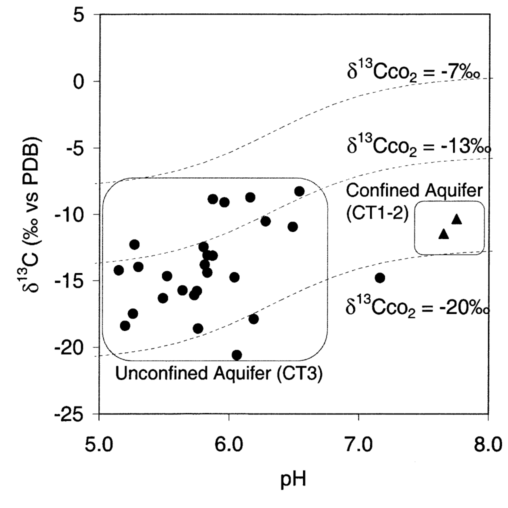

Fig. 3. Carbon-13 contents of the TDIC versus pH. The three curves represent the theoretical groundwater compositions in ^{13}C resulting from an isotopic equilibrium with gaseous CO_2 with different carbon-13 content (δ^{13}C of −7, −13 and −20 ‰ versus PDB).

4.2. Carbon-13 content and carbon-14 activity of TDIC

The carbon-13 data of groundwater from the unconfined aquifer show wide variations of the isotopic compositions, from −8 to −21‰ versus PDB. The carbon-13 contents of groundwater from the confined aquifer do not really differ from those of the shallow groundwater, but fall within a smaller range of values, between −10 and −11‰. There is no relationship between δ^{13}C and the total inorganic dissolved carbon contents; this lack of relationship is consistent with the silicate nature of the aquifer matrix and the very low alkalinity of the groundwater. Thus, the main source of inorganic dissolved carbon in this system is probably the soil gas CO_2. As shown by the carbon-13 contents versus pH diagram in Fig. 3, most water lies between the theoretical curves representing the carbon-13 contents of TDIC in isotopic equilibrium with soil CO_2 with $\delta^{13}C_{CO_2}$ ranging between −20 and −13‰ (Deines and Langmuir, 1974; Wigley, 1975). These $\delta^{13}C_{CO_2}$ values are in good agreement with those found in the literature for the area (Ousmane, 1988; Taupin, 1990; Joseph et al., 1990) where typical carbon-13 contents of soil gaseous CO_2 range from −19 to −13‰ versus PDB.

The CO_2 depleted in carbon-13 is related to native vegetation in which it ranges from −26 to −30‰, while the carbon-13 content of cultivated plants ranges from −10 to −13‰ (Taupin, 1990). Biogenic CO_2 respired by plants in the unsaturated zone is enriched in carbon-13 by 2.6–9.1‰ compared with the plant itself (Salomon and Mook, 1986). Under native vegetation, biogenic CO_2 is, therefore, likely to be depleted in carbon-13, with values ranging between −28 and −17‰. The final signature varies according to the contribution of each native species. Ousmane (1988) and Taupin (1990) measured the carbon-13 content of biogenic CO_2 at −19.2 and −20‰ under the area dominated by native vegetation. The soil gas enriched in carbon-13 is related to millet (−13‰, Taupin, 1990). Millet is extensively cultivated during the wet season when infiltration occurs and is, therefore, an important factor influencing the carbon-13 signature of soils in such areas.

The variations observed for carbon-13 contents of TDIC are thus related to different proportions of biogenic CO_2 produced under cultivated and/or native vegetation. In a high proportion of the samples, the TDIC appears to be in equilibrium with relatively enriched CO_2 gas. This suggests that recharge occurs preferentially in areas where millet is cultivated and during the wet season, when millet is grown. The signature of the groundwater TDIC, therefore, potentially provides a valuable tracer to identify groundwater recharge areas and to better describe groundwater flow paths. To fully investigate the potential of cabon-13 as a tracer of recharge areas, further investigation would be needed, including study of vegetation cover, its spatial and temporal variations, topography, carbon-13 signature of soil CO_2 and carbon-13 signature of the TDIC.

The carbon-14 activity measured on the TDIC varies between 44 and 126 pmc for groundwater from the shallow aquifer. However, most samples remain within the range of 69–126 pmc. This indicates relatively recent recharge, from both before and during the nuclear tests period. The carbon-14 activity of the confined aquifer is less than 2.3 pmc, which agrees with the palaeosignatures of the groundwater (Le Gal La Salle, 1994). Again, according to the carbon-14 activities and carbon-13 contents, most samples show no evidence of mixing, suggesting that carbon-14 activity variation is mainly due to radioactive decay and reflects the residence time of water within the aquifer.

An additional process that could affect the groundwater activity signal is the diffusion of CO_2 gas from the unsaturated zone to the groundwater (Fontes and Edmunds, 1989; Walker and Cook, 1991). Diffusion of modern CO_2 can increase the carbon-14 activity of groundwater and lead to an underestimation of the groundwater residence time. This can occur mainly in two situations: (i) in modern groundwater with high pH where dissolution of CO_2 gas is enhanced (Fontes and Edmunds, 1989; Wigley, 1975; Stumm and Morgan, 1981), and (ii) in older groundwater with low recharge rate where transport of carbon-14 is higher by diffusion than by recharge (Walker and Cook, 1991). The low pH of the groundwater in the studied area suggests that the first process is not important. The calculated recharge rates, as discussed later, are above the estimated threshold where diffusion becomes important, i.e. 0.1 mm a^{-1} (Walker and Cook, 1991). Thirdly, the good correlations between carbon-14 and tritium suggest that the carbon-14 signature has not been modified and still reflect groundwater renewal rate. All this allows us to consider that the diffusion process is not important.

5. Modelling of carbon-14 activity and the calculation of renewal rates

Annual renewal rates of groundwater into the unconfined aquifer can be estimated from both tritium and carbon-14 activities taking into account both (i) the annual input of the radio-tracers and (ii) the radioactive decay. Two models of groundwater mixing based on successive annual recharge within the aquifer have been considered. The model of a well-mixed reservoir assumes that a complete mixing of groundwater issued from successive recharge events occurs within the aquifer. This model was applied by Leduc et al. (1996) to tritium data. Based on mixing in the aquifer and radioactive decay, this model can be directly adapted to carbon-14 data. A second model, of mixing in equal proportions (Salem et al., 1980; Gonfiantini 1988), assumes (i) a vertical displacement of groundwater within the aquifer by piston flow resulting in stratification of groundwater issued from successive recharge events and (ii) a mixing of different groundwater layers during sampling. For both models, the aquifer is supposed to be at steady state, i.e. water loss equals water input.

Annual input of a radioisotope in the groundwater system depends on both the annual rainfall and the radioactive activity of the input water. These components represent the two parameters of the models.

The tritium variations in precipitation have been reconstructed by Leduc et al. (1996) from International Atomic Energy Agency (IAEA) records of the nearest stations (Bamako, Mali and Ndjamena, Chad) and from the records of Ottawa, Canada. Tritium fallout shows similar patterns throughout the Northern Hemisphere, with varying intensity at different sites. This allows for a reconstruction of the tritium chronicle in precipitation using appropriate stations with tritium records. The methods for reconstruction of tritium chronicles are discussed in Siegenthaler et al. (1972) and IAEA (1992). Bamako and Ndjamena are the IAEA stations closest to Niamey geographically and climatically, with similar latitudes, though Ndjamena is more humid. A few tritium analyses of Niamey's rainfall fall on the reconstructed curve validating the reconstruction method. Uncertainties inherent in the reconstitution led to sensitivity analyses of the tritium model as discussed in Leduc et al. (1996).

As the CO_2 is well homogenised in each hemisphere (Fontes, 1983), variations of atmospheric carbon-14 activity measured in the northern hemisphere can be used in the study area. Before 1905, the carbon-14 activity shows little variation (Stuiver et al., 1991), and the atmospheric activity is assumed to have been constant at 100 pmc. Departures from this initial value are discussed in Section 6. Between 1905 and 1950, the consumption of fossil fuel generated a slight decrease in the radiocarbon atmospheric activity from 99.5 to 97.5 (Suess, 1971). The main changes in the carbon-14 activity are due to the aerial thermonuclear tests between 1953 and 1963 when the radioactivity of the atmosphere rose dramatically, up to around 200 pmc in 1963 in the northern hemisphere. The curve of atmospheric carbon-14 from 1950 to 1980 appeared in a paper published by Levin et al. (1992). From 1980 to 1994, the mean annual radiocarbon activity of the atmosphere has been assumed to have decreased exponentially (Levin et al., 1995).

5.1. Model of well-mixed reservoir

For the model of a well-mixed reservoir with annual time steps, the radiocarbon activity of groundwater is calculated from the radioactive decay of carbon-14 in solution and the annual input of carbon-14 as follows:

$$A_{\mathrm{gw}_i} = (1 - R_{\mathrm{r}_i})A_{\mathrm{gw}_{i-1}}\mathrm{e}^{-\lambda} + R_{\mathrm{r}_i}A_{\mathrm{o}_i} \tag{1}$$

where R_r is the annual renewal rate, A_gw the ^{14}C activity of groundwater, A_o the ^{14}C activity of the input water, λ the radioactive constant ($1.21 \times 10^{-4}\ \mathrm{a}^{-1}$) and i the time by year, ranging from 0 to 90 (corresponding to the calendar year 1994–1905). Eq. (1) takes into account the annual variation of carbon-14 input since 1905. Before 1905, the system is assumed to have been at steady state, and the groundwater activity in 1905 is calculated for a constant input activity, A_o, of 100 pmc:

$$A_{\mathrm{gw}_{91}} = \frac{A_\mathrm{o}}{\left(\dfrac{\lambda}{R_\mathrm{r}} + 1\right)} \tag{2}$$

5.2. Model of mixing in equal proportions

In this second model, the measured activity corresponds to the sum of annual inputs corrected from radioactive decay (t) between the time of recharge and the time of sampling. The mean activity in the groundwater is given by

$$A_\mathrm{gw} = \frac{1}{t_\mathrm{max}} \sum_{0}^{t_\mathrm{max}} A_\mathrm{o}\mathrm{e}^{-\lambda t} \tag{3}$$

with t_max, referring to the time of the earliest recharge event and with the average annual renewal rate, R_r, equal to $\frac{1}{t_\mathrm{max}}$

As with the previous model, the system is assumed to have been at steady state before 1905, with an annual input activity, A_o, of 100 pmc. Contribution to the groundwater activity of the water infiltrated before that period is then given by:

$$A_{\mathrm{gw}_{91}} = \frac{R_\mathrm{r}A_\mathrm{o}(\mathrm{e}^{-\lambda t_{90}} - \mathrm{e}^{-\lambda t_\mathrm{max}})}{\lambda} \tag{4}$$

If the aquifer still contains groundwater recharged prior to 1905, the total activity of the groundwater in 1994 (after the rainy season) is then given by the sum

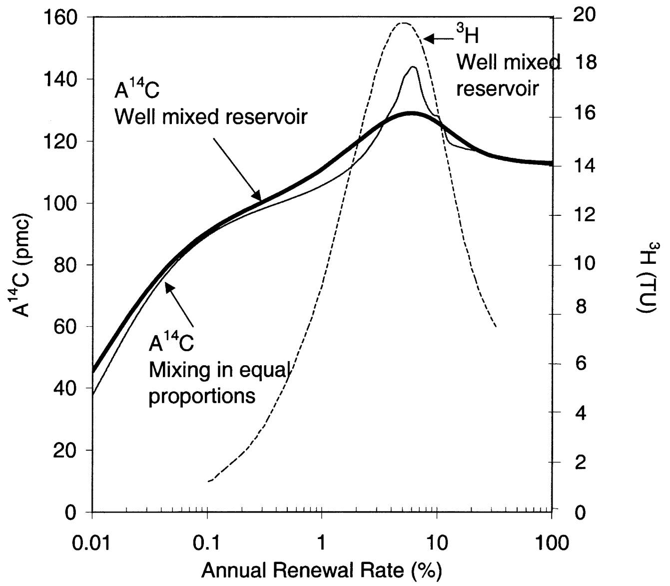

Fig. 4. Annual renewal rate calculated from the carbon-14 activity for both model of well mixed reservoir and model of mixing in equal proportion.

of (i) the contribution to the activity of the recharge prior to 1905 calculated at steady state (Eq. (4)) and (ii) the sum of the annual inputs after 1905 (Eq. (3)):

$$A_{\mathrm{gw}} = A_{\mathrm{gw}_{91}} + R_{\mathrm{r}} \sum_{0}^{90} A_{\mathrm{o}_t} \mathrm{e}^{-\lambda t}. \tag{5}$$

5.3. Variation in the annual recharge according to annual rainfall

For both models, the annual recharge is first assumed to be proportional to the annual rainfall. Therefore, the annual renewal rate, R_{r_i}, is given by the mean renewal rate, R_{r}, weighted by the annual rainfall, $P_i : R_{\mathrm{r}_i} = R_r \frac{P_i}{P_{\mathrm{m}}}$. However, the semi-arid nature of the study area necessitates an additional correction factor assuming a minimum threshold of rainfall amount, under which no recharge occurs. The annual renewal rate is then given by:

$$R_{\mathrm{r}_i} = \frac{R_{\mathrm{r}}(P_i - P_{\mathrm{t}})}{(P_{\mathrm{m}} - P_{\mathrm{t}})} \tag{7}$$

The threshold rainfall amount, P_{t}, has been estimated at 320 mm a^{-1} (Leduc et al., 1996). The annual rainfall record at Niamey between 1905 and 1991 has been provided by the National Meteorology of Niger. Before that period, the rainfall is assumed to have equalled the annual mean over the whole period (P_{m}).

6. Discussion

6.1. Sensitivity of models

As the two models represent two extreme views of the processes, which can occur in the aquifer system, the calculated renewal rates should represent extreme values. However, from one model to the other, the calculated renewal rates show little variation for a given carbon-14 activity (Fig. 4). These variations are negligible compared to the wide range of renewal rates suggested by the tritium (less than 0.05–5%) and carbon-14 data (0.03–3%). Moreover, most shallow groundwater samples show activities ranging from 70 to 105 pmc where the discrepancy between models is not greater than 1.2%. Thus, the estimation of the renewal rate is not affected much by the choice of model. In the following discussion, we will focus only on the model of a well-mixed reservoir.

The use of this model over a long period of time might be limited by the assumption of steady state.

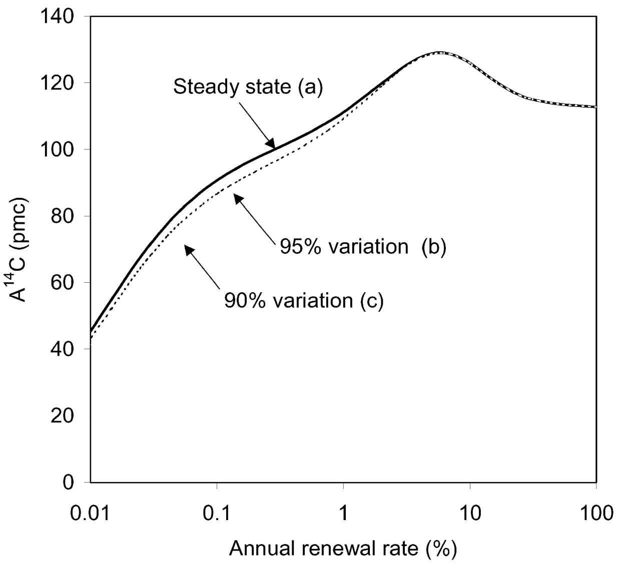

Fig. 5. Annual renewal rate calculated from the model of well mixed reservoir for initial groundwater activities in 1905 (a) at a steady state and lower than the steady state by (b) 5% and (c) 10%.

Variations in the initial conditions assumed for the model can include (i) changes in the atmospheric carbon-14 activity before 1905, as recorded by dendrochronology or suggested by changes in the terrestrial geomagnetic dipole (Stuiver et al., 1991; Guyodo and Valet, 1996; Kitagawa and Van der Plitch, 1998), (ii) changes in the tritium content in precipitation before 1953, (iii) variations in the annual amount of precipitation prior to 1905 and/or (iv) variations in the renewal rate itself or in the volume

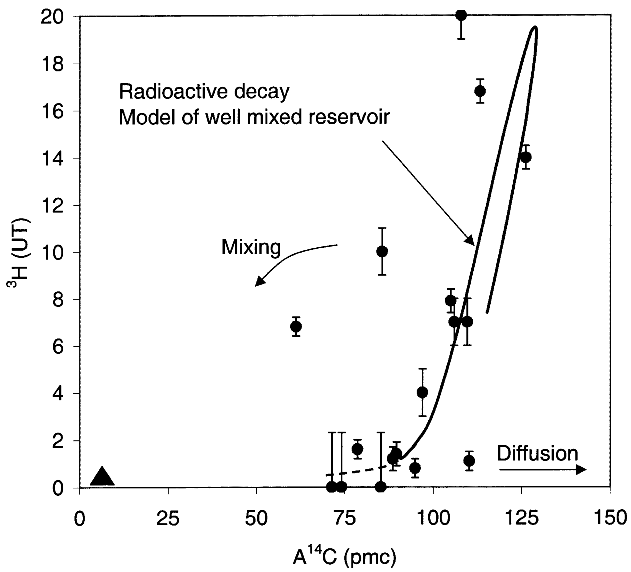

Fig. 6. Carbon-14 activity versus tritium content. The theoretical curve indicates a model of well-mixed reservoir for different renewal rates. Unconfined aquifer: ● Confined aquifer: ▲.

of the reservoir. The sensitivity of the model using the tritium data was discussed by Leduc et al. (1996). Variations of tritium activities prior to 1953 have little impact on the renewal rate estimation. The sensitivity of the models using carbon-14 is discussed below.

Groundwater with a low carbon-14 content, i.e. with a long residence time, is more sensitive to changes in the initial conditions. The model has been tested for initial groundwater activities in 1905 differing by 5 and 10% from those at a steady state (Fig. 5). For samples with activity of 70 pmc, such variations in the carbon-14 activity at steady state would correspond (i) to fairly large variations in the atmospheric carbon-14 activity before 1905, departing by 5–10 pmc from the assumed 100 pmc or (ii) to variations of up to 7–8% in the mean annual precipitation. Overall, the discrepancy remains reasonably low: for a large variation of the initial groundwater activity (10 pmc), the renewal rate changes from 0.03 to 0.04% for a groundwater activity of 70 pmc, or, at most, from 0.3 to 0.6% for a groundwater activity of 100 pmc. The estimated renewal rate remains within the same order of magnitude and is still negligible compared to the wide range in the renewal rate occurring in the aquifer.

6.2. Renewal rates results

The carbon-14 and tritium data are shown in Fig. 6, where the curve represents the theoretical composition of groundwater for different renewal rates following the model of a well-mixed reservoir. Most of the analytical data follow the theoretical curve, confirming that the model of well-mixed reservoir is a reasonable representation of the process existing within the aquifer. However, a few samples are either depleted of or enriched in carbon-14 showing evidence of other processes.

Two of the samples fall on a mixing curve between recent water and old groundwater of the confined aquifer (CT1-2) being devoid of carbon-14 and of tritium (Wankama and Banizoumbou, NGO-21 and NGO-64). For both samples, though not identified with oxygen-18 content and electrical conductivity values, a mixing process is clearly shown by the carbon-14 and tritium data. Mixing with a small amount of the highly carbonated confined groundwater would have a significant impact on the carbon-14 activity of the shallow groundwater with relatively little influence on the electrical conductivity or on the oxygen-18 content.

Samples showing a high carbon-14 activity compared to the predicted values could have been contaminated with modern atmospheric CO_2 in open wells, leading to an increase of the carbon-14 activity without changing the tritium activity. All these processes emphasise the necessity to consider both tritium and carbon-14 for an estimation of the renewal rate. The dual approach used in this study allows us to disregard non-representative samples.

Ongoing investigation showed that there is no upward leakage to the studied reservoir. Renewal rate can then be estimated using the model of a well-mixed reservoir. For most samples, the renewal rate calculated from the carbon-14 data varies by more than one order of magnitude, from 0.03 to 0.9%. For samples in which carbon-14 is not representative of the residence time, renewal rates obtained from tritium data complete the set of values. The overall range of the renewal rate extends up to 3%. However, since the groundwater was collected from open wells, sampled water could represent only the top meters of the aquifer, and if mixing within the aquifer is not complete, the renewal rate would be over-estimated. Nevertheless, these renewal rates correspond to periods of recharge ranging from 3500 a to now, which include recharge from the last small humid period in the Southern Sahara (around 4000–3500 a BP: Joseph and Aranyossy, 1985; Riser and Petit-Maire, 1986; Fontes et al., 1991; Andrews et al., 1994).

Assuming the geometric dimension of the aquifer (i.e. a mean aquifer thickness of 30 m and a porosity of 0.15), the apparent local recharge rate ranges from 1.3 to 117 mm a^{-1}. This again attests to the spatial heterogeneity of the recharge process. The results remain in good agreement with the previous data estimated from the tritium investigation (Leduc et al., 1996). However, the mean value of the renewal rate is 1% with a median of 0.1%, five times lower than the median based on tritium results. This is consistent with the difference between the tritium and carbon-14 methods, with carbon-14 allowing the time scale of the investigation to be extended.

There is a significant relationship between the depth to the water table and the groundwater renewal

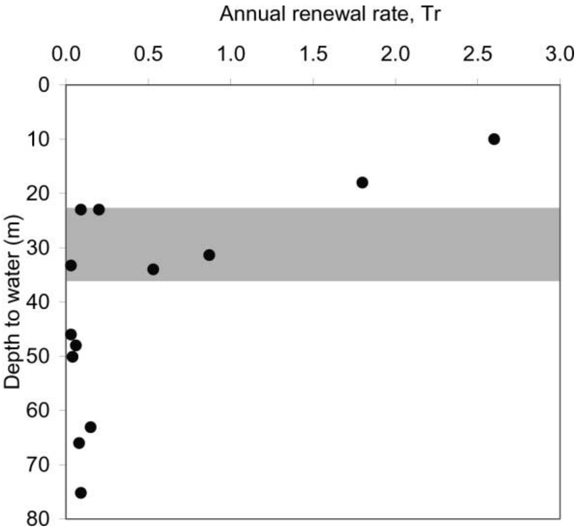

Fig. 7. Renewal rates calculated from the carbon-14 and tritium data according the model of well-mixed reservoir versus depth of water table. For depths to the water less than 35 m the recharge is high, when the depth to the water is higher the recharge is much lower.

rate, with renewal rate decreasing with depth to water (Fig. 7). When the depth to water is greater than 35 m, the renewal rate is always low and ranges from 0.2 to 0.03% ($t_{max} = 600$–3500 a). When the depth to water does not exceed 35 m, the renewal rate is variable and ranges from 3 to 0.2% and the recharge is relatively recent and rapid ($t_{max} =$ 20–500 a). In fact, as the water table is fairly flat, the depth to water table is a reflection of the topography. This confirms that little recharge occurs in areas below the lateritic plateaux, as shown by Desconnets et al., 1997, where the depth to water is great, while high recharge occurs below topographic depressions where the temporary pools and drainage systems form during the wet season. Also, little horizontal homogenisation seems to occur between areas of high and low recharge.

7. Conclusion

The oxygen-18 signature of the groundwater emphasises the heterogeneity of the recharge processes that is also shown by the highly variable carbon-14 activity in the unconfined aquifer. The variations of the carbon-13 signature show the relative influence of native vegetation and millet on the geochemical signature of soil carbon dioxide.

The model of a well-mixed reservoir is in good agreement with the experimental data of tritium and carbon-14 content of the groundwater in the studied aquifer, and the model sensitivity is suitable for the range of renewal rates investigated. The multi-tracers approach appears to be necessary to identify groundwater with carbon-14 data that are not representative of residence time.

The renewal rate of the Continental Terminal shallow aquifer in the Iullemeden basin shows very high variability, ranging from 0.03 to 3%. Recharge to the aquifer has taken place in the period between the last small humid period, which started 4000 years ago, and the present. The renewal rate distribution with depth confirms the importance of the temporary drainage system formed in topographic depressions where the recharge rate is high, in contrast to below the lateritic plateaux, where recharge is very low.

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