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Age and recharge area of thermal waters in Lądek Spa (Sudeten, Poland) deduced from environmental isotope and noble gas data

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Abstract

Environmental isotope and noble gas data are shown to identify the global parameters of the thermal water system in Lądek Zdrój (Spa), Sudeten, Poland. The mean ^{14}C age is about 5 ka, and the $\delta^{18}O$ and δD data are assumed to yield the mean altitude of recharge by making use of the altitude effect known for the Sudeten area. The natural discharge rate multiplied by the ^{14}C age gives the total volume of water, whereas the volume of the host rock is roughly estimated for three scenarios of system dimensions. For these scenarios, the ratio of these two volumes yield a total porosity in the range of about 0.005–0.014, which, in a good approximation, should be equal to the matrix porosity. Laboratory determinations of unaltered rock samples yield a matrix porosity of 0.008 ± 0.001. For the three scenarios, the tracer age and matrix porosity are shown to yield a hydraulic conductivity for the whole system in the range 0.8–1.6×10^{-8} m s^{-1}, without any knowledge of the fissure network parameters. Noble gas temperature compared with known temperature dependence yields the mean altitude of the recharge area equal to that found from the $\delta^{18}O$ and δD data. The radiogenic ^{4}He excess and $^{40}Ar/^{36}Ar$ ratios are consistent with the mid-Holocene age determined from the ^{14}C data.

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1. Introduction

Lądek Spa (Lądek Zdrój) is one of four known occurrences of thermal waters in the Polish Sudeten (Ciężkowski et al., 1992). Its thermal waters are known to have been used for therapeutical purposes since the eighteenth century. Early investigators (Finck et al., 1942) regarded the thermal waters of Lądek to be of juvenile origin and related to the last volcanic activities in the area. However, the total dissolved solid (TDS) content below 200 mg l^{-1} suggested their meteoric origin (Gierwielaniec, 1968), confirmed later by ^{14}C contents of about 15–35 pmc (Dowgiałło et al., 1974), and $\delta^{18}O$ and δD values close to the world meteoric water line (Dowgiałło, 1976). However, similarly to another Sudeten thermal system in Cieplice (Ciężkowski et al., 1992), a number of controversies exist with respect to the thermal water resources, age of water, position of recharge area, temperature of water in the system, and possible mixing with local cold waters. The aims of this paper are to clarify these controversies on the basis of environmental isotope and noble gas data as well as to demonstrate relations between global rock parameters and environmental tracer data.

2. Morphology and geology of the Lądek Spa area

Lądek Spa lies on the western slopes of Góry Złote (Golden Mountains) at an altitude of 430–500 m. The Góry Złote, Góry Bialskie (Bialskie Mountains) and the Śnieżnik Massif, which according to the geomorphological classification belong to the eastern Sudeten, represent a natural eastern boundary of the Ziemia Kłodzko (Kłodzko Land) as shown in Fig. 1.

The Spa area is within a geological unit called the metamorphic unit of Lądek and Śnieżnik (Fig. 2), which belongs to the north-eastern part of the Czech Massif. The unit consists of a number of synclines and anticlines with their axes radially converging, as shown in Fig. 2 by a radial convergence of ellipsoidal outcrops of gneisses. The western boundary of that unit is represented by the Cretaceous graben of the Nysa Kłodzka river, filled with sandstones and marls. The north-western boundary is at the granitoides of Kłodzko and Złoty Stok, the north-eastern boundary is at the Sudeten fault, and the south-eastern boundary at the Ramzow overthrust.

There are three young Proterozoic and old Palaeozoic rock complexes within the metamorphic unit: the Stronie schist series (mica schists, paragneisses, quartzites, calcitic and dolomitic marbles, amphibolites, porphyroids, metarhyolites), the Gierałtów gneisses, which dominate in the Spa area, and the Śnieżnik augen-gneisses (mainly fine-grained gneisses of metasomatic–migmatic character). Lamprophires and silica veins in Góry Złote are considered to result from the Carboniferous granitic intrusions found in Góry Bialskie and in the northern part of Góry Złote. Close to the Spa, four occurrences of Pleistocene basalts (age 690 ka) in the form of domes and surface covers are known (Birkenmajer et al., 1970). Two basalt veins were also struck by boreholes L-1 and L-2, drilled to depths of 600 and 700 m, respectively.

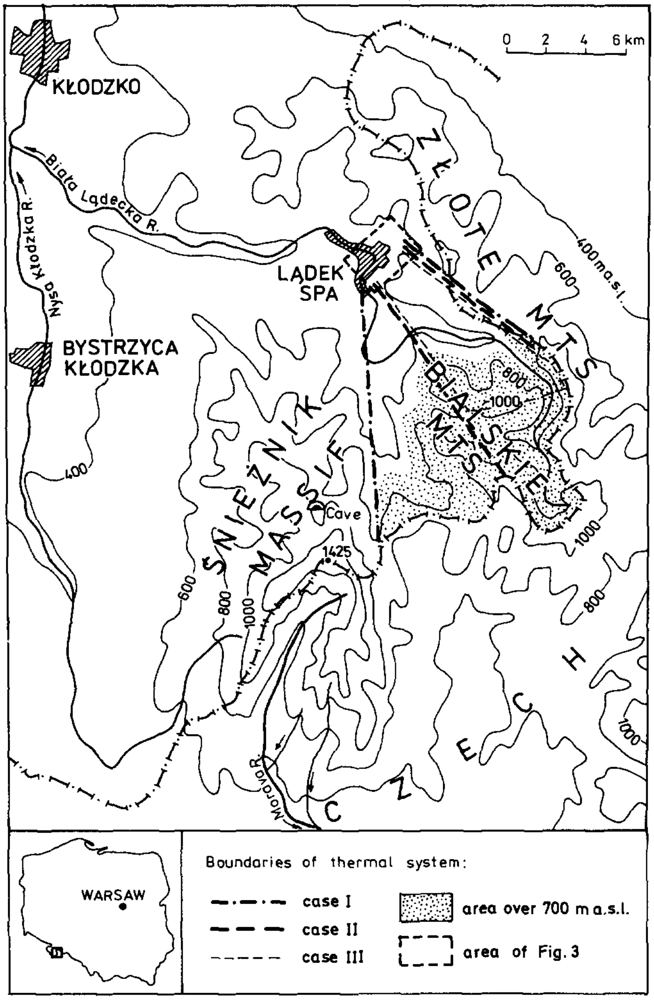

Fig. 1. Morphology of the eastern part of the Kłodzko Land with recharge areas (dotted) of thermal waters for the three scenarios assumed.

The metamorphic unit surfaces in 95% of the area and the rest is covered by Quarternary river sediments.

In Fig. 3, the geology of the Spa area is shown, and in Fig. 4 a schematic presentation of the faults is given. The Lądek fault, with a SW–NE direction, is shown at the surface by a contact zone between the Stronie schists and the Gierałtów gneisses. Strong tectonic activity in the past is evidenced in the contact zone of that fault by occurrences of mylonites. The L-2 borehole disclosed that the fault declines to the south-east at 45–50°. This fault is crossed by a number of transverse faults, which are directed in accordance with the Sudeten direction (NW–SE), and decline to the north-east at 50–65°, with shifts of up to 200 m. Particularly important is the

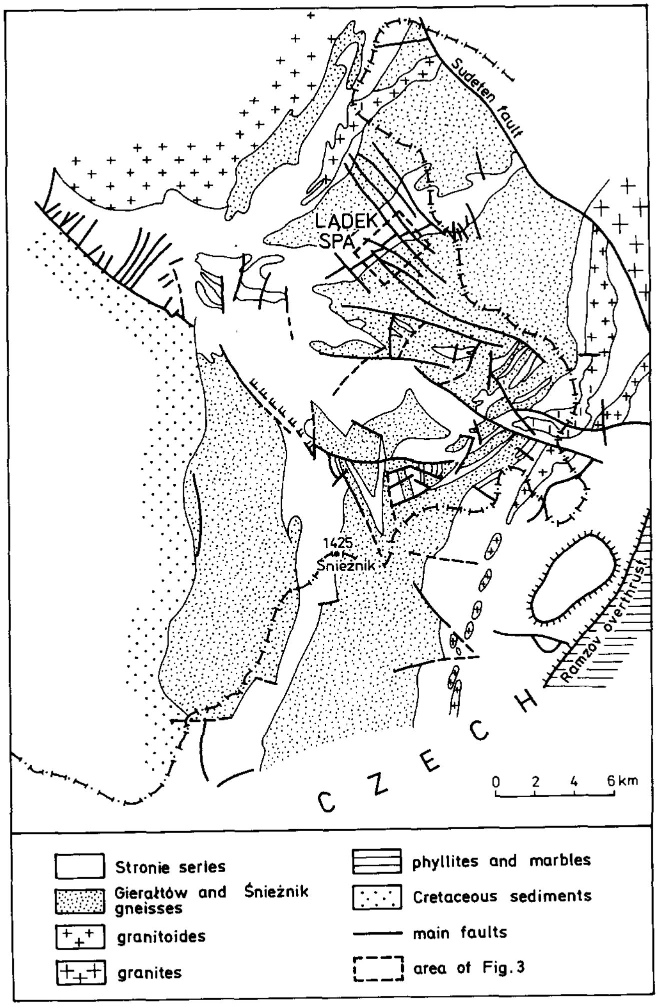

Fig. 2. Geology of the Lądek–Snieżnik methamorphic unit adapted from Don (1964) and Oberc (1972).

Lądek–Karpno fault, which has two parts (north and south) connected in the Spa area by a short and very steep fault (Gierwielaniec, 1971).

3. Hydrogeology and hydrochemistry of shallow waters

The hydrogeology and hydrochemistry of shallow waters in the metamorphic unit is best known in the Spa area (Ciężkowski, 1980). Waters from the Quaternary sediments formed by weathered material are omitted from this discussion because they are unrelated to the thermal water outflows. Cold waters are commonly exploited from weathered fractured rocks to depths of 30 m, which is supposed to

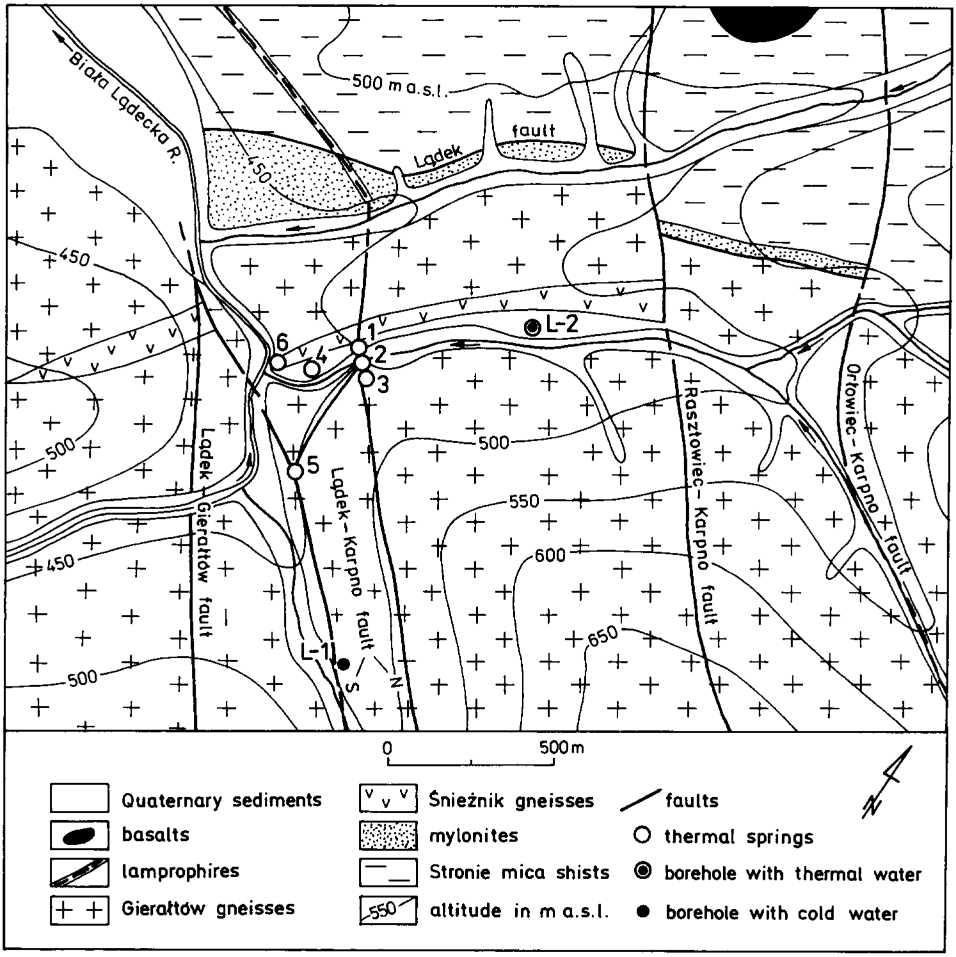

Fig. 3. Geology of the Lądek area after Gierwielaniec (1968).

correspond to the zone influenced by external changes in temperature. Four of the six natural outflows of the thermal water occur at the intersections of faults (Figs. 3 and 4). The flow path model of Fig. 4 is confirmed by observations performed in 30-m deep boreholes, which showed the presence of a small geothermal anomaly centred at the thermal outflows and related to the fork of the Lądek–Karpno fault (Ciężkowski, 1980). The L-2 borehole is within the anomaly, whereas the L-1 borehole is beyond it, and at a depth of 500 m yielded cold water (<20°C). However, the cold water system is not quite separated from the thermal one because when the L-1 borehole was pumped for a long time the thermal water outflow dropped.

In general, the water table follows the morphology, and the outflows occur mainly in shallow local depressions as area outflows (80%), or as point sources related to

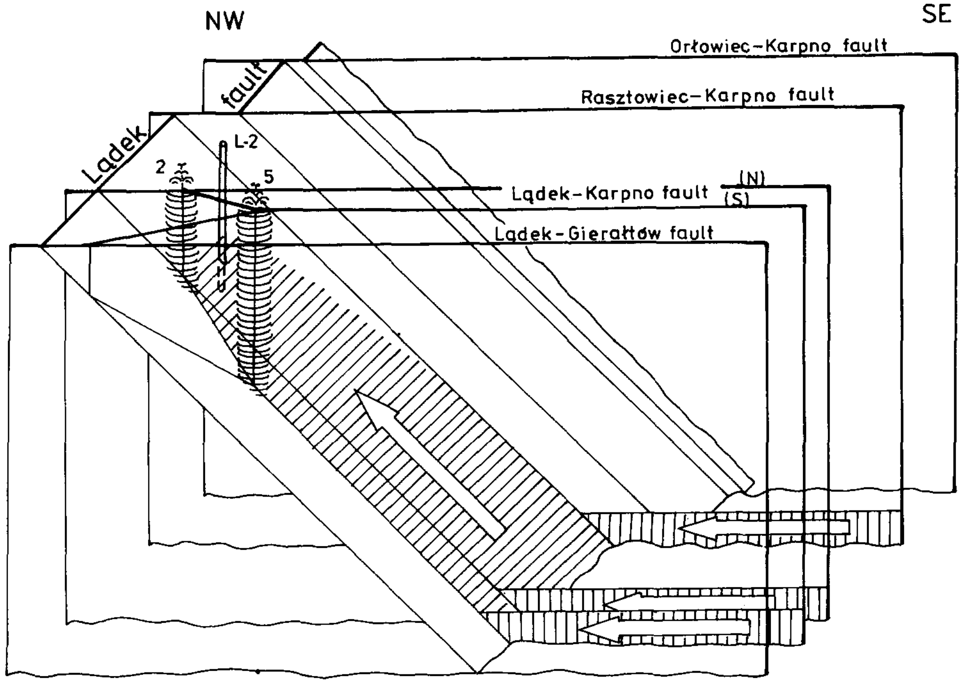

Fig. 4. Schematic presentation of faults and supposed thermal water flow paths.

fracture and fault zones. The temperature of springs varies from about 6.5°C to about 12°C and its mean altitude gradient is 0.5°C per 100 m.

The precipitation rate is 860–1000 mm year^{-1} (for the Lądek station, the highest monthly mean is 134 mm in July and the lowest 41 mm in February). The area of the Biała Lądecka basin down to the flow gauge in Lądek is 166 km^2, and the length of the river from its beginning at an altitude of about 1050 m to the gauge at an altitude of 420.5 m is about 30 km. That part of the basin is nearly completely covered by forest. The surface runoff from the basin is about 21 l s^{-1} km^{-2} (ODRA, 1987), and the mean low runoff is about 11 l s^{-1} km^{-2} (Kryza et al., 1989).

The yearly mean air temperature at the Spa area is about 6.6°C (with the highest monthly mean of 15.8°C in July and the lowest of −3.0°C in January). The air temperature dependence on the altitude (H, in m a.s.l.) in the Sudeten is given by (Ciężkowski, 1990; Ciężkowski et al., 1992):

$$t(^\circ\mathrm{C}) = -0.00558H + 9.22 \qquad (r = 0.99,\ n = 22) \tag{1}$$

where r is the correlation coefficient and n is the number of measurements.

This dependence was obtained from measurements taken in open areas, whereas the temperature of soil, and presumably of the recharge water, is lower by about 1°C in areas covered by forests (Schubert, 1930; Kappelmeyer, 1968). Therefore, for further consideration, the temperatures obtained from Eq. (1) were decreased by 1°C.

Shallow waters in fractured rocks differ from the thermal water mainly by lower F^- and higher Mg^{2+}, Ca^{2+} and SO_4^- contents (Fig. 5). In the vicinity of the thermal springs there are occurrences of waters with intermediate contents of the listed components, which suggests a possibility of some contribution of thermal water (see Fig. 5).

Deeper waters are related to fault zones which, probably owing to higher conductivity, act as the collectors of water. In the L-1 borehole, two such zones with cold water (< 20°C) were found at the depths of 150–180 m and 440–600 m. In the L-2 borehole, the following zones of increased inflows were found: at about 150 m (for a drawdown of 110 m, the flow rate was $4.5\,m^3\,h^{-1}$ with a temperature of 20.7°C), at about 400 m (artesian outflow of less than $1\,m^3\,h^{-1}$ with a temperature of 31°C), and below 608 m (outflow of about $180\,m^3\,h^{-1}$ with a temperature of about 45°C).

4. Isotope characteristics of meteoric waters in the Sudeten

The isotopic composition of shallow groundwaters in the Sudeten corrected for the continental effect to the meridian of Kłodzko is (Ciężkowski, 1990):

$$\delta D = (5.30 \pm 0.25)\delta^{18}O - (16.1 \pm 2.7) \qquad (r = 0.93, n = 76) \tag{2}$$

This equation is supposed to resemble the regional precipitation with strongly smoothed fluctuations of the seasonal effect. The altitude effect, corrected for the

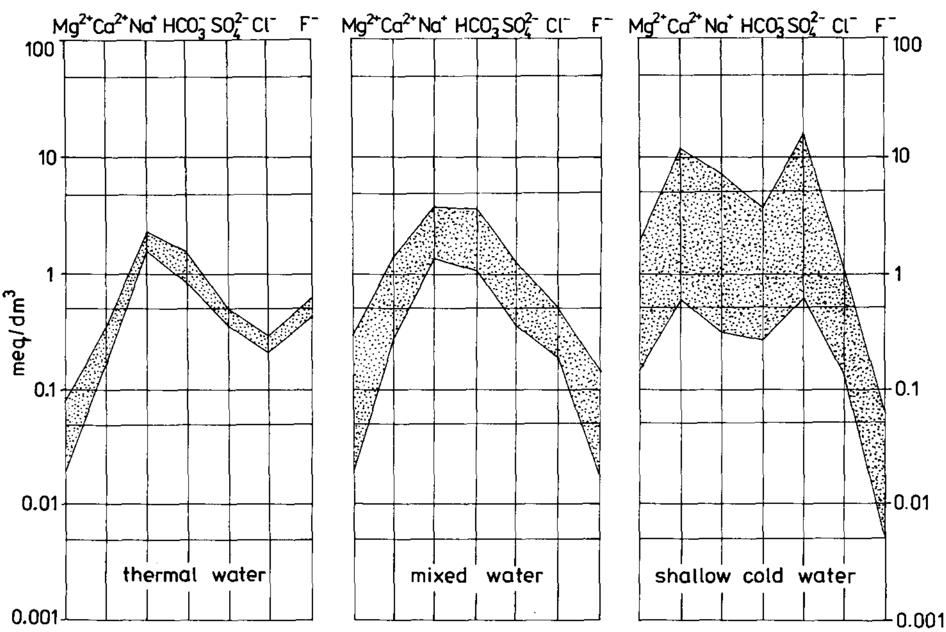

Fig. 5. Chemistry of thermal and cold waters in Lądek.

continental effect in the Kłodzko region, is as follows:

$$H(^{18}\mathrm{O}) = -(595.6 \pm 80.1)\delta^{18}\mathrm{O} - (5520 \pm 843) \quad (3)$$

$$(r = -0.86,\ n = 21)$$

$$H(\mathrm{D}) = -(85.8 \pm 18.6)\delta\mathrm{D} - (5405 \pm 1335) \quad (4)$$

$$(r = -0.75,\ n = 21)$$

Eqs. 2–4 were obtained from the same data and in the same way as similar equations for the Cieplice region (Ciężkowski et al., 1992).

5. Temperature and hydrogeology of thermal waters

The temperatures of the thermal water outflows are given in Table 1 together with chemical data. It seems that the temperature of about 45°C observed in the L-2 borehole can be accepted as the highest representative value at the system outlet. Lower temperatures observed in the natural outflows could be caused by cooling on the way to the surface of water flows having low discharge rates, especially as each spring has several outflows. It is interesting that spring 5 (Jerzy) has the same temperature as spring 2 (Wojciech) with about three times lower discharge. However, the latter is between three other springs (1, 3 and 4) which warm up the surrounding rocks.

Leśniak and Nowak (1993) found that Lądek thermal waters are generally undersaturated with respect to most minerals, and that all the chemical geothermometers are likely to fail. Only chalcedony shows an equilibration temperature of about 95°C, but a lack of equilibrium with other silicate phases suggests that chalcedony is not a good temperature indicator for Lądek waters. Results of their calculations for spring 5 (Jerzy) and the L-2 borehole suggest that the water temperature should be close to about 50–70°C.

The initial outflow from the L-2 borehole was about six times greater than the total discharge of the springs. However, the discharge of the springs and the water level in the unexploited L-1 borehole distinctly dropped, and it was decided to reduce strongly the discharge from the L-2 borehole (Ciężkowski, 1980). The present outflows are stable and reach about 660 l min^{-1} for the total outflow and 275 l min^{-1} for the springs. The undisturbed discharge from springs was about 520 l min^{-1} (H. Fresenius, unpublished data, 1938), which shows that a considerable increase in the exploitation rate is not possible, though exploitation by boreholes may yield higher temperature.

The exploitation of the L-2 borehole caused a decrease in the discharge from the springs and in the water level in the L-1 borehole, as well as a decrease in the yield of some shallow wells in Lądek. This means that although preferential flow routes exist in the fault zones, and the thermal system behaves in some respects as if it were confined, the shallow and cold waters cannot be regarded as being well separated from the thermal system in the Spa area.

Table 1
Chemical data of thermal waters in Lądek after Jarocka (1976) (mg l^{-1}, Rn in Bq l^{-1}, temperature, T in °C)

Site	pH	Rn	H_2S	H_2SiO_3	TDS	Na^+	K^+	Ca^{2+}	Mg^{2+}	F^-	Cl^-	SO_4^{2-}	HCO_3^-	CO_3^{2-}	T
L-2	9.2	152	3.1	39	196	52	1	3.1	0.5	10	8.9	16.5	42.6	21	45.5
1 (Chrobry)	9.1	126	2.6	41	188	47.5	0.8	3.9	0.7	10	8.9	20	39.7	15	26.8
2 (Wojciech)	8.9	144	2.6	39	187	47	1	4.3	0.5	9	8.9	22	39.6	15	28.3
3 (Skł.-C.)	8.9	270	1.7	39	183	45.5	1	4.3	0.5	9	8.9	19	40.3	15	24.8
4 (Dąbrówka)	9.1	133	3.6	39	182	44.5	0.8	5.0	0.5	9	8.9	16.5	42.6	15	20.3
5 (Jerzy)	8.9	1270	3.1	36	173	40	1	5.8	0.9	8	6.6	16.5	42.6	15	28.3
6 (Stare)[a]	8.6	48	0.4	34	173	39.5	0.8	6.7	0.2	8	7.1	28	42.1	6	

[a] Unexploited, chemical data of 1967.

Table 2
Environmental isotope data, ^{14}C age and undisturbed discharges of springs

Site	$\delta^{18}O$ (‰)	δD (‰)	Tritium (TU)	^{14}C (pmc)	$\delta^{13}C$ (‰)	^{14}C age (ka)	Discharge ($l\ min^{-1}$)
L-2	-10.52 ± 0.22	-72.9 ± 1.7			-15.0^b		
			0.0 ± 0.5^c	17.8 ± 1.0^c	-13.2^c	8.9	
1	-10.56 ± 0.14	-74.4 ± 1.7	4 ± 3^a	16 ± 5^a	-9.8^b		38
			0.0 ± 0.5^c	26.6 ± 1.7^c	-13.9^c	6.0	
2	-10.44 ± 0.12	-72.4 ± 0.2	1 ± 3^a	24 ± 8^a	n.a.		88
3	-10.58 ± 0.10	-73.6 ± 1.0	1 ± 3^a	32 ± 25^a	-12.8^b		67
4	-10.52 ± 0.14	-74.2 ± 1.1	3 ± 3^a	20 ± 6^a	-16.2^b		20
			0.3 ± 0.5^c	27.7 ± 1.5^c	-14.3^c	6.5	
			0.4 ± 0.5^c				
5	-10.50 ± 0.16	-73.7 ± 1.4	3 ± 3^a	32 ± 7	-15.0^b		292
			0.7 ± 0.5^c	34.9 ± 1.8^c	-15.0^c	4.5	17
			0.3 ± 0.5^c				

[a] Dowgiałło et al. (1974); [b] 1990; [c] 1993; $\delta^{18}O$ and δD taken five times between 1972 and 1993 and measured in Heidelberg, IAEA Vienna, Cracow (one series of ^{18}O and one of D rejected) and Neuherberg (one series of D rejected).

6. Chemistry and environmental isotopes of thermal waters

The first correct chemical analyses of the Lądek thermal waters were performed by Meyer (1863) and Fresenius (unpublished data, 1910, 1938). Since 1957, one full and one simplified analysis has been performed per year. The chemical compositions of all the thermal water outflows are given in Table 1 in the order of decreasing TDS content. In Table 2 the environmental isotope data, flow rates and ^{14}C ages are given, whereas Table 3 contains the noble gas data for both Lądek and Cieplice.

The amount of dissolved gases is about $11–16\ cm^3\ l^{-1}$ and the main components are: N_2 (about 90–96%), CO_2 (about 1.5–8%), Ar (about 1.6–2.2%) and CH_4 (about 0.1–0.3%) (Ciężkowski, 1980).

7. Interpretation of data and discussion

7.1. Origin and age of thermal waters

The meteoric origin and the Holocene age of the Lądek thermal waters is indicated by their isotopic content, which corresponds to the local meteoric water line, and by the ^{14}C ages given in Table 2. The piston-flow model was used to calculate the ^{14}C ages, i.e. $t_a = 8300 \ln(C_0/C)$, with the initial ^{14}C content (C_0) corrected by the simplified formula of Pearson and Hanshow (1970), i.e. $C_0 = 100\,\mathrm{pmc}(\delta^{13}C)/(-25)$. The $\delta^{13}C$ values given in Table 2 suggest the presence of carbonate material somewhere on the flow paths. In that correction, the carbonate material is assumed to be 0‰, which remains unverified. However, if the solid carbonate were isotopically heavier or lighter by about 4‰, the ^{14}C ages would differ from those given in Table 2 by less

Table 3
Noble gas temperatures (NGT), He excess, $^{3}He/^{4}He$, $^{40}Ar/^{36}Ar$ and ^{40}Ar in thermal waters of Lądek (for 850 m a.s.l.) and Cieplice (for 500 m a.s.l.)

Site	Date	NGT (°C)	He excess (10^{-6} cm^3 STP g^{-1})	$^{3}He/^{4}He$ (10^{-8})	$^{40}Ar/^{36}Ar$	A^{40} (10^{-6} cm^3 STP g^{-1})	Ref.
Lądek							
L-2	1990	3.1	9.8	8.1	297	2.4	2
	1991	3.7	13.8	8.6	296	0.0	2
	1992	3.6	12.5	9.8	296	0.0	2
	1992	3.5	13.4	n.a.	298	4.3	2
4	1991	4.0	7.0	6.3	298	3.9	2
5	1991	4.2	8.9	5.5	297	2.4	2
Cieplice							
4	1992	Sample partly degassed		33.2			2
5	1991	0.8	142	25.6	301	12	1
	1992	2.9	111	29.1	300	8.2	2
6	1990	1.1	85.7	25.8	302	9.5	1
6	1991	4.2	135	23.7	301	8.0	1
6	1992	1.6	86.7	27.7	300	7.7	2
C-2	1991	0.0	251	26.4	303	21	1
C-2	1992	2.6	143	27.0	300	8.8	2

[1] Ciężkowski et al. (1992); [2]Osenbrück et al. (1993), n.a.—not analysed.

than about 2 ka. For the L-2 borehole, the ^{14}C age cannot be greater than the obtained value of 9.8 ka, because otherwise the stable isotope content of water would be shifted owing to the climatic effect, which is not observed. On the other hand, noble gas data, discussed further below, suggest that the ages cannot be much smaller.

For a constant tracer input (as in the ^{14}C method), and for ages comparable with the half-life time, the use of the piston flow model is acceptable because other lumped-parameter models yield similar results (e.g. Małoszewski and Zuber, 1982). For instance, for sites 1, 2, 4 and 5, an extremely different model, the exponential model, which describes the residence time distribution in phreatic aquifers with an exponential distribution of flow times, yields ages of 16, 9.0, 8.8 and 6.0 ka, respectively. In that case only the age for site 1 differs drastically from the piston flow age, and corresponds to the glacial period. However, it would be very difficult to explain why waters of Holocene and glacial ages have the same isotopic composition. In conclusion, the piston flow model and Pearson's correction applied to the Lądek data are regarded as yielding internally consistent results. In addition to the above arguments one may add that the piston flow model is known to be applicable to the interpretation of ^{14}C data in confined aquifers. The thermal outflows do not contain the modern water component, and the L-2 borehole is artesian, which means that a part of the thermal system can be regarded as confined as far as the interpretation of tracer data is concerned. In conclusion, the mean ^{14}C piston-flow age for springs 1, 4 and 5, calculated for the results of 1993 sampling and weighted by the discharge rates, equal to about (4.8 ± 1.0) ka, is regarded as representative.

7.2. Recharge area

Stable isotope data from a large groundwater system in northern Poland with ages from recent to glacial showed a very low scatter for the Holocene range of ages (Zuber et al., 1990); except for some samples shifted by evident evaporation processes resulting from numerous palaeolakes which existed in the area (Ralska–Jasiewiczowa, 1988). On that ground, the mean isotope content of waters recharged in Poland during the Holocene is assumed to be constant, and, consequently, for the Holocene age of the thermal water in Lądek, its isotopic composition reflects the altitude effect. The mean altitude of recharge obtained from Eqs. (3) and (4) is about 825 m a.s.l., for the $\delta^{18}O$ and δD values given in Table 2. Considering both the morphology and the important role of the NW–SE faults shown in Figs. 3 and 4, the recharge area is most probably to the south-east of Spa in the Bialskie Mountains. The highest elevations are at about 1050 m a.s.l, and therefore it can be assumed that the recharge area should be somewhere between 700 m and the peaks, as indicated in Fig. 1. In Fig. 1, only the area within the Biała Lądecka basin is indicated, though for fractured rocks the subsurface run-off may also originate beyond the morphological boundary. Other directions of flow to the Spa are less probable because they do not conform to the main directions of the faults which are supposed to collect the thermal water. In addition, recharge in Góry Złote, north-east of the Spa, seems to be less probable because of the small area available within the basin at altitudes above 700 m. On the other hand, recharge in the Śnieżnik Massif, south of the Spa, also seems to be improbable owing to the higher altitudes of the massif and because of the preferential flow from that massif across the morphological watershed boundary as determined by dye tracer experiments, which showed the passage of water in karstic systems from the Bear Cave on the northern slopes of the Śnieżnik Massif to the Morava River on the southern slopes (Ciężkowski, 1989).

The estimated altitude of recharge allows consideration of at least three scenarios for the recharge areas, as shown in Fig. 1. For the first scenario, the recharge area is the whole Biała Lądecka river basin. For the second scenario, it is limited to a section with a western boundary estimated according to the position of faults shown in Fig. 2 and their role indicated in Figs. 3 and 4. For the third scenario, the southern part beyond the first peak is omitted under an assumption that it is completely drained by the river. For each recharge area scenario shown in Fig. 1, and for the discharge in Lądek, a large part of the thermal system can be regarded as confined, i.e. no flow lines recharged at that part reach the thermal springs.

7.3. Mixing hypothesis

The proportional pattern of changes in the TDS content and in the concentration of main components, especially of F^- (Table 1), led in the past to a suggestion of a possible mixing between the ascending thermal water and the local cold water (Ciężkowski, 1980). Changes in Na^+ and F^- concentrations indicate that the dilution by cold water cannot be larger than 20–25%. According to Eq. (3), the local water recharged at an altitude of about 500 m should have a $\delta^{18}O$ value heavier by

0.6‰ with respect to the water recharged at 825 m. The $\delta^{18}O$ value observed for the highest F^- content, which is assumed to represent the unmixed water, should be shifted to a heavier value by about 0.1–0.15‰ for a 20–25% fraction of the local water in the springs with the lowest F^- content. Such a shift is difficult to observe. However, considering a large number of samples taken at each site, and low differences between the data of each site, the admixture of local and isotopically heavier water does not seem to take place. The pattern of the δD values conforms to that of the $\delta^{18}O$ values, though their accuracy is lower due to the lower number of samples measured. The lack of tritium is also against a hypothesis of mixing with local groundwater, though some traces of tritium in springs 4 and 5 suggest that a small fraction of local water may perhaps be present in these two springs. However, in general, the mixing hypothesis must be rejected, and the slight differences in chemical composition shown in Table 1 most probably result from different flow line depths and temperatures.

7.4. Volume and porosity

The geothermal maps for the region of the Sudeten Mountains do not show any regional anomaly at depths of 1000 and 2000 m, and the geothermal gradient is about 2.5°C per 100 m (Majorowicz and Plewa, 1979), which suggests that the penetration depth of the hottest water (L-2) is 1600 m or more. On that basis, the thickness of the water system is assumed to be $h = (2500 \pm 750)$ m. The volume of the host rock system can be estimated as $V_r = xBh$, where x is the mean length of the system, and B is the mean width, both taken from Fig. 1 with reasonably assumed inaccuracies. For the three scenarios, the volumes are: $(2.6 \pm 1.1) \times 10^{11}\ m^3$, $(1.6 \pm 0.7) \times 10^{11}\ m^3$ and $(1.2 \pm 0.5) \times 10^{11}\ m^3$, respectively.

The volume of water (V_w) in a lumped system is given as the volumetric flow rate (Q) through the system multiplied by the mean tracer age (t_t) observed at the outlet. Assuming the natural discharge of the springs, observed before the construction of the L-2 borehole, to be representative at least for the second half of the Holocene, and assuming that the mean tracer age is sufficiently well represented by the radioisotope age (t_a), the volume of water is: $V_w = Qt_t = 0.27 \times 10^6\ m^3\ year^{-1} \times (4.8 \pm 1.0) \times 10^3\ year = (1.3 \pm 0.3) \times 10^9\ m^3$. This volume represents both the mobile water in fissures and the stagnant, or quasi-stagnant, water in the porous matrix, because solutes are transported at large scales as if the movement were also in the stagnant water, accessible for solutes only by molecular diffusion (Małoszewski and Zuber, 1985, 1991; Zuber and Motyka, 1994). For sparsely fissured rocks, a decaying tracer is unable to penetrate the whole matrix, and, in consequence, the age of such a tracer (t_a) can be somewhat lower than that expected for the total penetration of the matrix by a non-decaying tracer (t_t). However, for the ^{14}C tracer, such a difference is usually negligible, except for large fissure spacings, say, above 10 m (Neretnieks, 1980; Małoszewski and Zuber, 1991; Zuber and Motyka, 1994). In other words, the radioisotope age is assumed to represent the mean age of a conservative tracer, which would be observed for an instantaneous injection in the inlet to the system.

The total interconnected porosity (n) of the system can be estimated from the volumes of water and rock given above:

$$n = n_f + n_p = V_w/V_r \tag{5}$$

where n_f and n_p are the fissure and matrix porosities, respectively. For the three scenarios, the porosities calculated from Eq. (5) are: 0.0050 ± 0.0025, 0.008 ± 0.004 and 0.011 ± 0.006, respectively. As the matrix porosity of a fissured rock is usually much larger than the fissure porosity (Zuber and Motyka, 1994; Motyka et al., 1994), it can be assumed that $n \cong n_p$. The matrix porosity of unaltered Gierałtów gneisses was determined by the vacuum drying method on eight 50 g samples obtained from two rock blocks of several kilograms taken from a wall of a large excavation, 4 m from the wall surface and 12 m below the ground surface. The median value was 0.0074 ± 0.0011 and the arithmetic mean was 0.0089 ± 0.0010. Nine rock blocks which were taken from different excavations up to 4 m deep showed some signs of changes in colour, typical of weathering. The arithmetic mean of 0.018 ± 0.006 was obtained for 38 samples taken from these blocks. The last value is undoubtedly too high owing to the weathering, whereas the first two values can probably be accepted as limits for matrix porosity at larger depths, i.e. $n_p = 0.008 \pm 0.001$. Similar relations were observed for a large number of samples taken from quarries of Carboniferous granites in the Sudeten; for unaltered samples $n_p = 0.025$ and for slightly weathered samples $n_p = 0.071$ (J. Motyka, personal communication, 1992).

7.5. Tracer velocity, Darcy's velocity and hydraulic conductivity

The mean tracer velocity (v_t) is given by the ratio of the mean distance of travel (x_t) and the mean tracer age, i.e. $v_t = x_t/t_t$, where x_t is taken as the distance between the middle of the supposed recharge area and the Spa. The following tracer velocities were obtained for the three scenarios: 2.2 ± 0.6, 2.1 ± 0.5 and 1.7 ± 0.4 m year^{-1}, respectively.

Darcy's velocity (v_f) through the system, which had existed before the exploitation of the L-2 borehole was started, can be estimated as the ratio of the initial spring discharges ($Q = 0.27 \times 10^6$ m^3 year^{-1}) and the cross-sectional area normal to the flow, i.e. $v_f = Q/Bh$. The following values were obtained: 0.0135 ± 0.0053, 0.0216 ± 0.0085 and 0.027 ± 0.011 m year^{-1}, respectively.

For fissured rocks, Darcy's velocity can also be estimated from its relation to the velocity (v_t) of a conservative tracer, without any knowledge of the fissure porosity (Zuber, 1985; Małoszewski and Zuber, 1985):

$$v_f = (n_f + n_p)v_t \cong n_p x_t/t_t \tag{6}$$

The hydraulic conductivity (k) of the whole system can also be obtained in two ways. From Darcy's Law it is $k = v_f/(\Delta H/\Delta x)$, where $\Delta H/\Delta x$ is the hydraulic gradient taken in approximation from Fig. 1 as equal to the morphological slope. For the first two scenarios, the slope is 0.035 ± 0.007, and for the third one it is 0.052 ± 0.010. These gradients and Darcy's velocities given above yield the following

k values: $(1.5 \pm 0.6) \times 10^{-8}$, $(2.0 \pm 0.8) \times 10^{-8}$ and $(1.6 \pm 0.7) \times 10^{-8}\,\mathrm{m\,s^{-1}}$, respectively.

The hydraulic conductivity determined at regional scale from the tracer age and matrix porosity results directly from Eq. (6) (Małoszewski an Zuber, 1993a, b; Zuber and Motyka, 1994):

$$k = (n_\mathrm{f} + n_\mathrm{p})x_\mathrm{t}/[(\Delta H/\Delta x)t_\mathrm{t}] \cong n_\mathrm{p}x_\mathrm{t}/[(\Delta H/\Delta x)t_\mathrm{t}] \quad (7)$$

For the three scenarios, and for n_p taken from laboratory determinations, Eq. (7) yields the following k values: $(1.6 \pm 0.8) \times 10^{-8}$, $(1.5 \pm 0.7) \times 10^{-8}$ and $0.8 \pm 0.4) \times 10^{-8}\,\mathrm{m\,s^{-1}}$, respectively. If the fissure porosities estimated below are put into the full form of Eq. (7), the k values given above increase from 2.5%, for the lowest porosity, to 12.5%, for the highest fissure porosity. In the latter case the corresponding k values are: $(2.0 \pm 1.0) \times 10^{-8}$, $(1.9 \pm 0.9) \times 10^{-8}$ and $(1.0 \pm 0.5) \times 10^{-8}\,\mathrm{m\,s^{-1}}$, respectively. In conclusion, the best agreement between the porosity and k values found by different methods is observed for the second scenario.

7.6. Parameters of the fissure network and the mean water velocity in fissures

Fissure apertures and spacings were not investigated for the Gierałtów gneisses. However, a visual inspection of a large excavation, which was sampled for unaltered rock, suggests that typical spacings are in the range of 0.1–1 m whereas in shallow excavations even lower spacings can be observed. The spacings given, together with the known k value, can serve as rough estimates for the mean fissure porosity and aperture. The commonly known model of parallel fissures of equal aperture and spacing can serve this purpose. The model becomes more realistic if the tortuosity factor allowing for lengthening of flow lines owing to both the different directions of fissures and the roughness of their walls is introduced (Zuber, 1974; Małoszewski and Zuber, 1985, 1990, 1993a). For that model

$$k = (g\rho_\mathrm{w}/12\mu)n_\mathrm{f}(2b)^2/\tau_\mathrm{f}^2 \quad (8)$$

where g is the acceleration due to gravity, ρ_w is the density of water, μ is the dynamic viscosity, $2b$ is the fracture aperture, and τ_f is the tortuosity factor for fissures, which for smooth and parallel fissures is equal to 1.0 and for rough fissures without a preferable space orientation is equal to about 1.5.

For $k = 2.0 \times 10^{-8}\,\mathrm{m\,s^{-1}}$, $\tau_\mathrm{f}^2 = 2.25$, and the mean temperature of water in the system assumed to be equal to about 20°C, Eq. (8) yields $n_\mathrm{f}(2b)^2 = 7.2 \times 10^{-12}\,\mathrm{m^2}$, which in turn for $n_\mathrm{f} = 0.001$ yields $2b \cong 85\,\mu\mathrm{m}$, and for $n_\mathrm{f} = 0.0002$ yields $2b \cong 190\,\mu\mathrm{m}$. The mean fissure spacings ($L \cong 2b/n_\mathrm{f}$) are 0.085 and 0.95 m, respectively. These three pairs of values can be accepted as limits for the real values of n_f, $2b$ and L. In any case it is evident that the assumption of $n_\mathrm{p} \gg n_\mathrm{f}$ is relatively well satisfied. Zones of higher conductivities, e.g. those manifested by springs and high inflows to the boreholes, are very important for the exploitation of the drainage area, but they do not influence essentially the values of parameters obtained for the whole system.

For large scale solute transport, the average interstitial velocity (v_w), also called the mobile water velocity in order to distinguish it from the velocity of traced water molecules (Małoszewski and Zuber, 1985, 1993a,b; Zuber and Motyka, 1994), is related to the tracer velocity by a factor given as the ratio of the total interconnected porosity to the fissure porosity (Małoszewski and Zuber, 1985), i.e. $v_w = v_t(n_f + n_p)/n_f$. For the fissure porosities given above and for the tracer velocity of 2.2 m year^{-1}, that factor yields a water velocity of about 20–90 m year^{-1}. Similar mobile water velocities in fissures are obtained from Darcy's law ($v_w = k(\Delta H/\Delta x)n_f$), i.e. about 22 and 110 m year^{-1}, respectively. It should be noted that the effective porosity in Darcy's law is assumed for fissured rocks to be equal to the fissure porosity, and refers to the fluid transmission, whereas a number of authors define tacitly the effective porosity as that which is accessible to a solute during its transport. In such a case the effective porosity properly describes the solute transport but is poorly related to the water flux in fissures, and, consequently, to Darcy's law. For a critical discussion of this problem the reader is referred to Zuber and Motyka (1994).

7.7. Recharge rate and thermal water resources

For the second scenario, the recharge area is about 30 km^2, and the recharge rate is $0.27 \times 10^6\,\mathrm{m^3\,year^{-1}}/30 \times 10^6\,\mathrm{m^2} \cong 0.01\,\mathrm{m\,year^{-1}}$, which corresponds to 0.30 l s^{-1} km^{-2} and shows that only a very small fraction of groundwater runoff is related to the thermal system. A smaller value of 0.2 l s^{-1} km^{-2} was suggested by Dowgiałło (1976) who, however, arrived at thermal water reserves being four times larger by incorrectly assuming the thermal system to be recharged over the whole Biała Lądecka basin (183 km^2). A low value of the recharge rate, obtained in the case of a high precipitation rate, qualitatively confirms the low permeability of the system.

The total volumetric flow rate through the system and its thickness considered in the modelling are shown further not to be included in the validation process, and, therefore, no assurance with respect to the thermal water resources can be obtained given the present stage of knowledge. However, a large decrease in the spring outflow caused by the exploitation of the L-2 borehole suggests that the amount of available thermal water cannot be much larger than that actually exploited.

8. Noble gas data and their comparison with the Cieplice Spa data

The anlytical details of the noble gas sampling and measurements were given by Rudolph et al. (1984) and Weise and Moser (1987). For reasons explained in an earlier paper (Ciężkowski et al., 1992), only Ar, K and Xe were used for the noble gas temperature calculations whereas Ne was excluded.

The Cieplice thermal system was described in detail by Ciężkowski et al. (1992) whereas preliminary comparisons of the noble gas data of that system with those of the Lądek system were given by Zuber et al. (1993) and Osenbrück et al. (1993). Within this paper the verified data are presented and a refined interpretation is given. A comparison with the Cieplice system is interesting because it lies only about 120 km to the north-west of Lądek and discharges water from a granitic massif. The isotopic composition of the main Cieplice system ($\delta^{18}O \cong -10.3‰$ and $\delta D \cong -72‰$) is very close to that of Lądek (Table 2). However, according to Ciężkowski et al. (1992), the main thermal system in Cieplice discharges water of the last glacial period, which means that the shift in the isotopic composition to more negative values in comparison with the local waters at the altitude of discharge is caused mainly by the climatic effect. The Lądek system is shown within the present work to contain Holocene water, and, therefore, the isotopic shift can be interpreted in terms of the altitude effect. Undoubtedly, an evaluation of these hypotheses by another method would be very useful. Therefore, the noble gas methods were applied in both spas.

Low and nearly the same $^3He/^4He$ values for all sites sampled in Lądek and Cieplice suggest the dominant crustal origin of helium because the typical $^3He/^4He$ values for crustal and mantle gases are about 2×10^{-8} and 1.1×10^{-5}, respectively (Mamyrin and Tolstikhin, 1984).

The uranium content is about 5 ppm for the Stronie schists and Gerałtów gneisses, and about 11 ppm for the Śnieżnik gneisses (Przeniosło, 1970). For $n_p = 0.008$ and for the mean uranium content of about 8 ppm and thorium content assumed to be equal to 24 ppm, the in situ ^4He production in Lądek estimated from the formula of Andrews and Lee (1979) is about 5.4×10^{-10} cm^3 STP cm^{-3} year^{-1}.

Noble gases from larger depths are transported by a horizontal flow in an aquifer of a large length in comparison with thickness, according to the following formula which results from a two-dimensional transport model (Torgersen and Ivey, 1985; Stute et al., 1992):

$$J_{He} = [(C_{He}/t) - A_{is}]nh \tag{9}$$

where J_{He} is the flux of helium (cm^3 STP cm^{-2} year^{-1}), C_{He} is the mean helium content in water (cm^3 STP cm^{-3}) averaged over the whole aquifer thickness (h), and t is the tracer age (time of travel in years) at the observation site.

For in situ production as above and on age of 4.8 ka, the ^4He flux is 3.5×10^{-6} cm^3 STP cm^{-2} year^{-1}, which falls within the range of crustal flux reported in the literature (0.6×10^{-6} to 8.2×10^{-6} cm^3 STP cm^{-2} year^{-1}, according to Andrews and Lee (1979), Ozima and Podosek (1983), Mamyrin and Tolstikhin (1984) and O'Nions and Oxburgh (1988)), and close to the average values of 2.4×10^{-6} and $(1.0 \pm 0.4) \times 10^{-6}$ cm^3 STP cm^{-2} year^{-1} reported by Torgersen and Clarke (1985) and Torgersen (1989), respectively.

The in situ ^4He production for Cieplice is 2.8×10^{-10} cm^3 STP cm^{-3} year^{-1} and the crustal flux is 32×10^{-6} cm^3 STP cm^{-2} year^{-1} (for U and Th contents of 10 and 32 ppm, respectively, matrix porosity of about 0.02, age of about 15 ka, and thickness of 2500 m, after Ciężkowski et al. (1992)). Thus, the crustal flux in Cieplice is about one order of magnitude larger than the average and Lądek values.

Deeper mobile water systems remove a large fraction of the crustal flux. Therefore, for the He flux in Lądek being close to, and in Cieplice much higher than, the average crust value, the existence of deeper mobile water systems does not seem to be probable.

The $^{40}Ar/^{36}Ar$ ratios in Lądek are close to the atmospheric ratio of 295.5, and are distinctly lower than those observed for the glacial water of the Cieplice thermal system. Unfortunately, for low $^{40}Ar/^{36}Ar$ ratios, the accuracy of the ^{40}Ar-excess determinations is very low. However, a relative comparison for both systems may be worth considering. For an age of the order of 10^4 years, the radiogenic in situ production of ^{40}Ar can be neglected, and the crustal flux of ^{40}Ar is about 1.1×10^{-6} and 3×10^{-6} cm^3 STP cm^{-2} year^{-1}, for Lądek and Cieplice, respectively. The Lądek ^{40}Ar flux reasonably agrees with the average crustal flux value of 0.8×10^{-6} cm^3 STP cm^{-2} year^{-1} reported in the literature (Torgersen et al., 1989) whereas the Cieplice flux is about three times higher.

The high He flux in Cieplice can perhaps be in part an apparent value caused by uranium deposits, which existed in the recharge area at the western foot of the Rudawy Janowickie Mountains (Ciężkowski et al., 1992), but a high value of the ^{40}Ar flux cannot be explained in this way. Both high fluxes can be explained by an admixture of a small fraction of much older water. However, the hydrochemical and isotope data given by Ciężkowski et al. (1992) do not justify such a hypotheses.

The method of noble gas temperatures (NGT) was developed by Mazor (1972). In Fig. 6 the NGT values calculated as functions of altitude for both spas are given and compared with the line calculated from Eq. (1), decreased by 1°C as explained earlier.

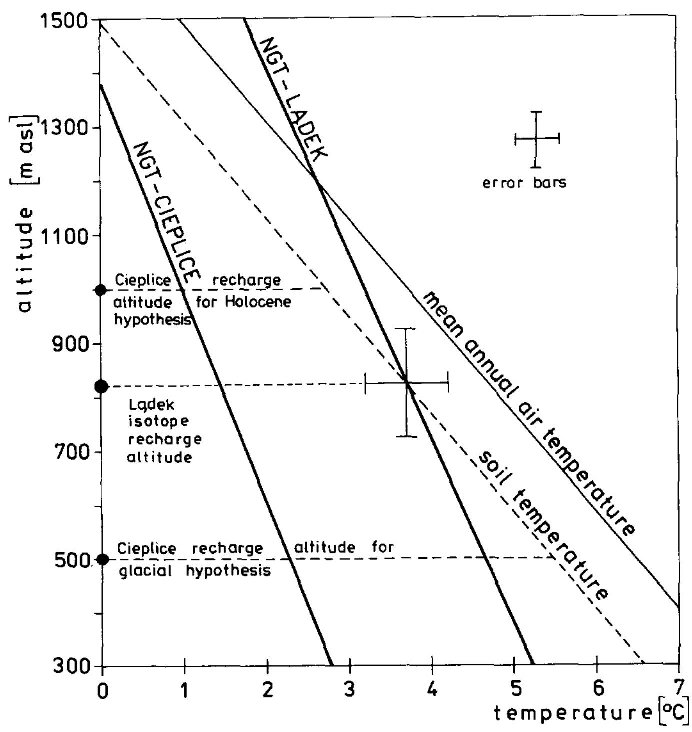

Fig. 6. Noble gas temperature (NGT) for the Lądek and Cieplice spas, real temperature versus altitude for the Sudeten, and the mean altitudes of recharge determined from $\delta^{18}O$ and δD values.

For the Lądek thermal system, the two lines cross at about 825 m, which agrees well with the altitude determined from the $\delta^{18}O$ and δD data. However, for the Cieplice thermal system, the two lines do not cross, and differ significantly from each other at an altitude of about 1000 m. This altitude would be expected from the interpretation of the $\delta^{18}O$ and δD data according to the altitude effect, if the water were recharged in the Holocene (Ciężkowski et al., 1992). In conclusion, the noble gas temperatures are consistent with the Holocene age of the Lądek thermal water and the glacial age of the main Cieplice system.

9. Validity of models

Model validation is questioned by a number of authors (e.g. Konikow and Bredehoeft, 1992; Oreskes et al., 1994) on philosophical grounds and because of either a common identification of the operational definition of validation with the calibrations process or imprecise definitions related to the common meaning of the word. Of course, in such cases the criticism is fully justified. However, according to the terminology proposed in the International Hydrocoin Project (1990), and adapted by Małoszewski and Zuber (1992, 1993b) for the tracer methods, validation is a process of obtaining assurance that a model is a correct representation of the process or system for which it is intended. It is a qualitative process based on the modeller's judgment, adapted to the modeller's needs, and applied in a restricted way by comparison of parameter values obtained by independent methods (Małoszewski and Zuber, 1993a). A validated model is understood as a confirmed model according to the terminology applied by Oreskes et al. (1994). Within this paper, a partial validation of the whole approach is attempted by comparison of the values of the parameters obtained in different ways and by sensitivity analysis.

The part of the model which led to the determination of the mean altitude of recharge can be regarded as validated because the same value was obtained from two completely independent methods. Similarly, the mid-Holocene age of water molecules, given by the ^{14}C piston flow model, can cautiously be regarded as validated by an independent method because a high ^{4}He excess and the $^{40}Ar/^{36}Ar$ ratio, a little higher than the atmospheric ratio, indicate that the water cannot be very young. On the other hand, the ^{14}C age cannot be distinctly larger owing to reasons explained earlier. For the estimated age, porosity and thickness, Eq. (9) yields ^{4}He and ^{40}Ar crust fluxes in agreement with the average global values, which gives an additional assurance in relation to the general approach.

However, considering the inaccuracies of the porosity and hydraulic conductivity estimations it is difficult to decide which scenario most adequately describes the investigated system. As mentioned, the third scenario can probably be rejected because it yields the largest difference between the porosity found from Eq. (5) and in the laboratory, as well as the largest difference between the hydraulic conductivities estimated from Darcy's law and Eq. (7). The other two scenarios yield similar values of parameters; however, the second scenario gives slightly better agreement of the porosity values.

Table 4
Changes in the values of n_p (Eq. 5), k (Darcy) and k (Eq. 7) caused by assumed changes in the values of Q, t_t, Bh, x and/or x_t

Assumed changes	Q (+/−)	t_t (+/−)	Bh (+/−)	x, x_t (+/−)
n_p, Eq. (5)	+/−	+/−	−/+	−/+
k, Darcy	+/−	0	−/+	0
k, Eq. (7)	0	−/+	0	+/−

The validation process can be questioned as a number of parameter sets in Eq. (5) may yield the same porosity. Similarly, both Eq. (5) and Darcy's law, in the way they are applied within this work, depend in part on the same parameters, and, consequently, are not quite independent of each other. Therefore, a sensitivity analysis is needed. Possible directions of changes in the matrix porosity determined from Eq. (5), and hydraulic conductivities estimated directly from Darcy's law and from Eq. (7) are shown in Table 4. An error in the estimate of volumetric flow rate through the system (Q), or in the tracer age (t_t), or in the Bh product, or in the length of the system (x), would lead to a difference between the laboratory and Eq. (5) porosities, and to a difference between the hydraulic conductivity found from Darcy's law and that calculated from Eq. (7). A positive (or negative) error in the Q value, and, at the same time, a negative (or positive) error in the t_t value would lead to no change in n_p, and to positive (or negative) changes in the both k values. The same relative errors committed at the same time for the Q and Bh values would cause no errors in the n_p and k values. Similarly, the same relative errors committed for the t_t and x (or x_t) values would cause no errors in the n_p and k values. A positive (or negative) error in the Bh value and a negative (or positive) error in the x and x_t values would cause no errors in the n_p value and would lead to negative (or positive) errors in the k values. However, owing to the reasons given earlier, the age cannot be characterised by a large error.

Similarly, the distances are relatively well defined by the morphology. Therefore, the only pair of parameters which cannot be validated by the comparison of the n_p and k values is that of the flow rate (Q) and the cross sectional area of flow (Bh), if they change in the same direction. In other words, if Q is larger owing to some underground outflow from the system, the thickness (h) can also be proportionally larger than that estimated. However, though no validation with respect to the flow rate and thickness can be obtained, a large drop in the spring outflows, which was caused by the exploitation of the L-2 borehole, suggests that the estimates given are probably acceptable. For instance, if the present total outflow from the L-2 borehole and springs (about 0.35×10^6 m^3 year^{-1}) is accepted as representative for the total outflow from the system in the Holocene period, a depth of about 3200 m is obtained, which is still within the accuracy assumed (2500 ± 750 m).

10. Conclusions

The environmental tracer data identified the age and the altitude of recharge of the thermal water system in Lądek Spa, and yielded the global values of the porosity and hydraulic conductivity.

Identification of the recharge area, though approximate, is undoubtedly important from a practical point of view. Similarly, suggestions related to limited thermal water resources are of practical importance, particularly as opposite opinions are often expressed. It is more difficult to specify the usefulness of the global hydraulic conductivity. However, together with other related parameters, it serves to provide a better understanding of the system, and its low value confirms in general low thermal water resources.

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