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# Late-Glacial pollen analysis, oxygen-isotope record, and radiocarbon stratigraphy from Rotsee (Lucerne), Central Swiss Plateau

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## ABSTRACT

Pollen analysis and stable isotope analysis for Rotsee sediment cores revealed three major climatic changes during the Late-Glacial period. AMS-$^{14}$C ages for 59 terrestrial plant macrofossil samples provide a high-resolution time scale for those climatic changes. The cores suggest two periods of constant radiocarbon ages: at 12,600 and at 10,000 yrs B.P. Both coincide with phases of climatic warming. A new correlation between chrono- and palynostratigraphy is proposed on the basis of the presented hardwater error free AMS-dates.

## ZUSAMMENFASSUNG

Die Resultate der Pollenanalyse sowie der Analyse stabiler Isotope ($\delta^{18}$O) lassen in den Sedimenten des Rotsees bei Luzern drei grössere Wechsel des Klimas (Temperatur) im Verlauf des Spätglazials erkennen. Mittels 59 Beschleuniger-$^{14}$C-Daten an fossilen Pflanzenresten terrestrischer Herkunft (hauptsächlich *Betula*-Früchte und -Kätzchenschuppen), wurde eine zeitlich hochauflösende Radiokarbonstratigraphie erarbeitet, um die durch Bio- und Isotopenstratigraphie aufgezeigten Klimawechsel in einen zeitlichen Rahmen zu stellen. Die Altersbestimmungen an den nicht durch einen Hartwasserfehler beeinflussten Makroresten zeigen zwei Abschnitte mit konstanten $^{14}$C-Altern (sog. Plateaux), welche um 12 600 B.P. und um 10 000 B.P. liegen. Aufgrund von palyno- und isotopenstratigraphischen Vergleichen mit andern Lokalitäten des Schweizer Mittellandes sowie aufgrund der Sedimentzusammensetzung kann eine kurzfristige Erhöhung der Sedimentationsrate während diesen beiden Abschnitten ausgeschlossen werden. Es lässt sich eine Koinzidenz zwischen den Phasen klimatischer Erwärmung und dem Auftreten dieser $^{14}$C-Plateaux feststellen. Aufgrund des konstanten Radiokarbonalters wird eine Abschätzung der Dauer dieser für die Vegetations- und Klimageschichte äusserst wichtigen Zeitabschnitte verunmöglicht; nur $^{14}$C-unabhängige Datierungsmethoden (Varven, Baumringe) können in diesen Zeitabschnitten weiterführen.

Auf der Basis der hochauflösenden Radiokarbonstratigraphie schlagen wir für das Schweizer Mittelland eine neue Korrelation zwischen Chrono- und Palynostratigraphie vor: Der Beginn der Bølling-Chronozone, 13 000 B.P., kommt aufgrund der hartwasserfehlerfreien Daten in die waldfreie Zwergbirkenphase der FIRBAS-Zone Ia zu liegen. Dadurch ergibt sich eine Diskrepanz von einigen Jahrhunderten zwischen dem Beginn der Bølling Chronozone (BØ) und der Bølling Biozone (Ib/c). Das Ende des Bølling und der Übergang zur Allerød-Chrono- (AL) und Biozone (II), 12 000 B.P., ist mit dem starken Anstieg der *Pinus*-Kurve gegeben. Der Übergang zur Jüngeren Dryas (DR3) Chronozone, 11 000 B.P., wird durch die Tephraschicht des Laacher Vulkans gebildet, währenddem die Biozone der Jüngeren Dryas (III) erst einige Jahrhunderte später, mit der Zunahme der NBP (hauptsächlich *Artemisia*) einsetzt. Aufgrund des $^{14}$C-Plateau um 10 000 B.P. kann das Ende der Jüngeren Dryas (DR3) Chronozone und somit auch der wichtige Übergang zum Holozän keinem biostratigraphischen Ereignis zugeordnet werden.

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## Introduction

The most interesting and dramatic changes in climate and environment of the last 15,000 years took place during the Late-Glacial. Therefore, many Quaternary geologists and palaeoecologists have been attracted by this rather short period, which begins after the retreat of the glacier (timing is different from region to region, probably ca 16,000 yrs B.P.) and ends 10,000 yrs B.P., after the last major climatic cooling. Lacustrine sediments contain different items (e.g. plant and animal remains, carbonates, stable isotopes) which store information about past environmental conditions (e.g. vegetation, climate, human impact) in their drainage basin. The chances of finding undisturbed Late-Glacial sedimentary records in Switzerland are much higher in a small peri-alpine lake basin with a low water renewal time than in the pre-alpine lakes with considerable meltwater influence and river activity.

The development of terrestrial vegetation not only depends on factors such as soil and light but also on close interaction between the prevailing climate (temperature, humidity) and the vegetation. Pollen analysis, in connection with stable isotope analysis of calcareous lake sediments can detect major climatic changes (Eicher & Siegenthaler 1976; Eicher 1987). Because of the synchroneity of the shifts in the isotope record at various sites, the $\delta^{18}O$-curve can be used as a correlation tool between sites with different biostratigraphies.

It has always been very important and desirable in Quaternary studies to have an absolute time scale for correlations of climatic and biotic changes in time and space and for the determination of the palaeoecologically interesting "rates of change" (Watts 1973). An accurate and high-resolution chronology is, however, one of the major problems for the Late-Glacial, since the sediments of this period commonly lack datable material. In a very few cases there are organic sediments (e.g. gyttja) which may be $^{14}C$-dated. However, these $^{14}C$-dates may include a hardwater error of unknown magnitude which can increase the radiocarbon age of the gyttja sample by up to 1,000 years (Andrée et al. 1986). This problem can be circumvented by dating only remains of terrestrial plants (Mielke & Müller 1981; Lister et al. 1984; Andrée et al. 1986; Lister 1988), which reflect the atmospheric $^{14}C$-concentration. But the scarcity of these plant remains in Late-Glacial lake sediments hampers the application of the conventional decay counting method for reasonably sized core segments. By means of the accelerator mass spectrometry (AMS) technique it has become recently possible to date very small samples that contain only a few milligrams of carbon, and thus obtain a very high time resolution (Sutter et al. 1984; Andrée et al. 1986).

This paper presents a step towards a dating of the Late-Glacial climatic and biotic changes and shows some of the many problems related to the establishment of a Late-Glacial chronology.

## Material and methods

The present study is part of a multidisciplinary palaeoecological and palaeolimnological investigation (Küttel & Lotter 1987; Zbinden 1987; Lotter 1988) of Rotsee (47°9'N, 8°20'E, 419 m a.s.l.). This elongated, 2.5 km long lake which is situated at the northern periphery of Lucerne, Central Swiss Plateau, was formed towards the end of

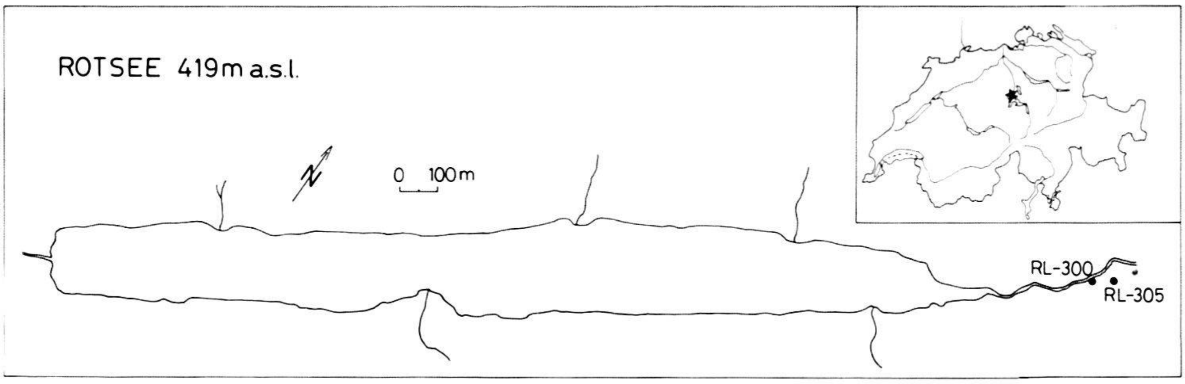


Fig. 1: Location of lake Rotsee and coring sites RL-300 and RL-305, used in this study.

the last glaciation and its sediment yields information on the continuous environmental history of the last 15,000 years.

The cores RL-300 and RL-305 were taken at the northeastern end of the lake (Fig. 1) in an area today overgrown and covered by tall herbs. A modified Livingstone piston sampler (Merkt & Streif 1970) with one meter core segments and a core diameter of 80 mm was used. The lithostratigraphy was described according to Troels-Smith (1955). Sediment analyses were carried out in 5 cm intervals according to Geyh et al. (1971) to determine the amount of organic matter, calcium carbonate, and minerogenic matter (including the CaO part). After sampling for pollen analysis, cores RL-300 and RL-305 were subsampled at intervals of 5 cm and 4 cm, respectively. While core RL-305 was contiguously sampled, only every second sample was analyzed for plant macrofossils in core RL-300. A detailed plant macrofossil analysis was carried out for both cores (Lotter 1988). The plant remains selected for AMS-dating were exclusively identifiable parts of terrestrial vegetation (mainly fruits and catkin scales of birch). $^{14}$C-sample preparation and data evaluation was carried out in Bern following Andrée et al. (1984) and the measurements proper were executed at the AMS facility, ETH Zürich (Bonani et al. 1986). All AMS-dates mentioned in this study refer to Zbinden et al. (in press) and are expressed as conventional $^{14}$C-years before present (B.P.), according to Stuiver & Polach (1977).

Stable isotope analyses ($\delta^{18}O$, $\delta^{13}C$) on lake-marl samples were carried out according to Siegenthaler & Eicher (1986) and the results are expressed as ‰ PDB (Craig 1957).

## Results

Figure 2 illustrates the vegetational history during the Late-Glacial and the earliest part of the Holocene at Rotsee. Detailed descriptions of the local pollen assemblage zones (PAZ) are given by Küttel & Lotter (1987) and by Lotter (1988). The regional biozonation follows Firbas (1948, 1954) and is marked by Roman numerals (Ia to V). The chronozones according to Mangerud et al. (1974) are abbreviated by DR1 (Oldest Dryas), BØ s.l. (Bølling), DR3 (Younger Dryas) and PB (Preboreal). Owing to

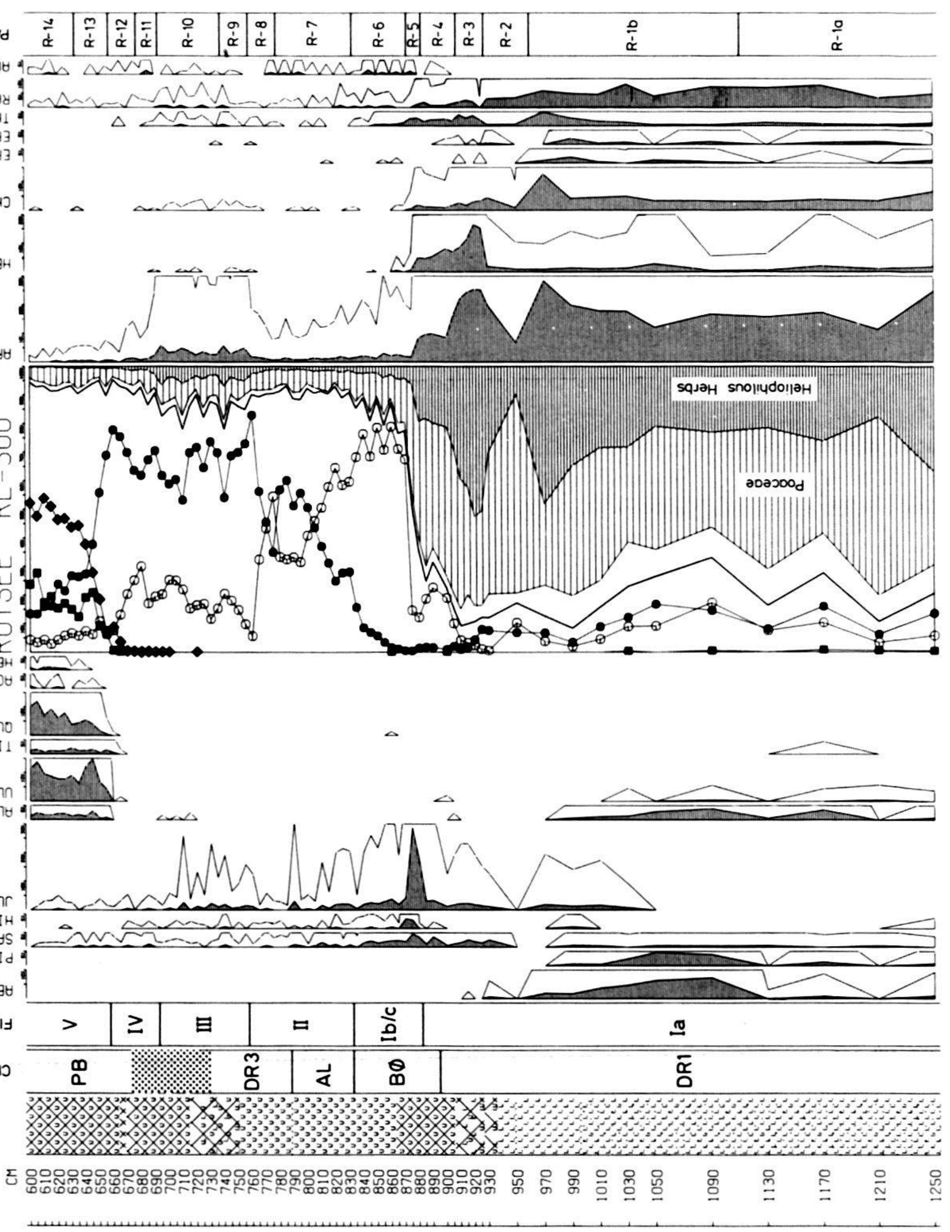


Fig. 2: Pollen diagram of core Rotsee RL-300, focusing on the development of several key taxa during the Late-Glacial and early Holocene.

the lack of evidence, both in time and climatic record (pollen, $\delta^{18}O$), we omit the Older Dryas chrono- (DR2) and biozone (Ic).

Pollen analysis at Rotsee revealed the main features of the Late-Glacial vegetational succession ("steppe tundra" → "shrub tundra" → juniper scrub phase → tree birch phase → pine – birch phase → hazel and mixed oak forest phase), which has already been outlined for many sites on the Swiss Plateau (cf. Rösch 1983; Ammann & Tobolski 1983; Gaillard 1984; Zoller 1987). We shall therefore proceed without describing these classical sequences in detail once more.

The results of the sediment analyses are illustrated in Figure 3 and corroborate the visual sediment description according to Troels-Smith (1955). The lithology of both cores is very similar: the lowermost 2–3 m consist of sandy glacial clay, which gradually changes first to a (calcareous) clay-gyttja and then to a calcareous gyttja. The carbonate

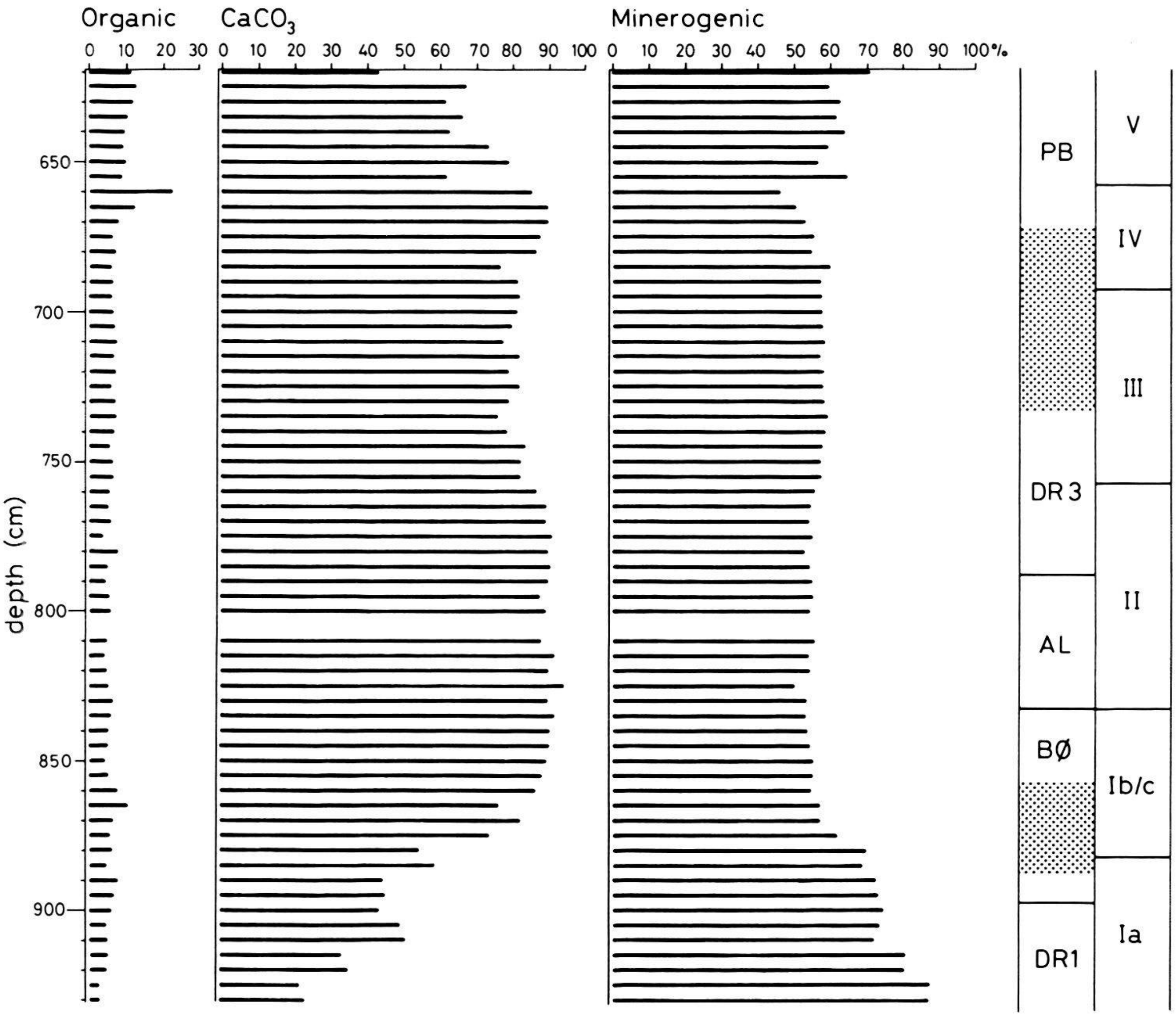


Fig. 3: Composition of the Late-Glacial sediment of Rotsee core RL-300. The values are expressed in % of dry sediment weight. The shaded areas in the chronozone column highlight the plateau phases in $^{14}C$-ages.

content of the sediment increases towards the end of the Oldest Dryas (Ia) and reaches up to 90% of dry weight during the Bølling (Ib/c) and Allerød (II), whereas the minerogenic part of the sediment is leveled down to between 50–60% of dry weight during Bølling (Ib/c). The Younger Dryas (III) is characterized by a slight decrease in $CaCO_3$ portion and minor increase in minerogenic sedimentary matter.

Oxygen isotope analysis at Rotsee was carried out on three different cores (Lotter 1988). Figure 4 illustrates the $\delta^{18}O$-record of core RL-300. Each core showed the three characteristic Late-Glacial shifts at the transitions from the biozones Oldest Dryas (Ia) to Bølling (Ib/c), Allerød (II) to Younger Dryas (III) and from Younger Dryas (III) to Preboreal (IV). The climatic implications of these shifts have already been discussed in detail by Eicher & Siegenthaler (1976), Eicher (1979, 1987) and others.

Figure 4 also shows radiocarbon ages plotted against core-depth. The error bars signify, according to Stuiver & Polach (1977), the measurement accuracy, to one

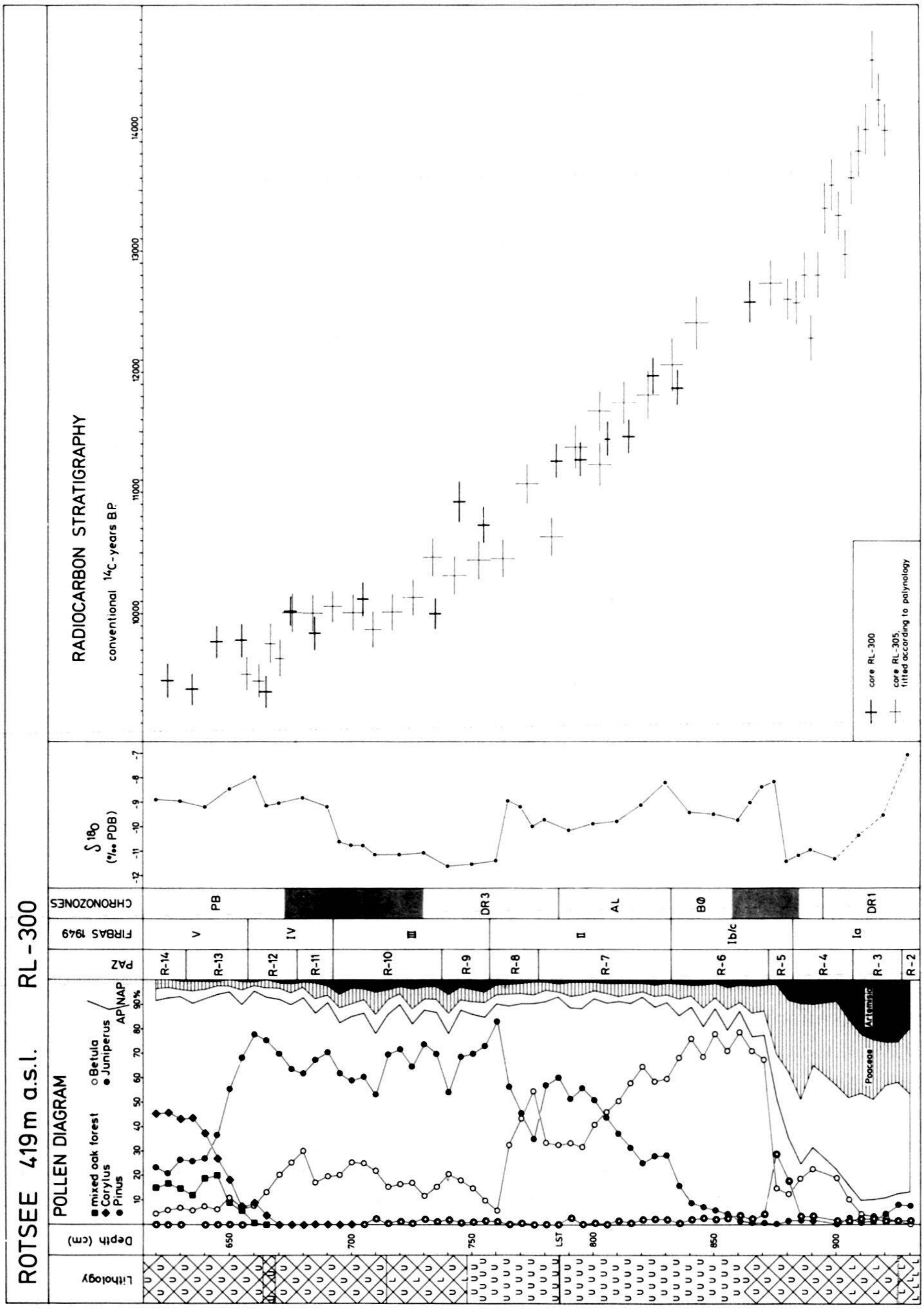


Fig. 4: Palyno-, isotope-, and radiocarbon stratigraphy of core Rotsee RL-300. The shaded areas highlight the plateau phases in $^{14}C$-ages.

standard deviation, involving the statistical errors of sample and standard measurements, and the long-time stability of the background, as well as the error of the estimated $\delta^{13}C$ value of plant material. Another error must be considered which originates in the distribution of the single plant macrofossils in the core segment used as sample. The actual depth of these plant remains in the core segment sampled cannot be determined precisely. This uncertainty in depth corresponds to an uncertainty in age within the range of 100 years, depending on the core segment size and the sediment accumulation rate. In order to exclude major errors due to sediment disturbance we measured two different cores. The results of the contiguously sampled core RL-305 were fitted into the depth scale of core RL-300 by correlation of palyno-, litho- and isotope-stratigraphy of the two cores (LOTTER 1988).

The oldest AMS-dates are situated in PAZ R-3 and give an age of approximately 14,500 yrs B.P. They decrease until, at the beginning of R-5 (Ib/c), they reach a zone of constant age ranging from 12,800 to 12,600 yrs B.P. A similar phase of constant $^{14}C$-age during the Bølling (Ib/c) biozone has been reported by ANDRÉE et al. (1986) from Lobsigensee. During the second part of the Bølling (Ib/c) the $^{14}C$-ages decrease again until the beginning of R-10 (III) where they remain at 10,000 yrs B.P. for several samples, including the second part of the Younger Dryas (III) and the fist part of the Preboreal (IV). This younger phase of constant age can also be observed at Lobsigensee (ANDRÉE et al. 1986) and in the peat bogs of Wachseldorn and Zugerberg (OESCHGER et al. 1980; WELTEN 1982), where lake level fluctuations, bioturbation and sediment reworking can be excluded.

## Discussion

### *$^{14}C$-Plateaux*

When compared with previously published $^{14}C$-stratigraphies the results of ZBINDEN et al. (in press) represent extreme dating resolution and high accuracy. If we examine the shape of the age-depth relationship in Figure 4, two extremely vertical steps, so called "plateaux" of constant $^{14}C$-age, are striking: the first occurs during the Bølling (Ib/c) and the second during the Younger Dryas (III) and Preboreal (IV). Possible explanations for this phenomenon include either unusually high sediment accumulation rates or a decrease in the atmospheric $^{14}C$-content. It seems, however, unlikely that major changes such as the reforestation in Bølling or the disappearance of heliophilous plants at the end of the Younger Dryas occured in such a short time as suggested by the radiocarbon ages. There are distinct changes both in vegetation and in climate ($\delta^{18}O$) to be observed during these plateau phases. We concluded that the atmospheric $^{14}C$-concentration decreased considerably, probably due to changes in the global carbon cycle (ZBINDEN et al. in press).

What are the implications of such plateaux of constant radiocarbon age? All events dated within the two plateaux seem to be synchronous, but, in fact, temporal differences may be considerable. Our radiocarbon stratigraphy illustrates that $^{14}C$-ages in the range of 12,800 to 12,600 and around 10,000 yrs B.P. must be interpreted cautiously. Radiocarbon ages across these plateau periods cannot be unconditionally interpreted as synchronous. On the one hand this impedes a comparison among different coring sites (Lotter 1988; Ammann & Lotter in press) and therefore a correlation in time and space. On the other hand these plateaux hamper or even prevent the calculation of (sediment, pollen, macrofossil) accumulation rates (flux) which would contribute to our understanding of precisely such periods of high vegetation dynamics as the Bølling and the transition Younger Dryas/Preboreal.

The only way to resolve these plateaux problems will be a detailed study over the appropriate time interval using archives that provide a $^{14}$C-independent, high-resolution relative time control (e.g. annually laminated sediments).

*Chronology*

For Switzerland, Welten (1982) adapted the Mangerud et al. (1974) chronozonation and coupled it with the Firbas (1949, 1954) pollen zonation, thus producing a mixed chrono- and biozonation. Because of the dating problems mentioned above (hardwater error, $^{14}$C-plateaux) a discussion of the Late-Glacial chronozonation is required. On the basis of our high-resolution radiocarbon stratigraphy we propose a new correlation between chrono- and biostratigraphy. We continue to use the Firbas-pollen zones but strictly as a biostratigraphy. Furthermore, we have adapted the chronozone-system of Mangerud et al. (1974) although we are well aware of the confusion already created by using the same zone-names in both systems. This implies that we must always specify whether we are referring to the biological or the chronological content of any zone we mention.

Hitherto the transition from the Oldest Dryas (DR1) to the Bølling (BØ) chronozone (13,000 yrs B.P.) was attributed to the rise or peak of the juniper pollen values. This event, which also corresponds to the transition of Ia to Ib/c in the Firbas-system, is usually dated on gyttja-samples to between 13,500 and 13,000 yrs B.P. (Welten 1982; Rösch 1983; Gaillard 1984; Lotter 1985). But these dates are unsatisfactory because they are too old, most probably owing to a hardwater error (Lotter 1988; Ammann & Lotter in press). Based on our AMS-dates on terrestrial vegetation we locate the beginning of the Bølling (BØ) chronozone for the Swiss Plateau during the maximum development of the dwarf-birch phase (PAZ R-4) of the Oldest Dryas (Ia) biozone. There is, however, no other distinct event in the palynostratigraphy that could help to correlate this chronozone boundary from one Swiss Plateau site to another. The juniper phase of the early Bølling has an age of between 12,700 and 12,400 yrs B.P., but unfortunately these dates are located within the first plateau of the $^{14}$C-ages. Furthermore there is a discrepancy of several centuries between the beginning of the Bølling (BØ) chronozone and the Bølling (Ib/c) biozone.

According to Welten (1982) the transition from the Bølling (BØ) to the Allerød (AL) chronozone at 12,000 yrs B.P. coincides with the first major *Pinus*-increase which also marks the beginning of the Allerød (II) biozone on the Swiss Plateau. Our dates confirm this correlation of chrono- and biozonation very well and we therefore retain this event as a marker for the beginning of the Allerød (AL) chronozone.

The transition from the Allerød (AL) to the Younger Dryas (DR3) chronozone (11,000 yrs B.P.) is given by the occurrence of the Laacher See Tephra (LST), an ash-layer from one of the latest eruptions of the Laacher See volcano in Germany. It has been dated at 11,000 yrs B.P. (Wegmüller & Welten 1973; Rösch 1983; van den

Bogaard 1983). Although Welten (1982) was well aware of the age attributed to this excellent stratigraphic time marker he considered the increasing NAP-values as the onset-index for the Younger Dryas (DR3). We propose fixing the beginning of the Younger Dryas (DR3) chronozone at the LST-level, whereas the Younger Dryas (III) biozone starts with the increasing values of NAP (mainly *Artemisia*). If the LST cannot be detected visually or by sediment analysis it is possible in many cases to locate the end of the Allerød (AL) close to a minor increase in the *Betula*-values (Fig. 2, see also Rösch 1983; Lotter 1988). Again there is a discrepancy of 200–300 years between the beginning of the Younger Dryas chrono- (DR3) and biozone (III).

Due to the second plateau of the $^{14}C$-ages, it is very difficult to position the important boundary between the Younger Dryas (DR3) and the Preboreal (PB) chronozone (10,000 yrs B.P.). This age also forms the boundary between the Late-Glacial and the Holocene. In core RL-300 the second plateau encompasses about 50 cm of sediment in which the end of the Younger Dryas (III) biozone, marked by the decrease of the NAP-values occurs. Given the spread of available dates, it is not possible to correlate this boundary with a specific biostratigraphic event.

### *Palaeoclimate*

The lowermost clayey samples of core RL-300 derive from the influence of the meltwater of the Reuss glacier. Palynostratigraphically this includes the PAZ R-1 and R-2, which contain considerable reworked pollen (*Ulmus, Abies, Picea, Alnus*, trilete spores). The vegetation is typical for the cryocratic stage (Iversen 1958): a poor "steppe-tundra" with many grasses and sedges indicating a treeless, open vegetation. The $\delta^{18}O$-values of carbonates in PAZ R-1 to R-3 merely reflect the isotopic composition of allochthonous detrital carbonate minerals derived from the geology of the drainage basin.

At the end of R-2 sediment lithology changes from clayey silt to calcareous clay-gyttja and documents the end of meltwater influences on the lake as well as the onset of biogenically-induced carbonate sedimentation. During this biozone, submerged *Chara* vegetation spread in the lake, encouraged by increased water transparency. Another sediment transition from clay-gyttja to calcareous gyttja at the end of R-3 indicates decreasing soil erosion due to the more effective vegetation cover by plant species characteristic of the protocratic stage (Iversen 1958): heliophilous pioneer plants (*Artemisia, Helianthemum, Saxifraga oppositifolia, Ephedra*, Chenopodiaceae) and shrubs (*Salix, Juniperus, Betula nana*). The $\delta^{18}O$-record of the bulk samples of autochthonous carbonate for this period is consistent with cold temperatures. Investigations on fossil insect remains at Lobsigensee suggest mean July temperatures of 10–12 °C (Elias & Wilkinson 1983) for the dwarf birch phase of Ia. Despite numerous analyses (Emiliani 1966; Dansgaard & Tauber 1969; Shackleton & Opdike 1973) oxygen isotope ratios of carbonates cannot be directly interpreted in terms of absolute temperatures. The first increase of 3‰ in the $\delta^{18}O$-curve coincides with the increase in *Juniperus* and *Hippophaë* at the beginning of the Bølling (Ib/c) biozone. Both, isotope ratios and vegetation development suggest a change in temperature towards a warmer climate: the minimum July temperatures for *Juniperus* and *Hippophaë* are above 10 °C, possibly not lower than 11–12 °C (Kolstrup 1980).

This climatic improvement allowed the onset of reforestation initiated by tree birch and then followed by pine, both fast spreading and fast growing trees requiring abundant light. The fossil insect remains at Lobsigensee indicate mean July temperatures in a range from 14–16 °C (ELIAS & WILKINSON 1983) and plant macrofossils of *Chara tomentosa* and *Najas marina* from Rotsee (LOTTER 1988) suggest mean July temperatures of 15 °C or more (LANG 1981) for the Bølling (Ib/c) and Allerød (II) period. During the first part of the mesocratic stage (IVERSEN 1958) pioneer trees such as birch and pine produce a fertile humus layer that is needed for the establishment of the slow spreading and slow reproducing climax trees.

At the transition from the Allerød (II) to the Younger Dryas (III) biozone the $\delta^{18}O$-values shift 2‰ lower, accompanied by a regressive development of the vegetation. Due to this climatic deterioration, the pine-birch forest became less dense thus promoting an expansion of heliophilous herbs (*Artemisia*, Chenopodiaceae, *Thalictrum*) and shrubs (*Juniperus*). This open vegetation led to an increased soil erosion, which can be detected as an increase in the clastic mineral input (Fig. 3).

At the end of the Younger Dryas (III) biozone the $\delta^{18}O$-values increase again. This correlates well with the decreasing values of heliophilous plants, suggesting that the pine-birch forests became denser again. Furthermore the first thermophilous, shade tolerant trees (*Corylus, Ulmus, Quercus, Tilia, Alnus*) migrated into the area during the Preboreal (IV) and spread during the Boreal (V) biozone. This event represents the onset of the second part of the mesocratic stage which is characterized by the displacement of the pioneer trees by more shade tolerant climax trees. The oxygen-isotope ratios remain at a level of −9‰ throughout the Holocene and therefore give no further evidence of any major climatic changes. The occurrence of *Hedera*, which appears in the pollen record during the Boreal (V), indicates a minimum average January temperature of not lower than −1.5 °C (IVERSEN 1944).

Until recently the first shift in $\delta^{18}O$, which is commonly synchronous with the increase in juniper (EICHER 1979, 1987), has been dated between 13,400 and 13,000 yrs B.P. by pollen analytical correlation with $^{14}C$-dated sediment cores. Because of the first $^{14}C$-plateau it is difficult to give a precise age for this event in the Rotsee sediment record. However, we can state that the first significant oxygen isotope shift during the Late-Glacial is definitely younger than 13,000 yrs B.P., and most probably occurred around 12,600 yrs B.P. The results of LISTER (1988) from Zürichsee obtained on ostracod and mollusc oxygen isotope ratios, AMS-dating the first $\delta^{18}O$-shift between 12,800 ± 250 and 12,400 ± 250 yrs B.P. thus corroborate perfectly.

The second oxygen-isotope shift at the end of the Allerød (II) signifies a cooling and is dated between 10,900 and 10,600 yrs B.P. The climatic warming at the end of the Younger Dryas (III) biozone, indicated by increasing $\delta^{18}O$-values, occurs around 10,000 yrs B.P. within the second plateau of constant $^{14}C$-age.

It is interesting to note that there is a coincidence between periods of climatic warming and phases showing constant $^{14}C$-ages, i.e. each increase in the $\delta^{18}O$-record occurs within a plateau of constant $^{14}C$-age, or vice versa.

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