



## Environmental isotopes and noble gases in the deep aquifer system of Kazan Trona Ore Field, Ankara, central Turkey and links to paleoclimate

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### ABSTRACT

Environmental isotopes and noble gases in groundwater samples from the Kazan Trona Ore Field are studied to establish the temperature change between the Holocene and the late Pleistocene. Noble gas temperatures (NGTs) presented in this study add an important facet to the global paleotemperature map in the region between Europe and North Africa. The groundwater system under investigation consists of three different aquifers named shallow, middle and deep in which  $\delta^{18}\text{O}$  and  $\delta^2\text{H}$  vary from  $-8.10\text{‰}$  to  $-12.80\text{‰}$  and from  $-60.89\text{‰}$  to  $-92.60\text{‰}$  VSMOW, respectively. The average isotopic depletion between unconfined and confined parts of the system is  $-2.5\text{‰}$  in  $\delta^{18}\text{O}$  and  $-20\text{‰}$  in  $\delta^2\text{H}$ . It is not possible to explain this depletion solely with the elevation effect. Recharge temperatures derived from dissolved atmospheric noble gases reflect the current average yearly ground temperatures ( $13^\circ\text{C}$ ) for samples collected near the recharge area but are  $3$  to  $8^\circ\text{C}$  lower than today's temperatures in the deep aquifer system. Low  $^{14}\text{C}$  activities and high He excesses in the confined parts of the aquifer system suggest that the water in the deep aquifer was recharged during the last Pleistocene under considerably cooler climatic conditions.

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### Introduction

Understanding regional sensitivity to global climate change is crucial for predicting the response to warming caused by human-induced increases in greenhouse gas concentrations. The reconstruction of past climate variations can help to answer this question (Alley et al., 2003).

The “groundwater noble gas thermometer” is based on the temperature dependence of the solubility of noble gases in water. Noble gas concentrations in groundwater from confined aquifers have been used to reconstruct paleotemperature records primarily for the last glacial period (Mazor, 1972; Andrews and Lee, 1979; Stute and Deak, 1989; Stute et al., 1992, 1995a, 1995b; Stute and Schlosser, 1993; Clark et al., 1997; Beyerle et al., 1998; Aeschbach-Hertig et al., 2002; Hall et al., 2005; Castro et al., 2007; Klump et al., 2008; Blaser et al., 2010; Alvarado et al., 2011; Varsanyi et al., 2011). These studies have

shown that selected confined aquifers can be considered as low-pass filtered continental paleoclimate archives. The length of the records is limited by the available dating methods (primarily radiocarbon), as well as diffusion and dispersion within the archive.

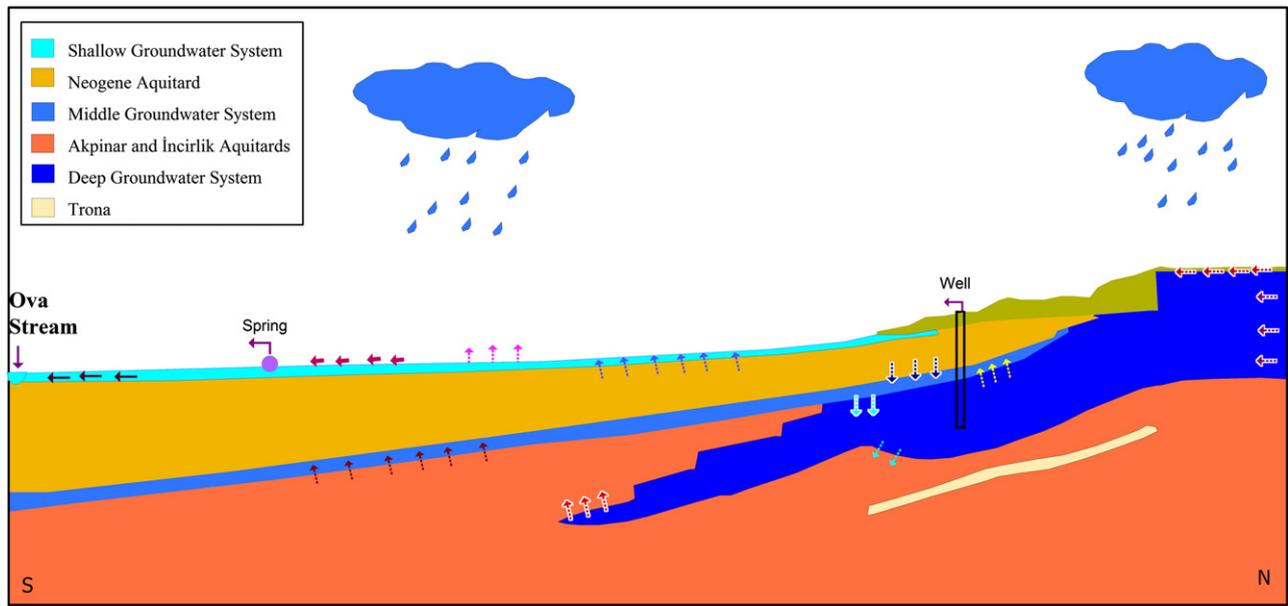
A confined aquifer system near Ankara, Turkey was studied in the context of exploitation of a trona deposit. Trona ( $\text{Na}_2\text{CO}_3 \cdot \text{NaHCO}_3 \cdot 2\text{H}_2\text{O}$ ) is a rare, naturally occurring mineral refined to soda ash used mainly in glass manufacturing. Kazan trona is found in the Eocene deposits of the Kazan Basin and will be extracted by using solution mining technology. After the discovery of the deposit, groundwater resources in the area were characterized hydrogeologically and hydrogeochemically to assess the potential impacts of solution mining on the aquifer system (Yazicigil et al., 2001, 2009; SRK, 2004; Camur et al., 2008). To provide supporting evidence for the mixing mechanisms within the complex hydrogeologic system, consisting of three different aquifers above the deposit, a multi-tracer approach was utilized ( $\delta^{18}\text{O}$ ,  $\delta^2\text{H}$ ,  $\delta^{13}\text{C}$  and  $^{14}\text{C}$  of DIC,  $^3\text{H}$ , CFC-11 and CFC-12, He and Ne) and groundwater residence times were established (Arslan, 2008; Arslan et al., 2012). Arslan et al. (2012) used the residence times to derive information needed for the development of strategies for sustainable groundwater management throughout the area, since significant amounts of water from local aquifers would be required during solution

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**Figure 3.** Cross-section showing the shallow, middle and deep groundwater systems (arrows indicate the groundwater flow directions). Modified from Yazıcıgil et al. (2009).

saline zone in the fractured rock system affects the groundwater quality in this system (SRK, 2001).

## Methods

### Field sampling

Three field trips were carried out between the summers of 2006 and 2007 in the Kazan Basin. A total of 28 representative samples were collected from the groundwater wells monitoring different depths ranging from 4 m to 540 m below ground surface. In situ pH, temperature, electrical conductivity and dissolved oxygen measurements were carried out and the samples were collected after the field parameters reached equilibrium to ensure that they represent the water in the aquifer. 60-ml Boston round clear glass bottles with 20-mm sized poly seal screw caps were used to collect samples for stable isotope analysis, 250 ml Boston round clear glass bottles with 24 mm screw caps were used to collect samples for  $^3\text{H}$ ,  $^{13}\text{C}$  and  $^{14}\text{C}$  analysis. 0.2 ml of saturated  $\text{HgCl}_2$  solution was added to  $^{13}\text{C}$  and  $^{14}\text{C}$  samples immediately after collection to prevent post-collection biological activity from changing the carbon isotopic composition. During noble gas sampling, 50 cm long, 1-cm outer diameter copper tubes fixed in aluminum channels were sealed on both ends by stainless steel pinch-off clamps (Weiss, 1968). The formation of gas bubbles was reduced by attaching a back-pressure valve assembly; the details of the procedure are explained in Stute et al. (1995a).

Another field trip was carried out in May 2010 to collect samples from the geological units hosting the deep aquifer system (Eocene Mulk Formation) and a total of 13 samples were collected and analyzed for U and Th concentrations. The sampling locations are presented in Figure 2.

### Laboratory analysis

Stable isotope ratios  $\delta^{18}\text{O}$  and  $\delta\text{D}$  were measured in the Environmental Isotope Laboratory of University of Waterloo (Drimmie and Heemskerk, 2001; Drimmie et al., 2001), Canada, by using a Micromass ISOPRIME Continuous Flow Isotope Ratio Mass spectrometer (CF-IRMS). Analytical uncertainties of the individual measurements are estimated to be within  $\pm 0.2\%$  for  $\delta^{18}\text{O}$  and  $\pm 0.8\%$  for  $\delta\text{D}$ .

Tritium concentrations were determined by the  $^3\text{He}$ -ingrowth method (Clarke et al., 1976). The samples were degassed and flame-sealed in glass bulbs where they were stored for more than 4 months. After storage, the  $^3\text{He}$  which had accumulated from tritium decay was extracted and admitted to a VG-5400 He isotope mass spectrometer at Lamont-Doherty Earth Observatory (Ludin et al., 1997). The analytical precision and the detection limit of the tritium measurements was about  $\pm 0.1$  TU (tritium units, 1 TU is equivalent to a  $^3\text{H}/^1\text{H}$  ratio of  $10^{-18}$ ).

$^{13}\text{C}/^{12}\text{C}$  and  $^{14}\text{C}/^{12}\text{C}$  ratios were measured at the National Ocean Sciences Accelerator Mass Spectrometry Facility (NOSAMS) of Woods Hole Oceanographic Institution by converting Dissolved Inorganic Carbon (DIC) to  $\text{CO}_2$ .  $^{13}\text{C}/^{12}\text{C}$  analyses were carried out by using a VG PRISM Stable Mass Spectrometer with a precision of  $\pm 0.05\%$ .  $^{14}\text{C}/^{12}\text{C}$  analyses were performed by using an accelerator mass spectrometry (AMS) system following the procedures described in WHOI (1989). The results are given as percent modern carbon (pmc) and the analytical precision and detection limits are reported to be within  $\pm 1\%$  and 0.07 pmc, respectively.

The dissolved gases were extracted from the copper tubes on a vacuum line and the individual noble gases were separated and analyzed by using a multi-purpose noble gas mass spectrometer (MAP 215–50) at L-DEO. The procedure has been described in detail by Stute et al. (1995a). The precision of the measurements is about  $\pm 1$  to 2% for noble gas abundances and helium isotope ratios.

Concentrations of total U and Th in aquifer material samples were measured at Brooklyn College (City University of New York) with inductively coupled plasma mass spectrometry (ICP-MS). The samples were prepared as follows: 0.05 g of powdered rock sample was placed in Teflon vessel and dissolved in 3 ml HF + 1 ml  $\text{HNO}_3$  + 1 ml  $\text{H}_2\text{O}$  solution for 2 days on a hot plate ( $180^\circ\text{C}$ ). The solution was then completely evaporated and 2.5 ml  $\text{HNO}_3$  and 2.5 ml DI water were added. This new solution was left on the hot plate at  $100^\circ\text{C}$  for 2 days. This process was repeated and 0.5 ml of the final solution was added to 9.5 ml DI water followed by measurement in ICP-MS.

## Results

Results of the chemical data collected in the field, isotopic analyses and noble gas isotope measurements are reported in Tables 1, 2 and 3,

respectively. Electrical conductivity (EC) values range from 526 to 23,902  $\mu\text{Scm}^{-1}$ , the highest values observed in both shallow and deep systems (Table 1). pH values vary between 7.53 and 12.29, with the highest values found in the middle aquifer system. Dissolved oxygen contents of the samples from the shallow system range from 0.1 to 10  $\text{mg l}^{-1}$ , and for the deeper systems it is  $<0.05 \text{ mg l}^{-1}$  (Table 1).

Stable isotope ratios of the water molecule within the aquifer systems cover a wide range of values from  $-8.10\%$  to  $-12.80\%$  and from  $-60.89\%$  to  $-92.60\%$  V-SMOW for  $\delta^{18}\text{O}$  and  $\delta\text{D}$ , respectively. Isotopically lighter values are found in the deep aquifer system (Table 2). An isotopic contrast is observed between the shallow and deep aquifer systems (Fig. 4). Most of the  $\delta^{18}\text{O}$  and  $\delta\text{D}$  samples from the deep aquifer system fall along the global meteoric water line (GMWL) whereas most samples from the shallow unconfined aquifer and a few of the confined aquifer plot to the right of the meteoric water line (Fig. 4).

$\delta^{13}\text{C}$  values for DIC show an enrichment from shallow to deeper aquifers and range from  $-8.65\%$  near the recharge area to values up to  $+3.52\%$  in the deep aquifer system (e.g., samples D-57A, D-47 and D-8; Table 2).  $^{14}\text{C}$  activities are close to the detection limits (0.07 pmc) for some of the samples from the deep system (e.g., samples D-57A, D-47; Table 2).

The concentrations of noble gases were measured in a total of 15 samples, 8 of which are from the shallow system and the rest from the deep, confined aquifer system (Table 3). Helium concentrations vary between  $5 \times 10^{-8}$  and  $1.5 \times 10^{-5} \text{ cm}^3 \text{ STP g}^{-1}$ .  $^3\text{He}/^4\text{He}$  ratios range between  $0.13 \times 10^{-6}$  and  $1.71 \times 10^{-6}$ . The highest  $^3\text{He}/^4\text{He}$  ratios were found in water from the shallow aquifer, the lowest in the deep aquifer (Table 3).

Results of the U and Th concentrations in aquifer rocks are presented in Table 4. U and Th concentrations range from 0.85 ppm to 2.54 ppm and from 2.02 ppm to 5.20 ppm, respectively.

## Calculation of noble gas temperatures

The interpretation of noble gases in groundwater requires correction of the “excess air” component (Heaton and Vogel, 1981). There are three models proposed to explain the formation of excess air, which are total dissolution (TD) of trapped bubbles (Andrews and Lee, 1979; Stute and Schlosser, 1993), partial re-equilibration (PR) with the atmosphere after total dissolution (Stute et al., 1995b), and closed-system equilibration (CE) of groundwater with partially dissolved entrapped air which are discussed in detail in Aeschbach-Hertig et al. (2000). In this study, these models were fitted by the inverse algorithm of Aeschbach-Hertig et al. (1999) to the noble gas data of 15 samples obtained from the aquifers above the Kazan Trona ore field. Here, only the CE model equation presented by Aeschbach-Hertig et al. (2000) is given (Eq. (1)).

$$C_i(T, S, P, A, F) = C_i^*(T, S, P) + \frac{(1-F)Az_i}{1 + FAz_i/C_i^*} \quad (i = \text{He, Ne, Ar, Kr, Xe}) \quad (1)$$

where  $C_i^*(T, S, P)$  defined as the moist air solubility equilibrium concentrations as functions of temperature, salinity ( $\text{g kg}^{-1}$ ) and atmospheric pressure (atm) (Weiss, 1970, 1971; Weiss and Kyser, 1978; Clever, 1979),  $z_i$  are the noble gas volume fractions in dry air.  $F$  shows the fractionation of excess air and  $A$  denotes the initial amount of dry entrapped air per unit mass of water.

Equation (1) is not applicable to the observed He for the aquifer system in this study due to the presence of non-atmospheric He. The parameter  $S$  can be estimated as:  $S = 0\%$  for the meteoric recharge water although in two of the samples (S-4 and D-47) EC values are up to  $14,069 \mu\text{S cm}^{-1}$  corresponding to a salinity value of  $\sim 10 \text{ g kg}^{-1}$ . A salinity correction should be carried out for these two samples, assuming that during the measurements there is salt in the samples

**Table 1**  
Monitoring well and field chemical data. Aquifer SU = shallow unconfined, SNU = shallow Neogene unconfined, MC = middle confined, DC = deep confined, IA = Incirlik Aquitard, N.D. = not determined.

Sample	Depth to screened zone [m]		Aquifer	Coordinates		Cond. [ $\mu\text{S/cm}$ ]	Temp. [ $^{\circ}\text{C}$ ]	pH	DO [mg/l]
	Top	Bottom		Northing	Easting				
S-2	8	20	SU	458866.8	4439755.7	742	13.4	7.94	2.0
S-3	8	20	SU	459986.4	4438608.1	726	14.3	7.53	3.0
S-4	5	9	SU	461000.0	4439113.0	14069	15.2	7.64	0.7
S-5	8	20	SU	460231.6	4440080.6	625	13.9	7.96	10.0
S-9	7	15	SU	461520.5	4439964.2	2673	13.6	8.32	0.1
S-11	4	12	SU	463142.2	4441246.7	816	14.6	7.71	0.4
S-16	7	23	SU	459264.3	4441443.7	725	11.1	7.62	5.5
S-19	4	20	SNU	457306.5	4439795.9	610	13.3	7.90	5.0
S-20	4	20	SNU	457763.0	4439650.9	526	12.1	7.75	5.0
S-22A	13	21	SNU	457680.0	4439963.3	684	12.8	7.67	5.0
D-8	158.99	188.74	DC	458803.4	4442308.8	3130	18.4	9.61	0.0
D-13	361.79	391.54	DC	461261.3	4444346.2	920	13.5	8.91	0.0
D-20	191.96	221.71	DC	458931.9	4444438.2	879	15.0	7.57	N.D.
D-33	151.43	163.33	DC	457656.3	4439808.0	1759	15.7	8.74	0.0
D-37	211.78	241.53	DC	460556.8	4444364.9	880	16.9	8.96	0.0
D-47	72	372	DC	458608.2	4441977.3	5486	17.2	9.24	0.05
D-53	546	576	DC	461207.0	4442103.7	6354	12.3	9.20	N.D.
D-57A	300.71	318.62	DC	458876.9	4439740.0	6707	17.2	8.01	0.0
M-57B	132.09	144.03	MC	458875.4	4439742.4	2360	12.8	8.81	0.0
D-60A	280	292	DC	457814.6	4440764.3	525	13.0	8.77	N.D.
M-60B	114	126	MC	457814.6	4440764.3	573	12.7	9.34	0.0
D-63A	199	211	DC	459207.3	4443039.4	606	18.1	7.91	0.0
D-63B	110.5	122.5	DC	459207.3	4443039.4	641	14.0	8.87	0.0
M-65B	101	117	MC	458254.2	4441658.1	4175	13.9	12.24	N.D.
D-68R	63	83	DC	458768.0	4442660.0	739	15.7	8.53	0.0
I-74A	520	544	IA	459431.0	4441041.0	6919	10.7	8.06	N.D.
D-74B	282	294	DC	459431.0	4441041.0	23902	7.8	8.41	N.D.
M-74C	78	96	MC	459431.0	4441041.0	6455	12.4	12.29	N.D.

**Table 2**  
Isotopic composition of groundwater from the Kazan Trona aquifers. N.D. = not determined.

Sample	$\delta^{18}\text{O}$ VSMOW [‰]	$\delta^2\text{H}$ VSMOW [‰]	Rech. elev. (m)	$^3\text{H}$ TU	DIC [mmol/l]	$\delta^{13}\text{C}$ VPDB [‰]	$\text{A}^{14}\text{C}$ [pmc]	Alkalinity [mg/l]	$\text{a}_0^{14}\text{C}_{\text{DIC}}$ [pmc]	q	Corrected age <sup>a</sup> ( $^{14}\text{C}$ ka BP)
S-2	-9.39	-69.21	1050	2.47	N.D.	N.D.	N.D.	366	N.D.	N.D.	N.D.
S-3	-9.56	-68.61	1050	6.05	N.D.	N.D.	N.D.	431	N.D.	N.D.	N.D.
S-4	-8.1	-64.57	1050	0.13	9.07	-8.65	29.79	499	68	0.41	-0.6
S-5	-9.67	-68.55	1050	6.96	N.D.	N.D.	N.D.	392	N.D.	N.D.	N.D.
S-9	-10.04	-73.88	1050	0.02	16.39	-8.17	24.00	747	68	0.39	0.9
S-11	-9.86	-69.13	1050	N.D.	N.D.	N.D.	N.D.	388	N.D.	N.D.	N.D.
S-16	-9.88	-69.42	1050	6.58	N.D.	N.D.	N.D.	416	N.D.	N.D.	N.D.
S-19	-8.1	-60.89	1050	14.22	N.D.	N.D.	N.D.	464	N.D.	N.D.	N.D.
S-20	-9.3	-67.91	1050	4.96	N.D.	N.D.	N.D.	374	N.D.	N.D.	N.D.
S-22A	-8.86	-65.51	1050	4.44	N.D.	N.D.	N.D.	463	N.D.	N.D.	N.D.
D-8	-12.31	-88.60	1250	0.05	12.79	2.28	0.17	2163	68	0.06	25.8
D-13	-12.30	-90.07	1250	0.11	12.51	-2.30	0.45	674	68	0.20	28.3
D-20	-10.60	-74.70	1250	0.32	8.06	-7.01	27.81	409	68	0.36	-1.2
D-33	-12.90	-90.80	1250	N.D.	N.D.	N.D.	N.D.	2702	N.D.	N.D.	N.D.
D-37	-12.37	-89.60	1250	0.07	12.01	-3.00	0.29	622	68	0.23	32.8
D-47	-12.50	-92.12	1250	N.D.	69.50	2.76	0.15	N.D.	68	0.04	23.8
D-53	-8.59	-68.20	1250	N.D.	N.D.	N.D.	N.D.	5639	N.D.	N.D.	N.D.
D-57A	-12.80	-90.30	1250	0.00	95.72	3.52	0.08	5406	68	0.02	21.9
M-57B	-12.70	-92.60	1250	N.D.	N.D.	N.D.	N.D.	2542	N.D.	N.D.	N.D.
D-60A	-12.50	-90.99	1250	0.07	N.D.	N.D.	N.D.	398	N.D.	N.D.	N.D.
M-60B	-12.54	-90.60	1250	0.09	7.07	-6.64	0.39	485	68	0.34	33.9
D-63A	-11.00	-78.70	1250	0.05	7.80	-7.74	3.70	386	68	0.38	16.0
D-63B	-11.40	-82.23	1250	N.D.	N.D.	N.D.	N.D.	453	N.D.	N.D.	N.D.
M-65B	-9.72	-72.90	1250	N.D.	N.D.	N.D.	N.D.	2324	N.D.	N.D.	N.D.
D-68R	-12.25	-89.00	1250	0.04	7.71	-6.33	0.80	391	68	0.33	27.7
I-74A	-11.90	-90.10	1250	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.
D-74B	-8.65	-67.80	1250	N.D.	N.D.	N.D.	N.D.	991	N.D.	N.D.	N.D.
M-74C	-10.00	-76.90	1250	N.D.	N.D.	N.D.	N.D.	2408	N.D.	N.D.	N.D.

<sup>a</sup>  $^{14}\text{C}$  ages are calculated according to Pearson and Hanshaw (1970) assuming  $\delta^{13}\text{C}$  values of -27‰ and +4‰ for soil  $\text{CO}_2$  and carbonates, respectively and  $^{14}\text{C}$  activity of 0 pmc for carbonates.

besides water. Nevertheless, the salinity is ignored due to the fact that this correction would only change the temperatures by less than 0.27°C.

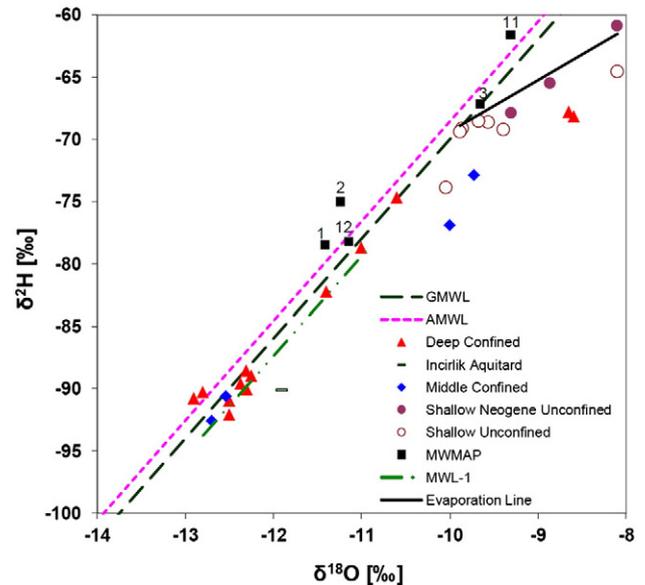
In Eq. (1), the parameter P depends on the recharge elevations. For the samples from the deep system the recharge elevations vary between 1150 m and 1400 m where Eocene units outcrop corresponding to P values between 0.871 atm and 0.845 atm, respectively. For the shallow aquifer, recharge elevations change between 1050 and 900 m where alluvium units outcrop corresponding to P values between 0.882 atm and 0.898 atm. The sensitivity of the NGTs to elevation is small, a change in assumed elevation of the recharge area from 1150 to 1400 m (250 m) results in a 0.7°C lower recharge temperature equivalent to an altitude effect of 0.3°C/100 m. Therefore, the

recharge elevations are assumed to be the average elevations where units outcrop and the calculations were carried out based on these average elevations (1050 m for the shallow system, 1250 m for the deep system).

Equation (1) was solved for the unknown parameters by inverse modeling using four measured concentrations (Ne, Ar, Kr, Xe) with

**Table 3**  
Noble gas data.

Sample	He $10^{-8} \text{ cm}^3$ STP $\text{g}^{-1}$	Ne $10^{-7} \text{ cm}^3$ STP $\text{g}^{-1}$	Ar $10^{-4} \text{ cm}^3$ STP $\text{g}^{-1}$	Kr $10^{-8} \text{ cm}^3$ STP $\text{g}^{-1}$	Xe $10^{-8} \text{ cm}^3$ STP $\text{g}^{-1}$	$^3\text{He}/^4\text{He}$ $10^{-6}$
S-2	5.042	2.008	3.540	7.948	1.133	1.71
S-3	1204.265	2.128	3.411	7.815	1.109	0.28
S-4	22.216	2.034	3.264	7.275	1.020	0.43
S-5	4.586	1.992	3.269	7.504	1.007	1.39
S-9	754.540	2.289	3.681	8.405	1.182	0.22
S-11	131.669	2.408	3.637	8.007	1.108	1.15
S-16	5.082	2.078	3.678	8.067	1.117	1.43
S-19	5.391	2.182	3.337	7.872	1.050	1.78
D-8	31.906	3.058	4.635	10.019	1.409	0.40
D-13	37.983	2.853	4.342	9.716	1.401	0.32
D-33	86.419	2.668	4.208	9.224	1.280	0.13
D-37	33.305	2.815	4.340	9.767	1.410	0.34
D-47	1531.444	2.843	4.254	9.589	1.284	0.20
D-63A	11.247	3.018	4.313	9.532	1.473	0.90
D-68R	12.136	3.044	4.548	10.119	1.517	0.76



**Figure 4.**  $\delta^2\text{H}$  vs  $\delta^{18}\text{O}$  diagram. GMWL=Global Meteoric Water Line (Craig, 1961); AMWL=Ankara Meteoric Water Line equation calculated from data in IAEA/WMO, 2004 ( $\delta^2\text{H}=8*\delta^{18}\text{O}+11.42$ ); MWL-1=Meteoric Water Line equation  $\delta^2\text{H}=8*\delta^{18}\text{O}+8.6$  calculated from data in this study; MWMAP (monthly weighted mean of Ankara precipitation; the numbers 1, 2, 3, 11 and 12 refer to calendar months January, February, March, November and December, respectively).

**Table 4**  
Aquifer material U and Th concentration data.

Geologic unit	Lithology	Sample ID	UTM coordinates		Th (ppm)	U (ppm)	
			x	y			
Asmalidere member	Consolidated green mudstone	F-7	459226	4445303	5.20	1.51	
	Loose green mudstone	F-8	459224	4445300	4.68	1.36	
	Green mudstone	F-9	459229	4445308	3.89	1.00	
	Organic laminated mudstone	F-10	459229	4445308	3.23	1.25	
	Nodular limestone	F-11	459229	4445308	2.95	1.11	
	Clayey limestone	F-12	459229	4445308	4.07	2.54	
	Green mudstone with organic matter	F-13	459229	4445308	5.61	1.31	
	Green mudstone	F-14	459229	4445308	3.65	1.37	
	Fethiye member	Yellowish green mudstone	IN-3	458541	4442937	2.02	0.85
		Cream beige mudstone	IN-4	458541	4442937	3.89	1.68
Yellow-cream-beige sandy mudstone		IN-5	458614	4442908	3.31	1.38	
Yellow-colored limestone		IN-6	458655	4442935	2.37	1.59	
Yellow-colored mudstone		IN-7	458659	4442914	4.86	1.69	

three free parameters ( $T, A, F$ ) (Aeschbach-Hertig et al., 2000). According to this non-linear least squares method, a  $\chi^2$ -test is used to test the quality of fit of different models to describe the data within their uncertainty, and the model parameters that minimize  $\chi^2$  are found (Aeschbach-Hertig et al., 1999; Ballentine and Hall, 1999). One sample yielding a probability  $p < 0.01$  was rejected (i.e., sample D-63A);  $p$  is the probability to obtain a  $\chi^2$ -value equal or higher than the observed value with a correct model; see Table 5). All other samples provided acceptable fits with the PR- and CE-models (14 samples out of 15,  $\chi^2 = 17.4$ ;  $p = 0.23$  for CE-model and  $\chi^2 = 22.6$   $p = 0.07$  for PR-model). The CE model better describes the noble gas data set from the Kazan aquifer system with a minimum  $\chi^2$  value. In Table 5, as defined by Aeschbach-Hertig et al. (2000),  $F$  is equal to  $v/q$  where  $v$  is the ratio of the entrapped gas volumes in the final and initial stage and  $q$  is the ratio of the dry gas pressure in the trapped gas to that

**Table 5**  
Results of fitting Eq. (2) to the measured concentrations of Ne, Ar, Kr, Xe.

Sample	$\chi^2$	$p$	$T$ [°C]	$A$ [cm <sup>3</sup> ·STP/g]	$F$	$q$	$v$	$\Delta\text{Ne}$ [%]	Radiogenic He [10 <sup>-8</sup> cm <sup>3</sup> ·STP/g]
S-2	0.31	0.58	11.63 ± 0.6	0.02	0.82	1.11	0.91	16.06	0.33 ± 0.12
S-3	3.25	0.07	11.76 ± 0.4	0.00	0.00	1.11	0.00	23.20	1195 ± 24.1
S-4	0.09	0.77	14.26 ± 0.4	0.00	0.13	1.11	0.14	20.06	17.26 ± 0.46
S-5	1.52	0.22	14.70 ± 0.5	0.01	0.73	1.12	0.81	16.47	-0.18 ± 0.11
S-9	0.17	0.68	10.19 ± 0.4	0.01	0.34	1.17	0.40	30.41	750 ± 15.1
S-11	0.01	0.93	12.40 ± 0.5	0.01	0.39	1.24	0.48	40.14	126 ± 2.63
S-16	1.04	0.31	13.89 ± 4.2	0.11	0.81	1.20	0.97	21.40	0.17 ± 0.12
S-19	5.93	0.02	13.44 ± 0.4	0.00	0.00	1.14	0.00	26.73	0.09 ± 0.16
D-8	0.29	0.59	6.16 ± 0.5	0.02	0.40	1.45	0.58	69.74	24.62 ± 0.65
D-13	0.05	0.82	5.42 ± 0.4	0.01	0.27	1.32	0.35	56.85	31.11 ± 0.77
D-33	0.00	0.95	8.54 ± 0.5	0.02	0.51	1.36	0.69	51.80	80.08 ± 1.73
D-37	0.03	0.87	5.18 ± 0.4	0.01	0.27	1.30	0.36	54.78	26.49 ± 0.68
D-47	2.81	0.09	8.45 ± 0.5	0.02	0.42	1.40	0.59	61.33	1523.2 ± 30.6
D-63A	19.68	0.00	4.11 ± 0.3	0.01	0.00	1.29	0.00	63.82	4.05 ± 0.29
D-68R	1.93	0.16	3.18 ± 0.3	0.01	0.01	1.31	0.01	63.17	4.58 ± 0.3

$\chi^2$  is the sum of the weighted squared deviations between modeled and measured concentrations. (Expected value is one, the number of degrees of freedom, since there are four constraints (Ne, Ar, Kr, Xe) and three free parameters ( $T, A, F$ ) for each sample).

$p$  is the probability for  $\chi^2$  to be equal to or larger than the actual value due to random errors, although the model is correct. Fits with  $p < 0.01$  are rejected (values in italics).

$T$  is the model parameter for the equilibration temperature (noble gas temperature, NGT).

$A$  is the model parameter for the STP-volume of initially entrapped air per mass of water.

$F = v/q$  is the fractionation parameter, showing the reduction of the entrapped gas volume.

$v$  is the remaining fraction of the entrapped gas volume after partial dissolution.

$q$  is the dry gas pressure in the entrapped gas relative to that in the atmosphere.

$\Delta\text{Ne}$  (%) =  $(\text{Ne}_{\text{meas}}/\text{Ne}_{\text{eq}} - 1) \times 100\%$  is the relative Ne excess above solubility equilibrium.

Radiogenic He is the excess He above the atmospheric He components predicted by the model.

in the free atmosphere. The noble gas temperatures (NGTs) calculated by using the CE-model range from 3.2°C to 14.7°C, the lower ones belonging to the samples from the deep aquifer system (Table 5).

Helium excess values are presented as radiogenic <sup>4</sup>He (excess He above the atmospheric He components predicted by the model; Table 5). <sup>4</sup>He excesses vary from  $-0.17 \times 10^{-8}$  cm<sup>3</sup> STP g<sup>-1</sup> to  $1523 \times 10^{-8}$  cm<sup>3</sup> STP g<sup>-1</sup>.

## Discussion

### Noble gas temperatures

The samples from the recharge area (S-4, S-5, S-16 and S-19) show an average noble gas temperature of  $14.1 \pm 0.4$ °C. This value is close to today's ground temperature for the recharge area (~13°C) at an elevation of 1050 m. The average ground temperature is calculated by using the daily soil temperature measurements taken 50 cm below ground between 1941 and 2006 from a station in Ankara (ground temperature: ~14°C; elevation: 891 m; Turkish State Meteorological Service, personal communication), after correcting this temperature for the decrease in temperature from 891 m to 1050 m. In Ankara station, ground temperatures are 2.2°C warmer than mean annual air temperatures.

NGTs calculated for samples S-2, S-3 and S-9, which are from the shallow system but under the influence of mixing with deeper waters as suggested by the geochemical and tracer data (SRK, 2004; Arslan et al., 2012), are 1.2 to 2.8°C lower than the ground temperatures at 1050 m (13°C).

The samples from the deep system are about 3 to 8°C lower than today's ground temperature for the recharge area (11.7°C for a recharge elevation of 1250 m), indicating that some groundwater recharge occurred to the confined parts of this system under colder climatic conditions than at present.

### Stable isotopes

For the shallow aquifer samples, the deviation from the GMWL is most likely caused by evaporative enrichment as the depth to groundwater is between 2 and 15 m in those wells (less than 2 m in well S-9 and around 15 m in S-5). Such evaporative enrichment

frequently occurs during or after recharge in arid regions (e.g., Clark and Fritz, 1997).

According to the relationship between  $\delta^{18}\text{O}$  and elevation obtained by Apaydin (2004) for Bey pazari Trona Ore field, which is located 50 km west of the study area, the depletion in  $\delta^{18}\text{O}$  is about 0.44‰ per 100 m rise in altitude. In our study area, the average isotopic depletion in  $\delta^{18}\text{O}$  between shallow and deep systems is 2.5‰ and it is not possible to explain these isotopically lighter values in samples from the deeper groundwater systems solely with the altitude effect based on Apaydin's (2004) study. In fact, the highest topographical elevation in the recharge area is not as high as the elevation that would be obtained from  $\delta^{18}\text{O}$  data if their minimum values would be explained by the altitude in the recharge area of the shallow aquifer only. Therefore, it has been concluded that the lighter stable isotope values reflect that the recharge elevation of the deep aquifer system is higher than the shallow system (altitude effect) and the recharge temperatures in the deep aquifer system are lower than that in the shallow system (temperature effect). Since the elevation difference between shallow and deep systems is around 200 m, about 0.9‰ of the depletion would be due to the elevation changes and the remainder (about 1.6‰) is a lower limit for the climate-induced change. Consequently, the differences in stable isotope ratios of groundwater from the shallow and deeper systems indicate that deep groundwaters were recharged under climatic conditions different from today's.

The temperature effect is one of the climate-related factors influencing the  $\delta^{18}\text{O}$  values and has to be taken account when interpreting  $\delta^{18}\text{O}$  data. This effect can be estimated using modern values obtained from the monitoring station of the IAEA/WMO network located in Ankara. Monthly mean values of  $\delta^{18}\text{O}$  and temperature at this station are correlated ( $r=0.57$ ) with a slope of  $0.32 \pm 0.01\%$   $^{\circ}\text{C}^{-1}$  (IAEA/WMO, 2004). Such strong relationships between  $\delta^{18}\text{O}$  and temperature can also be observed in other regions (e.g., Kabul, Tashkent, Irkutsk, Zagreb, and Krakow) that are under the influence of continental climate conditions similar to that of Ankara (IAEA, 1992).

A meteoric water line (MWL-1) with a deuterium excess (d) of 8.7‰ can be constructed by using the stable isotope data of samples D-13, D-37, D-47, D-60A, and D-63B from the deep aquifer system with an  $r^2$  value of 0.98 (Fig. 4). The d value is somewhat different from that of modern Ankara precipitation, which is about 11.4‰. On a global basis the d-value averages about 10‰ (Craig, 1961). However, it changes regionally due to variations in source of humidity, wind speed and sea-surface temperature (SST) (Clark and Fritz, 1997). According to Clark and Fritz (1997), when humidity is about 85% stable isotope values for precipitation plot very close to the global meteoric water line. It is hard to determine the deuterium excess values in precipitation of the Holocene or Pleistocene epochs with our present data set. However, in general, the deuterium excess seems to be more or less uniform throughout the record, suggesting a constant circulation regime of the atmosphere over the study area for the past 35  $^{14}\text{C}$  ka BP as suggested by Rozanski (1985). The general cooling during the late Pleistocene might be accompanied by a change in source of water vapor recharging the aquifer.

#### Radiocarbon dating

The radioactive isotope of carbon,  $^{14}\text{C}$ , is extensively used in dating of dissolved inorganic carbon (DIC) in groundwater ( $\text{DIC} = \text{CO}_{2(\text{aq})} + \text{HCO}_3^- + \text{CO}_3^{2-}$ ) as the atmospheric  $^{14}\text{C}$  dissolved in the precipitation reaches the groundwater table and starts decaying to nitrogen in the groundwater system. Its relatively long half-life of 5730 yr allows dating of ages up to about 40,000 yr. To successfully date a groundwater sample by radiocarbon dating, the initial activity of  $^{14}\text{C}$  should be known (Clark and Fritz, 1997). Estimation of the initial activity of  $^{14}\text{C}$  ( $a_0^{14}\text{C}$ ) requires

radiocarbon data of a sample from the recharge area, which is proven to be modern but tritium-free pre-bomb water (Kazemi et al., 2006). During the hydrogeology studies for conceptual understanding of the study area, Yazicigil et al. (2001) gathered isotope data from some of the springs and wells. One of the sampled springs, SP-2, is a seasonal spring responding to precipitation immediately and ceases to flow during the dry season. For this spring, the  $^3\text{H}$  amount,  $^{14}\text{C}$  activity, and the  $\delta^{13}\text{C}$  value were reported in February, 2001 to be 7.58 TU, 68 pmc and  $-8.96\%$ , respectively (Yazicigil et al., 2001). This modern sample contains tritium and there would be a slight overestimation of the initial condition for older groundwaters if 68 pmc is used as  $a_0^{14}\text{C}$ , because of the presence of bomb  $^{14}\text{C}$ . Being aware of this fact,  $a_0^{14}\text{C}$  was assumed to be 68 pmc for all the samples during calculations (Table 2).

To simply use the first-order kinetic rate law for decay to date a groundwater sample, the final concentration of  $^{14}\text{C}$  in groundwater should have only been affected by radioactive decay, although it can be modified by isotopic exchange from recrystallization and precipitation reactions. In our study area, there is evolution of DIC in the groundwaters and the influence of this evolution on  $\delta^{13}\text{C}$  values can be observed in Fig. 5. The evolution of DIC and the enrichment in  $\delta^{13}\text{C}$  between shallow and deep aquifer systems implies isotope exchange between groundwater and the aquifer matrix. There is possible dissolution of carbonate minerals adding dead carbon to the groundwater, resulting in a lower  $^{14}\text{C}$  ratio for the sample. Therefore,  $^{14}\text{C}$  ages need to be corrected for carbonate dissolution. There are different approaches to correct apparent  $^{14}\text{C}$  ages and the most appropriate approach depends on the geochemical system and the data available.  $^{13}\text{C}$  is a good tracer of open and closed system evolution of DIC in groundwater; the large difference in  $\delta^{13}\text{C}$  between the soil-derived DIC and carbonate minerals in the aquifer provide a reliable measure of  $^{14}\text{C}$  dilution by carbonate dissolution. Therefore, in this study a  $\delta^{13}\text{C}$  mixing model was used (Pearson, 1965; Pearson and Hanshaw, 1970) to correct our  $^{14}\text{C}$  ages. This model is strongly dependent on recharge conditions affecting the  $^{13}\text{C}$  enrichment factor during dissolution of  $\text{CO}_2$  and the evolution of  $\delta^{13}\text{C}_{\text{DIC}}$ . Any process that adds to, removes from, or exchanges carbon with the DIC pool altering the  $^{14}\text{C}$  concentrations will also affect its  $^{13}\text{C}$  values. A dilution factor (q) as suggested by Pearson and Hanshaw (1970) was employed during our calculations as a measure of carbonate dissolution in the aquifer. This factor can be calculated by using Eq. (2). In this equation,  $\delta^{13}\text{C}_{\text{DIC}}$  is measured  $^{13}\text{C}$  in groundwater;  $\delta^{13}\text{C}_{\text{soil}}$  is  $\delta^{13}\text{C}$  of the soil  $\text{CO}_2$  (depends on the vegetation,  $\text{C}_3$  plants  $-27\%$ ;

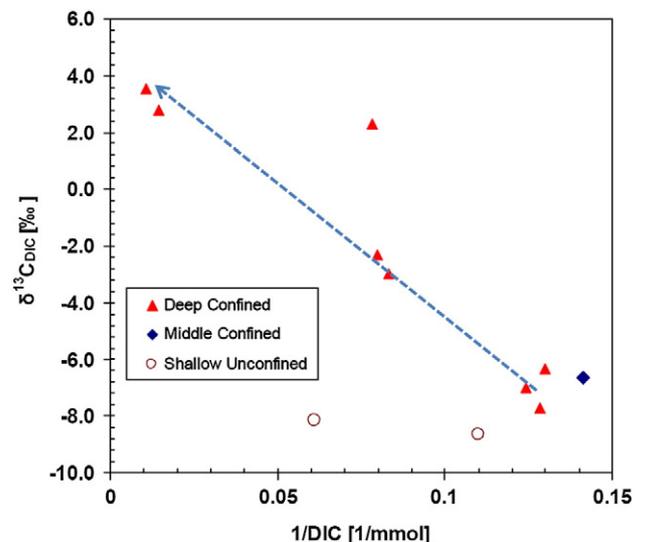


Figure 5.  $\delta^{13}\text{C}_{\text{DIC}}$  vs  $1/\text{DIC}$  (dissolved inorganic carbon) graph showing the evolution of DIC in the groundwaters of Kazan Trona Basin.

C<sub>4</sub> plants approximately  $-12.5\%$ ); and  $\delta^{13}C_{carb}$  is  $\delta^{13}C$  of the calcite being dissolved (usually close to  $0\%$ ;  $2\%$  for marine carbonates).

$$q = \frac{\delta^{13}C_{DIC} - \delta^{13}C_{carb}}{\delta^{13}C_{soil} - \delta^{13}C_{carb}} \quad (2)$$

We do not have any information regarding  $\delta^{13}C_{carb}$  values of the carbonates in the study area; however, they should have a value similar to that measured for the carbonates located near the study area and deposited at the same time with the Eocene carbonates in the area. In a study carried out by Genc (2006) in Kirsehir, located in central Anatolia, Turkey,  $\delta^{13}C$  values of Lutetian carbonates ranged from  $+2.1\%$  to  $+2.2\%$ . Kirsehir is located in central Anatolia, just like the Kazan Basin, and the carbonates in question are of the same age. This value can be greater than  $+2.2\%$  for some other carbonates deposited in the central Anatolia because the Mesozoic marine dolomitic limestones in Konya Closed Basin have  $\delta^{13}C$  values of  $+4.21\%$  (Bayari et al., 2009). Therefore, it is not unreasonable to assume a value of  $+4\%$  for  $\delta^{13}C_{carb}$  for the carbonates in the study area. Such an assumption also can explain the positive  $\delta^{13}C_{DIC}$  values of samples D-57A, D-47 and D-8.

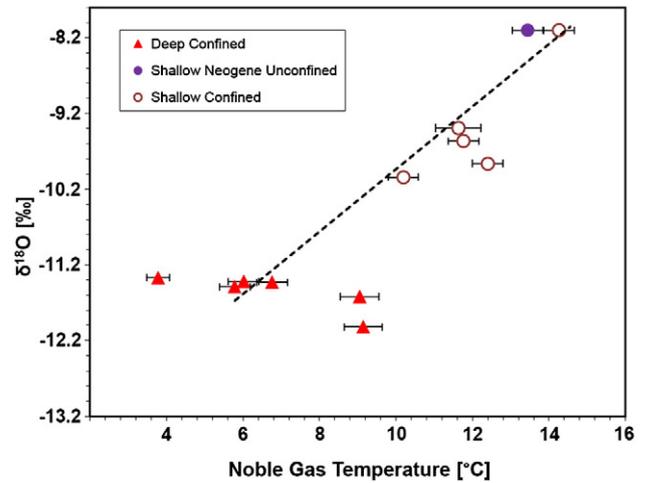
From field observations and other studies carried out around Ankara (Cetin et al., 2002; Elci and Erik, 2005) the modern plant species should follow a C<sub>3</sub> pathway as their principal photosynthesis cycle. Therefore, the modern  $\delta^{13}C_{soil}$  value should be around  $-27\%$ ; however,  $\delta^{13}C_{soil}$  values might not be constant over time. Botteme and van Zeist (1981), Prentice et al. (1992) and Wick et al. (2003) pointed out a low content of arboreal pollen and relatively high content of Chenopodiaceae pollen in the late Pleistocene record (especially from 30 ka to 15 ka) in lakes Karamik and Sogut from western Turkey, Lake Ioannina from northern Mediterranean, and Lake Van from eastern Turkey, respectively. During this period, cold arid climate caused the dominance of the steppe vegetation (C<sub>3</sub> plants) which is a type similar to the one around the study area today. Pollen evidence from Lake Eski Acigol and Lake Van region shows that arboreal vegetation re-established itself slowly in central Turkey during the Holocene (Wick et al., 2003; Woldring and Bottema, 2003). Therefore, the  $\delta^{13}C_{soil}$  value might have changed in the past due to the change in vegetation cover; however, it probably was close to  $-27\%$  especially between 30 ka and 15 ka.

$q$  values calculated accordingly are presented in Table 2 together with the estimates of ages through  $^{14}C$  dating obtained by using Eq. (3). According to Table 2, all the samples from the middle and deep aquifer systems, except for sample D-20, entered the system before the Holocene. This observation is consistent with the findings extracted from the noble gas data. It should be noted that our age estimates derived from  $^{14}C$  dating are likely lower limits because most measurements are only slightly above the detection limit. Moreover, the calculated  $^{14}C$  ages should not be considered as absolute ages due to the uncertainties associated with the amount of dead carbon contribution, the vegetation cover in the area and the  $\delta^{13}C$  value for the carbonate rocks in the area.

$$t = -8267 \ln \left( \frac{a_t \cdot ^{14}C}{q * a_0 \cdot ^{14}C} \right) \quad (3)$$

#### Climate signals in noble gas and stable isotope records

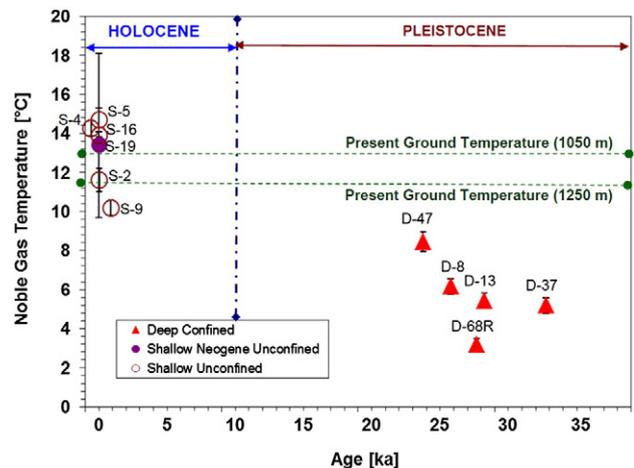
NGTs from the deep aquifer system of Kazan Trona Basin together with the stable isotope data suggest the presence of water from the last glacial period (LGP). The plot of  $\delta^{18}O$  vs NGT (Fig. 6) shows a linear correlation ( $r=0.84$ ) with a slope of  $0.35 \pm 0.07\%$   $^{\circ}C^{-1}$ . This correlation suggests that there is a climate signal in the  $\delta^{18}O$  data obtained in this study. As stated above, modern monthly mean values



**Figure 6.** The correlation between  $\delta^{18}O$  and NGT. There is a good correlation between  $\delta^{18}O$  and NGT ( $r=0.84$ ) with a slope of  $(0.35 \pm 0.07)\%$   $^{\circ}C^{-1}$  (the error is  $\pm 1$  SE). The  $\delta^{18}O$  and NGT data for the deep aquifer samples are modified to eliminate the elevation effect. See text for details.

of  $\delta^{18}O$  and temperature at Ankara station are correlated with a slope of  $0.32 \pm 0.0\%$   $^{\circ}C^{-1}$  (IAEA/WMO, 2004), which is similar to the one obtained by using the NGTs and  $\delta^{18}O$  data from our groundwater samples ( $0.35 \pm 0.07\%$   $^{\circ}C^{-1}$ ). Thus, it can be concluded that the  $\delta^{18}O$  signal reflects a temperature effect close to that of the modern  $\delta^{18}O$  temperature relationship. It should be noted that since data from different aquifers are compared in this study, an elevation correction had to be carried out for samples from the deep aquifer system since using different recharge elevations introduces a systematic offset between shallow and deep samples. For this purpose, the deep aquifer samples  $\delta^{18}O$  and NGT data were modified in Figure 6 and  $0.88\%$  was added to  $\delta^{18}O$  data (as there is  $0.44\%$  decrease in  $\delta^{18}O$  per 100 m rise in altitude) and  $0.6^{\circ}C$  was added to NGT data (as there is  $0.3^{\circ}C$  decrease in NGT per 100 m rise in altitude) to eliminate the effects of the recharge elevation difference between shallow and deep aquifer systems.

Data are clustered around three points with low  $\delta^{18}O$  and NGT values centered at  $-11.50\%$  and  $6^{\circ}C$ , high  $\delta^{18}O$  and NGT values centered at  $-9.60\%$  and  $11^{\circ}C$ , and  $-8.20\%$  and  $13^{\circ}C$ , respectively (Fig. 6). The samples from the deep aquifer system display a wide range of NGTs ranging from  $3.8^{\circ}C$  to  $9.1^{\circ}C$  although the  $\delta^{18}O$  values



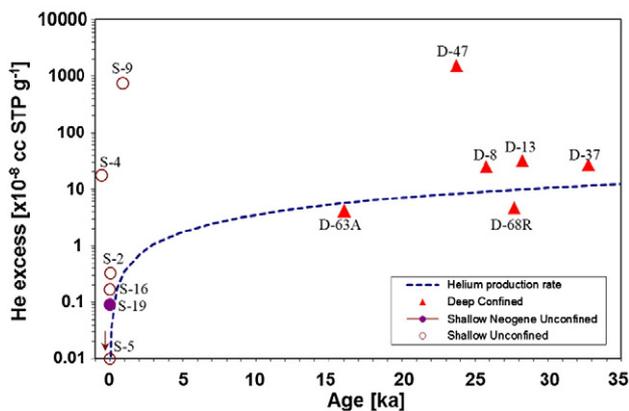
**Figure 7.** The ages obtained by  $^{14}C$  and  $^3H/^3He$  dating vs noble gas temperatures. Modern samples from the shallow system have NGTs close to today's ground temperatures ( $13^{\circ}C$  for 1050 m implied by soil temperatures,  $11.7^{\circ}C$  for 1250 m predicted by using an environmental lapse rate of  $6.5^{\circ}$  in 1000 m) whereas the samples with high  $^{14}C$  ages have lower NGTs.

are more or less similar. There is also a strong relationship between the ages obtained by  $^{14}\text{C}$  and  $^3\text{H}/^3\text{He}$  dating and NGTs (Fig. 7). Details of  $^3\text{H}/^3\text{He}$  age calculations can be found in Arslan et al. (2012). Recently recharged waters with the isotopic signature of modern precipitation from the shallow system have NGTs close to today's mean ground temperatures ( $13^\circ\text{C}$ ) whereas the isotopically lighter samples with high  $^{14}\text{C}$  ages have lower NGTs. The depletion in heavy isotopes of the groundwaters with  $^{14}\text{C}$  ages of about 15 cal ka BP and older, in comparison with more recently recharged waters, indicates the presence of paleorecharge to the deeper systems (Fig. 7).

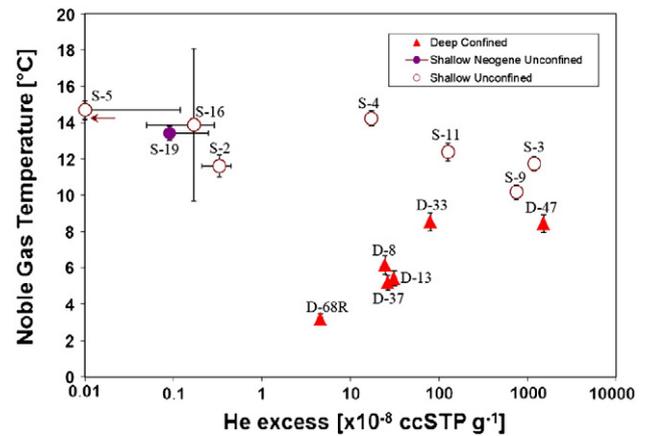
The radiogenic helium values presented in Table 5 are compared with the age estimates based on  $^3\text{H}/^3\text{He}$  and radiocarbon and NGTs (Figs. 8 and 9). Non-atmospheric  $^4\text{He}$  concentrations can be used as a relative age indicator (He concentrations increase with groundwater age due to accumulation of radiogenic  $^4\text{He}$  produced U and Th decay series elements in the rocks and released into groundwater or  $^4\text{He}$  accumulated from crustal or mantle helium fluxes; e.g., Torgersen and Stute, in press). The internal rate of  $^4\text{He}$  production from U/Th-series decay were calculated roughly by using an average density of  $2.5\text{ g cm}^{-3}$ , 20% rock porosity and 1.43 ppm and 3.82 ppm U and Th concentrations, respectively, in aquifer material. U and Th concentrations were obtained by averaging a total of 13 measurements carried out in aquifer material samples. The production rate was calculated to be  $3.5 \times 10^{-12}\text{ cm}^3\text{ STP g}^{-1}\text{ yr}^{-1}$ , which is slightly below the typical in situ production rate reported as  $5 \times 10^{-12}\text{ cm}^3\text{ STP g}^{-1}\text{ yr}^{-1}$  (e.g., Torgersen and Clarke, 1985). Samples from the deep aquifer system except for sample D-47 are consistent with the accumulation of internally produced He (Fig. 8). Sample D-47 is different from the rest of the samples, as indicated above, and its He excess indicates that the mean residence time of this sample is higher than the estimated radiocarbon age.

When plotted as a function of helium excess, the NGTs show a smooth trend (Fig. 9). NGTs increase with increasing He excess values, which is interpreted as indication of gradual cooling from the pre-LGM to the LGM period. In the same figure, the shallow samples show varying He excess values, although the NGTs are similar. The modern samples (S-2, S-5, S-16, S-19) have very little excess He while some of the shallow samples (S-3, S-4, S-9, S-11) have high NGTs and high He excess values likely reflecting mixing between mostly groundwater of Holocene origin with some older glacial groundwater (Fig. 9).

There is a clear distinction between the shallow and deep samples in terms of excess air (Table 5) and the deep samples have distinctly



**Figure 8.** Helium excess (terrestrial He) vs. estimates of ages through  $^3\text{H}/^3\text{He}$  dating (Arslan et al., 2012) and radiocarbon dating. Most samples from the deep aquifer system follow the trend that He-excess increases with age. The calculated helium production rate is  $3.5 \times 10^{-12}\text{ cm}^3\text{ STP g}^{-1}\text{ yr}^{-1}$ .



**Figure 9.** Noble gas temperatures (NGTs) vs. helium excesses. There is a correlation between noble gas temperatures and He-excesses for the deep aquifer system. For the shallow system, although the NGTs are more or less the same, He-excess values vary due to the mixing of older groundwater.

higher  $\Delta\text{Ne}$  values. The amount of excess air is linked to the amplitude of groundwater table fluctuations due to intermittent rainfall patterns (Heaton and Vogel, 1981). Therefore, higher  $\Delta\text{Ne}$  values observed in deep groundwater samples indicate that the recharge dynamics in the deep system were different in the past and the recharge and precipitation amounts were more variable, causing large fluctuations of the water table.

In the study area, there is not much information on late Pleistocene paleoclimate. In fact, our NGT record is the first data set of this nature from central Turkey. Besides, no study carried out in central Anatolia describes the temperatures in Pleistocene extending beyond the LGM period that can be used for comparison. However, there is a study by Affek et al. (2008) in Soreq cave speleothems (Israel) that reports temperatures  $6\text{--}7^\circ\text{C}$  colder than modern-day temperatures. Affek et al. (2008) stated that about 56 ka ago the temperatures were  $3^\circ\text{C}$  colder during pre-LGM than the modern temperatures. Also, NGT records obtained from different locations in Europe and Southeastern Arabia (Great Hungarian Plain–Hungary, Northern Oman, Ledo–Paniselian Aquifer–Belgium, Pannonian Basin–Hungary and Bohemian Cretaceous Basin–Czech Republic) revealed similar temperature changes ( $5\text{--}9^\circ\text{C}$ ) between glacial periods and the Holocene (Stute and Deak, 1989; Weyhenmeyer et al., 2000; Blaser et al., 2010; Alvarado et al., 2011; Varsanyi et al., 2011).

## Conclusions

The stable isotope, noble gas and  $^{14}\text{C}$  data from the Kazan Trona Ore Field verify the existence of paleowater in the deep aquifer system. The noble gas temperatures (NGTs) of the shallow aquifer system are consistent with the average annual ground temperatures whereas in the deep groundwater system NGTs  $4\text{--}8^\circ\text{C}$  lower than at present are observed. The estimated radiocarbon ages are modern for the samples from the shallow system and up to 34 cal ka BP in the middle and deep aquifer systems. Helium excess values are consistent with the accumulation of internally produced He. In one sample the He excess value indicates that the mean residence time of the sample is higher than the estimated radiocarbon age. The mean residence times, together with the stable isotope data, provide strong evidence of recharge to the middle and deep aquifer systems under colder climatic conditions, as compared to the present. Our data support previous studies on paleoclimate in central Turkey indicating that glacial periods were  $5\text{--}8^\circ\text{C}$  cooler than present. Moreover, a different deuterium excess value ( $8.7\text{‰}$ ) than today's

(11.42‰ for Ankara) indicates that the general cooling during the last Pleistocene might be accompanied by a change in source of water vapor recharging the aquifer.

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