where the opposite trend was usually observed (e.g. Talbot 1990). It has been proposed (Różański et al. 1992) that the observed anticorrelation between $\delta^{18}O$ and $\delta^{13}C$ might be caused by changes in the catchment area of the lake, induced by climatic fluctuations. In through-flow lakes controlled by underground inflows, which is the case of Lake Gościąż, the $\delta^{13}C$ of dissolved carbonates and consequently of the deposited calcite is largely controlled by the carbon-13 content of the inflowing water. It has been demonstrated that $\delta^{13}C$ of newly formed groundwater, both under open and closed-system conditions, is a function of partial pressure of soil CO_2 (Duliński & Różański 1990). Higher pCO_2 levels in the soil lead to more negative $\delta^{13}C$ values of the dissolved carbonates in the infiltrating water. The climatic control of CO_2 levels in the soil zone was confirmed in numerous studies (e.g. Dörr & Münnich 1987). Higher soil temperatures during summer lead to enhanced root respiration and accelerated decomposition of soil organic matter through microbial activity. Consequently, significantly higher partial pressures of CO_2 in the soil are observed during summer months. A detailed study of the present-day carbon cycle in Lake Gościąż (Wachniew unpubl.) revealed a distinct seasonal cycle of $\delta^{13}C$ in the dissolved carbonates in springs feeding the lake, with more negative values observed during summer months. This provides a strong argument in favour of the above-mentioned hypothesis.

Another feature of the isotope data is an apparent correlation between the depth of sedimentation and the isotopic signature of the deposited carbonates: calcite originating in the shallower parts of the lake is depleted in oxygen-18 and enriched in carbon-13, when compared with the material deposited in the central deep. During the Younger Dryas, this difference was about 1‰ for oxygen-18 and about 5‰ for carbon-13.

The hypothesis that might explain the observed shifts in the isotope signals preserved in calcite calls for lateral differences in water temperature of the epilimnion. It might well be that shallow, littoral part of the lake (Tobyłka Bay) was not well mixed with the major body of water and thus experienced, on the average, slightly higher temperature during summer stratification. The locally higher temperature would likely lead to more intense biological activity in these parts of the lake and consequently to higher consumption of the dissolved inorganic carbon and associated carbon-13 enrichment of the deposited calcite.

Modelling results suggest (Wachniew unpubl.) that to account for 5‰ shift in $\delta^{13}C$ the increase of primary production by 3 to 4 times in a period of 1 month is needed.

It is clear that the isotope records originating from sediments deposited in the central deep better reflect behaviour of the entire lake system and its response to climate fluctuation.

7.7. VARIATIONS OF ATMOSPHERIC ^{14}C CONCENTRATIONS AT THE PLEISTOCENE/HOLOCENE TRANSITION, RECONSTRUCTED FROM THE LAKE GOŚCIĄŻ SEDIMENTS

Tomasz Goslar, Maurice Arnold & Mieczysław F. Pazdur†

Because of natural variations of atmospheric concentration of ^{14}C, the radiocarbon calibration curve showing the relation of radiocarbon to calendar age is not a straight line. The characteristic feature of the calibration curve are the plateaux, periods of constant radiocarbon age a few hundred years long, reflecting the rapid drops of ^{14}C concentration in the atmosphere. One such plateau occurs at the radiocarbon age of ca. 10,000 ^{14}C BP (Kromer & Becker 1993), which is also the worldwide accepted date of the Pleistocene/Holocene or Younger Dryas/Preboreal (YD/PB) boundary (Mangerud et al. 1974, Ammann & Lotter 1989, Wright 1989, Peteet 1992). The plateau complicates an absolute dating and synchronization of this event among different sites if the time scale is reconstructed by only a few radiocarbon dates. One may overcome this problem using independent calendar time scales provided by tree-rings or annual laminations (varves) in ice or lacustrine sediments. However, even for the best absolute German tree-ring and Greenland ice-core chronologies available till now, the difference among age estimates of YD/PB boundary (Becker et al. 1991, Johnsen et al. 1992, Taylor et al. 1993) remains significant. It can be interpreted as a real delay in climatic warming in Europe with respect to North Atlantic as well as a result of underestimated error of counting the annual ice layers and/or incorrect match between German pine and oak chronologies (Kromer & Becker 1993). Alternatively, the beginning of slow change in the isotopic composition of German wood does not necessarily reflect the main YD/PB warming, which, according to other reconstructions (Lotter et al. 1992a, Goslar et al. 1993) was as abrupt in Europe as in Greenland. Another method of synchronization independent of the error of varve counting over the whole Holocene is to compare the position of YD/PB boundary with respect to worldwide synchronous radiocarbon plateau. This cannot be done in ice cores, but it is possible in varved lacustrine sediments. The time relation between warming and the ^{14}C plateau is also important for recognition of the causes of warming, since one possible mechanism of abrupt warming, the turning-on of thermohaline circulation (Broecker & Denton 1989), should produce a large drop in atmospheric concentration of radiocarbon. The laminated sediment of Lake Gościąż, having the YD/PB boundary very well documented, is excellent material for such a study.

Radiocarbon dates and synchronization of Lake Gościąż varve chronology and German pine chronology

The reliability of radiocarbon dating of lacustrine sediment using terrestrial macrofossils has been discussed in Chapter 6.2. In that Chapter we also presented the AMS datings of samples from the Holocene part of the Lake Gościąż (LG) sediments and discussed the match of LG varve chronology to the absolutely dated German oak chronology. Relying on the ^{14}C match alone we could date the top of the LG floating varve chronology to 3140±120 cal BP and consequently the Younger Dryas/Preboreal transition to 11,440±120 cal BP. The correlation of laminae thickness and tree-ring widths, discussed in Chapter 6.4, pointed to 3211 cal BP as the most probable age of the top of the floating varve chronology, giving consequently the YD/PB transition at 11,510±50 cal BP. Both estimates agree quite well with the age of the Pleistocene/Holocene boundary recorded in Greenland (Johnsen et al. 1992, Alley et al. 1993).

However, the correlation of laminae and tree-ring thickness is not completely sure, so the Lake Gościąż data in this chapter are presented with only the ^{14}C match, unless stated otherwise.

The AMS radiocarbon dates of terrestrial macrofossils from the Late-Glacial and early-Holocene parts of the Lake Gościąż sediments are listed in Table 7.4, and presented in Fig. 7.43 together with the calibration data from German oaks (GO) and pines (GP) (Kromer & Becker 1993), and corals from Barbados (Bard et al. 1990, 1993) and Huon Peninsula (Edwards et al. 1993). The early-Holocene ^{14}C dates from Lake Gościąż (Goslar et al. 1995a) clearly show a shift of Lake Gościąż varve chronology with respect to that of German pines. The wiggle-match of Gościąż to pine ^{14}C dates (Fig. 7.44) is excellent, with a very smooth narrow minimum of S^2 = 1.2 (expected value = 1; for the discussion of the wiggle-matching procedure see Chapter 6.4), enabling synchronization of LG and GP chronologies with a precision better than 50 years. The ^{14}C ages of two early-Holocene plateaux shown by LG data agree very well with those reconstructed in high-precision dendrocalibration (Fig. 7.45). This indicates that systematic errors in AMS radiocarbon dating and problems of rebedding of Gościąż macrofossils are negligible. However, the synchronization requires either ’compression’ of Lake Gościąż varve chronology in the period between ca. 9.7–10.5 kyr BP by ca. 200 yr, or the revision of tentative dendro-match between German oak and pine chronologies. The rejection of 200 varves from the sequence 800 yr long seems unlikely, since the chronology is replicated in several varved cores from two separate lake deeps. The revision of dendro-match and shift of pine chronology, on the other hand, would expand the plateau of 8900 ^{14}C BP to ca. 600 yr. In fact, such a long plateau is suggested by data from Soppensee (Hajdas et al. 1993), where 5 dates of 8900–9000 ^{14}C BP in the time span of 700 yr occur. It must be stressed that because of uncertainties of ±120 yr and ±50 yr, respectively, the two wiggle-matches may be nearly reconciled with no shift (i.e. 200±170 yr). In such a case,

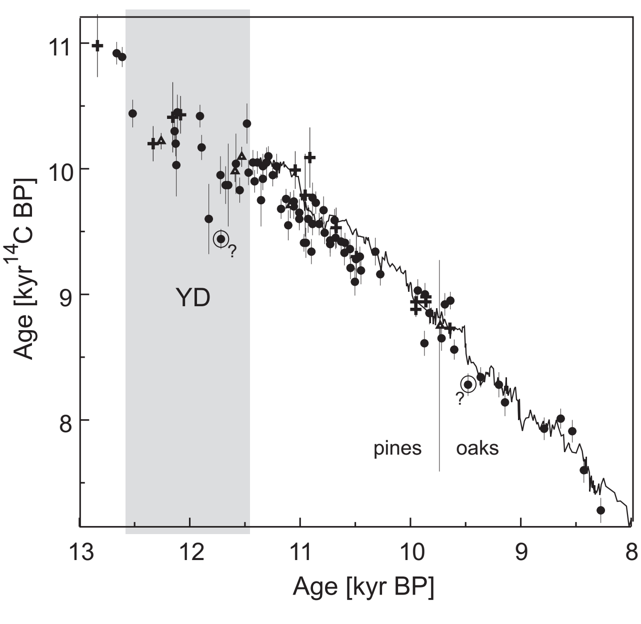

Fig. 7.43. AMS radiocarbon dates of terrestrial macrofossils obtained from the Lake Gościąż sediments (Goslar et al. 1995a), superimposed on the radiocarbon calibration data. Circles – LG macrofossils; triangles – Barbados corals (Bard et al. 1993); crosses – Huon Peninsula corals (Edwards et al. 1993); continuous line – German oaks and pines (Kromer & Becker 1993). The question marks indicate two samples (encircled) assumed to be contaminated. The calendar time scale of LG points (±120yr) is based on the ^{14}C wiggle-match to German oaks.

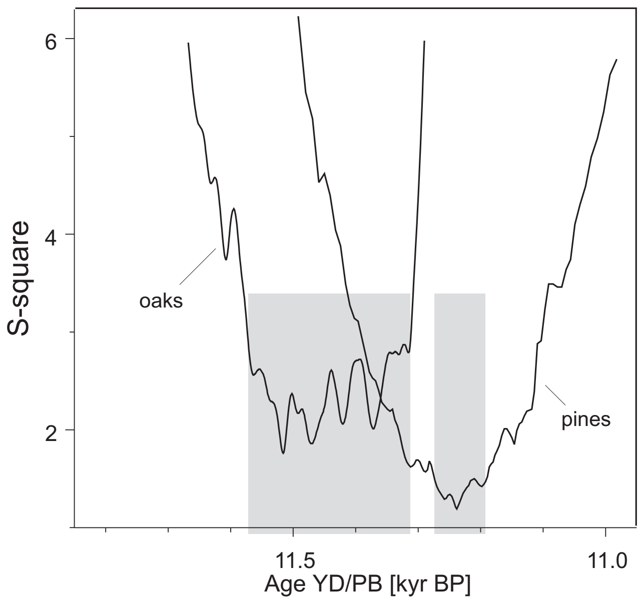

Fig. 7.44. The result of wiggle matching of Gościąż ^{14}C dates to German oak and pine chronologies. Both curves show the mean square differences between ^{14}C dates of LG samples and the radiocarbon calibration data, as a function of the age of LG floating varve chronology. The age of floating varve chronology is represented by the age of varve no. 1072 (i.e. the YD/PB boundary, see Tab. 7.4). The ranges of age allowed by both matches are shaded.

Table 7.4. AMS radiocarbon dates of terrestrial macrofossils from the Late-Glacial and early-Holocene part of annually laminated sediments of Lake Gościąż. The YD/PB boundary corresponds to varve no. 1072. The calendar age (±120yr) is given according to the match to German oak chronology. S, P – samples of bud scales and pine peridermis from the same horizon. 1 – samples dated in the first run, already published (Goslar et al. 1993); 2 – samples of scales and peridermis from section of irregular lamination in core T1/90, not dated by varve chronology; 3 – sample probably contaminated with modern carbon.

Sample code	Mass mgC	Varve no.	Age cal BP	Age ^{14}C BP	Comment
93 388	0.56	-170÷ -130	12 665	10 920 ± 90	
93 389	1.68	-130÷ 70	12 615	10 890 ± 80	
93 377	0.84	-65÷ 55	12 520	10 440 ± 110	
93 401	0.56	370÷ 390	12 140	10 300 ± 100	
93 400	0.51	350÷ 420	12 130	10 200 ± 100	
G201	0.17	380÷ 400	12 124	10 030 ± 250	1
G202	0.76	389÷ 409	12 115	10 450 ± 140	1
93 370	1.19	595÷ 615	11 910	10 420 ± 90	
93 371	0.96	605÷ 635	11 895	10 170 ± 100	
G223+224+225	0.15	654÷ 714	11 830	9 600 ± 280	1
G233	0.59	780÷ 800	11 724	9 950 ± 150	1
93 324	1.00	755÷ 835	11 720	9 440 ± 80	3
G236+237	0.36	816÷ 856	11 678	9 870 ± 150	1
G238	0.17	850÷ 870	11 654	9 870 ± 330	1
G244+245	0.23	900÷ 960	11 584	10 040 ± 240	1
G252+255	0.33	980÷1 080	11 484	10 360 ± 160	1
					YD/PB
93 342	0.80	1 105÷1 145	11 390	10 050 ± 90	
G264	0.23	1 147÷1 167	11 357	9 750 ± 210	1
93 395	0.84	1 150÷1 200	11 340	9 920 ± 90	
93 343	0.56	1 145÷1 205	11 340	10 020 ± 100	
G268	0.83	1 197÷1 217	11 307	10 050 ± 120	1
93 336	2.06	1 205÷1 245	11 290	10 100 ± 80	
93 344	1.24	1 245÷1 285	11 250	9 950 ± 90	
93 365	1.10	1 275÷1 325	11 215	10 020 ± 100	
93 366	0.86	1 310÷1 370	11 175	9 680 ± 80	
93 323	0.96	1 365÷1 405	11 130	9 760 ± 80	
93 326	0.35	1 380÷1 430	11 110	9 550 ± 120	
93 367	1.01	1 425÷1 485	11 060	9 700 ± 100	
93 396	1.24	1 425÷1 485	11 060	9 740 ± 90	
93 319	0.97	1 475÷1 535	11 010	9 600 ± 90	S
93 320	2.60	1 475÷1 535	11 010	9 650 ± 110	P
93 312	1.20	1 527÷1 567	10 968	9 410 ± 70	
93 402	0.31	1 540÷1 590	10 950	9 410 ± 120	
93 403	0.40	1 565÷1 605	10 930	9 600 ± 110	
93 346	0.78	1 595÷1 635	10 900	9 340 ± 100	
93 310	0.57	1 595÷1 655	10 890	9 560 ± 90	S
93 331	0.66	1 595÷1 655	10 890	9 770 ± 120	P
93 347	2.63	1 645÷1 665	10 860	9 730 ± 90	
93 432	1.69	1 655÷1 715	10 830	9 560 ± 90	
93 348	1.46	1 685÷1 765	10 790	9 670 ± 110	
93 406	0.88	1 710÷1 760	10 780	9 490 ± 90	
93 338	2.77	1 765÷1 805	10 730	9 430 ± 100	P
93 337	0.77	1 765÷1 805	10 730	9 400 ± 100	S
93 410	1.47	1 805÷1 845	10 690	9 590 ± 90	
93 316	2.40	1 815÷1 855	10 680	9 450 ± 90	
93 352	2.09	1 865÷1 905	10 630	9 420 ± 100	
93 411	1.40	1 895÷1 935	10 600	9 330 ± 100	
93 325	0.85	1 895÷1 945	10 595	9 410 ± 80	
93 349	0.69	1 935÷1 995	10 550	9 360 ± 80	

Table 7.4. *Continued.*

Sample code	Mass mgC	Varve no.	Age cal BP	Age ^{14}C BP	Comment
93 412	1.51	1 945÷1 995	10 545	9 210± 90	
93 408	0.69	1 985÷2 035	10 505	9 100±110	
93 317	1.93	2 000÷2 040	10 495	9 280± 90	
93 350	1.02	2 045÷2 065	10 460	9 300± 80	
93 409	0.74	2 045÷2 085	10 450	9 190±110	
93 353	1.24	2 175÷2 215	10 320	9 340±110	
93 354	2.27	2 215÷2 265	10 275	9 160± 90	
93 382	0.90	2 530÷2 630	9 935	9 030± 90	
93 355	0.77	2 575÷2 705	9 875	8 610±100	
93 383	0.90	2 625÷2 665	9 870	9 000± 90	
93 384	0.91	2 665÷2 705	9 830	8 850± 90	
93 332	1.67		ca. 10 430	9 340±120	S[2]
93 335	1.74		ca. 10 430	9 480±120	P[2]
93 416	0.57		ca. 10 330	9 050±130	S[2]
93 417	1.74		ca. 10 330	8 980±100	P[2]

however, we must allow that either both the onset and termination of cold Younger Dryas period in Poland were delayed with respect to those recorded in Greenland by more than 300 yr, or a systematic error in counting of annual layers in both GRIP and GISP2 ice cores occurred.

The plateaux of radiocarbon age in the Late-Glacial and early Holocene

Independently of the absolute age, our data allow us to determine the position of YD/PB boundary at the plateau

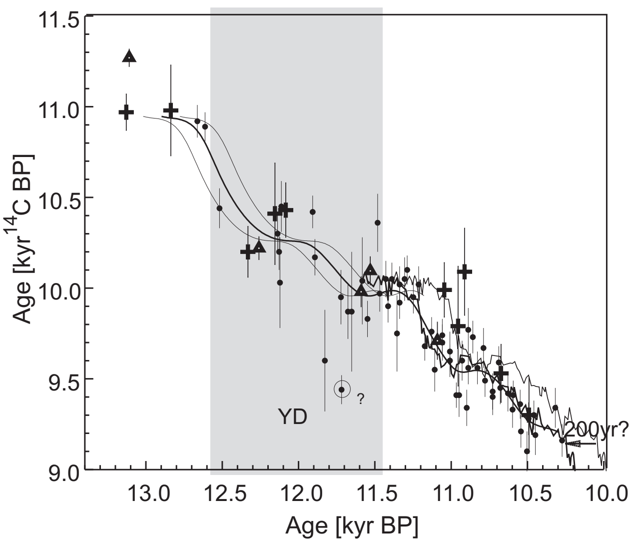

Fig. 7.45. AMS radiocarbon dates of Lake Gościąż macrofossils (circles) from the Late-Glacial and early-Holocene, compared with the radiocarbon calibration data. The smoothed curve is a cubic spline fitted to LG dates. The uncertainty of calendar age of LG chronology is ±120 yr. Triangles – Barbados corals (Bard et al. 1993); crosses – Huon Peninsula corals (Edwards et al. 1993); light and heavy jagged lines – German pines (Kromer & Becker 1993) in original position, and shifted by 200 yr, respectively.

of 10,000 ^{14}C BP (Fig. 7.46). The warming in Poland, which was as abrupt as that recorded in Greenland, happened 250 yr before the end of plateau. The beginning of plateau is not reconstructed in detail; presumably its total length is ca. 500 yr. The precise synchronization of LG and GP chronologies enables also the direct comparison of YD/PB transitions recorded in both archives. The beginning of slow rises of δ^{13}C and δD in German pines, defined as the YD/PB boundary (Becker et al. 1991) is thus younger by ca. 200 yr than the main δ^{18}O increase in the LG sediments, and surely younger than the termination of Younger Dryas cold period in Poland, as evidenced by the continuous *Ulmus* pollen curve (Ralska-Jasiewiczowa et al., Chapter 7.4). Since both regions are situated only 1000 km apart at the common direction of westerly winds being the main present air circulation heating central Europe from the North Atlantic, it is very unlikely that the warming farther east happened earlier than farther west. Therefore we can conclude that the rises of δ^{13}C and δD in German pines do not mark the abrupt climatic warming at the termination of Younger Dryas. The case presented above illustrates the advantage of synchronizing the Late-Glacial/early-Holocene events in different sites by matching radiocarbon plateaux, as being independent of an uncertainty of long calendar chronologies. The scarcity of AMS radiocarbon dates from laminated sediments of Soppensee makes the similar direct synchronization in the period considered impossible. On the other hand, the AMS radiocarbon datings of YD/PB boundary in the non-laminated sediments of Swiss lakes (Ammann & Lotter 1989) seem to show the warming in the middle of plateau, in agreement with our data.

The radiocarbon ages of LG macrofossils in the late

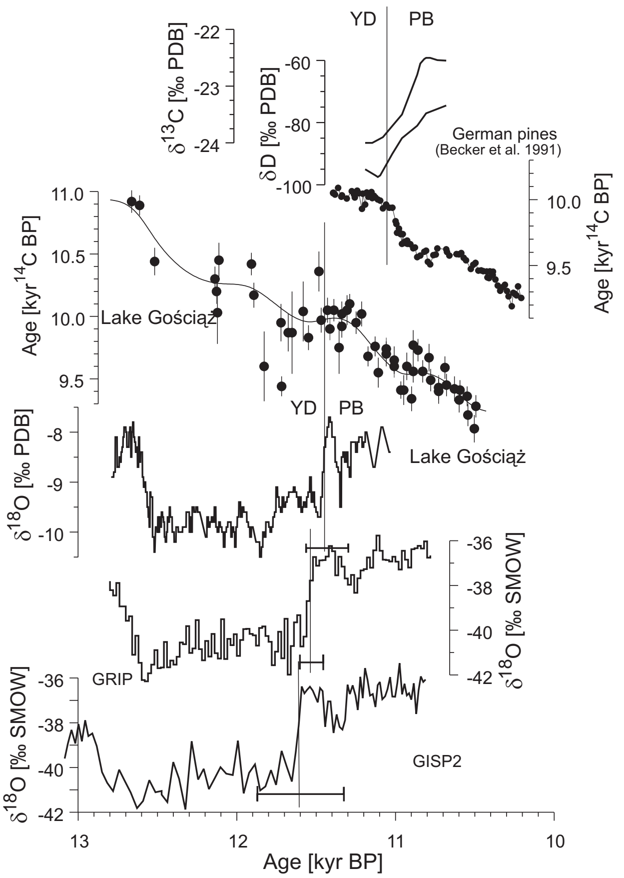

Fig. 7.46. Illustration of synchronization of ^{14}C calibration curve and stable-isotope records at the Late-Glacial/Holocene boundary (after Goslar et al. 1995a). The only archive providing both ^{14}C and ^{18}O data sets is the sediment of Lake Gościąż. The synchronization of Lake Gościąż and German pine chronologies requires shift of one of chronologies by ca. 200 yr. The smoothed lines at $\delta^{13}C$ and δD scales represent the changes of stable-isotope content in German pines (Becker et al. 1991). The uncertainties of absolute dates of the YD/PB transition in Lake Gościąż and in ice cores are represented by horizontal bars.

Allerød and Younger Dryas (Fig. 7.45) generally confirm the U/Th calibration (Bard et al. 1993, Edwards et al. 1993). They show another plateau 300–500 yr long at 10,400 ^{14}C BP, and a rapid decline of radiocarbon age at the onset of Younger Dryas. The correlation with coral data, however, does not favour any particular age of the Lake Gościąż floating varve chronology in the range allowed by the match to German oaks.

The calendar age of Younger Dryas boundaries

The age of AL/YD and YD/PB boundaries, set, according to the convention proposed by Broecker (1992), at the middle of rapid $\delta^{18}O$ changes in the LG sediments, is directly related to the age of floating varve chronology. According to the correlation of varve thickness to the tree-ring widths, the ages of AL/YD and YD/PB boundaries could be determined to 12,650±90 cal BP and 11,510±50 cal BP, respectively. The quoted errors correspond only to the uncertainty of varve counting of floating chronology. According to the AMS radiocarbon dates alone, the corresponding ages can be determined to 12,580±140 cal BP and 11,440±120 cal BP.

The calendar age of YD/PB boundary as determined in the Lake Gościąż study is compared in Table 7.5 with the estimates from other investigations. The definitions of transition differ among archives. The boundary is defined most sharply in the change in the accumulation rate of Greenland ice (completed in 20–30 yr). The changes of oxygen isotope ratios in Greenland ice and Lake Gościąż carbonates (50–70 yr) were as rapid as changes of fluxes of calcium and magnesium. The transitions in vegetation cover (Gościąż, Soppensee, Holzmaar) responding to the climate warming took a longer time, but major changes were completed in 100–200 yr. The slowest change was that observed in the isotope composition of carbon and hydrogen in German pines (ca. 500 yr). The duration of the major change in the Swedish study is difficult to determine. It must be stressed that the durations of major change, when reconstructed by proxy data of the same type, are similar, but the calendar ages of major change are different, and the differences are well beyond the durations of individual transitions. For that reason the delay between climate warming recorded in Lake Gościąż, Lake Holzmaar, and Greenland Summit and those recorded in Swiss lake, Swedish varves, and German pines must be regarded real unless the error of calendar-age estimates in appropriate archives is found. The same problem concerns the climate cooling recorded at the transition between Allerød and Younger Dryas (Tab. 7.6). There is always some possibility that the uncertainties of chronologies, constructed by counting thousands of varves, are underestimated. The best method to verify the non-synchronism of YD/PB and AL/YD boundaries is to use independent, undoubtedly synchronic markers. Here, either the layers of volcanic tephra or the worldwide synchronous changes of radiocarbon age can be used. The only doubtless synchronization of ^{14}C curves, between German pine and Lake Gościąż chronologies, was discussed earlier.

In Fig. 7.47 we compare the radiocarbon dates from all discussed archives (except Greenland ice) with calibration data based on Barbados and New Guinea corals. In the upper part of the figure the calendar ages of boundaries of Younger Dryas are compared. We also show the ages of Laacher See tephra (Van den Bogaard & Schmincke 1985) found in Holzmaar and Soppensee. The ages of YD boundaries in different archives are different, but the discrepancies among radiocarbon data are also large. The differences between AMS radiocarbon dates of adjacent samples from a single archive are not too large, so these data seem reliable. Although the Lake Gościąż data generally fit the coral ones, the ^{14}C dates from Holzmaar and Soppensee are older. Apparently not all calendar chronologies of Lake Gościąż, Holzmaar, Soppensee, and corals are synchronous. The differences in age esti-

Table 7.5. Comparison of calendar-age estimates of the Younger Dryas/Holocene transition (after Goslar et al. 1995b).

Archive	Definition of boundary	Age (yr cal BP)	Reference
GRIP ice core	Abrupt increase of $\delta^{18}O$	11,550±90	Johnsen et al. 1992
GISP2 ice core	Abrupt increase of $\delta^{18}O$, abrupt increase of snow accumulation rate, drop of fluxes of calcium, magnesium etc.	11,640±250	Alley et al. 1993 Taylor et al. 1993 Mayewski et al. 1993
Lake Gościąż	Abrupt increase of $\delta^{18}O$, changes in terrestrial and lacustrine vegetation	11,510±50[1] 11,440±120[2]	this work this work
Lake Soppensee	Changes in terrestrial vegetation	10,986±69	Hajdas et al. 1993
Lake Holzmaar	Changes in terrestrial vegetation	10,630±180[3] 11,510±180[4]	Zolitschka et al. 1992 Hajdas 1993
Swedish varves	Onset of rapid retreat of ice margin, second drainage of Baltic ice lake	10,940 10,980	Strömberg 1994
German pines	Increase of $\delta^{13}C$ and δD in wood	(10,970)[5]	Becker et al. 1991
		11,045[6]	Kromer & Becker 1993

[1] – based on the match of varve thickness sequence to the German dendroscales (see Chapter 6.4),
[2] – based on the match of AMS radiocarbon dates to the ^{14}C calibration data (see Chapter 6.2),
[3,4] – the varve chronology of Lake Holzmaar corrected according to the match of AMS radiocarbon dates to the ^{14}C calibration data,
[5] – the boundary set originally in the pine chronology,
[6] – the boundary set originally in pine chronology, shifted with a tentative tree-ring match to the oak master chronology.

mates of the AL/YD and YD/PB transitions may partly be due to errors in calendar chronologies. An even higher offset is shown by recent Swedish data (Wohlfarth et al. 1995).

We therefore tried to “correct” the calendar chronologies to obtain the ages of YD boundaries similar to that recorded in Greenland, and to synchronize the level of Laacher See Tephra (after Goslar et al. 1995b). This required an addition of 450 varves in the chronology of Soppensee below 10.4 kyr BP, and 600 varves to the sequence from Holzmaar below 11.8 kyr BP. The age of the Lake Gościąż chronology was also adjusted, using the match of the sequence of laminae thickness to the German dendroscales. The German pine chronology was shifted to synchronize with the Gościąż chronology.

The radiocarbon dates of “corrected” chronologies are compared in Fig. 7.48. We observe that the data from different archives are much more consistent than in Fig. 7.47. The plot in Fig. 7.48 clearly demonstrates that the differences of ages of Younger Dryas boundaries are produced mostly by inadequate calendar chronologies. Some doubts may be connected with the two samples from the YD/PB boundary in Soppensee, distinctly younger than the plateau of 10,000 ^{14}C BP. However, the AMS data from non-laminated sediment of the adjacent lake Rotsee (Ammann & Lotter 1989), with a pollen diagram very similar to that from Soppensee, show the YD/PB boundary in the middle of distinct 10,000 ^{14}C BP plateau, traced by as many as 11 dates (Fig. 7.49). Therefore the two critical samples from Soppensee could be regarded as contaminated. As shown by Wohlfarth et al. (1993), the contamination of small macrofossils by modern carb-

Table 7.6. Comparison of calendar-age estimates of the Allerød/Younger Dryas transition (after Goslar et al. 1995b).

Archive	Age cal BP	Reference
GRIP ice core	12,700±100	Johnsen et al. 1992
GISP2 ice core	12,820±260	Alley et al. 1993[1]
Lake Gościąż	12,650± 90[2] 12,580±130[3]	this work this work
Lake Soppensee	12,125± 86	Hajdas et al. 1993
Lake Holzmaar	11,080±210[4] 11,960±210[5]	Zolitschka et al. 1992 Hajdas 1993
Swedish varves	11,800	Wohlfarth et al. 1993

[1] – age reported by Alley et al. (1993) was based on the changes in accumulation rate; the quoted age is that of the midpoint of major drop of $\delta^{18}O$ (Grootes, personal communication),
[2] – based on the match of varve thickness sequence to the German dendroscales (see Chapter 6.4),
[3] – based on the match of AMS radiocarbon dates to the ^{14}C calibration data (see Chapter 6.2),
[4] – varve chronology of Lake Holzmaar,
[5] – varve chronology of Lake Holzmaar corrected according to the match of AMS radiocarbon dates to the ^{14}C calibration data.

on may sometimes alter a ^{14}C age by many hundred years. Obviously contaminated is one sample from Lake Gościąż sediment (indicated in Fig. 7.43 by a question mark).

The only non-synchronous YD/PB boundary is that in German pines which, without any doubt, is delayed with respect to that in Lake Gościąż by ca. 200 yr. This problem has been discussed in detail in the previous section.

The plot in Fig. 7.48, though demonstrating the close synchronism of boundaries of Younger Dryas, cannot identify their real calendar ages, because one could argue that, along with the chronologies of Soppensee and Holzmaar, the U/Th chronology of corals and varve chronology of Lake Gościąż are inadequate. The correction of

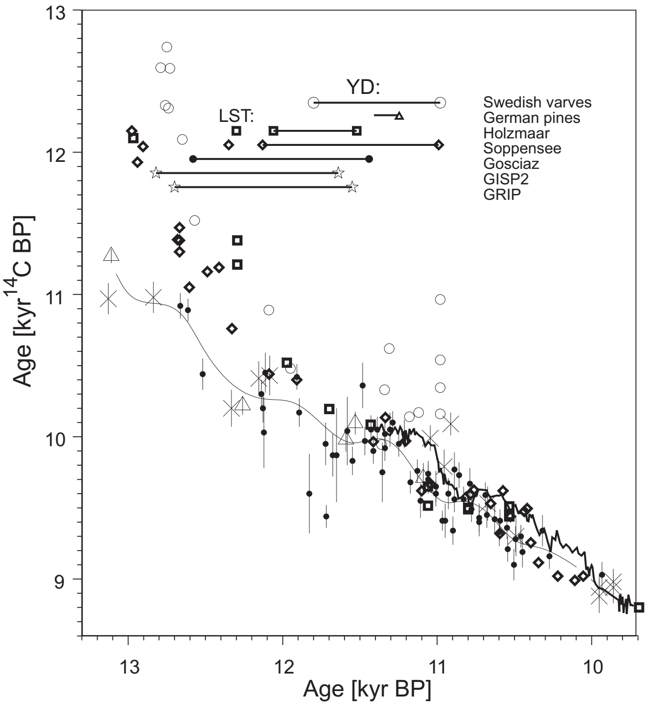

Fig. 7.47. Comparison of ^{14}C and calendar ages, derived from U/Th dates, dendrochronology, and varve chronologies in the Late-Glacial and early Holocene, and of the age of the YD boundaries reconstructed from different archives (after Goslar et al. 1995b). In the upper part of the figure, the boundaries of YD are shown by points connected with heavy lines (except the YD/PB boundary in German pines). The age of the Laacher See tephra (LST) is shown to the left of the Holzmaar and Soppensee bars. Heavy line: German pines (Kromer & Becker 1993); squares: Lake Holzmaar macrofossils (Hajdas 1993); diamonds: Soppensee macrofossils (Hajdas et al. 1993); black circles and thin smoothed line: Lake Gościąż macrofossils (this work); open triangles: Barbados corals (Bard et al. 1993); crosses: Huon Peninsula corals (Edwards et al. 1993).

Holzmaar and Soppensee chronologies would require some hundred varves missing from the sequences, while the error of Lake Gościąż would require the fragment of some hundred varves to be doubled. It must be stressed that, on the basis of AMS radiocarbon dates Hajdas (1993) demonstrated the lack of ca. 880 varves in the Lake Holzmaar sequence from the 4th millennium BP.

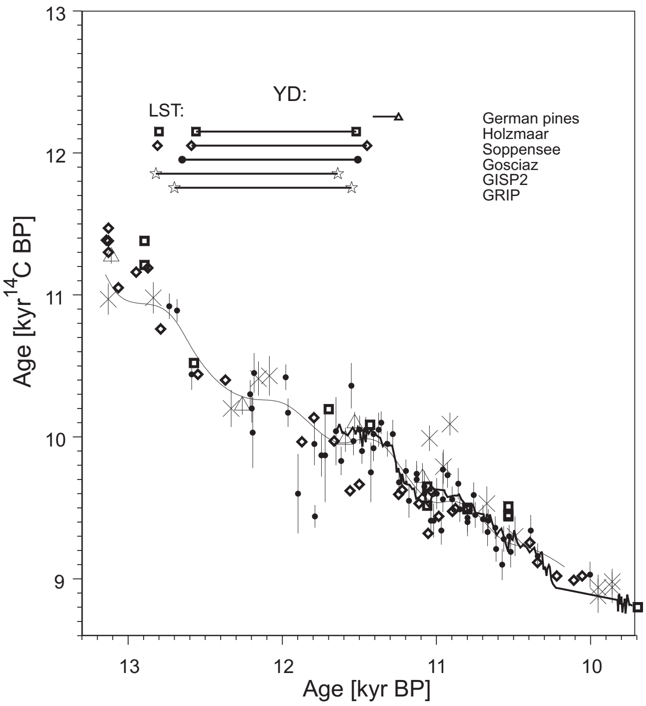

Fig. 7.48. Revised comparison of ^{14}C dates and boundaries of the YD cold period, reconstructed in archives considered in Fig. 7.47 (after Goslar et al. 1995b). The dates are modified after correction the varve chronologies of Lake Holzmaar and Soppensee to synchronize the boundaries of the YD in the North Atlantic region, the adjustment of the Lake Gościąż chronology, and the shift of the German pine chronology to synchronize with that of Lake Gościąż (for further explanations see the text). The description of symbols is the same as in Fig. 7.47.

This gap was not detected when the varve structures were analysed. On the other hand, the doubling of the laminated sequence (by a slump?) with no serious disturbance

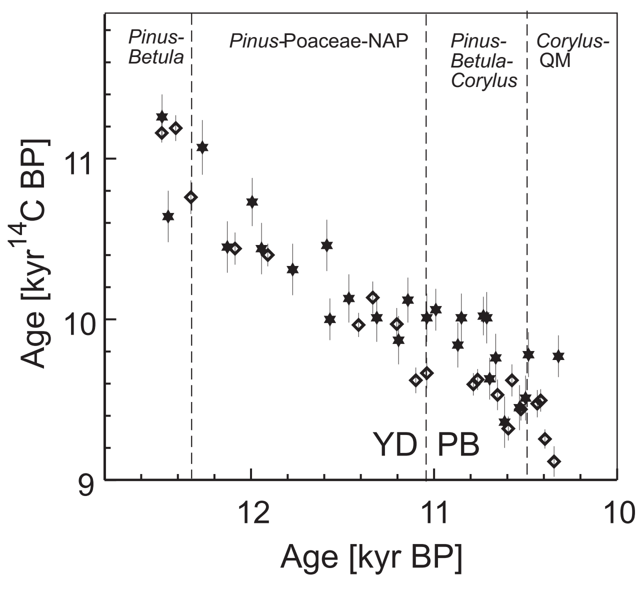

Fig. 7.49. Comparison of ^{14}C dates of macrofossils from Soppensee (diamonds) and non-laminated sediments of Rotsee (stars; Ammann & Lotter 1989). The calendar ages of Soppensee samples are as published by Hajdas et al. (1993). The time scale for Rotsee was obtained by synchronization of boundaries between corresponding biozones in both lakes (Lotter et al. 1992b) and linear interpolation between boundaries according to sample depth.

of the laminated structure seems unlikely. Further, close varve-to-varve correlation of laminated sequences from two separate lake deeps also seems to preclude the occurrence of any slump. The serious revision of Lake Gościąż chronology would additionally require the revision of U/Th chronology of corals, which is difficult to justify. The validity of Lake Gościąż chronology is also supported by the agreement of the ages of YD boundaries with those recorded in Greenland. The above-mentioned arguments seem to indicate that some varves in the chronologies of Soppensee and Holzmaar are missing rather than the chronology of Lake Gościąż is erroneous. At any rate, the interpretation presented needs proof by further studies.

Variations of atmospheric radiocarbon concentration

The plateaux of radiocarbon ages are consequences of declines of atmospheric concentration of ^{14}C. The concentration of radiocarbon is usually expressed as Δ^{14}C, the difference between the actual specific ^{14}C activity and the activity in the standard of modern biosphere:

$$\Delta^{14}C(t) = \frac{A_s\,(t) - A_{st}}{A_{st}} \cdot 1000‰$$

The variations of atmospheric Δ^{14}C, reconstructed using Lake Gościąż data are presented in Fig. 7.50. Our data seem also to confirm the large increase of radiocarbon concentration at the onset of YD, from ca. 180–200‰ to ca. 240–250‰ within 300 yr. This is the largest increase of atmospheric ^{14}C concentration in the whole Late-Glacial and Holocene record. On the other hand, the drop of Δ^{14}C by 50–70‰ at the YD/PB boundary is the largest and most rapid of the series of declines occurring regularly between 12.3 and 9.8 kyr BP. This general long-term decline, by ca. 150‰ within ca. 2.5 kyr, is not comparable to any other in the whole Holocene. The coincidence of large climatic and radiocarbon variations suggests some common causes, which needs to be discussed in detail.

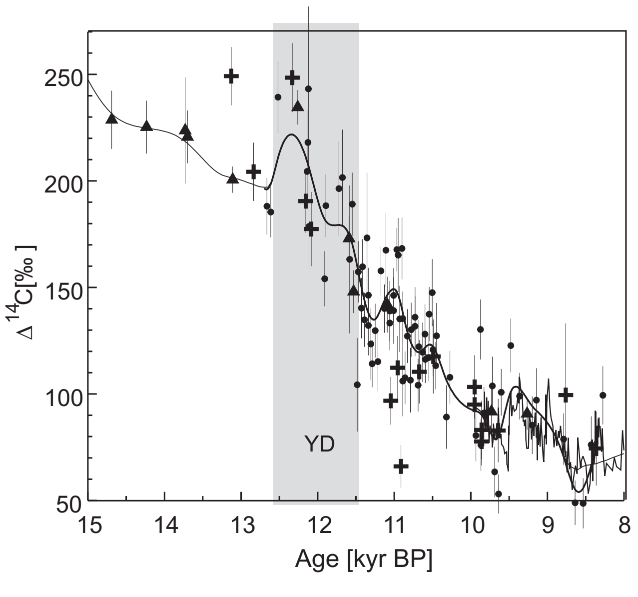

Fig. 7.50. The reconstruction of atmospheric ^{14}C concentration in the Late-Glacial and early-Holocene (after Goslar et al. 1995a). For explanation of symbols see Fig. 7.43. The cubic splines fitted to LG data and, before 12.6 kyr BP, to the coral data, were the basis of model calculations.

The possible causes of Δ^{14}C variations are changes of ^{14}C production rate and reorganizations in global carbon cycle. It is commonly accepted that the secular Δ^{14}C trend in the Holocene reflects the changes of geomagnetic dipole moment, while the variations of a period of a few hundreds years are connected to changing solar activity (Beer et al. 1988, Stuiver et al. 1991, Stuiver & Braziunas 1993). It has also been proposed that the strongest changes in the global carbon cycle, which could perhaps affect significantly the atmospheric radiocarbon concentration, happened at the Late-Glacial/Holocene interface. To recognize whether the production rate or carbon cycle were responsible for the observed variations of Δ^{14}C, the independent data on factors affecting the concentration of ^{14}C should be considered.

Variations of the radiocarbon production rate

The steady-state mean ^{14}C/^{12}C ratio on the Earth is a result of equilibrium between production by cosmic rays and radioactive decay. The changes in flux of galactic protons, solar-wind magnetic properties, or the dipole moment of the Earth may all influence the production rate and thus contribute to the variations of Δ^{14}C. The present knowledge about possible variations of the cosmic rays on the timescale of interest is rather poor. As shown below, the Δ^{14}C variations can be explained by changes of the Earth's magnetic field and changes in the carbon cycle. The cosmic-ray flux can thus be assumed to be constant until reliable data show the contrary. Changes in the solar-wind modulation are not directly controlled either. However, the regular occurrence of Δ^{14}C maxima 180–220 yr wide during the Holocene allows us to believe that they are induced by the solar events, similar to those responsible for the Spörer and Maunder minima (Stuiver et al. 1991, Stuiver & Braziunas 1993, Eddy 1976). This seems to be additionally confirmed by the fairly good agreement between observed Δ^{14}C and that calculated from the production rate derived from analyses of ^{10}Be in polar ice (Beer et al. 1988, Siegenthaler & Beer 1988). Unfortunately, the concentrations of ^{10}Be in Late-Glacial ice are strongly affected by changes in accumulation rate. Therefore, unless contradictory evidence is shown, we can only assume that the Late-Glacial solar-wind-induced variations of the ^{14}C production rate were similar to those in Holocene.

The best recognized factor influencing the radiocarbon production is the magnetic moment of the Earth

(McElhinny & Senanayake 1982, Constable & Tauxe 1987, Tric et al. 1992, Meynadier et al. 1992). Since the atmospheric $\Delta^{14}C$ levels after 18 kyr BP are placed within the range of $\Delta^{14}C$ allowed by uncertainty of palaeomagnetic data, the observed $\Delta^{14}C$ variations appeared fully explained by the variations of geomagnetic field (Bard et al. 1990, Mazaud et al. 1991). However, because the carbon cycle acts as a low-pass filter, short-term variations within the calculated $\Delta^{14}C$ range may require much larger changes of the production than the uncertainty of geomagnetic dipole moment (GDM) would allow. To recognize this, the models should be run in the deconvolution mode, calculating production rate from known ^{14}C concentrations. The simple box models (Houtermans et al. 1973, Siegenthaler et al. 1980, Oeschger et al. 1975, Siegenthaler & Oeschger 1978) were run in such a mode, starting from the steady state with $\Delta^{14}C$ = 230‰ and using as input data the ^{14}C concentrations represented in Fig. 7.50 by the smoothed line. The results given by different models are very similar. In Fig. 7.51a we compare the mean curve of required production rate with the production rate derived from palaeomagnetic data compiled by Tric et al. (1992), using the relations of Lal (1988). The required production rate falls out of the range allowed by GDM reconstruction. The most prominent discrepancies are the short increase at the onset of Younger Dryas and ca. 1.5-kyr-long lowering of the production in the early Holocene. The 30% minimum at the YD/PB transition is apparently coincident with the period of large uncertainty of Tric's reconstruction, but other GDM data also confirm the significance of this discrepancy. It must be mentioned that the age control of GDM reconstruction is worse than that of $\Delta^{14}C$, and direct comparison of individual peaks is rather speculative. It must also be mentioned that the calculations probably underestimate the required variations of production rates since the input $\Delta^{14}C$ record was much smoothed.

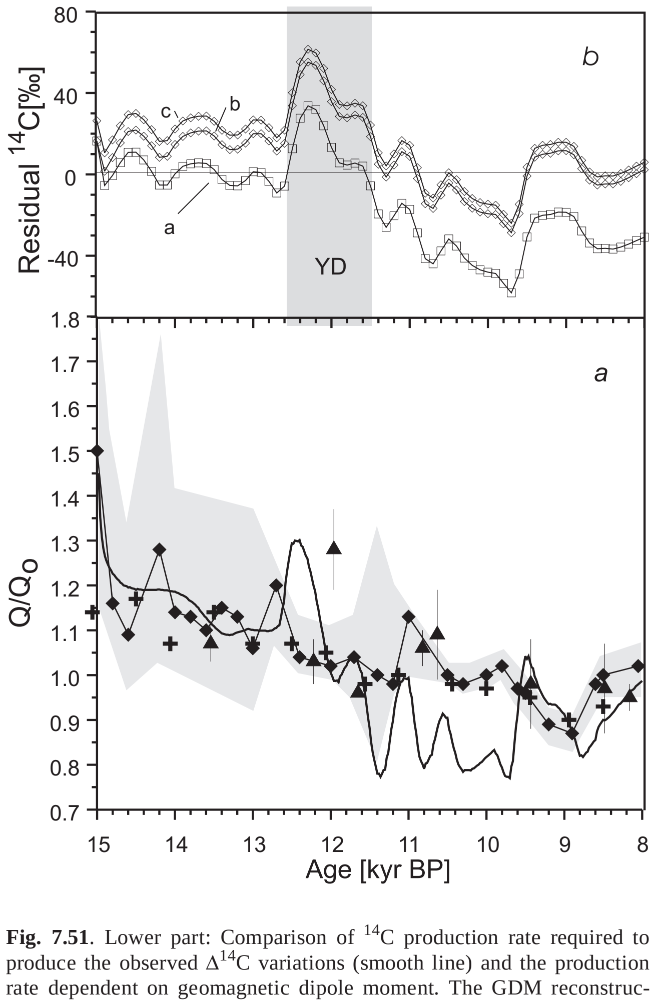

Fig. 7.51. Lower part: Comparison of ^{14}C production rate required to produce the observed $\Delta^{14}C$ variations (smooth line) and the production rate dependent on geomagnetic dipole moment. The GDM reconstructions come from: diamonds – Mediterranean sedimentary records (Tric et al. 1992), triangles – volcanic data compiled by Tric et al. (1992), crosses – box cores in the North Pacific (Constable & Tauxe 1987). The range of untercainty of the GDM-dependent production rate is shaded. Upper part: Differences between reconstructed ^{14}C concentrations and those calculated from geomagnetically induced production rates. a – $\Delta^{14}C$ = 230‰, production rate from Tric et al. (1992), b – production rate systematically lowered by 5%, c – $\Delta^{14}C$ lowered by 10‰.

The residuals between the smoothed curve of observed $\Delta^{14}C$ and those calculated using geomagnetically induced production rates are shown in Fig. 7.51b. The results of model calculations of $\Delta^{14}C$ are sensitive to the absolute value of ^{14}C production and to the choice of initial conditions. The run starting at steady state of $\Delta^{14}C$ = 230‰ reconstructs well the pre-YD radiocarbon concentration, but it leads to Holocene $\Delta^{14}C$ levels on average too high by ca. 30‰. The better reconstruction of mean early-Holocene $\Delta^{14}C$ levels was obtained by the lowering of the production rate by 5%. In such a case, however, the model predicted too low ^{14}C concentrations prior to the YD. The change of initial ^{14}C level by 10‰ has little influence on the early-Holocene $\Delta^{14}C$ values. It seems that the change of mean $\Delta^{14}C$ level between 15–12.5 kyr BP and 9.5–8 kyr BP and its large variations between 12.5 and 9.5 kyr BP are both independent of changes of Earth's magnetic field and must be explained by other mechanism.

The difference between observed and geomagnetically induced changes of $\Delta^{14}C$ values might be explained by variations of solar activity. Although the amplitude of $\Delta^{14}C$ maximum during the Younger Dryas is similar (30–40‰), but its duration (1000–2500 yr) is distinctly larger than those induced by solar activity in the Holocene (Stuiver et al. 1991). Therefore another possible cause of Late-Glacial/early-Holocene $\Delta^{14}C$ variations should be sought in the changes of global carbon cycle.

Variations in the global carbon cycle

The general property of global carbon cycle is that nearly all radiocarbon is produced in the atmosphere, while most of it decays in the deep ocean. For that reason, the concentration of radiocarbon in the atmosphere is higher than in the deep ocean and depends on the size of the atmospheric carbon reservoir and the rate of carbon exchange with the ocean. Generally, the smaller the atmospheric CO_2 inventory and its exchange with the deep sea are, the higher is the specific ^{14}C activity in atmosphere.

The large increase of atmospheric CO_2 from ca. 200 to 280 ppm between 17 and 10 kyr BP is well documented (Barnola et al. 1987, Neftel et al. 1988, Staffelbach et al. 1991), though the mechanisms of this increase are not completely known (Leuenberger et al. 1992, Marino et al. 1992). The model calculations show that change of CO_2 from 200 to 280 ppm might decrease steady state $\Delta^{14}C$ by 25–75‰ (Lal & Revelle 1984, Keir 1983, Siegenthaler et al. 1980), so the 20‰ decrease between mean pre-YD and Holocene $\Delta^{14}C$ residuals could be explained by the change of CO_2 concentration. On the other hand, the changes of CO_2 during the short periods of $\Delta^{14}C$ variations at the boundaries of YD and about 9.5 kyr BP were presumably too small to alter $\Delta^{14}C$ more than by 10‰. Another factor influencing atmospheric $\Delta^{14}C$ is the rapidity of carbon exchange between atmosphere and ocean surface. It is generally accepted that the winds were stronger in glacial time (Petit 1981), raising the exchange by even 50% (Bard 1988), and due to this effect one could expect the lowering of $\Delta^{14}C$ by 20‰ (Siegenthaler et al. 1980). This mechanism would cause the lowering of $\Delta^{14}C$ during cold periods, and therefore it cannot explain observed $\Delta^{14}C$ variations.

The other possible mechanism is the exchange with the deep ocean. It is especially important, since a significant part of total exchange is through the formation of North Atlantic Deep Water (NADW), which is a driving force of thermohaline circulation heating the North Atlantic region (Manabe & Stauffer 1988). As was hypothesized by Broecker & Denton (1989), switching NADW formation on and off would be a plausible mechanism of abrupt climatic changes. According to box or box-diffusion models, the instant drop of exchange with the deep ocean by a factor of two would raise $\Delta^{14}C$ by 100‰ within a few hundred of years (Siegenthaler & Beer 1988). This, however, should be accompanied by an increase of radiocarbon age of deep water by more than 1000 yr. The growing evidence from radiocarbon dating of contemporaneous planktonic and benthic foraminifera (Andree et al. 1986, Shackleton et al. 1988, Broecker et al. 1988, 1990) says that such an increase did not occur. Therefore the Late-Glacial/early-Holocene decline of $\Delta^{14}C$ by 150‰ cannot be explained by changes of deep-sea ventilation alone. Fortunately most of this decline could be explained by the GDM-induced changes of radiocarbon production and changes of CO_2 concentration. The residual variations, the 40‰ increase of $\Delta^{14}C$ at the onset of YD and its 30‰ decline at the beginning of the Holocene, would require the changes of deep-sea ventilation less than 20%. The corresponding ^{14}C age difference between benthic and planktonic foraminifera would then change at the beginning and termination of YD by only 250–300 yr. The difference of ventilation index between Glacial Maximum and Holocene by 75–325 yr seems well documented (Broecker et al. 1990), but the data available now are too scarce to support or to rule out the postulated short-term $\Delta^{14}C$ variations during YD and early Holocene. The $\Delta^{14}C$ record from YD, if interpreted in terms of ventilation rate, suggests the lowest oceanic circulation in the earlier part of Younger Dryas. This seems to be consistent with the amelioration of climate in the second half of YD, reconstructed in the Lake Gościąż sediment (Goslar et al. 1993) and other regions in Europe (Björck & Digerfeldt 1984, Pennington 1977, Lowe & Walker 1980, De Groot et al. 1989).

The direct reconstructions of NADW formation during YD and early Holocene do not give clear results. Many palaeooceanographic reconstructions, based on faunal, chemical and isotopic data (Boyle & Keigwin 1987, Boyle 1988, Keigwin et al. 1991, Keigwin & Lehman 1994) suggest that the NADW formation was significantly reduced during both YD and Last Glacial Maximum (LGM). Other studies (Jansen & Veum 1990, Charles & Fairbanks 1992) suggest that to some extent the formation of NADW was continued throughout the YD. These discrepancies may be reconciled if only the deeper part of NADW circulation was weakened while the upper belt persisted, though at different depths during LGM and YD (Lehman & Keigwin 1992). The reconstruction of distribution of different ocean-water masses in the past is then a very complicated task and is beyond the scope of the present work. Obviously the radiocarbon concentration in the atmosphere is not sensitive to the details of oceanic circulation but rather integrates the effects of vertical mixing over the whole ocean.

7.8. DISCUSSION OF THE LATE-GLACIAL RECORDED IN THE LAKE GOŚCIĄŻ SEDIMENTS

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The extended studies on sediments of Lake Gościąż and its environmental setting aimed to reconstruct the history of the lake and surrounding area during the last 15 kyr. They brought a large set of data concerning changes of different environmental components, which, for the Late-Glacial period, were described in individual Chapters 7.1–7.7. However, the components of environmental system respond jointly to external forcing as well as interact with one another, so they cannot be treated separately. In this chapter we attempt to confront the more significant data from individual chapters (Tab. 7.7) and discuss them together. For reasons outlined in the In-