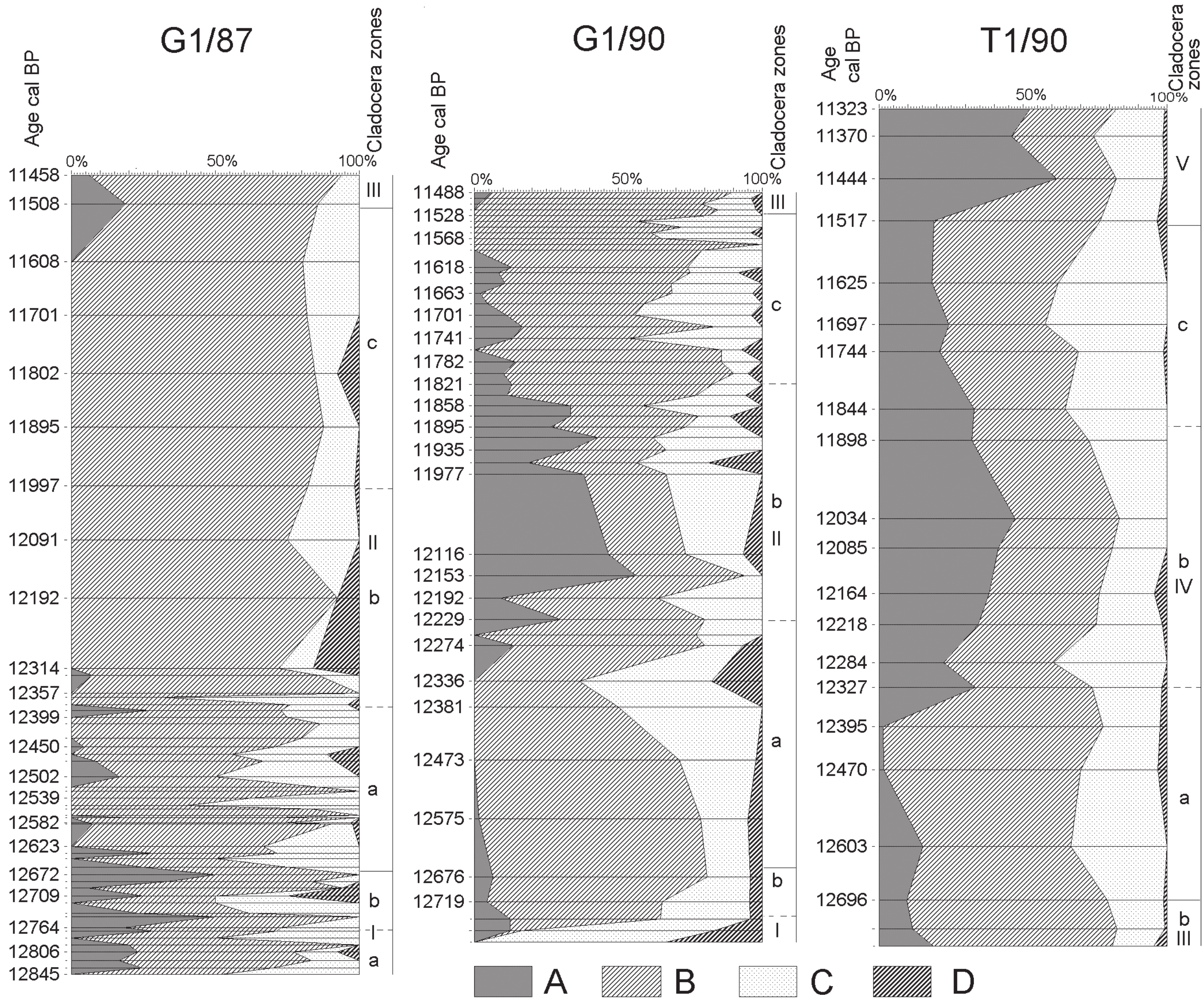

Fig. 7.36. The planktonic/littoral Cladocera ratio in profiles: G1/87, G1/90, and T1/90 (Late-Glacial). A – *Bosmina longirostris,* B – planktonic exl. *Bosmina longirostris,* C – littoral exl. *Chydorus sphaericus,* D – *Chydorus sphaericus.*

7.6. ISOTOPIC INDICATORS OF THE LATE-GLACIAL/HOLOCENE TRANSITION RECORDED IN THE SEDIMENTS OF LAKE GOŚCIĄŻ

*Tadeusz Kuc, Kazimierz Różański & Marek Duliński**

Climate changes recorded in stable isotopes composition

Over the past three decades, a large number of palaeoclimatic studies was carried out for both marine and continental environments, using stable isotopes as proxy indicators of climatic changes. Of particular importance are detailed oxygen-18 profiles obtained for two recently drilled ice cores in central Greenland (Taylor et al. 1993, Grootes et al. 1993). They cover the full glacial/interglacial cycle. The records of proxy climatic indicators of comparable (annual) resolution are not easily available for continental environments.

The isotopic studies of lacustrine sediments performed to date have demonstrated the usefulness of stable isotopes as a powerful tool in reconstructing past climatic and environmental changes on the continents (e.g. Dean & Stuiver 1993, Gasse et al. 1991, Kuc et al. 1993). Lake sediments often contain authigenic carbonates and fossil shells whose oxygen isotope composition is mainly controlled by that of lake water and by temperature. For open lakes with fast water turnover, the oxygen isotope composition of the lake water corresponds to that of precipitation over the lake basin, which in turn is temperature-dependent. The periods of cold and mild climate are reflected by minima and maxima in the oxygen-18 content of the deposited carbonate. For closed lakes, the

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isotopic composition of water is largely determined by evaporation and exchange with atmospheric moisture; dry periods are marked by high oxygen-18 content due to enhanced evaporation. Thus interpretation of $\delta^{18}O$ records from lacustrine sediments will require an evaluation of palaeohydrological conditions.

Variations in carbon-13 content of lake carbonates are usually associated with changes in the carbon balance of the lake; increase of carbon-13 content is usually interpreted as an indication of enhanced biological activity of the lake and/or better ventilation (changes in the water level), both processes being climatically controlled. However, the interpretation of $\delta^{13}C$ records is seldom so straightforward. In many cases additional processes, either in the lake itself (diagenesis, methanogenesis) or in the catchment area (changes in the vegetation cover, chemical erosion), may influence the carbon-13 content in the deposited carbonates and make interpretation of the $\delta^{13}C$ more complex and equivocal in many cases.

Sampling for stable isotope measurements

Over the past eight years, several cores were retrieved from both deep and shallow parts of Lake Gościąż (Fig. 7.37). The cores were stored in a controlled environment and isolated from ambient air. Isotopic composition ($\delta^{18}O$ and $\delta^{13}C$) of bulk carbonate fraction was analysed in selected cores. Samples obtained from the cores usually encompassed 2 to 10 varves. The sampling was continuous for the basal part of the sediments (Allerød, Younger Dryas, Preboreal). The Holocene portion of the cores was sampled at 50-year intervals (Różański et al., Chapter 8.6). Extensive cross-correlation of varve sequences in different cores and the AMS radiocarbon dating of macrofossils extracted from the cores enabled conversion of the varve chronology to the absolute chronology (Goslar, Chapter 7.2). Consequently, the age of samples in isotopic profiles can be expressed versus calendar years. For more technical details see Chapter 4.4 (Methods).

Isotopic profiles of the cores

Isotope analyses of bulk carbonate fraction in the Lake Gościąż sediments were performed to date for five cores (Fig. 7.37): two cores (twin) representing the central deep (G1/87, G2/87), one core pulled from the western deep (G1/90), and two short cores retrieved from the Tobyłka Bay (T1/90) and adjacent shore-fan (G28/92).

The lowest parts of the cores retrieved from the western deep of the lake (G1/90) represent the younger phase of the Allerød warm episode. The oldest non-laminated part (ca. 20 cm long, with a 5 cm layer of peat/humus), shows gradually decreasing $\delta^{13}C$ values, from +0.4‰ below the peat layer to -5.0‰ at the transition to laminated segment above (Fig. 7.38). The $\delta^{18}O$ values in the

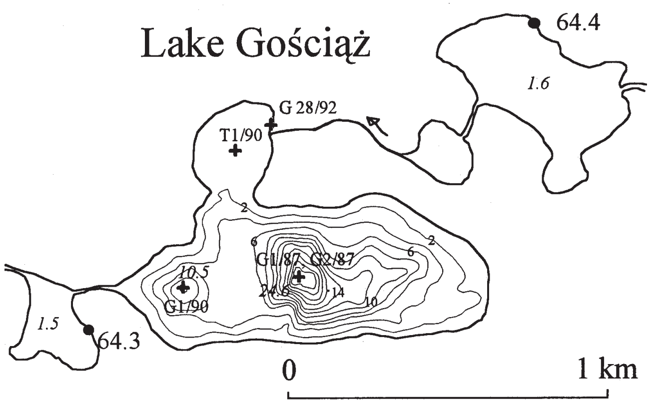

Fig. 7.37. Bathymetric map of Lake Gościąż with indicated coring points from which carbonates were analysed.

non-laminated lower part scatter between -7.9‰ and -9.5‰. The observed isotopic variability most probably reflects the early stages of formation of the lake: decreasing $\delta^{13}C$ of the precipitated carbonates might indicate gradually increasing water depth and a weakening exchange with atmospheric CO_2, whereas the relatively large scatter of $\delta^{18}O$ may show isotopic variability of the inflow and/or unstable hydrological regime of the relatively young part of the lake and associated changes in the degree of the evaporative enrichment of the lake water.

The Allerød/Younger Dryas transition (AL/YD) is very well marked in all five cores analysed so far. The "twin" cores G1/87 and G2/87, for which high-resolution sampling has been performed and shown in Fig. 7.39 as a combined profile, reveal a drop of $\delta^{18}O$ at the AL/YD transition of approximately 2‰ (from peak to valley) and a corresponding increase of $\delta^{13}C$ of about 1.5‰ (Fig. 7.39). Similar isotopic changes are observed in the cores:

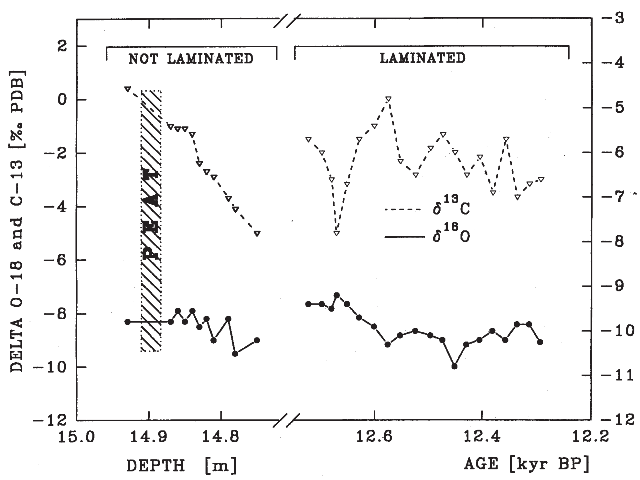

Fig. 7.38. $\delta^{18}O$ and $\delta^{13}C$ records from the bottom part of the G1/90 core (western deep). The deepest samples represent non-laminated Allerød sediment with ca. 5 cm thick peat/humus layer.

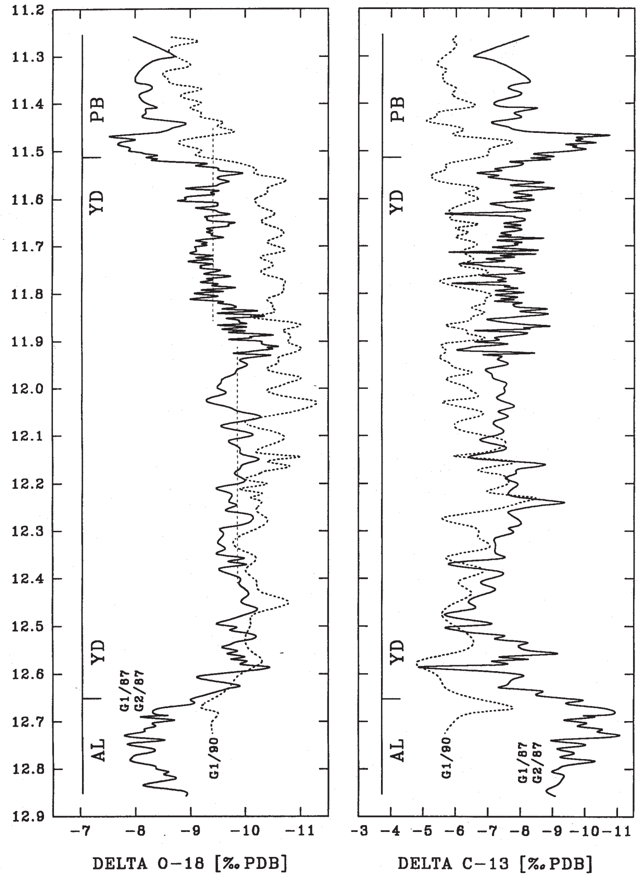

Fig. 7.39. $\delta^{18}O$ and $\delta^{13}C$ combined records for the Younger Dryas segment of the G1/87 and G2/87 cores (solid line) and the G1/90 core (short dashed line). The curves were smoothed by cubic spline method. Vertical dashed lines show average values of $\delta^{18}O$ for subperiods.

T1/90 (Fig. 7.41), G28/92 (Fig. 7.42), and G1/90 (western deep), but in this last case the Allerød part is short and without a distinct maximum. Duration of the AL/YD transition, was estimated to be around 150 years (between maximum and minimum $\delta^{18}O$ values), or around 80 years if calculated between the average $\delta^{18}O$ values (see Fig. 7.40). The cores T1/90 and G28/92 also indicate a rapid transition from Allerød to Younger Dryas. The AL/YD transition is also well marked in the $\delta^{13}C$ records: the $\delta^{13}C$ values increase with the cooling of climate. For the shallow part of the lake (cores T1/90 and G28/92) the observed isotope shift is about 2‰, whereas for the deepest part of the lake (cores G1/87, G2/87) it increases by approximately 3‰.

Stable-isotope composition of the bulk carbonate fraction was analysed in the above mentioned five cores for the entire Younger Dryas period. The largest number of analyses was performed on the cores retrieved from the central deep (G1/87, G2/87), for which the selected parts were sampled with high resolution (2 to 5 varves per sample). The obtained $\delta^{18}O$ record reveals characteristic fluctuations, which are more pronounced in the younger part of the Younger Dryas (Ralska-Jasiewiczowa et al. 1992). The estimated periodicity is about 80 years and the range of variations close to 0.5‰. Moreover, this younger part, lasting approximately 300 years, reveals slightly enriched $\delta^{18}O$ values as compared to the older part (on the average by ca. 0.3‰), suggesting the existence of two subperiods within the Younger Dryas with slightly different climate, documented also by palynological investigations (Ralska-Jasiewiczowa et al., Chapter 7.4). The G1/90 core (western deep), analysed with lower resolution, shows a record similar to that of G1/87 and G2/87, with majority of peaks coinciding. However, the average values are systematically shifted for both isotopes, and the subperiods within the Younger Dryas are not well marked (Fig. 7.39). This may be due to local conditions of sedimentation (e.g. difference in water temperature). The laminated core T1/90, originating from the present shallow part of the lake (Tobyłka Bay), shows a stepwise decrease of $\delta^{18}O$ approximately 200 years after the onset of the Younger Dryas, and a gradual increase about 250 years before the termination of this period (Fig. 7.41). The profile G28/92, which is relatively short and not laminated (Demske 1995), displays a very sharp AL/YD transition with an almost constant $\delta^{18}O$ value during the Younger Dryas. This core also shows less negative values of $\delta^{18}O$ during the late Younger Dryas, when compared to the G1/87 and G2/87 (Fig. 7.42). The $\delta^{13}C$ values within the Younger Dryas are relatively constant, not correlated with $\delta^{18}O$ values. An apparent gradual enrichment in carbon-13 with the decreasing water depth is observed: from about -8.0‰ in the deepest part

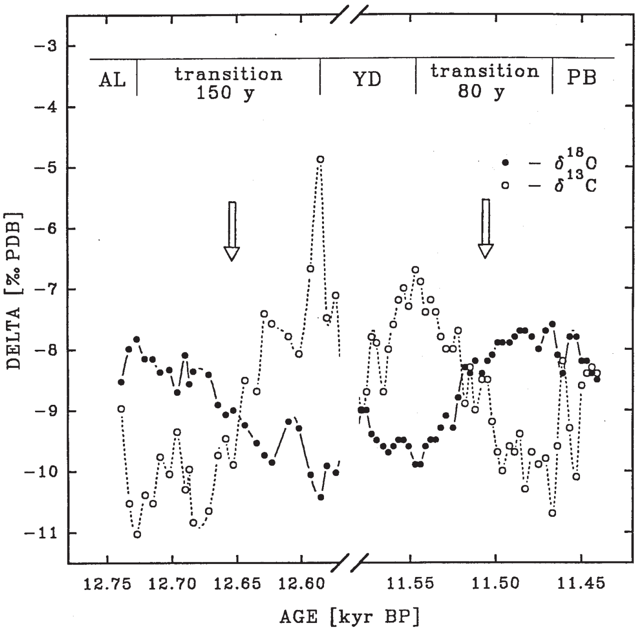

Fig. 7.40. Transition zones: AL/YD and YD/PB in the isotopic records for G2/87 core. Onset and termination of the Younger Dryas were set in the middle of zones (arrows), and corresponding dates (12,650 BP and 11,510 BP) were calculated according to the calendar year scale (Goslar, Chapter 7.2).

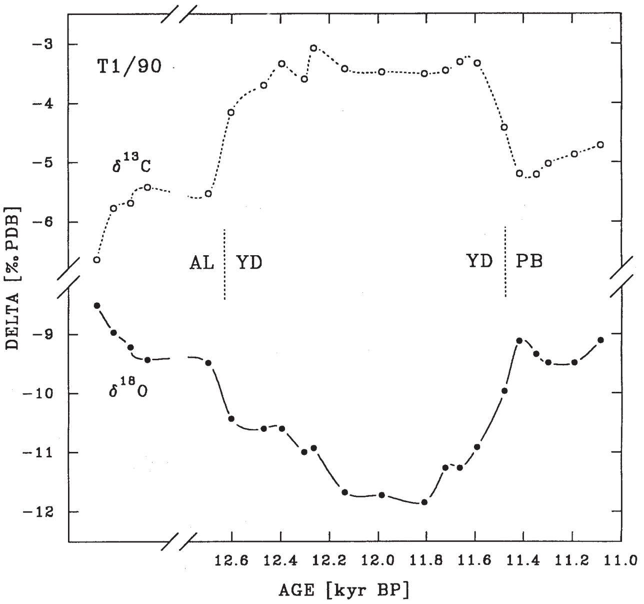

Fig. 7.41. $\delta^{18}O$ and $\delta^{13}C$ records for the Younger Dryas part of the T1/90 core, Tobyłka Bay, Lake Gościąż.

of the lake (cores G1/87 and G2/87) to around -2.2‰ in its shallow part (core G28/92).

The Younger Dryas/Preboreal transition (YD/PB) is also very well marked in all the analysed cores. The duration of this transition, based on the $\delta^{18}O$ profiles for the cores from the central deep, was estimated to be around 80 years (between minimum and maximum $\delta^{18}O$ values) or only about 20 years when the average values are considered (Fig. 7.40). Very close duration is found in the G1/90 core from the western deep. The change of $\delta^{18}O$ values across the transition amounts to approximately 2‰, and is similar also for the cores from shallower parts (G1/90, T1/90, G28/92).

Isotopic boundaries of climate transitions

The Late-Glacial/Holocene transition after the Younger Dryas cold episode, is very well marked in oxygen-18 and carbon-13 isotope records preserved in authigenic carbonates of the Lake Gościąż sediments. The $\delta^{18}O$ profiles in all analysed cores suggest a short duration of both onset and termination of the Younger Dryas. The transition from Younger Dryas to Preboreal appears in the isotope profiles substantially faster than the AL/YD transition; it was completed within several decades. Having well defined transition zones in the isotopic records, the boundaries of the Younger Dryas were set conventionally in the middle of each transition zone (Fig. 7.40). The absolute varve chronology developed for the lower part of the sediment column enabled reporting the age of transitions in calendar years: 12,650 cal BP and 11,510 cal BP, for the AL/YD and YD/PB transitions, respectively (Goslar, Chapter 7.2). The $\delta^{18}O$ profiles available from the central deep of the lake (cores G1/87, G2/87, G1/90) are very similar to the $\delta^{18}O$ profiles available from the GISP2 and GRIP ice cores recently drilled in central Greenland (Taylor et al. 1993, Grootes et al. 1993). The characteristic structure of the YD/PB transition, with a very fast rise of oxygen-18 content at the boundary, preceded by a relatively small but distinct minimum in $\delta^{18}O$ and followed by a brief, temporary drop and further gradual increase of $\delta^{18}O$ values, is well recognized in all profiles. As in the Greenland profiles, the lowest $\delta^{18}O$ values in the Lake Gościąż cores are observed in the early part of Younger Dryas. The similar Late-Glacial/early Holocene $\delta^{18}O$ variations were also recognized in other Greenland ice cores (Johnsen & Dansgaard 1992) and in the lake sediments of Gerzensee, Switzerland (Siegenthaler et al. 1984). The amplitude of $\delta^{18}O$ change at YD/PB transition recorded in Lake Gościąż does not differ substantially from the change observed in the Gerzensee sediments. The duration of the Younger Dryas, defined by oxygen-18 isotope boundaries (Fig. 7.40) and by counting of the annual lamination preserved in the Lake Gościąż sediments (Goslar, Chapter 7.2), was estimated to be around 1140±40 years, in very good agreement with the estimates derived from ice-core studies.

A comparison of $\delta^{18}O$ and $\delta^{13}C$ profiles in the analysed cores reveals a distinct anticorrelation between these two parameters observed during the AL/YD and YD/PB transitions ($r = 0.8$), and a lack of significant correlation within the Younger Dryas ($r = 0.2$). This lack of correlation points to an open lake system with fast turnover of water, and the long-term stability in its water balance. However, the apparent anticorrelation is rather surprising in view of numerous published examples

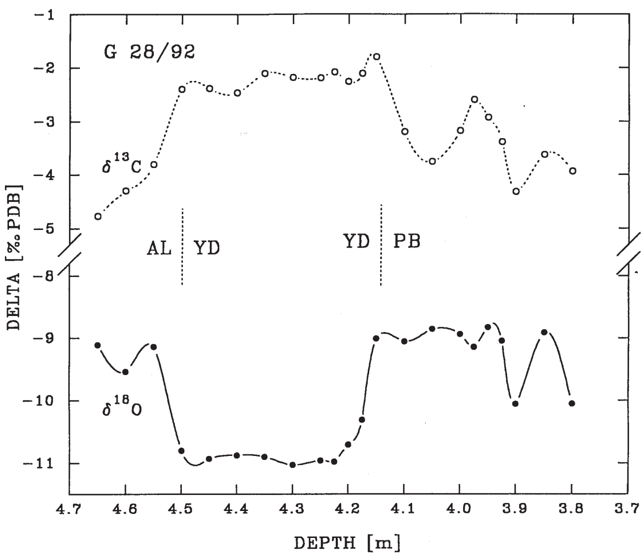

Fig. 7.42. $\delta^{18}O$ and $\delta^{13}C$ records for the lowest part of the non-laminated core G28/92, peat bog, close to Tobyłka Bay.

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.