



Geochemistry and residence time estimation of groundwater from Miocene–Pliocene and Upper Cretaceous aquifers of Southern Tunisia



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ABSTRACT

Miocene–Pliocene and Upper Cretaceous formations in Tunisia contain one of the most productive artesian aquifers in the country. They consist of Mio-Pliocene sands and Senonian/Turonian carbonates. Intensive pumping in Kébili and Nefzaoua regions over the past decades resulted in gradual lowering of water table at the rate of approximately 0.85 m y^{-1} . Forty-two groundwater samples have been collected from different sectors of the studied aquifers for chemical and isotopic analyses. The observed large spatial variability of chemical composition of groundwater in the study area is most probably linked to two processes: (i) dissolution of dolomite and gypsum, combined with calcite precipitation (dedolomitisation), and (ii) partial evaporation of water. The first process plays an important role in the study area due to abundance of evaporites. Partial evaporation occurs in the upper part of the unsaturated zone during infiltration, especially for groundwater sampled in the Kébili and Djerid regions. Apart from these processes, there are others which influence the salinity of the aquifers. In the Mio-Pliocene aquifer, which behaves as an open system to gases and which receives inputs of CO_2 gas derived from intensive tectonic activity in the area, the interaction of carbon dioxide with carbonate matrix of the aquifer produces an increase in the alkalinity of water. In the Senonian and Turonian aquifers, the process of dedolomitisation evolves in a closed system with respect to CO_2 gas. $\text{Ca}^{2+}/\text{Na}^+$ cation exchange and halite dissolution processes are also important. Stable isotope composition of water ($\delta^{18}\text{O}$, $\delta^2\text{H}$) indicates that the recharge occurs from the Dahar upland. The ^{14}C activity varies between $89.5 (\pm 1.5)$ and $3.7 (\pm 2.1)$ pmc. The ^{13}C content in the total dissolved inorganic carbon (TDIC) range between -13.9 and -3.6‰ . The calculated concentrations of ^{13}C in the CO_2 gas in equilibrium with the TDIC vary between -22 and -11‰ , indicating two sources of carbon in the solution: carbonate matrix ($\delta^{13}\text{C} = -2\text{‰}$) and soil CO_2 ($\delta^{13}\text{C}$ from -25 to -21‰ for the cultivated areas). Mean residence times of water have been determined after correction of the initial ^{14}C activities for ^{14}C -dilution processes including carbonate dissolution, calcite precipitation and cation-exchange. The dilution processes were quantified on the basis of geochemical and ^{13}C mass balance equations. The calculated mean residence times of water confirm modern recharge from Dahar upland and the mountains surrounding depressions, and indicate the presence of paleowaters in the east and south-west region, and in the discharge zone.

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1. Introduction

Sedimentary basins form important groundwater resources for the Saharan-platform area of Southern Tunisia (Mamou, 1990; Yahyaoui, 1996; Edmunds et al., 1997; OSS, 2003). Such basins are frequently the only available water resources in many semi-arid and arid regions in the world (Robertson, 1989; Mukhopadhyay et al., 1994; Birkle et al., 1998; Rosen and Jones, 1998; Guendouz et al., 2002). Most of the agricultural area in Southern Tunisia is

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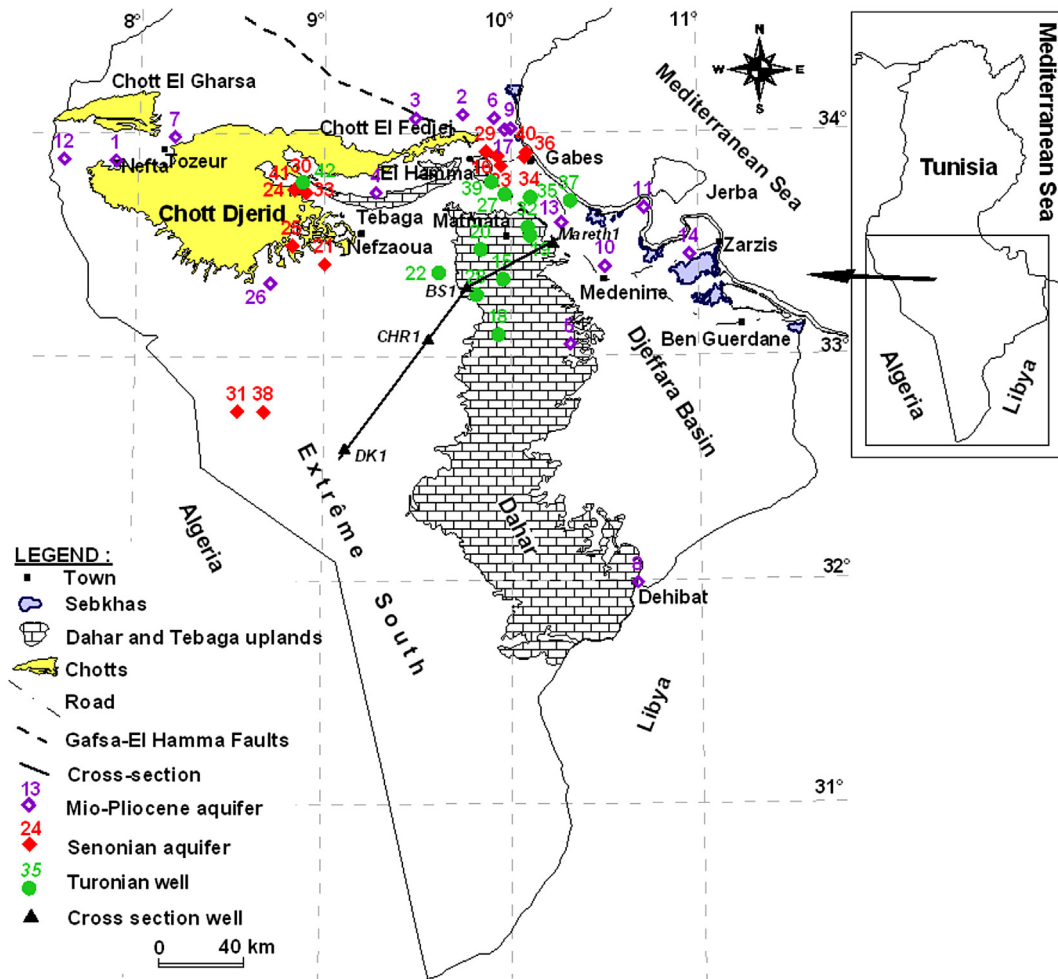


Fig. 1. Location map of study area showing position of sampled wells.

situated on top of sedimentary systems, and the problem of groundwater quality deterioration has become urgent there, mostly because of the increasing exploitation of aquifers. This situation requires better knowledge of the functioning of these groundwater systems.

The studied aquifers are composed mainly of Mio-Pliocene sands and Senonian and Turonian limestones. Using geochemistry and isotopic tools the presented study was addressing the following questions related to the functioning of these aquifers: (i) identification of main geochemical reactions which control mineralization of groundwater (ii) identification of recharge areas, and (iii) providing constrains on the mean residence time of groundwater using radiocarbon.

Several major studies (ERESS, 1972; OSS, 2003) have been already carried out on the studied aquifers. They were focused mainly on modelling of groundwater flow and prediction of its long-term behaviour. Chemical and isotopic information on groundwater in Tunisia was obtained as part of many other studies in the region (Edmunds et al., 1997; Guendouz et al., 2002; Edmunds et al., 2003). Although they confirmed the existence of paleowaters in these aquifer systems, the mean residence time of groundwater has not been well quantified until now.

2. Study area

The study area is situated in northern Africa and covers the region limited by Algerian frontier in the west, the Mediterranean

Sea in the east, the Atlas Saharan Mountains in the north, and the Libyan-Algerian frontier in the south ($30^{\circ}10'–34^{\circ}10' N$; $7^{\circ}30'–10^{\circ}30' E$, Fig. 1). The present climate in the study area ranges from semi-arid in the northeast to arid or Saharan type in the southwest. Precipitation rate in the region is highly irregular; the maximum (250 mm/y) is recorded along the Gulf of Gabes and Dahar upland, whereas the southern part of the basin receives less than 100 mm/y. The annual evaporation is even more than 2500 mm and the mean annual temperature is of the order of $21^{\circ}C$.

3. Geology and hydrogeology

Southern Tunisia is marked by extensive outcrop series of Permian-Mesozoic and deep faults. Attempts to explain the structure and evolution of this domain were presented by Busson (1970) and Bouaziz (1986, 1995). Subsurface studies in the study area began in the 1950s by sequence seismic stratigraphy, well-to-well correlations, isochron and isopach mapping and subsurface basin dynamics analyses of Mesozoic series since the Triassic (Ben Ferjani et al., 1990).

Sedimentary series of lower Cretaceous, which outcrop in Chott El Fedjej area, consist of alternating sandstones, clays, evaporites and dolomites, defined as Merbah El-Asfer (Upper Jurassic-Lower Aptian), Douiret (Aptian s.s.) and Ain El Guettar (Lower Albian) Formations (Busson, 1967; Bouaziz, 1986). These series are of varying thickness (Fig. 2).

Upper Cretaceous sedimentary sequences are well exposed in the outcrops of Dahar upland (Fig. 1) from Upper Senonian to Lower Turonian (Burolet, 1956; Busson, 1967; Bouaziz, 1986). They are composed of alternating thick shallow and deep marine clays, shales, and carbonate packages, and are assembled in the Albian–Cenomanian Zebbag Formation, Cenomanian–Turonian Guettar Formation, Turonian–Campanian Aleg Formation and Campanian–Maastrichtian Abiod Formation (Burolet, 1956). Tectonic processes are important and are responsible for the genesis of numerous horsts and grabens in the region. In addition, they lead to the fracturing of the platform facies which constitute potential reservoir rocks. The Mio-Pliocene series in southern Tunisia are mainly siliciclastics. There is substantial uncertainty in the lateral facies and their age relationships in the Neogene section in the study area. In the northern part of the study area (Nefzaoua and Djerid), the Miocene is represented by two formations, sands and clays of the Beglia Formation and the overlying clays and conglomerates of the Segui Formation.

In southern Tunisia, the major aquifer systems are characterized by the superposition of two principal levels: the “Complexe Terminal” (CT) which forms the upper aquifer and the “Continental Intercalaire” (CI) which constitutes the deeper one. Three main hydrogeological units can be distinguished in the CT aquifer system: the Mio-Pliocene detrital aquifer and Senonian and Turonian carbonate aquifers.

The Mio-Pliocene aquifer reaches its maximum thickness in Chotts basin (300 m) which diminishes towards the edges of the basin. Lateral continuation of the CT aquifer in the coastal plain is the Djefara aquifer. Transmissivities range between 10^{-3} and $24 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}$ (Mamou, 1990; OSS, 2003). Tectonic movements have largely influenced hydrodynamic functioning of these aquifers.

Senonian and Turonian limestone aquifers extend over the Saharan platform and outcrop on the surrounding uplands due to

the presence of several N–S and W–E faults. In the southeast of the study area (Djeffara basin), these series are missing. Stratigraphic gaps in these areas are due to several factors such as irregularity of the paleotopography, eustatic movements and tectonic instabilities (Castany, 1954; Burolet, 1956; Ellouze, 1984; Bouaziz, 1995). Transmissivities range between 0.5×10^{-3} and $15 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$ (Mamou, 1990). The Cretaceous limestones are considered the most important aquifers in Nefzaoua and Kebili regions. Most of groundwater used comes from over 600 wells and boreholes in the CT aquifer system. With a view to increase water supply for agriculture, various boreholes have been drilled to exploit resources from the deeper sectors of the CT aquifer. However, groundwater in some compartments of the basin appeared to be highly saline and therefore unsuitable for irrigation purposes.

The water-bearing sandy formations of the deep CI aquifer constitute the largest confined aquifer in southern Tunisia and contain large groundwater reserves. Groundwater resources of the CI aquifer system are contained in the continental formations of Lower Cretaceous (Neocomian, Barremian, Aptian and Albian), formed by alternations of detrital levels with clayey silt horizons and gypsum intercalations. Thickness and lithology are highly variable laterally.

4. Sampling and analytical methods

For this study, 42 groundwater and surface water sampling points were chosen (Fig. 1, Table 1). They represent hydrochemical types of water and hydrogeological conditions typical of different sectors of the Mio-Pliocene, Senonian and Turonian aquifers. The measurements of water temperature, pH, electrical conductivity (EC) and total alkalinity were carried out on-site. Analyses of major ions and isotopic composition ($\delta^{18}\text{O}$ and $\delta^2\text{H}$ of water, $\delta^{13}\text{C}$ and ^{14}C content of TDIC reservoir) have been performed on all samples. Major cations (Ca, Mg, Na and K) and anions (Cl , SO_4 and NO_3) concentrations were measured by liquid ion chromatography (HPLC) in the Laboratory of Radio-Analyses and Environment of the National Engineering School of Sfax (ENIS), Tunisia. The total alkalinity (as HCO_3) was determined by titration with 0.01 or 0.1 HCl against methyl orange indicator.

Radiocarbon analyses were also done at ENIS laboratory. Dissolved inorganic carbon was precipitated in the form of BaCO_3 in the field from 160 L of water. The ^{14}C content in BaCO_3 samples was determined using benzene synthesis and liquid scintillation spectrometry. Analyses of stable isotope ratios ($^{13}\text{C}/^{12}\text{C}$, $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$) were performed at the isotope hydrology laboratory of the International Agency of Atomic Energy (IAEA) in Vienna using the CO_2 equilibration method for $\delta^{18}\text{O}$ (Epstein and Mayeda, 1953) and isotopic exchange with H_2 gas for $\delta^2\text{H}$ (Colplien et al., 1991), followed by mass spectrometry measurements. Water samples for $\delta^{18}\text{O}$ and $\delta^2\text{H}$ determinations were collected in 30 ml polyethylene bottles with poly-seal caps. Stable carbon isotope composition ($\delta^{13}\text{C}$) was measured by mass spectrometry at the IAEA isotope hydrology laboratory. The results of stable isotope analyses are expressed as per mill deviations from V-SMOW standard (Vienna-Standard Mean Oceanic Water) for ^{18}O and ^2H content and from V-PDB standard (Vienna Pee Dee Belemnite) for ^{13}C content. Concentrations of ^{14}C are reported as percent of modern carbon (pmc). Analytical precision of isotope analyses was equal $\pm 0.1\text{‰}$ for $\delta^{18}\text{O}$, $\pm 1\text{‰}$ for the $\delta^2\text{H}$, $\pm 0.3\text{‰}$ for $\delta^{13}\text{C}$ and 1 pmc for ^{14}C .

5. Results and discussion

The results of chemical and isotope analyses of the collected water samples are listed in Table 1. Water temperature ranges from 18.2 to 41.6 °C. The reported pH values measured at well head are

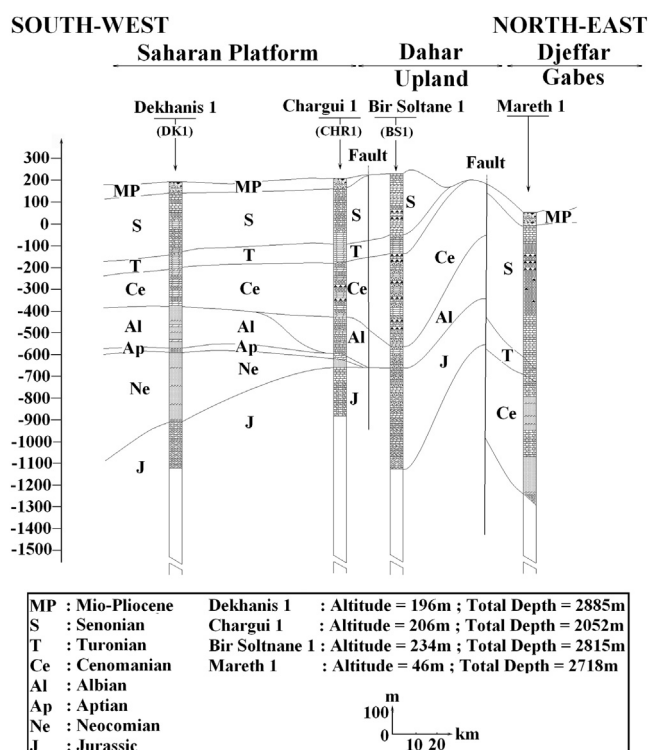


Fig. 2. Hydrogeological cross section of the study area.

Table 1
Isotopic and geochemical data for the studied aquifers.

Well N°	Names	Aquifer names	Screen depth (m)	EC (μS/cm)	T (°C)	pH	Na ⁺ (Mg/l)	Ca ²⁺	Mg ²⁺	K ⁺	Cl ⁻	SO ₄ ²⁻	HCO ₃ ⁻	TDS	SI-calcite	SI-dolomite	Ion balance (%)	δ ¹⁸ O (‰ VSMOW)	δ ² H	D-excess	δ ¹³ C (‰ VPDB)	¹⁴ C (pmc)
1	Mbarek Tatta	Mio-Pliocene	4	7950	23.0	7.3	403.0	780.0	300.0	27.3	750.0	2688.0	183.0	5131.3	0.48	0.30	1	-4.82	-49.2	-10.6	-12.25	85.5
2	P.public El Feji	Mio-Pliocene	50	9050	21.6	8.1	1863.0	704.0	336.0	27.0	2414.0	3000.0	250.0	7750	1.21	1.85	3	-4.31	-27.4	7.1	-10.76	86
3	Mghatta	Mio-Pliocene	26.7	3840	23.0	8.8	257.2	591.0	143.7	<0.05	241.3	2123.7	60.0	3468	1.24	1.62	1	-2.56	-18.2	2.3	-10.27	88.6
4	Source Hzeim 1	Mio-Pliocene		8130	18.2	7.1	1212.0	656.8	121.9	<0.05	2573.4	861.7	147.0	5640	0.12	-0.80	1	-7.61	-53.1	7.8	-10.58	51.5
5	Abdallah Mansouri	Mio-Pliocene	15.5	2530	20.2	7.0	310.7	211.4	128.2	7.0	524.6	698.6	176.9	1880	-0.22	-0.93	3	-6.34	-36.1	14.6	-10.77	89.5
6	Taher Abbas	Mio-Pliocene		9200	22.9	7.6	1127.0	784.0	153.0	16.0	2059.0	1800.0	164.0	7010	0.70	0.45	1	-5.25	-37.5	4.5	-9.23	85
7	Kamel Rhouma	Mio-Pliocene	5	10,000	23.0	7.1	1104.0	640.0	312.0	39.0	1953.0	2688.0	226.0	6962	0.25	-0.06	-4	-4.23	-44.7	-10.9	-9.96	87.5
8	Merzougui Dehibi	Mio-Pliocene	13	3680	21.8	8.1	237.9	245.9	102.3	19.3	330.9	763.1	183.0	2094	0.96	1.28	4	-7.57	-54.4	6.2	-8.55	53.8
9	Aouinet 4	Mio-Pliocene		5430	24.1	7.9	575.0	384.0	153.0	24.0	1136.0	1363.0	189.0	3960	0.89	1.15	-5	-7.51	-53.7	6.4	-7.89	10.5
10	Amra	Mio-Pliocene	400	5670	25.8	8.0	804.5	413.1	174.2	<0.05	984.4	1654.9	146.6	4000	0.87	1.16	3	-4.56	-38.2	-1.7	-7.92	16
11	Jorf Aquaculture	Mio-Pliocene	300	7310	29.8	7.6	1205.6	465.2	111.8	<0.05	1435.4	1972.7	97.6	5300	0.39	0.01	1	-5.06	-43.0	-2.5	-8.33	10.3
12	Bir Touil	Mio-Pliocene	20	6500	20.0	8.3	1420.2	850.4	162.0	36.6	2321.5	2310.0	73.2	7700	0.99	0.98	2	-3.85	-36.6	-5.8	-6.96	75
13	Chettaoui Med	Mio-Pliocene	35	2960	22.4	8.2	350.4	256.0	82.0	<0.05	498.7	954.2	122.0	2220	0.81	0.88	-2	-6.40	-41.1	10.1	-7.21	38
14	Khalfallah	Mio-Pliocene		9090	28.3	7.5	1201.6	356.4	134.4	29.8	1652.9	1893.3	213.5	6257	0.45	0.31	-4	-6.42	-45.1	6.3	-8.24	3.7
15	Zmertem	Turonian		1150	27.2	7.5	234.93	99.40	44.06	<0.05	204.04	402.18	225.7	700	0.18	-0.18	3	-6.70	-39.1	14.6	-8.08	11.4
16	Od khalifa Mah.	Senonian	105	7280	29.1	7.8	499.3	302.0	59.8	50.1	666.7	1137.7	122.0	3270	0.63	0.39	-2	-7.77	-57.1	5.1	-7.88	18.5
17	Godfa 2 bis	Mio-Pliocene	90	4250	22.0	7.9	458.8	392.3	75.1	31.1	663.4	1119.0	128.0	3210	0.71	0.45	2	-7.84	-58.8	3.9	-6.88	16.15
18	Zridib	Turonian		2300	22.7	8.1	640.11	221.76	110.55	<0.05	639.55	1108.25	170.8	2900	0.80	1.05	4	-6.81	-46.9	7.5	-6.73	16.9
19	Beni Zelten 3	Turonian	50	3450	23.7	7.2	268.12	382.60	119.58	<0.05	376.63	1386.13	176.9	2820	0.16	-0.42	-3	-6.05	-32.5	15.9	-7.78	37
20	Oued Sandoug	Turonian	350	2500	27.2	8.1	227.5	149.1	44.6	<0.05	296.1	473.1	158.6	1824	0.84	0.98	1	-6.63	-41.2	11.9	-6.98	25.7
21	Hsay 5bis	Senonian	60	9260	22.3	8.1	1205.9	732.2	156.9	44.2	2733.9	1184.7	122.0	7000	1.06	1.21	0	-6.00	-49.5	-1.5	-6.47	20.7
22	Zoumit	Turonian		3560	26.7	7.4	387.76	284.73	111.88	<0.05	559.20	1152.70	134.2	2730	0.19	-0.21	-2	-7.24	-54.0	3.9	-7.32	30.2
23	Rmathie 2	Senonian	250	4200	31.0	7.6	446.0	433.6	75.8	31.5	708.9	1050.9	134.2	3456	0.65	0.41	4	-7.00	-55.3	0.6	-7.43	15.8
24	Negga Sonède	Senonian	102.7	2750	28.0	8.3	273.5	241.9	63.1	<0.05	543.5	539.3	122.0	1950	1.01	1.28	0	-5.87	-48.2	-1.2	-6.78	16.4
25	Negga 7	Senonian	94	1960	25.0	7.7	251.2	209.1	60.6	12.0	460.1	645.0	146.4	1968	0.46	0.16	-4	-5.15	-51.1	-9.9	-6.70	11.6
26	Sidi Marzoug	Mio-Pliocene	13.4	2850	18.2	7.7	278.3	418.6	68.5	<0.05	601.3	758.9	128.1	2264	0.57	0.03	4	-5.16	-42.9	-1.6	-5.89	52
27	Henchir Jehha	Turonian	571	3740	25.0	7.6	394.27	302.85	111.08	<0.05	567.23	1263.00	122	2870	0.29	-0.07	-4	-7.01	-51.6	4.5	-6.49	9.50
28	El Mthinin	Turonian	350	3120	21.5	7.8	369.6	256.68	98	53.6	528.8	882.4	128.1	2452	0.52	0.35	3			0.0	-5.88	29.7
29	Chenchou 1bis	Senonian	66	4250	41.6	7.7	459.2	420.6	68.2	51.5	731.6	1104.7	116.0	3170	0.78	0.72	2	-6.90	-54.5	0.7	-7.84	8.8
30	Oued zira 2	Senonian	58.5	4630	26.3	7.9	415.9	723.8	100.1	<0.05	800.3	1594.3	115.9	3680	0.96	0.87	4	-6.54	-49.2	3.1	-6.21	6.9
31	Sabria Mol	Senonian	118	1490	24.0	7.9	162.9	100.0	42.1	14.5	221.5	376.8	137.3	1132	0.38	0.16	-2	-5.62	-51.5	-6.5	-5.90	6.3
Well N°	Names	Aquifer names	Screen depth (m)	EC (μS/cm)	T (°C)	pH	Na ⁺ (Mg/l)	Ca ²⁺	Mg ²⁺	K ⁺	Cl ⁻	SO ₄ ²⁻	HCO ₃ ⁻	TDS	SI-calcite	SI-dolomite	Ion balance (%)	δ ¹⁸ O (‰ VSMOW)	δ ² H	D-excess	δ ¹³ C (‰ VPDB)	¹⁴ C (pmc)
32	Beni Zelt. S.	Turonian	103	2610	22.7	7.2	243.12	257.22	110.14	<0.05	400.54	891.48	183	2040	0.09	-0.44	1	-5.80	-31.9	14.5	-6.56	37
33	Zarcine 4	Senonian	154	2400	25.5	7.9	255.5	181.8	60.6	12.0	443.0	573.9	137.3	1974	0.58	0.49	-3	-5.35	-49.1	-6.3	-5.77	11.6
34	Rass el Aïoun	Senonian	233	3100	31.9	7.3	408.9	345.8	88.0	19.6	565.6	1133.8	134.0	2990	0.24	-0.24	1	-7.27	-53.8	4.4	-6.85	23.5
35	Garet Ltaifa	Senonian	146	3480	23.3	7.2	370.00	274.48	102.58	<0.05	500.18	1129.25	115.9	2640	-0.09	-0.85	-2	-6.97	-49.3	6.5	-6.22	12.50
36	Ambar	Senonian	324	3550	31.9	8.0	356.2	371.9	86.0	10.9	439.9	1118.1	146.0	2770	1.01	1.25	4	-7.04	-49.0	7.3	-6.01	16.7
37	Matmata 5	Turonian	88	4690	20.2	7.5	550.42	382.60	157.76	<0.05	939.92	1402.16	122	3750	0.22	-0.22	-2	-6.92	-49.7	5.6	-5.21	11.40
38	El Faouar 4	Senonian	550	2070	23.0	8.1	234.1	145.5	60.6	10.7	426.0	419.6	115.9	1634	0.61	0.60	0	-4.52	-46.5	-10.3	-4.59	9.6
39	Gourai	Turonian	199	4060	31.0	7.5	397.1	388.2	87.4	16.0	547.1	1131.3	134.2	3134	0.51	0.24	4	-7.34	-52.2	6.5	-5.46	11
40	Sidi Daoud	Senonian	60	3820	31.5	8.0	408.6	412.4	85.0	24.3	467.9	1220.4	134.0	3010	0.94	1.07	5	-7.01	-51.0	5.1	-5.21	12.5
41	Jemma 1bis	Senonian		2240	23.0	8.4	224.7	187.0	50.9	11.2	508.6	319.1	115.9	1495	1.00	1.21	0	-5.83	-48.1	-1.5	-3.60	10.8
42	Ziret Touati	Turonian	103	4830	25.3	7.3	545.93	438.20	144.56	<0.05	829.52	1703.96	146.4	3880	0.18	-0.34	-3	-6.80	-47.5	6.9	-4.49	11.2

generally between 7 and 8.78 units, with an average of 7.74, thus indicating a slight tendency towards a basic composition. Electrical conductivity varies between 1150 and 10,000 $\mu\text{S}/\text{cm}$, with an average of 4546 $\mu\text{S}/\text{cm}$; 31% of the samples gave values between 1150 and 3000 $\mu\text{S}/\text{cm}$, 40.5% gave values between 3000 and 5000 $\mu\text{S}/\text{cm}$ and 28.5% were between 5000 and 10,000 $\mu\text{S}/\text{cm}$. This relatively high salinity is due to presence of major ions SO_4^{2-} , Cl^- , Na^+ and Ca^{2+} (Table 1), which correlate well with the electrical conductivity (Fig. 3).

5.1. Hydrochemistry

The groundwater of the studied aquifer systems presents a wide range of hydrochemical types (Fig. 4). The Mio-Pliocene aquifer is mostly dominated by Na–Cl and Ca– SO_4 waters. The Na–Cl waters are characterized by high TDS content (mean value of ca. 7700 mg/l). They are recharged through outcrops of sands and silts of marine origin. The Senonian aquifer is characterized by relatively low TDS content (mean value of ca. 2800 mg/l), with Ca– SO_4 and mixed water types dominating. Relatively high Ca^{2+} and SO_4^{2-} content results from dissolution of evaporitic minerals present in this aquifer. Most of the sampled waters from Turonian aquifer are of Ca– SO_4 –Cl and Na–Cl types. They are somewhat richer in Na^+ when compared to Senonian waters, although have very similar Ca^{2+} and Mg^{2+} contents (between 99–740 mg/l and 42–158 mg/l, respectively).

The Mio-Pliocene aquifer works as an open system with respect to carbon dioxide, as calculated from the total dissolved inorganic carbon pool (TDIC) using the Diagrammes computer code (Simler, 2006). The openness of this system stems from significant variations in thickness of the aquifer and, more importantly, from recent neo-tectonic processes in the study area. Fig. 5 shows the calculated pCO_2 as a function of measured HCO_3^- content for all three studied aquifers. In Fig. 6, the measured pH values are plotted as a function of HCO_3^- content. Distinct anti-correlation observed between both variables for the samples representing the Mio-Pliocene aquifer suggest that the observed changes of pH are largely due to changes in alkalinity in this system.

Fig. 7 shows concentration diagrams for selected parameters measured in the collected groundwater samples (electrical conductivity, SO_4 , HCO_3 , Ca, Cl and Na content). The concentration diagram of a solute Y is defined as the relationship between $\log[Y]$ and $\log(\text{CF})$ where CF stands for the concentration factor (Valles et al., 1997). Under natural conditions, CF can be estimated by using a reference tracer whose behavior has been checked in the context of the studied system (Gac, 1980; Barberio, 1994; Ribolzi et al., 1996). The values of the concentration factor are calculated as the ratio of a reference tracer content in sample i and in the most diluted sample j in the analysed set of samples. The concentration diagrams were constructed by using electrical conductivity (EC) as a reference tracer:

$$\text{CF}_i = \text{EC}_i / \text{EC}_j$$

where EC_i and EC_j are the respective values of electrical conductivity in the analysed set of samples. Chloride was not used as a reference tracer, although this ion is generally considered to be conservative in groundwater systems, because a significant amount is believed to originate from halite dissolution.

It is apparent from Fig. 7a that sulphate concentration increases steadily during the process of groundwater mineralization, although not so much as EC. The slope of the calcium curve is smaller than that of sulphate. On the other hand, the Ca^{2+} versus SO_4^{2-} plot (Fig. 8) reveals more pronounced losses of Ca^{2+} with respect to SO_4^{2-} . This indicates removal of calcium from the aqueous

phase, most probably due to calcite precipitation. This hypothesis is supported by apparent over-saturation with respect to calcite (cf. Table 1).

As seen in Fig. 7a, alkalinity remains roughly constant for the values of the concentration factor $\log(\text{CF}) < 0.75$. Above it, the measured alkalinity gradually increases, most probably due to stronger impact of incongruent dissolution of dolomite and gypsum, leading to partial precipitation of calcite in the system (Back et al., 1983; Plummer et al., 1990; Cardenal et al., 1994) and increase of SO_4^{2-} content in the solution.

High Na^+ and Cl^- contents detected in the studied aquifers may suggest dissolution of halite as the governing process. The Na–Cl relationship, which has often been used to identify the mechanisms of acquiring salinity in arid and semi-arid regions, shows parallel enrichment in both ions for majority of the analyzed waters (Fig. 7b) pointing to dissolution of this mineral and/or enrichment of dissolved phase due to partial evaporation of water. These two mechanisms dominate in the studied region. The data points in Fig. 9 with Na^+/Cl^- ratios significantly higher than one may indicate presence of cation-exchange processes in the studied aquifers, in addition to dissolution of halite.

5.2. Stable isotopes of water

Stable isotope composition of the analysed groundwater samples reveals large variability, particularly in the Mio-Pliocene aquifer, suggesting significant differences in the origin as well as in groundwater evolution. Oxygen-18 and deuterium content in this system varies between -2.6 and -7.8‰ and from -18.2 to -58.8‰ , respectively, with the mean values equal -5.1 and -40.5‰ . For Senonian and Turonian aquifers the corresponding mean values are more negative (-6.3 and -51‰ and -6.8 and -45.1‰ , respectively). Senonian and Turonian aquifers reveal substantially smaller spread of $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values. For the Senonian system they range from -4.5 to -7.8‰ and from -46.5 to -57.1‰ , respectively. The corresponding ranges for the Turonian aquifer are equal: -5.8 to -7.2‰ , and -31.9 to -54‰ (Table 1).

The $\delta^{18}\text{O}$ and $\delta^2\text{H}$ data are presented in Fig. 10 in relation to the global meteoric water line (GMWL) and the Sfax meteoric water line (SMWL). Sfax meteorological station is the nearest IAEA/WMO GNIP station for which there is a continuous record of monthly mean $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values in precipitation covering the period 1992–2010 (IAEA/WMO, 2006). The long-term weighted mean values of Sfax precipitation are -4.4‰ for $\delta^{18}\text{O}$ and -24.9‰ for $\delta^2\text{H}$ (Abid, 2010). The local meteoric water line is defined by the equation: $\delta^2\text{H} = 8 \cdot \delta^{18}\text{O} + 13$ (Abid et al., 2009).

The $\delta^2\text{H}$ – $\delta^{18}\text{O}$ diagram shows that almost all data points representing the Mio-Pliocene aquifer are situated below the GMWL, except two points (wells No 5 and 13). Water samples from these wells were collected in the Mio-Pliocene sands in the Dahar and Gabes regions. Their position on the $\delta^2\text{H}$ – $\delta^{18}\text{O}$ diagram suggests the presence of an admixture of recently recharged water (Fig. 10). The data points representing other samples from this aquifer plot along an estimated evaporation line with a slope of 5.5. The intersection of this line with GMWL gives the values of -8 and -54‰ for $\delta^{18}\text{O}$ and $\delta^2\text{H}$, respectively, suggesting that precipitation recharging this system has isotopic signature significantly depleted in heavy isotopes when compared to Sfax rainfall.

As seen in Fig. 10, all data points representing Senonian aquifer are located below the GMWL and cluster along the regression line of $\delta^2\text{H} = 2.4 \cdot \delta^{18}\text{O} - 36$ ($R^2 = 0.52$). Very low slope of this regression line suggest specific conditions during evaporation, such as evaporation from the water table through unsaturated zone in discharge areas (Chotts).

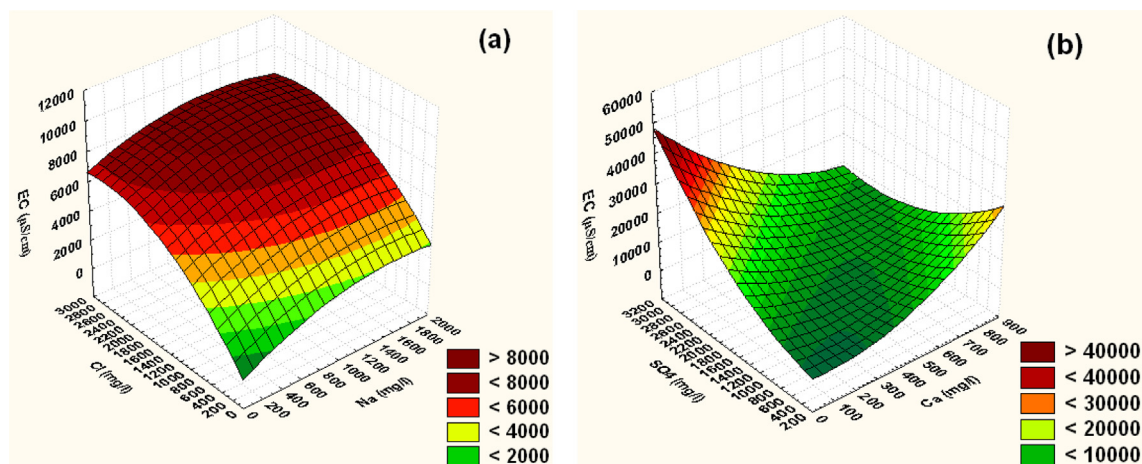


Fig. 3. 3D plot of EC ($\mu\text{S}/\text{cm}$) against Cl^- and Na^+ (a); and SO_4^{2-} and Ca^{2+} (b).

The Turonian samples plot slightly below or on GMWL and SMWL lines in the $\delta^2\text{H}$ – $\delta^{18}\text{O}$ diagram. The samples originating from the Dahar upland reveal more positive $\delta^{18}\text{O}$ and $\delta^2\text{H}$ values falling into the range of recent recharge (dashed-line ellipse in Fig. 10). Admixture of young local component probably originates from inflows via faults and/or fractures. This is confirmed by measurable tritium content (1.3 TU in well No. 19) and by a relatively high ^{14}C content (37 pmc in well No. 19).

The deuterium-excess (d-excess), defined as $d = \delta^2\text{H} - 8 \cdot \delta^{18}\text{O}$ (Dansgaard, 1964) is a useful proxy for identifying secondary processes influencing isotopic composition of water in the evaporation–condensation cycle in nature (e.g. Craig, 1961; Merlivat and Jouzel, 1979; Gat et al., 1994; Machavaram and Krishnamurthy, 1995). The d-excess values of the analysed groundwater samples range from -10.9 to 15.9‰ (cf. Table 1). Relatively high d-excess values of Turonian samples indicate rapid infiltration of rainwater and absence of evaporation processes. In contrast, the d-excess values of Senonian and Mio-Pliocene samples scatter widely,

pointing to importance of evaporation processes, both during infiltration and discharge. The observed variability of d-excess value may partly stem also from spatial and temporal changes of this parameter in precipitation. As the analysed groundwater samples cover wide temporal scale, from modern to glacial time, significant changes in the isotopic composition of rainfall in the study area in the past are very likely.

5.3. Assessing timescales of groundwater flow

Assessment of the mean residence time of groundwater in the studied aquifers has been made using radiocarbon. The use of radiocarbon for this purpose requires careful examination of the origin of carbon in the total dissolved inorganic carbon pool (TDIC). The dissolved carbon is mainly in the form of HCO_3^- .

As the aquifers contain carbonates, at least two sources of dissolved carbon in the solution have to be taken into account: (i) biogenic CO_2 with $\delta^{13}\text{C}$ values characteristic for C_3 vegetation in the study area (range from ca. -26‰ to -22‰ – Guendouz, 1985), and

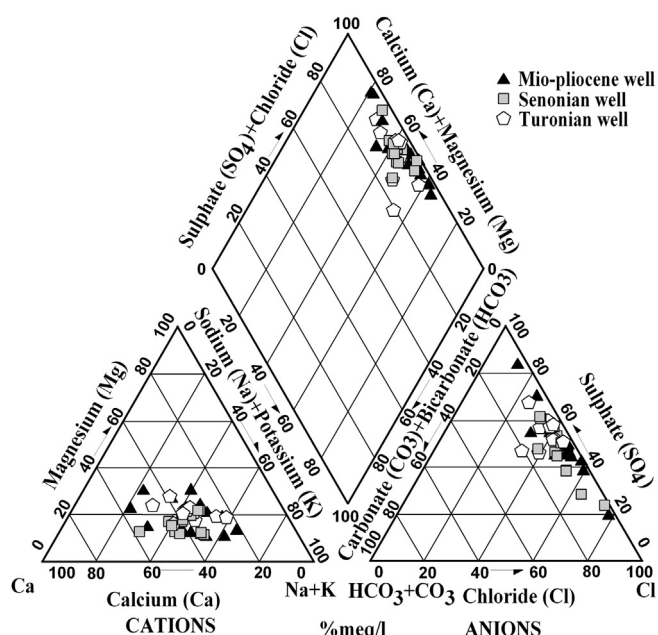


Fig. 4. Piper diagram for groundwater samples.

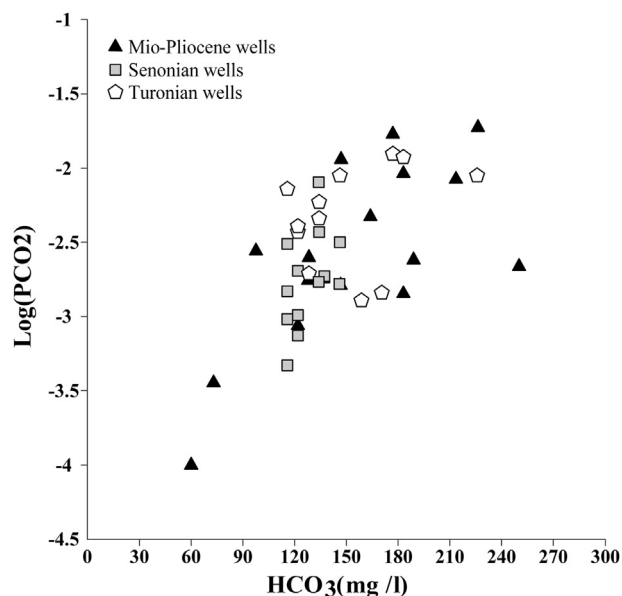


Fig. 5. Relationship between $\log(\text{pCO}_2)$ calculated using Diagrammes computer code (Simler, 2006) and measured HCO_3^- content in the analysed groundwater samples.

(ii) mineral carbon with a $\delta^{13}\text{C}$ content of approximately -2‰ . While marine carbonates should have ^{13}C content close to that of V-PDB reference ($\delta^{13}\text{C} \approx 0\text{‰}$), this slightly lower value was selected due to possible contribution of secondary carbonates. Water in the studied aquifers evolves from a state close to saturation in calcite towards a distinct oversaturation (cf. Table 1). This oversaturation results from dissolution of dolomite and gypsum accompanied by precipitation of calcite (dedolomitisation reactions). This secondary cement should have $\delta^{13}\text{C}$ values ranging from $\delta^{13}\text{C}$ of TDIC pool (-10 to -11‰) to approximately 0‰ (Edmunds and Smedley, 2000; Mook and Vries, 2000; Aeschbach-Hertig et al., 2002). An additional potential source of carbon for TDIC could be fossil organic matter in clastic rocks of the Mio-Pliocene reservoir.

In order to assess the origin of carbon in the solution, $\delta^{13}\text{C}$ of CO_2 in equilibrium with the measured TDIC pool ($^{13}\text{C}_{\text{gas eq}}$) has been calculated using the following expression:

$$\delta^{13}\text{C}_{\text{gas eq}} = \delta_T - \frac{a \cdot \epsilon_{a-g} + b \cdot \epsilon_{b-g} + c \cdot \epsilon_{c-g}}{\text{mTDIC}} \quad (1)$$

where δ_T is the ^{13}C content of the TDIC pool; a, b and c stand for molarities of H_2CO_3 , HCO_3^- , and CO_3^{2-} respectively; ϵ_{a-g} , ϵ_{b-g} , ϵ_{c-g} are the fractionation factors between the carbon species at equilibrium (Mook, 1968); mTDIC is the total dissolved inorganic carbon content (mol l^{-1}).

Three groups of groundwater can be distinguished taking into account the range of calculated values for $\delta^{13}\text{C}_{\text{gas eq}}$, TDIC content and its carbon isotope composition (Table 2). In the first group $\delta^{13}\text{C}_{\text{TDIC}}$ and TDIC content do not correlate and $\delta^{13}\text{C}_{\text{gas eq}}$ varies from -16.5 to -19.5‰ , close to the carbon isotope signature of soil CO_2 . The carbon isotopic composition of TDIC pool in these waters should result from equilibrium with biogenic CO_2 gas under open system conditions. For this group, the calculation of the mean residence times of water needs only the application of the radioactive decay. This group includes shallow waters which have relatively high ^{14}C activities, between 51.5% and 89.5% of modern carbon.

The second group is characterized by less negative $\delta^{13}\text{C}$ values of TDIC (from -3.6 to -7.8‰), and the calculated $\delta^{13}\text{C}_{\text{gas eq}}$ higher than -14‰ . This higher value of $\delta^{13}\text{C}_{\text{gas eq}}$ is an indication of

Table 2
Carbon isotope composition of TDIC pool in the analysed groundwater samples.

N°	Wells	T°C	pH	log pCO ₂	H ₂ CO ₃ (mmol/l)	HCO ₃ (mmol/l)	CO ₃ (mmol/l)	$\delta^{13}\text{C}$ (‰ VPDB)	$\delta^{13}\text{C}_{\text{gas eq}}^a$ (‰ VPDB)	$\delta^{14}\text{C}$ (pmc)
Open System										
1	Mbarek Tatta	23.0	7.3	-2.15	0.26	2.32	0.002	-12.25	-19.49	85.5
2	P.public El Feji	21.6	8.1	-2.76	0.06	3.22	0.017	-10.76	-18.88	86
3	Mghatta	23.0	8.8	-3.89	0.00	1.20	0.032	-10.27	-18.38	88.6
4	Source Hzeim 1	18.2	7.1	-2.05	0.37	1.87	0.001	-10.58	-17.66	51.5
5	Abdallah Mansouri	20.2	7.0	-1.83	0.57	2.50	0.001	-10.77	-17.46	89.5
6	Taher Abbas	22.9	7.6	-2.44	0.13	2.06	0.003	-9.23	-16.85	85
7	Kamel Rhouma	23.0	7.1	-1.84	0.52	2.82	0.002	-9.96	-16.69	87.5
8	Merzougui Dehibi	21.8	8.1	-2.85	0.05	2.95	0.017	-8.55	-16.67	53.8
Mixed System										
9	Aouinet 4	24.1	7.9	-2.70	0.07	2.54	0.010	-7.89	-15.68	10.5
10	Amra	25.8	8.0	-2.79	0.05	2.37	0.011	-7.92	-15.57	16
11	Jorf Aquaculture	29.8	7.6	-2.59	0.08	1.49	0.003	-8.33	-15.34	10.3
12	Bir Touil	20.0	8.3	-3.44	0.01	1.18	0.010	-6.96	-15.33	75
13	Chettaoui Med	22.4	8.2	-3.10	0.03	1.81	0.012	-7.21	-15.28	38
14	Khalfallah	28.3	7.5	-2.18	0.21	2.74	0.004	-8.24	-15.23	3.7
15	Zmertem	27.2	7.5	-2.11	0.25	3.20	0.004	-8.08	-15.16	11.4
16	Od khalifa Mahmoud	29.1	7.8	-2.62	0.07	2.38	0.008	-7.88	-15.13	18.5
17	Godfa 2 bis	22.0	7.9	-2.81	0.06	1.87	0.006	-6.88	-14.86	16.2
18	Zridib	22.7	8.1	-2.93	0.04	2.26	0.012	-6.73	-14.75	16.9
19	Beni Zelten 3	23.7	7.2	-1.96	0.38	2.54	0.002	-7.78	-14.66	37
20	Oued Sandoug	27.2	8.1	-2.91	0.04	2.52	0.017	-6.98	-14.55	25.7
21	Hsay 5bis	22.3	8.1	-3.10	0.03	1.52	0.008	-6.47	-14.53	20.7
22	Zoumit	26.7	7.4	-2.31	0.16	1.81	0.002	-7.32	-14.37	30.2
23	Rmathie 2	31.0	7.6	-2.52	0.09	1.79	0.004	-7.43	-14.35	15.8
24	Negga Sonède	28.0	8.3	-3.15	0.02	1.86	0.017	-6.78	-14.29	16.4
25	Negga 7	25.0	7.7	-2.56	0.09	2.09	0.005	-6.70	-14.25	11.6
26	Sidi Marzoug	18.2	7.7	-2.69	0.08	1.71	0.003	-5.89	-14.12	52
Closed System										
27	Henchir Jehha	25.0	7.6	-2.50	0.11	1.70	0.003	-6.49	-13.89	9.5
28	El Mthinin	21.5	7.8	-2.75	0.07	1.90	0.006	-5.88	-13.88	29.7
29	Chenchou 1bis	41.6	7.7	-2.55	0.07	1.74	0.006	-7.84	-13.83	8.8
30	Oued zira 2	26.3	7.9	-2.90	0.04	1.58	0.006	-6.21	-13.78	6.9
31	Sabria Mol	24.0	7.9	-2.78	0.06	2.03	0.007	-5.90	-13.69	6.3
32	Beni Zelten Sonède	22.7	7.2	-2.01	0.36	2.49	0.002	-6.56	-13.60	37
33	Zarcine 4	25.5	7.9	-2.78	0.06	1.99	0.008	-5.77	-13.41	11.6
34	Rass el Aïoun	31.9	7.3	-2.14	0.21	1.99	0.002	-6.85	-13.29	23.5
35	Garet Ltaifa	23.3	7.2	-2.23	0.21	1.56	0.001	-6.22	-13.25	12.5
36	Ambar	31.9	8.0	-2.83	0.04	2.13	0.012	-6.01	-13.06	16.7
37	Matmata 5	20.2	7.5	-2.47	0.13	1.68	0.002	-5.21	-12.97	11.4
38	El Faouar 4	23.0	8.1	-3.07	0.03	1.68	0.010	-4.59	-12.58	9.6
39	Gourai	31.0	7.5	-2.42	0.11	1.80	0.003	-5.46	-12.29	11
40	Sidi Daoud	31.5	8.0	-2.79	0.05	2.06	0.010	-5.21	-12.28	12.5
41	Jemma 1bis	23.0	8.4	-3.40	0.01	1.58	0.018	-3.60	-11.67	10.8
42	Ziret Ouled Touati	25.3	7.3	-2.15	0.24	1.92	0.002	-4.49	-11.40	11.2

^a $\delta^{13}\text{C}$ value of CO_2 in equilibrium with the measured TDIC pool, calculated using eq. (1).

isotopically heavy carbon introduced into the TDIC pool through the dedolomitisation processes. For this group, specific corrections of the initial ^{14}C content of the TDIC pool are necessary before the age based on radioactive decay of radiocarbon is calculated (Fontes and Garnier, 1979; Fontes, 1992). This group comprises the deepest groundwater, located in the northern and southern parts of the study area.

The samples classified in Table 2 as “mixed system” reveal intermediate values of the discussed parameters suggesting that groundwater evolved under intermediate conditions with respect to CO_2 gas, or that these samples represent true mixtures of waters of different origin.

The relationship between ^{14}C and ^{13}C content in TDIC pool of the analysed groundwater samples is shown in Fig. 11. The data points clearly illustrate a general tendency: the decreasing ^{14}C activity is typically accompanied by an increase of $\delta^{13}\text{C}$ value. The data points are spread between the two end-members in the ^{14}C – ^{13}C space: the cluster of points representing samples collected in the recharge area and the range of $^{14}\text{C}/^{13}\text{C}$ values representing carbonates associated with the matrix of the studied aquifer systems. This diagram clearly illustrates gradual evolution of carbon isotopic composition of the TDIC pool, both in terms of ^{14}C and ^{13}C content, resulting from interactions with aquifer matrix and radioactive decay of ^{14}C . Apparent shift of some data points out of the evolution path in Fig. 11, towards more negative $\delta^{13}\text{C}$ values (arrows), can be explained by dissolution of secondary carbonate cements depleted in ^{13}C , present in the system, and/or bacterial decomposition of fossil organic matter (lignite).

Any attempt to quantify the evolution of TDIC pool in groundwater from the perspective of radiocarbon content should consider processes which may potentially add “dead” carbon to the solution, reducing the concentration of ^{14}C in the system, in addition to radioactive decay. In the context of the studied aquifer systems, the most important process of this type is incongruent dissolution of dolomite. The processes in which dead carbon is added to the system, which is equivalent to removal of ^{14}C from DTIC pool, are described in terms of appropriate correction factors (q) applied to the initial radiocarbon concentration in the soil CO_2 gas in the recharge areas of the studied systems. The age of analysed

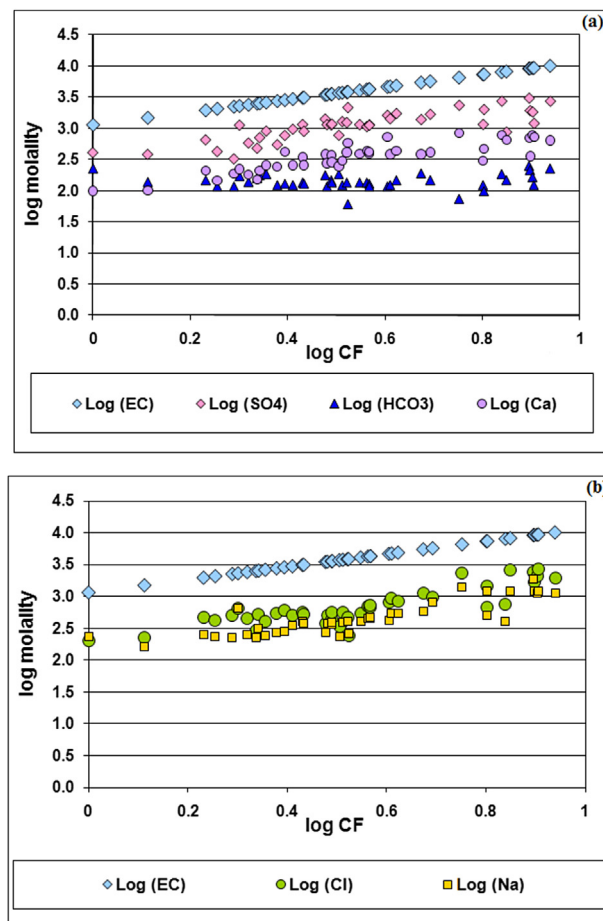


Fig. 7. Concentration diagrams for selected physico-chemical parameters of the analysed groundwater samples (see text for details).

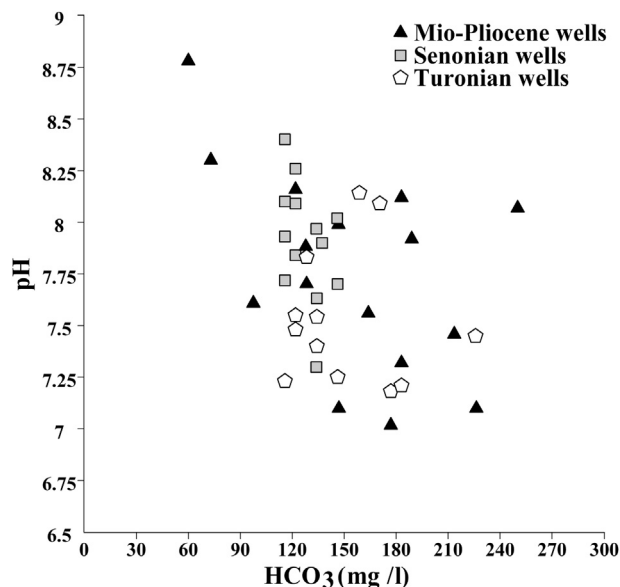


Fig. 6. Relationship between measured pH and HCO_3 content in the analysed groundwater samples.

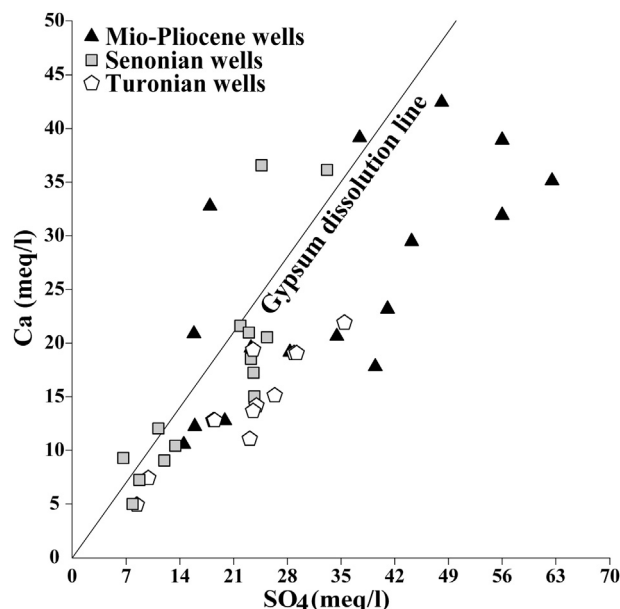


Fig. 8. Relationship between Ca^{2+} and SO_4^{2-} content in the analysed groundwater samples.

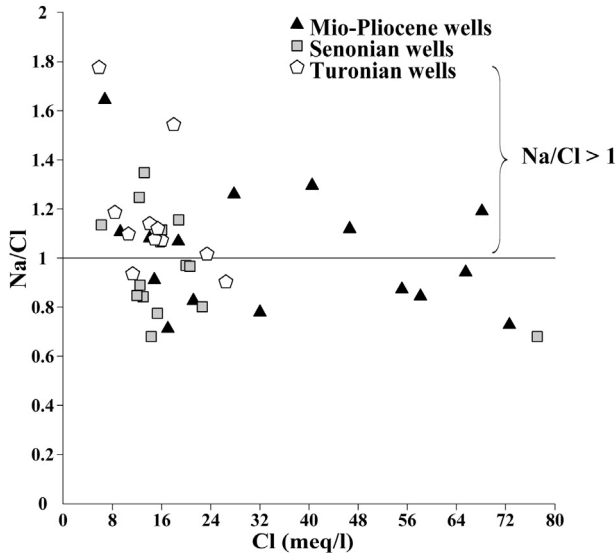


Fig. 9. Relationship between the ratio of Na^+/Cl^- and Cl^- content in the analysed groundwater samples.

groundwater samples can then be expressed by the following equation (Clark and Fritz, 1997):

$$t = -8267 \cdot \ln \left(\frac{a^{14}\text{C}_{\text{TDIC}}}{q \cdot a_0^{14}\text{C}} \right) \quad (2)$$

Where $a^{14}\text{C}_{\text{TDIC}}$ is the measured radiocarbon content in the analysed sample and $a_0^{14}\text{C}$ is the radiocarbon activity in the soil CO_2 in the recharge area(s). It is usually assumed to be equal 100 pmc. q stands for the correction factor accounting for ^{14}C -dilution processes in the system.

The analysed groundwater samples generally reveal excess of Mg^{2+} over that recorded in the potential recharge areas, pointing to importance of incongruent dissolution of dolomite as a source of this ion. Consequently, appropriate ^{14}C correction factor associated with this process can be defined (Clark and Fritz, 1997):

$$q_{\text{dol}} = \frac{\text{mTDIC}_{\text{rech}}}{\text{mTDIC}_{\text{rech}} + 2 \cdot \text{mMg}_{\text{excess}}} \quad (3)$$

where $\text{mTDIC}_{\text{rech}}$ stands for molar fraction of TDIC pool in the recharge water while $\text{mMg}_{\text{excess}}$ is the excess of Mg defined as $\text{mMg}_{\text{excess}} \approx \text{mMg}_{\text{meas}} - \text{mMg}_{\text{rech}}$ where mMg_{rech} stands for average Mg content in the recharge.

An independent approach to assess radiocarbon age of groundwater relies on intimate link between ^{14}C and ^{13}C signatures of TDIC pool during geochemical evolution of this reservoir (cf. Fig. 11). This approach utilizes the existing isotope contrast between TDIC pool and carbonate reservoir present in the aquifer matrix. Appropriate age correction factor based on ^{13}C signatures of the interacting reservoirs of carbon has been proposed (Ingerson and Pearson, 1964; Pearson, 1965; Clark and Fritz, 1997):

$$q_{^{13}\text{C}} = \frac{\delta^{13}\text{C}_{\text{TDIC}} - \delta^{13}\text{C}_{\text{carb}}}{\delta^{13}\text{C}_{\text{rech}} - \delta^{13}\text{C}_{\text{carb}}} \quad (4)$$

Where $\delta^{13}\text{C}_{\text{TDIC}}$ stands for the measured ^{13}C content in TDIC pool, $\delta^{13}\text{C}_{\text{carb}}$ is the ^{13}C signature of solid carbonate interacting with the solution and $\delta^{13}\text{C}_{\text{rech}}$ is the isotopic composition of TDIC in the

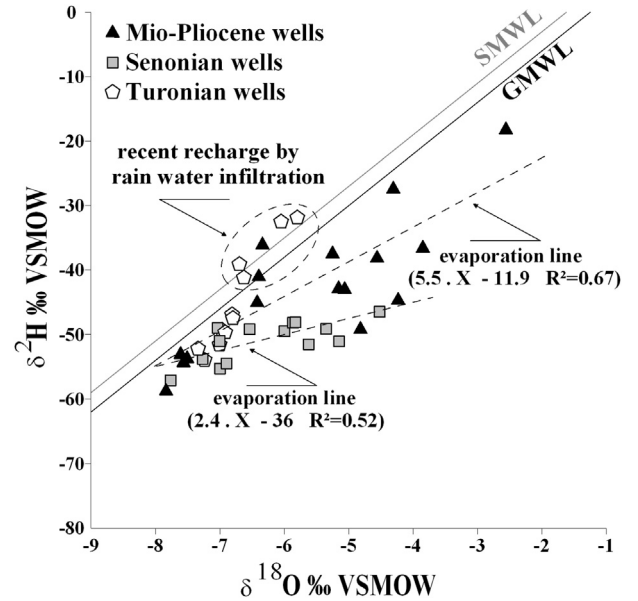


Fig. 10. $\delta^2\text{H}-8$ $\delta^{18}\text{O}$ relationship for the analysed groundwater samples.

recharge water. This composition can be approximated by the following equation:

$$\delta^{13}\text{C}_{\text{rech}} = \delta^{13}\text{C}_{\text{soil}} + \varepsilon^{13}\text{C}_{\text{TDIC}-\text{CO}_2(\text{soil})} \quad (5)$$

where $\delta^{13}\text{C}_{\text{soil}}$ represents ^{13}C content of soil CO_2 and $\varepsilon^{13}\text{C}_{\text{TDIC}-\text{CO}_2(\text{soil})}$ is the pH-dependent isotope enrichment between soil CO_2 and TDIC pool.

The $\delta^{13}\text{C}$ mixing model can handle open- and closed-system conditions. Under open-system conditions in the recharge area, both ^{14}C and ^{13}C signature of TDIC reservoir will fully equilibrate with soil CO_2 and will make the initial radiocarbon content of TDIC

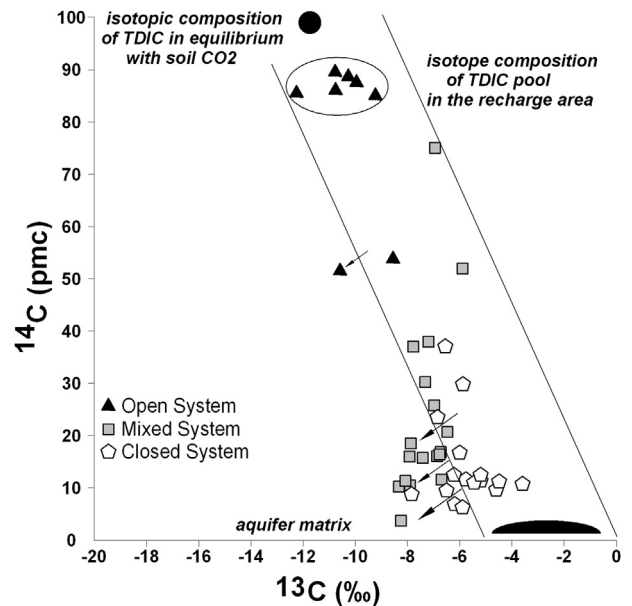


Fig. 11. Relationship between ^{14}C and ^{13}C isotope composition of TDIC pool in the analysed groundwater samples.

pool practically indistinguishable from its concentration in the atmospheric CO₂. In such situations, the correction factor defined by equation (4) will describe subsequent ¹⁴C-dilution effect due to dissolution of carbonates under closed-system conditions taking place in the aquifer. The weak point of this approach is its sensitivity to the adopted value of the isotope enrichment between soil CO₂ and TDIC pool in the recharge area (Clark and Fritz, 1997).

The apparent radiocarbon ages calculated using eqs. (2) and (4) are listed in Table 3 along with the values calculated using eqs. (2) and (3). It has been assumed that the TDIC pool in the recharge area has been fully equilibrated with soil CO₂ having the radiocarbon content of 100 pmc (open-system conditions). There is a general agreement between both approaches, although in some cases the differences in the estimated ages reach several thousand years.

than 5000 years in the recharge area of Dahar upland can be explained by the difference in hydraulic heads of the CI and the Turonian aquifer.

6. Conclusions

Comprehensive study of isotopic and chemical composition of groundwater from Mio-Pliocene, Senonian and Turonian aquifer systems in southern Tunisia has been carried out in order to gain insight into groundwater dynamics and the processes governing water quality in the region.

Deuterium and ¹⁸O isotope composition of the analysed water samples reveal substantial variability of these parameters across the region. In particular, stable isotope data showed that waters

Table 3
Comparison of radiocarbon ages of the analysed groundwater samples calculated assuming different correction models.

N°	Wells	a ¹⁴ C pcm	δ ¹³ C ‰ vs VPDB	q _{dol}	q _{13c}	¹⁴ C(q _{13c}) age (10 ³ years)	¹⁴ C(q _{dol}) age (10 ³ years)	Average ¹⁴ C age (10 ³ years)
1	Mbarek Tatta	85.5	−12.25	0.31		Recent	Recent	Recent
2	P.public El Feji	86	−10.76	0.35		Recent	Recent	Recent
3	Mghatta	88.6	−10.27	0.48		Recent	Recent	Recent
4	Source Hzeim 1	51.5	−10.58	0.91		0.07	0.19	0.13
5	Abdallah Mansouri	89.5	−10.77	0.68		Recent	Recent	Recent
6	Taher Abbas	85	−9.23	0.66		Recent	Recent	Recent
7	Kamel Rhouma	87.5	−9.96	0.37		Recent	Recent	Recent
8	Merzougui Dehibi	53.8	−8.55	0.85		Recent	Recent	Recent
9	Aouinet 4	10.5	−7.89	0.60		11.0	8.9	9.9
10	Amra	16	−7.92	0.51		7.4	4.0	5.7
11	Jorf Aquaculture	10.3	−8.33	0.93		11.2	12.9	11.5
12	Bir Touil	75	−6.96	0.51		Recent	Recent	Recent
13	Chettaoui Med	38	−7.21	1.00		Recent	24	1.2
14	Khalfallah	3.7	−8.24	0.78		20.0	20.0	20.0
15	Zmertem	11.4	−8.08	1.00		10.7	12.8	11.7
16	Od khalifa Mahmoud	18.5	−7.88	1.00		6.0	8.5	7.2
17	Godfa 2 bis	16	−6.88	1.00		6.1	9.6	7.8
18	Zridib	16.9	−6.73	0.82		5.6	7.5	6.6
19	Beni Zelten 3	37	−7.78	0.78		4.0	1.4	2.7
20	Oued Sandoug	25.7	−6.98	1.00		2.5	5.6	4.0
21	Hsay 5bis	20.7	−6.47	0.55		3.4	2.5	3.0
22	Zoumit	30.2	−7.32	0.75		1.3	2.4	1.8
23	Rmathie 2	15.8	−7.43	1.00		6.8	9.9	8.3
24	Negga Sonède	16.4	−6.78	1.00		5.8	9.3	7.6
25	Negga 7	11.6	−6.7	1.00		8.6	12.4	10.5
26	Sidi Marzoug	52.00	−5.89	1.00		Recent	0.1	Recent
27	Henchir Jehha	9.5	−6.49	0.76		9.9	11.9	10.9
28	El Mthinin	29.7	−5.88	0.91		Recent	3.8	1.9
29	Chenchou 1bis	8.8	−7.84	1.00		12.0	14.7	13.3
30	Oued zira 2	6.9	−6.21	1.00		12.1	16.6	14.3
31	Sabria Mol	6.3	−5.9	1.00		12.6	17.4	15.0
32	Beni Zelten Sonède	37	−6.56	0.79		Recent	1.5	0.7
33	Zarcine 4	11.6	−5.77	1.00		7.4	12.3	9.8
34	Rass el Aioun	23.50	−6.85	1.00		2.9	7.0	4.9
35	Garet Ltaifa	12.5	−6.22	0.85		7.2	11.0	9.1
36	Ambar	16.70	−6.01	1.00		4.7	9.2	6.9
37	Matmata 5	11.4	−5.21	0.49		6.6	6.9	6.7
38	El Faouar 4	9.6	−4.59	1.00		7.0	13.8	10.4
39	Gourai	11	−5.46	1.00		7.2	13.0	10.1
40	Sidi Daoud	12.5	−5.21	1.00		5.9	11.6	8.7
41	Jemma 1bis	10.8	−3.6	1.00		3.9	12.7	8.3
42	Ziret Ouled Touati	11.2	−4.49	0.60		5.6	9.0	7.3

Based on mean radiocarbon ages reported in Table 3, a contour map has been constructed (Fig. 12). It allows some insight into the dynamics of groundwater flow in the study area. The map suggests the existence of recent water in the central part of the basin. Wells located in this part of the basin clearly show increase in ¹⁴C activities and a decrease in ¹³C content. This isotopic signature seems to be related to the seepage of recent component to the shallower aquifer. The existence of groundwater with residence time higher

representing the Senonian aquifer have some significant evaporation history related to evaporation during recharge and/or during discharge, via direct evaporation from the water table. Analysis of chemical data showed that chemical composition of groundwater in the studied aquifers is controlled by several processes: calcite and dolomite dissolution, calcite precipitation, evaporite dissolution and cation-exchange. The ¹³C content of the TDIC pool shows evidence of specific chemical and isotopic

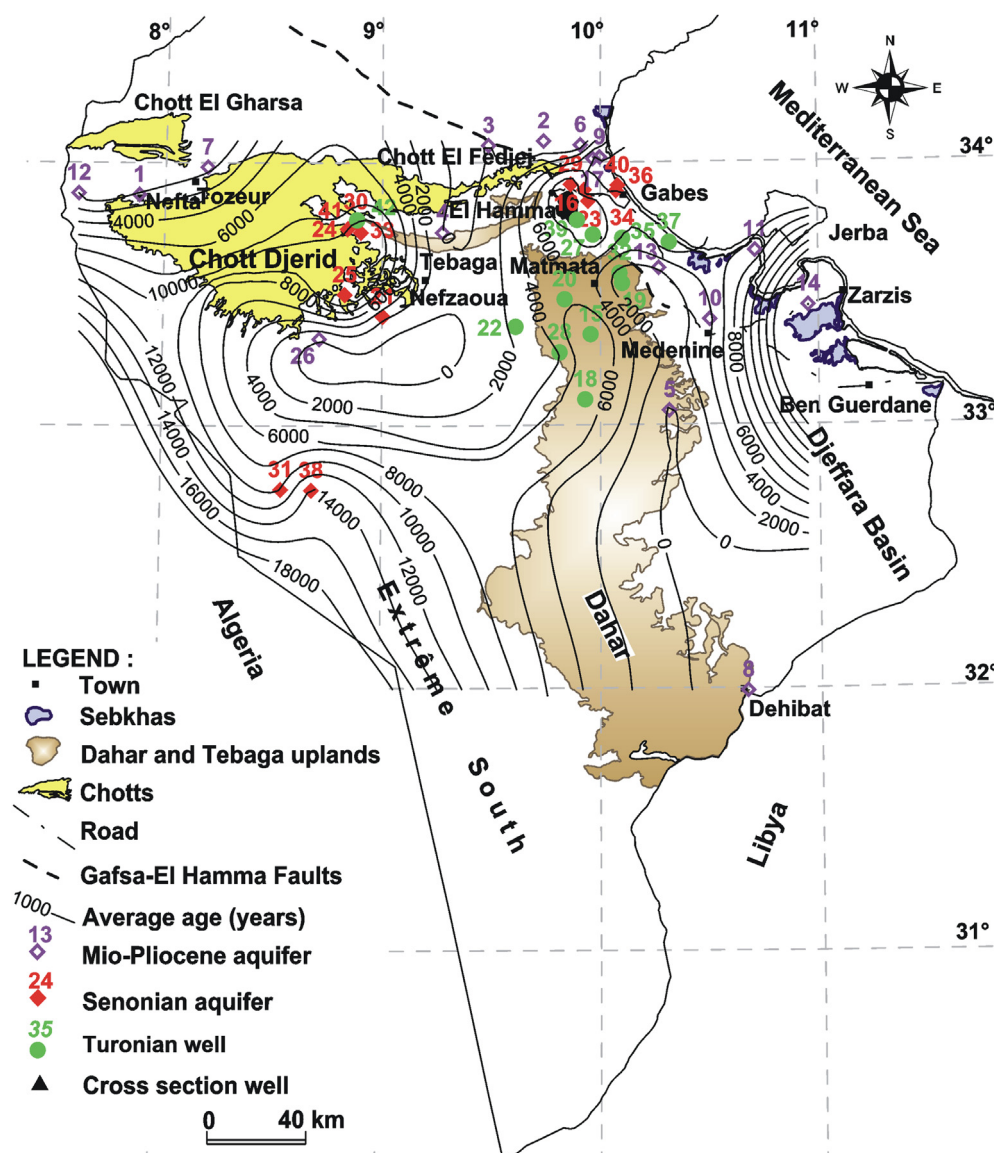


Fig. 12. Average age of the Mio-Pliocene, Senonian and Turonian aquifers (years).

evolution of groundwater within the studied aquifers, both with respect to water–rock interaction and mineral equilibrium in groundwater.

The studied aquifers are characterized by a wide range of groundwater ages, from present-day to the Late Pleistocene. Old groundwaters with radiocarbon ages in excess of ca. 4 ka are the least mineralised and evaporated. Recent groundwater with radiocarbon ages of less than ~4 ka, are more evaporated and mineralised.

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