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Late-glacial climatic oscillations as recorded in Swiss lake sediments

A. F. LOTTER EAWAG, Swiss Federal Institute for Water Resources and Water Pollution Control, CH-8600 Dübendorf, Switzerland; and Geobotanical Institute, University of Bern, Switzerland
U. EICHER and **U. SIEGENTHALER** Physikalisches Institut, University of Bern, Sidlerstrasse 5, CH-3012 Bern, Switzerland
H. J. B. BIRKS Botanical Institut, University of Bergen, Allégatan 41, N-5007 Bergen, Norway; and Environmental Change Research Centre, Dept. of Geography, University College, London, England

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ABSTRACT: Regional pollen assemblage zones for the late-glacial period of the Swiss Plateau are introduced and defined. They include four major zones (*Artemisia*, *Juniperus–Hippophaë*, *Betula*, *Pinus* PAZ) with several subzones. Pollen and oxygen-isotope analyses on lacustrine sediments from several lakes in the area reveal four distinct phases of climatic oscillation in the time period of 13 000–9500 yr BP.

The first oscillation, termed the Aegelsee fluctuation, occurs shortly before 12 000 yr BP and varve counts suggest its duration was ca. 100 yr. It is characterised by a short decrease in the oxygen isotopes as well as a short increase in NAP associated with a depression in birch pollen values. The second oscillation, which occurs in the $\delta^{18}O$ record shortly before the deposition of the Laacher See Tephra (ca. 11 000 yr BP), is termed the Gerzensee fluctuation. It occurs during a pine-dominated phase and its vegetational effects cannot be determined palynologically. The most prominent regressive phase is the Younger Dryas biozone (ca. 10 700–10 000 yr BP) characterised by an increase in heliophilous NAP and low $\delta^{18}O$ values. The Younger Dryas biozone can often be subdivided palynologically into two parts: a first part rich in grasses and juniper and a second part with higher *Filipendula* and birch values. During the Preboreal biozone another distinct oscillation is evidenced only in the oxygen isotope ratios.

Comparison of the Swiss oxygen isotope profiles with the Greenland Dye 3 record suggests that not only the three major shifts in the $\delta^{18}O$ curves but also the minor ones are closely comparable, suggesting some common climatic control.

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KEYWORDS: Pollen, oxygen isotopes, Younger Dryas, palaeoclimate, Switzerland.

Introduction

Climatic change in connection with global warming is currently a major issue. There is an urgent need to understand more about the mechanisms controlling climate and its dynamics. It is a major challenge for scientists to predict biotic reactions to changes in temperature and/or precipitation. However, it is equally important to know more about past climatic changes in order to have a better understanding of the dynamics and amplitude of the climate system. This information may be obtained from historical sources (e.g. Pfister, 1985), as well as from proxy data archives, such as tree rings (e.g. Hughes et al., 1982), ice-cores (Dansgaard et al., 1982), or marine and lacustrine sediments (e.g. Imbrie and Kipp, 1971; Coope, 1977). By its incorporation and integration of a variety of biotic and abiotic parameters, the sedimentary record contains some direct and indirect information about past climates.

The most important broad-scale climatic oscillations since the last ice-age occurred during the late-glacial period and the most prominent period of climatic change of the last 13 000 yr is certainly the Younger Dryas. Its abrupt climatic cooling is associated with the southward movement of the polar front in the North Atlantic (Ruddiman et al., 1977). The Younger Dryas has not only been evidenced in the marine stable-isotope record (Shackleton and Opdyke, 1976) and in the Greenland ice-cores (Dansgaard, 1987) but also in the pollen and fossil insect records of many lacustrine sediments in northwestern Europe (e.g. Pennington, 1977; Lemdahl, 1988) as well as in central Europe and the Alps (e.g. Welten, 1982; Bortenschlager, 1984; Ammann et al., 1985; Pons et al., 1987). New evidence indicates an amphi-Atlantic or even a global occurrence of its impact (e.g. Mott et al., 1986; Peteet, 1987; Wright, 1989; Kundrass et al., 1991; Walker et al., 1991). Besides the prominent Younger Dryas oscillation there are also indications for smaller late-glacial climatic fluctuations, for example the Older Dryas.

In many regions the number of palynological investigations provides a dense geographical coverage of pollen profiles.

This is especially true for northwestern Europe (see e.g. Huntley and Birks, 1983) and for the Alps (Lang, 1985a; Schneider, 1985a). On the Swiss Plateau within an area of ca. 11 000 km^2 between 400 and 1000m elevation there are today more than 60 pollen profiles available that include the late-glacial period; some of these cores have also been analysed for stable istopes (e.g. Eicher, 1987).

In the light of the high geographical density of pollen and oxygen-isotope stratigraphies, we take a closer look at the time-window 13 000–10 000 yr BP in an attempt to detect any consistent fine-scale patterns of past climatic change.

Investigated area and sites

The area investigated corresponds to the central part of the Swiss Plateau and the southern edge of the calcareous Prealps (see Lang, 1985b). It is mainly characterised by glacial landforms and all the sites mentioned in this study were ice-covered during the last glaciation. Depending on elevation the climate is characterised by a mean annual precipitation of between 1000 and 1400mm. The mean January temperature ranges between −5 and 0°C, whereas the mean July temperature is between 10 and 18°C. The natural vegetation of the region would consist of beech and oak forests at elevations between 400 to 800m and of fir–beech forests in the montane belt. It has, however, been largely replaced by intensive agriculture and pasturing, or by spruce plantations.

Aegelsee (7°32′30″E, 46°39′N, Fig. 1) is situated at 995m a.s.l. in the montane belt of the northern calcareous Prealps. This formerly small dystrophic lake was enlarged in 1955 with the construction of a hydroelectric power plant. As a consequence, large parts of the natural lake basin as well as parts of the adjacent mire were destroyed (Lotter and Fischer, 1991). The site was first investigated palynologically by Welten (1952). For an intensive reinvestigation of the mire (Wegmüller and Lotter 1990), several cores were taken along two transects with a modified Livingstone corer (Merkt and Streif, 1970). The results of cores AE-1 and AE-3 (Figs 2 and 3) are discussed in this study. Their lithology is very comparable: the oldest sediments consist of glacial clays that eventually change into a calcareous clay–gyttja and then gradually into a biogenically precipitated lake-marl. A short episode of increased minerogenic input within this marl can be observed. The Laacher See Tephra (LST) occurs as a grey layer 1 mm thick within the marl matrix (AE-1: 702.3 cm; AE-3; 801.5 cm). After this isochronous marker horizon the sediment changes into a clayey marl 20–30 cm thick. Thereafter, the sediment eventually changes again into a homogeneous lake-marl with a gradually increasing organic content.

Faulenseemoos (7°41′30″E, 46°40′45″N, Fig. 1) lies at 590 m a.s.l., close to the transition from the Swiss Plateau to the Alps. It is an ancient lake, about 700 m long and 200 m wide. Welten (1944) carried out his classical studies on geochronology and vegetation history using the laminated sediments at this site. The core presented here was taken in a marginal part of the ancient basin with a Hiller sampler (Eicher and Siegenthaler, 1976; Eicher, 1979; Welten, 1982; Fig. 4). Its lithology consists, in the deepest part, of a calcareous clay that eventually changes into a homogeneous lake-marl, with the LST at a depth of 433 cm.

Gerzensee (7°33′E, 46°50′N, Fig. 1) is situated at an altitude of 605 m a.s.l. on the Swiss Plateau. The lake is 1100 m long and 300 m wide with a surface area of 0.27 km^2. Its catchment drains 2.6 km^2. Several cores have been analysed

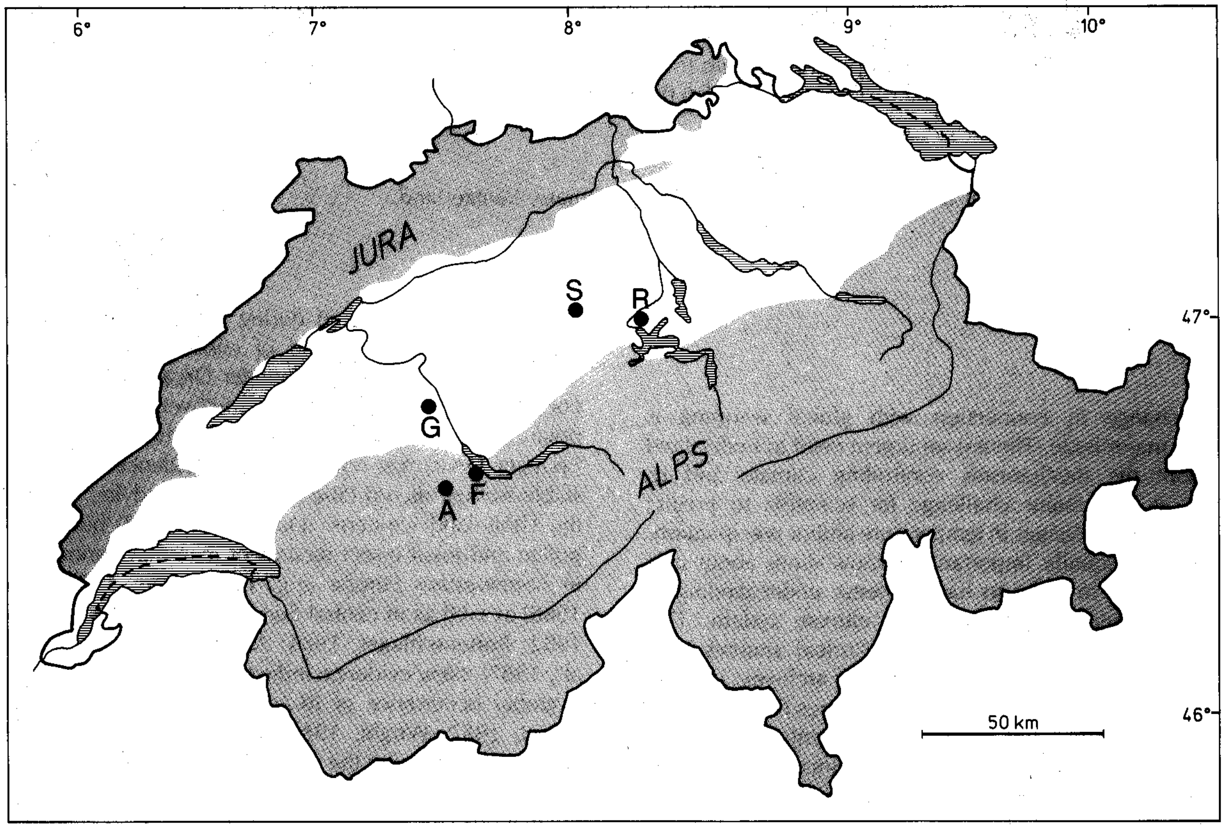

Figure 1 Map of Switzerland showing the location of the investigated sites. A, Aegelsee; F, Faulenseemoos; G, Gerzensee; R, Rotsee; S, Soppensee.

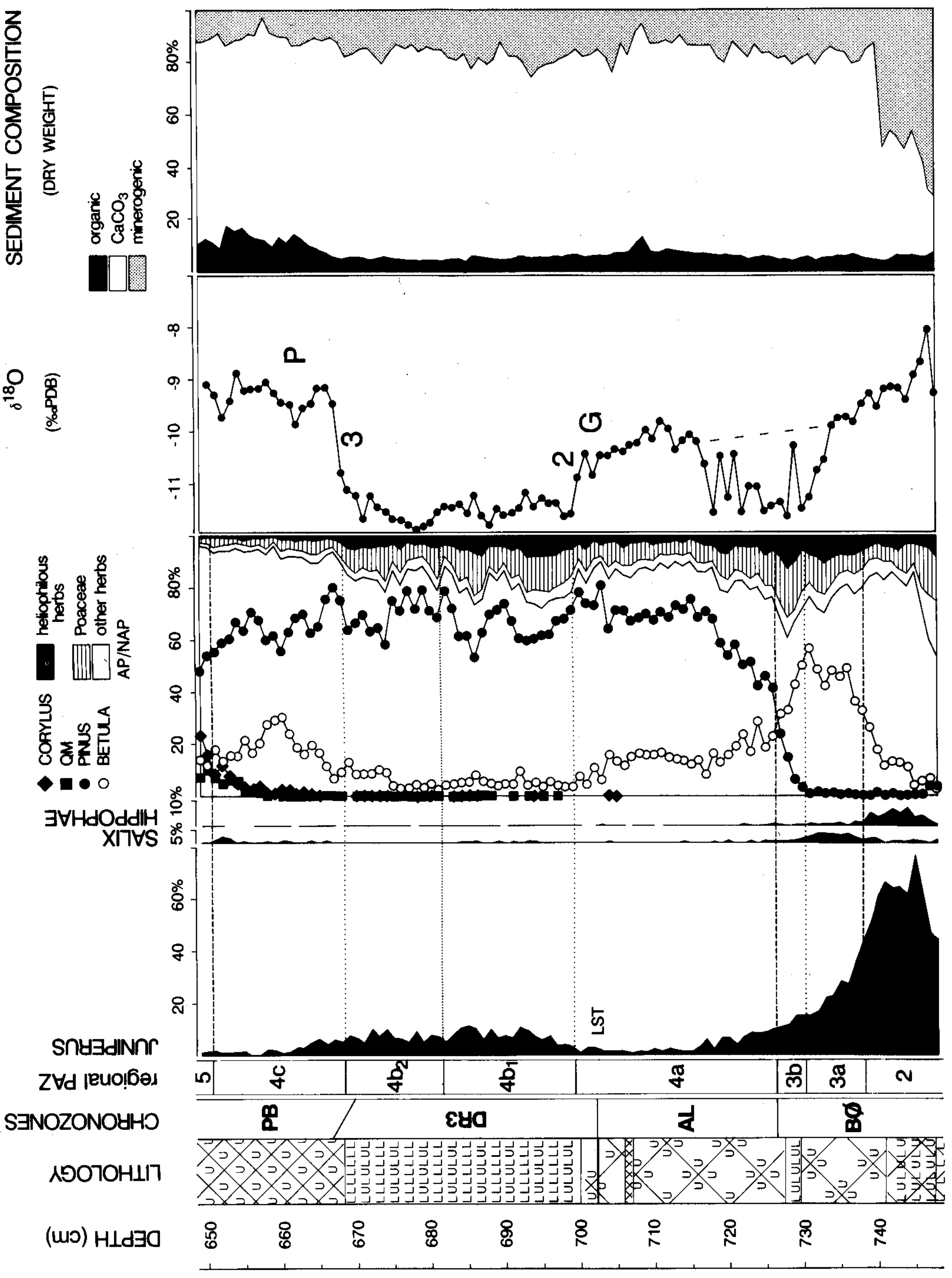

Figure 2 Pollen, oxygen-isotope stratigraphy, and sediment composition of Aegelsee core AE-1 (after Wegmüller and Lotter 1990).

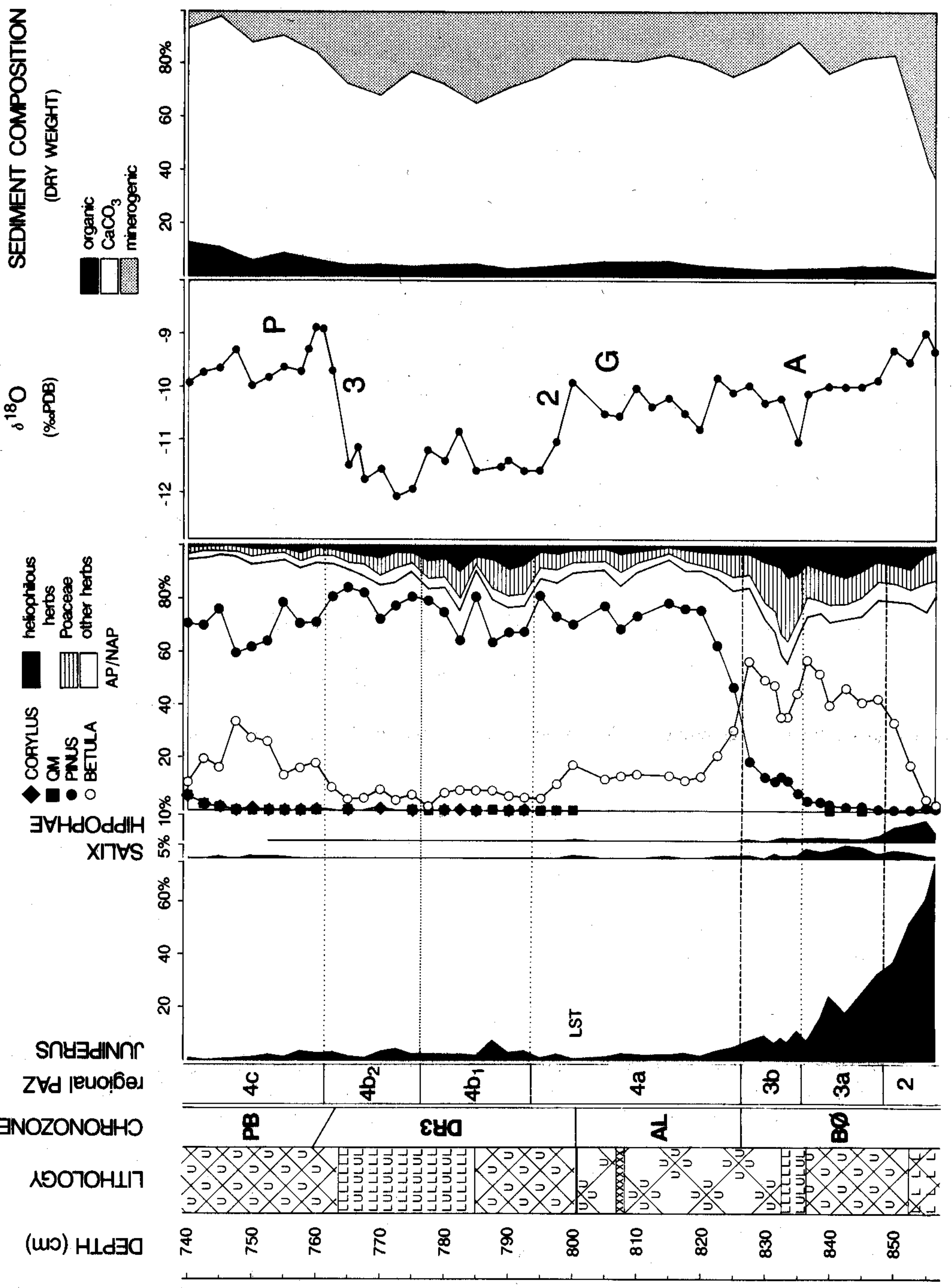

Figure 3 Pollen, oxygen-isotope stratigraphy, and sediment composition of Aegelsee core AE-3 (after Wegmüller and Lotter 1990).

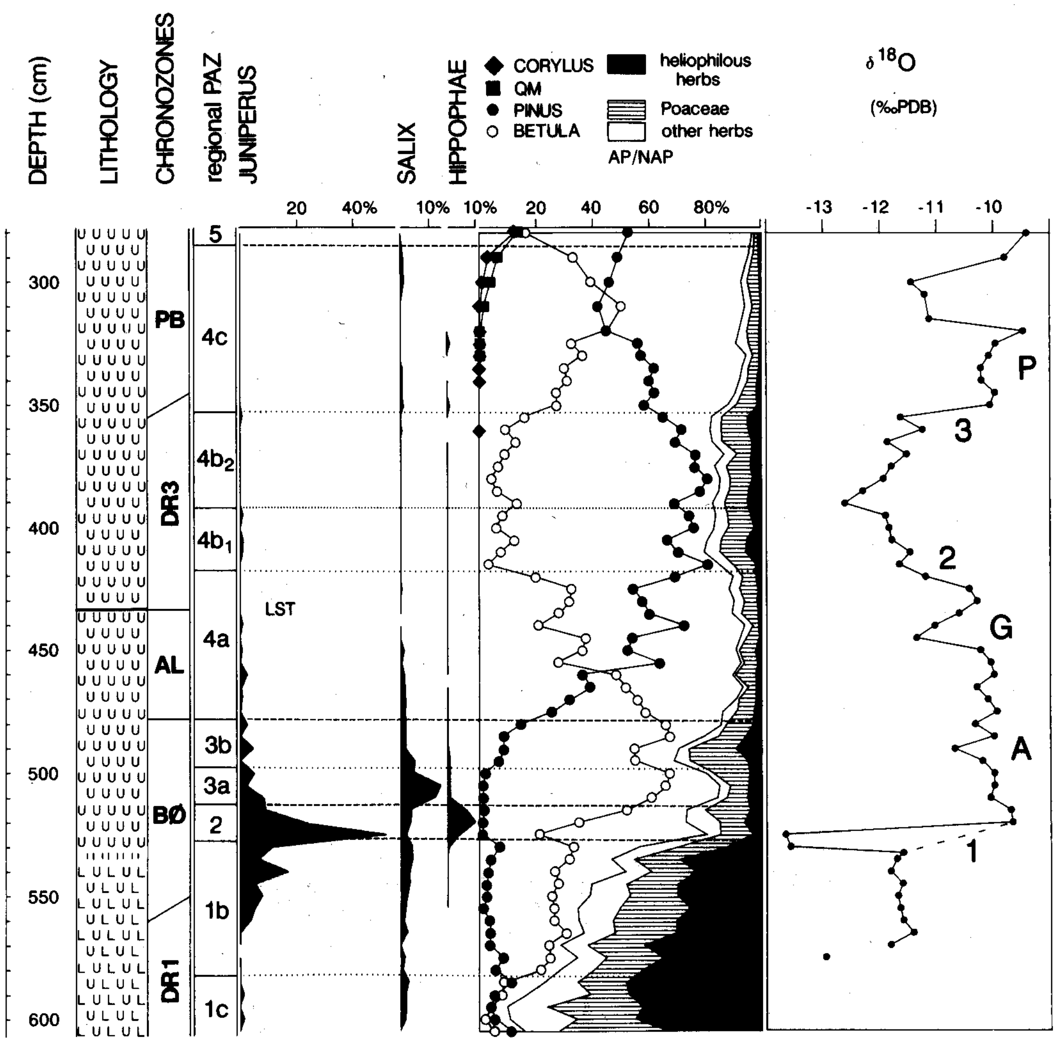

Figure 4 Pollen and oxygen-isotope stratigraphy of Faulenseemoos (after Eicher and Siegenthaler 1976).

for pollen and stable istopes (Eicher and Siegenthaler, 1976; Eicher, 1979, 1980). Core G-III (Fig. 5) was taken with a modified Livingstone corer on the eastern shore of the lake. The lithology of this core consists in its lowermost part of a clayey marl and changes into a homogeneous lake-marl. After the LST, which is present as a 5 mm layer at a depth of 227.5 cm the sediment changes again into a clayey marl for ca. 60 cm.

Rotsee (8°20'E, 47°9'N, Fig. 1) lies at 420 m a.s.l. on the Swiss Plateau. This lake is today 2.5 km long and 200 m wide. It has a surface area of 0.46 km^2 and its catchment drains 4.6 km^2. For palaeoecological and palaeolimnological studies of this lake (Lotter, 1988), several cores were taken on a transect with a modified Livingstone corer. Core RL-250 (Fig. 6) was taken at the southeast end of the lake in a water depth of 4.5 m. The base of the core consists of several metres of clay, which eventually changes into a calcareous clay-gyttja and then into a lake-marl. The LST is present at a depth of 776.1 cm.

Soppensee (8°5'E, 47°5'30"N, Fig. 1) is situated at 595 m a.s.l. on the Swiss Plateau. It is 800 m long, 400 m wide, and has a surface area of 0.227 km^2. It's catchment drains 1.6 km^2. Several cores were taken with a modified Kullenberg sampler (Kelts et al., 1986). Core SO86–14 (Fig. 7) was taken in the deepest part of the basin at a water depth of 27 m. Its basal part consists of silty clay gradually changing into a calcareous clay-gyttja and then into a laminated fine-detritus gyttja (Lotter, 1989). The LST is present at 481.5 cm. The laminated gyttja is interrupted by two distinct homogeneous layers containing coarse littoral carbonates.

Material and methods

Sample preparation for pollen analysis was carried out as described in Lotter (1988). A minimum of 500 AP (arboreal pollen) grains were counted for each level analysed. Spores and pollen of aquatics and of local mire vegetation were excluded from the calculation sum (100% = ΣAP + ΣNAP). The pollen diagrams have been subdivided into regional pollen assemblage zones (see Figs 2–7), whereas the chronozonation follows Ammann and Lotter (1989).

Organic matter and carbonate content were estimated by combustion at 525° and 900°C, respectively, and expressed as percentage of dry sediment weight.

Stable isotope ratios were determined using standard techniques (Siegenthaler and Eicher, 1986). CO_2 was obtained by treating the bulk carbonate (typically 40–100 mg per sample) at 50°C with 95% orthophosphoric acid for 60 min. The CO_2 was then analysed mass-spectrometrically for its $^{13}C/^{12}C$ and $^{18}O/^{16}O$ ratios. $\delta^{18}O$ values are expressed with respect to the international standad PDB.

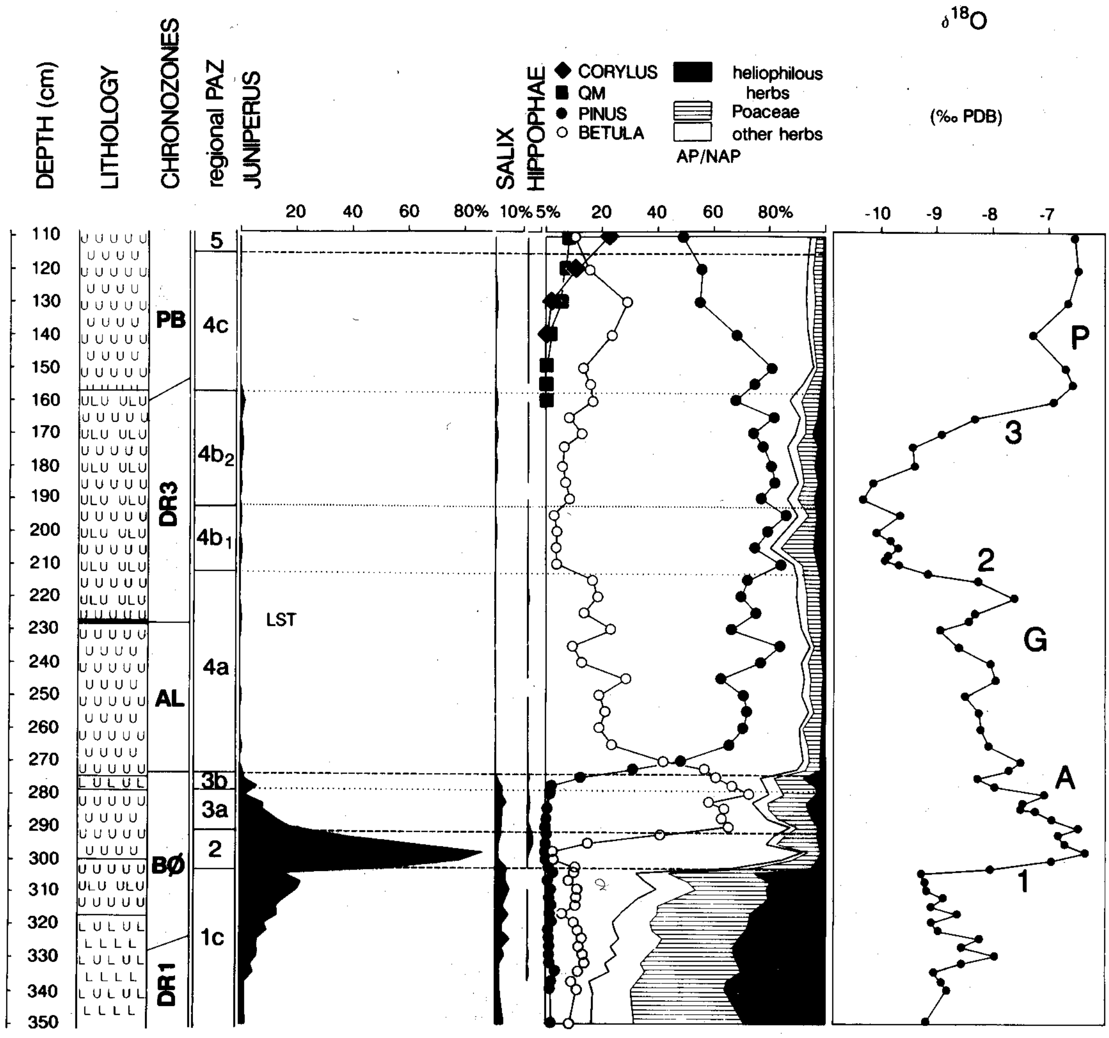

Figure 5 Pollen and oxygen-isotope stratigraphy of Gerzensee core G-III (after Eicher and Siegenthaler 1976).

All ages or radiocarbon dates are expressed in conventional radiocarbon years BP (before 1950, according to Stuiver and Polach, 1977).

Multivariate numerical analyses of the pollen-stratigraphical and the oxygen-isotope data from Aegelsee (AE-1, AE-3), Faulenseemoos, Gerzensee, and Rotsee were performed in an attempt to answer the following questions:

1. Are there any statistically significant divisions (zones) within the pollen-stratigraphical data deposited during the Younger Dryas?
2. Is there a statistically significant relationship between the pollen stratigraphy and the stable oxygen-isotope record for the late-glacial period?
3. How do the stable oxygen-isotope stratigraphies at the different sequences correlate between themselves and with the Dye 3 $\delta^{18}O$ profile from the Greenland ice sheet?

For question 1 there are several numerical techniques for partitioning stratigraphical sequences into smaller units or zones (e.g. Birks and Gordon, 1985; Grimm, 1987). Although they will all propose partitions, there is no means of testing the statistical validity of the suggested partitions (Birks, 1987). One approach that we have not implemented is chronological clustering (Legendre et al., 1985; Legendre, 1987) in which the statistical significance of each division is evaluated by randomisation tests (for examples see Legendre et al., 1984; Bell and Legendre, 1987; Galzin and Legendre 1987). The second approach, and the one we have adopted, involves Walker and Wilson's (1978) sequence-splitting, with its approximate statistical testing of the significance of splits derived by simulation. Sequence splitting was originally developed for partitioning sequences of pollen influx of individual taxa. Such sequences are not constrained by percentage calculations and thus the independent behaviour of different taxa could be investigated. In this study we do not have pollen influx data. To derive statistically independent 'sequences', we therefore transformed the pollen percentages (after square-root transformation) to their principal components. We then applied sequence-splitting to each principal component in turn. Only pollen spectra from the Younger Dryas biozone, plus immediately adjacent spectra, were transformed to principal components. As expected, statistically significant splits were confined to the first one or two principal components at each sequence, as these components represent up to 62% of the total variance in the pollen data (see Table 1). Significant splits were identified by calculating the likelihood ratio Λ_t for the split at level$_t$ with a χ^2 distribution at a significance level of $0.05/n$, where n is the number of samples. This is a stringent significance level and is the probability of rejecting the null hypothesis of no splits when it is in fact true (Birks and Gordon, 1985, p. 213).

Principal components analysis using a covariance matrix between variables (after square-root transformation) was implemented by means of CANOCO 3.11 (ter Braak, 1990).

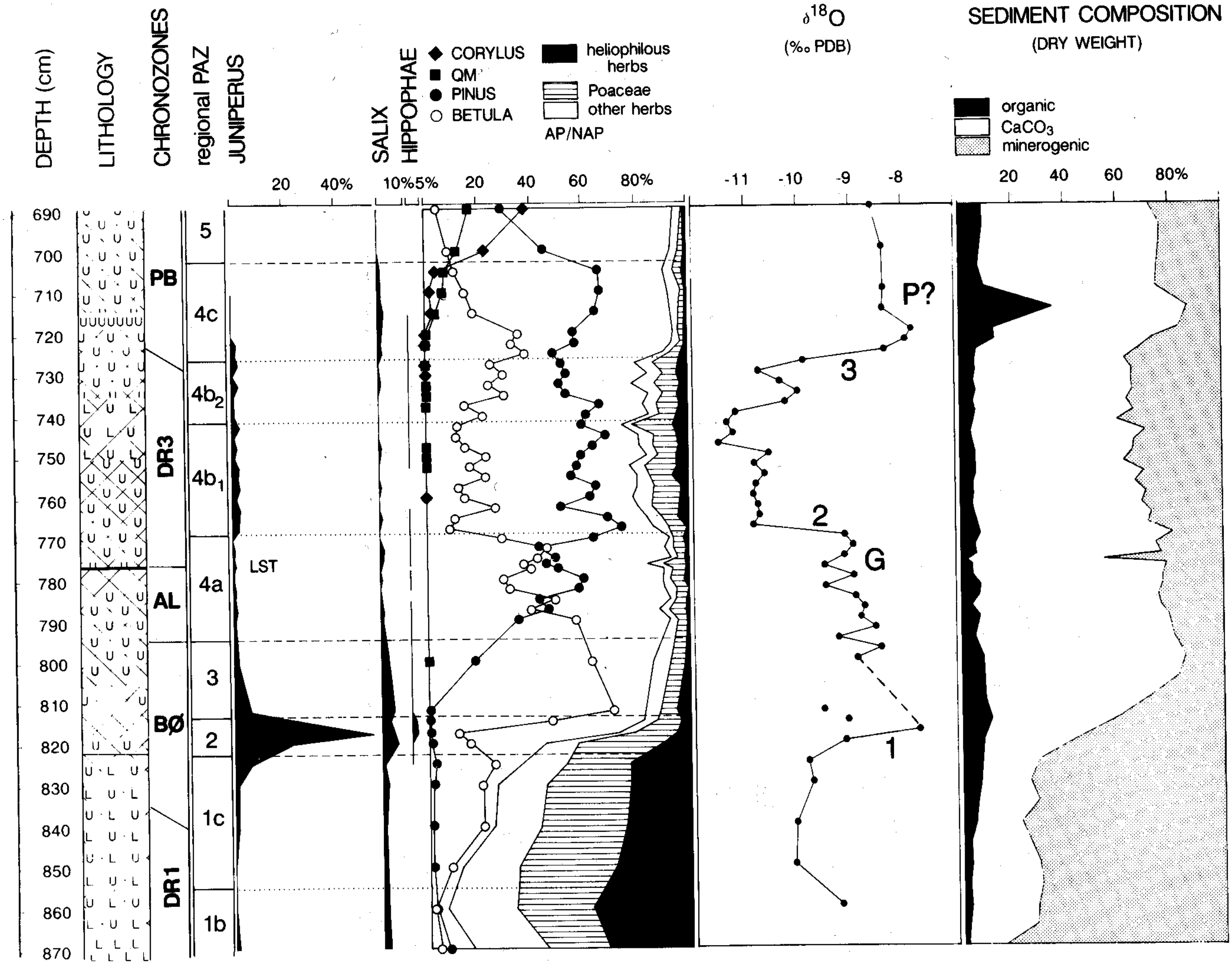

Figure 6 Pollen, oxygen-isotope stratigraphy, and sediment composition of Rotsee core RL-250 (after Lotter 1988, refined).

Sequence splitting was done by means of SPLIT2 (J. M. Line and H. J. B. Birks, unpublished program, see Birks, 1986).

In an attempt to answer question 2, we first reduced the highly multivariate pollen-stratigraphical data at each site to correspondence analysis axes by means of detrended correspondence analysis (DCA; Hill and Gauch, 1980). This serves to concentrate the major gradients of palynological variation or 'signal' within the data to the first few major axes and to relegate the minor variation or 'noise' to the later minor axes (Gauch, 1982). We retained for later analysis all DCA axes that represented 5% or more of the variance in the pollen data (Table 2). These axes are considered to reflect and summarise the major patterns of variation in the pollen data. The relationship between these major patterns of variation and the oxygen-isotope stratigraphy at each site was then explored by means of redundancy analysis (RDA; ter Braak, 1987) with depth and oxygen-isotope composition as predictor variables and the pollen data (summarised as the major DCA axes) as response variables.

Detrended correspondence analysis and redundancy analysis were implemented by CANOCO 3.11 (ter Braak, 1990). In the DCA rare taxa were downweighted and detrending was by segments. Default options were used for all other choices. The RDA was based on a covariance matrix between the response variables.

In approaching question 3, we modified Gordon's (1980) and Gordon and Reyment's (1979) constrained sequence-slotting procedure so that the oxygen-isotope sequences of interest are compared solely on the basis of their relative shapes rather than their absolute magnitudes because each site, particularly Dye 3, differs in their absolute values (see the sequence means in Table 3). Each sequence is first standardised to zero mean and unit variance, so that the sequence-slotting becomes a comparison of the shapes of the isotope curves with similar means and dispersion about these means. We modified Gordon's (1980) SLOTSEQ program to implement (constrained) sequence slotting using standardised oxygen-isotope data (CONSLOXY), and did unconstrained and constrained sequence slotting. The three constraints used are the major and unambiguous isotope shifts labelled 1, 2, and 3 in Fig. 9.

Results and discussion

Regional pollen assemblage zones (PAZ)

Because of terminological and semantic problems (see e.g. Ammann and Lotter 1989; Lotter et al., 1992a, b), we decided to subdivide the diagrams into regional PAZ rather than into Firbas (1949, 1954) pollen zones, as is traditionally done in central Europe. The proposed late-glacial and early

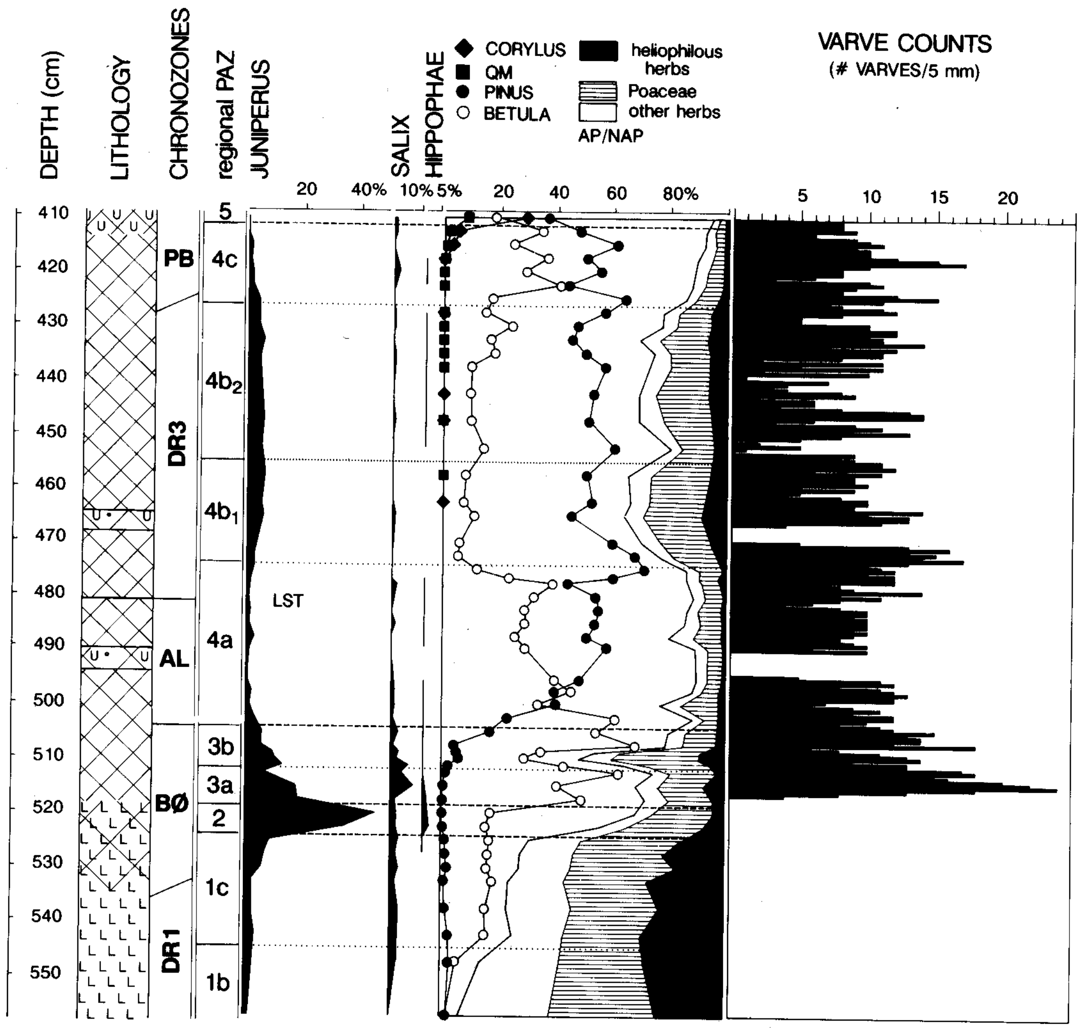

Figure 7 Pollen stratigraphy and varve counts of Soppensee core SO86-14 (after Lotter 1991a, refined).

Table 1 Summary of the results from principal components analysis (PCA) of Younger Dryas pollen spectra from five sequences

Site	Number of samples	Number of taxa	Cumulative percentage of PCA axes 1	2	3	4
Aegelsee AE-1	41	26	33.0	51.1	64.5	69.0
Aegelsee AE-3	18	25	36.0	58.5	67.2	75.3
Gerzensee G-III	9	24	35.9	54.6	66.5	77.1
Faulenseemoos	13	19	49.1	61.4	71.7	80.0
Rotsee RL-250	19	23	45.9	61.9	70.5	76.0

Table 2 Summary of the results from detrended correspondence analysis (DCA) of late-glacial pollen spectra from five sequences. The percentage variance represented by each DCA axis is listed

Site	Number of samples	Number of taxa	DCA Axis 1	2	3	4
Aegelsee AE-1	100	26	57.2	12.0	2.3	1.4
Aegelsee AE-3	54	32	44.3	3.3	1.5	1.4
Gerzensee G-III	65	28	37.6	4.0	1.2	0.9
Faulenseemoos	62	25	44.1	18.8	5.0	3.8
Rotsee RL-250	44	23	38.2	13.3	3.1	2.3

Holocene regional pollen zonation for the central part of the Swiss Plateau is based largely on the regional PAZ defined by Ammann and Tobolski (1983), Gaillard (1985), Lotter (1988), and Ammann (1989a). All diagrams presented in this study have been newly zoned and our regional biozonation does not necessarily correspond to the zonation by the original author of the diagram. Following the subdivision of Switzerland into IGCP 158 type regions (Ralska-Jasiewiczowa, 1986), the regional prefix CHb is used to label these regional PAZ.

For the definition of the regional PAZ the dominant and major taxa in the pollen diagrams are mainly considered. These PAZ describe the succession of pollen assemblages for the whole of the Swiss Plateau, whereas some sub-PAZ do not apply for all diagrams of the Swiss Plateau, depending on sampling resolution and/or local factors.

CHb-1 *Artemisia* PAZ

Description: high values of heliophilous NAP, especially of *Artemisia*.

Upper limit: rational limit or steep rise in *Juniperus*.

Age: > 14000 to 12700–12600 yr BP (^{14}C plateau).

Table 3 Psi values for pair-wise sequence slotting of the stable oxygen-isotope stratigraphy at five Swiss late-glacial sites and the Dye 3 site in Greenland. Values above the diagonal are constrained slotting, using the three major shifts shown in Fig. 9; values below the diagonal are for sequence slotting in the absence of any external constraints. The mean $\delta^{18}O$ and standard deviation for each sequence is also listed

			Constrained			
	AE-1	AE-3	G-III	FSM	RL-250	Dye 3
Aegelsee AE-1		1.365	2.065	1.126	0.994	1.725
Aegelsee AE-3	1.342		1.780	1.073	1.210	2.041
Gerzensee G-III	1.389	1.670		1.031	0.844	0.520
Faulenseemoos FSM	1.054	1.044	0.982		1.147	1.228
Rotsee RL-250	0.980	1.210	0.650	1.064		1.043
Dye 3	1.665	2.039	0.516	1.169	0.890	
			Unconstrained			
Means	−10.472	−10.451	−8.242	−10.857	−9.601	−32.490
Standard deviation	1.012	0.813	1.099	1.666	1.048	1.957

Subdivision

CHb-1a *Pinus* sub-PAZ
Description: many reworked pollen types, mainly AP of thermophilous (e.g. *Ulmus, Quercus*) and coniferous taxa (*Abies*). High *Pinus* values.
Upper limit: rise in NAP, decline in AP.
Age: > 14 000 yr BP.
CHb-1b *Helianthemum* sub-PAZ
Description: Heliophilous NAP (*Artemisia, Helianthemum, Thalictrum, Saxifraga oppositifolia*, Chenopodiaceae), Poaceae, and Cyperaceae dominant.
Upper limit: *Betula* (*nana*) increase.
Age: > 14 000 to ca. 13 500 yr BP.
CHb-1c *Betula nana* sub-PAZ
Description: expansion of *B. nana* and other dwarf shrubs (*Salix* spp., *Juniperus*).
Dominance of heliophilous NAP, Poaceae, and Cyperaceae.
Upper limit: rational limit or steep rise in *Juniperus*.
Age: ca. 13 500 to 12 700–12 600 yr BP (^{14}C plateau).

CHb-2 *Juniperus–Hippophaë* PAZ

Description: *Juniperus* peak (30–70%), then a *Hippophaë* peak (3–10%). Decrease of NAP and steady increase of *Betula*.
Upper limit: decline of *Juniperus* and increase of *Betula*.
Age: 12 700–12 600 to 12 500–12 400 yr BP.
Remark: At some sites absolute pollen analyses show that the rapidly increasing juniper values suppress birch percentages even though birch is also increasing. At these sites reforestation by birch took place during this zone.

CHb-3 *Betula* PAZ

Description: *Betula* dominates with high values and reaches its optimum during this zone.
Upper limit: rational limit of *Pinus* (> 20%)
Age: 12 500–12 400 to 12 000 yr BP.
Subdivision:

CHb-3a *Salix* sub-PAZ
Description: *Salix* peak (5–10%).
Upper limit: increase in NAP (Poaceae and *Artemisia*) and decrease in *Betula*.
Age: 12 500–12 400 to ca. 12 100 yr BP.
CHb-3b Poaceae–*Artemisia* sub-PAZ
Description: increase of NAP, mainly of Poaceae and *Artemisia* at the onset of the rational limit of the *Pinus* curve. A short increase or stabilisation in the declining *Juniperus* curve often occurs.
Upper limit: rise in *Pinus* > 20%.
Age: ca. 12 100 to 12 000 BP.
Remark: this subzone can be recognized only if there is a high temporal sample resolution. Moreover, it seems to be more prominent at elevations above 600 m. Two facies can be observed: a birch oscillation or a NAP oscillation during the birch decline.

CHb-4 *Pinus* PAZ

Description: *Pinus* is dominant throughout the whole zone, reaching its highest values.
Upper limit: *Corylus* increase > 10%, *Pinus* decline.
Age: 12 000 to ca. 9500 yr BP (^{14}C plateau).
Subdivision

CHb-4a *Betula* sub-PAZ
Description: *Pinus* values increase steadily while *Betula* is subdominant. Antagonism between the curves of *Pinus* and *Betula*.
Upper limit: NAP increase, often *Artemisia* > 2%.
Age: 12 000 to 10 800–10 700 yr BP.
Remark: a minor *Betula* peak is often observed shortly before or after the LST.
CHb-4b Poaceae-NAP sub-PAZ
Description: *Pinus* often reaches its highest values during the first part of this subzone. Increase of *Juniperus* and of the NAP values, especially Poaceae and heliophilous herbs (e.g. *Artemisia, Thalictrum*, Chenopodiaceae).
Upper limit: NAP decrease, decrease of *Artemisia* < 2%.
Age: 10 800–10 700 to 10 000 yr BP (^{14}C plateau).
Subdivision: At some sites this subzone can be further subdivided.
CHb-4b$_1$ Poaceae-NAP–*Juniperus* sub-PAZ
Description: one (sometimes two) *Pinus* depressions are characteristic, usually in connection with higher values of *Juniperus* and Poaceae.

Upper limit: end of *Pinus* depression and decrease in Poaceae and *Juniperus*.
CHb-4b$_2$ Poaceae-NAP–*Betula* sub-PAZ
Description: the values of *Juniperus* and Poaceae are slightly lower than during CHb-4b$_1$ whereas *Filipendula* values are usually slightly higher, especially towards the end of this subzone. Increase of *Betula* during this subzone.
Upper limit: NAP decrease, decrease of *Artemisia* $< 2\%$.
CHb-4c *Betula–Corylus* sub-PAZ
Description: NAP decrease while *Pinus* values increase with some oscillations or stabilise at a high level before they start to decline gradually. *Betula* reaches a minor peak. Rational limit of *Corylus* towards the end of this subzone. *Quercus, Ulmus,* and *Tilia* are present with values below 5%.
Upper limit: *Corylus* increase $> 10\%$, *Pinus* decline.
Age: 10000 to ca. 9500 BP (^{14}C plateaux).

CHb-5 *Corylus*–QM PAZ

These regional PAZ can be largely correlated with the Firbas (1949, 1954) system of pollen zonation: the Oldest Dryas biozone (Ia) includes regional PAZ CHb-1, the Bølling and Older Dryas biozones (Ib/c) include CHb-2 and CHb-3, whereas the Allerød biozone (II) corresponds to subzone CHb-4a. The Younger Dryas biozone (III) corresponds to subzone CHb-4b and the Preboreal biozone (IV) to subzone CHb-4c.

Chronology

The chronostratigraphic subdivision of late-glacial pollen records from the central part of the Swiss Plateau is based mainly on several high-resolution AMS ^{14}C dating series of terrestrial plant macrofossils at Lobsigensee and Rotsee (Zbinden et al., 1989) and follows Ammann and Lotter (1989). The problems and implications due to the same nomenclature for bio- and chronozones are discussed by Lotter *et al.* (1992a, b). It has been suggested that if these ambiguous zone names are used then it is necessary to specify whether they are used in the bio- or chronostratigraphic connotation.

The onset of the Bølling chronozone (BØ) at 13000 yr BP is situated within the *Betula nana* phase of PAZ CHb-1c. The transition to the Allerød chronozone (AL) at 12000 yr BP coincides with the rational limit of the *Pinus* curve (PAZ CHb-4). The LST forms an excellent isochronous time horizon in all the profiles investigated. The eruption which took place in the Eifel mountains (Germany), was dated to ca. 11000 yr BP (Wegmüller and Welten, 1973; Van der Bogaard and Schmincke, 1985) and therefore can be used as a good time-marker for the onset of the Younger Dryas chronozone (DR3).

The correlation between the late-glacial chrono- and biozonation is given in Table 4. Special attention should be paid to the periods of constant radiocarbon age (^{14}C plateaux, see Ammann and Lotter, 1989; Zbinden et al., 1989; Lotter, 1991a) at 12700, 10000 and 9500 yr BP and their implications for chronozonation and palaeoecological interpretation (see Ammann and Lotter, 1989; Lotter et al., 1992a, b). Owing to the plateau phase being dated at 10000 yr BP we provisionally correlate the chronostratigraphic transition from the late-glacial period to the Holocene with the transition from PAZ CHb-4b to CHb-4c (see also Ammann and Lotter, 1989; Lotter, 1991a).

Oxygen-isotope record

The $\delta^{18}O$ records of the cores presented here generally show the features typically observed in late-glacial calcareous European lake sediments. The oxygen-isotope values are thought to reflect approximately the mean annual temperature. As the oxygen isotope ratios in freshwater carbonates are influenced by several factors (e.g. isotopic composition of rain water, water temperature, evaporation, water residence time in the lake) they are not related in an unequivocal way to air temperature. However, several attempts have been made to reconstruct mean annual air temperatures, using $\delta^{18}O$ values measured on ostracods and molluscs (e.g. Lister, 1989; von Grafenstein et al., 1992). Provided there are no sediment disturbances (e.g. bioturbation), shifts in the $\delta^{18}O$ values of freshwater carbonates may reflect temperature changes without time lag. Moreover, as these shifts have been found to be synchronous at each site they represent excellent stratigraphical markers and consequently can be used as good tools for correlation between sites with different biostratigraphies.

The core from Gerzensee, the classic site for stable isotope studies on lake sediments, as well as the cores from Faulenseemoos and Rotsee, include three major shifts (numbered 1–3 in Fig. 9). Shift 1 coincides more or less with the transition from PAZ CHb-1c to CHb-2 (i.e. at the beginning of the Bølling biozone), shift 2 occurs shortly before the transition from CHb-4a to 4b (i.e. near the beginning of the Younger Dryas biozone), and shift 3 at the transition from CHb-4b to 4c (i.e. at the end of the Younger Dryas biozone).

Comparison of oxygen-isotope records as well as the pollen stratigraphies indicates that biogenic carbonate precipitation at Aegelsee started later than at the other sites investigated, i.e. after shift 1. Moreover, the $\delta^{18}O$ values of core AE-1 are anomalously low between 716 and 733 cm depth compared with the values for core AE-3 or for many other cores from central Europe (see Eicher, 1987). Since the other core (AE-3) from the same lake shows the expected isotopic trend during this period, we assume that anomalously low values in core AE-1 result from some local perturbation. The origin of this perturbation is not clear, as pollen, carbonate content, and $\delta^{13}C$ show no anomalous patterns in this part of the sediment record. As we are unable to trace the cause of this anomaly, we are forced to exclude this part of core AE-1 from further discussion.

Besides the three major shifts we can often detect several minor oscillations in the late-glacial and early Holocene $\delta^{18}O$ record. They have been labelled by letters in Fig. 9: Decrease A, which we shall refer to as the Aegelsee oscillation, coincides with regional PAZ CHb-3b. Decrease G, the Gerzensee oscillation (see Eicher, 1980), occurs before the deposition of the LST during PAZ CHb-4a. A slight resurgence of *Betula* pollen occurs with or shortly after this $\delta^{18}O$ decrease. Depression P occurs during PAZ CHb-4c i.e. the Preboreal.

Late-glacial climatic oscillations

On the basis of numerous pollen profiles the main features of the late-glacial vegetational history of the Swiss Plateau are now well known (see e.g. Zoller, 1987). Due to the coarse sample resolution in many profiles, small or very short oscillations in the pollen curves may, nevertheless, often be missed between two samples. As a result of this coarse sampling, the existence of minor large-scale climatic oscil-

Table 4 Regional pollen assemblage zones for the Central Swiss Plateau and their relationship to Firbas pollen zones, chronozones, and oxygen-isotope stratigraphy

Time (k yr) conventional ^{14}C years B.P.	^{14}C plateaux	$\delta^{18}O$	Chronozones Mangerud *et al.* (1974)	Chronozones Ammann & Lotter (1989)	Pollen zones Firbas (1949, 1954)	regional PAZ for the Central Swiss Plateau: PAZ	sub-PAZ	inferred vegetation
	▨	P	PB	PB	V	CHb-5 *Corylus*-QM		mixed deciduous forest
					IV	CHb-4 *Pinus*	4c *Betula-Corylus*	hazel, elm, oak, lime spread into pine-birch forest
10	▨	↑3	YD	DR3	III		4b Poaceae-NAP: 4b2 *Betula*	open pine forest with some birch and juniper stands
		↓2					4b Poaceae-NAP: 4b1 *Juniperus*	
11		G	AL	AL	II		4a *Betula*	pine-forest with some birch
			OD					pine expansion into birch forest
12		A	Bø	Bø	I b/c	CHb-3 *Betula*	3b Poaceae-*Artemisia*	opening of birch-forest
							3a *Salix*	birch-forest; reforestation by birch
	▨	↑1				CHb-2	*Juniperus-Hippophaë*	pioneer scrub of juniper, sea buckthorn and willow
13				DR1	I a	CHb-1 *Artemisia*	1c *Betula nana*	dwarf shrubs and meadows
							1b *Helianthemum*	meadows with alpine and steppe herbs
14							1a *Pinus*	treeless, sparse pioneer vegetation on open ground

lations during the late-glacial period has often been questioned for central Europe, especially for the Alps (e.g. Watts, 1980).

Aegelsee oscillation

Regional PAZ CHb-3 includes one (or often even two) regressive phase in vegetation development after reforestation. The expansion of herbs (mainly Poaceae and *Artemisia*) as well as the temporary decline of the birch curve indicates a climatic oscillation. This is often further emphasised by lithological evidence with a short change to clayey sediments (Figs 2, 3, and 5), suggesting increased soil erosion due to a reduced plant cover.

This oscillation has been found in many circum-alpine pollen profiles and was often attributed to the Older Dryas (Ic *sensu* Firbas, see e.g. Bertsch, 1961; Welten, 1972, 1982; Beug, 1976; Wegmüller, 1977; Bortenschlager, 1984). With a high temporal sample resolution it has also been found on the Swiss Plateau at elevations between 400 and 600 m a.s.l. (e.g. Gaillard, 1985; Rösch, 1985; Ammann, 1989a) where it always occurs shortly before or during the first rise in the *Pinus* curve.

Several independent radiocarbon datings on peat and terrestrial plant macrofossils suggest an age of 12000 yr BP for the rise in the pine curve (e.g. Welten, 1982; Ammann and Lotter, 1989). Interpolation of radiocarbon dates would yield a rough age in the range of ca. 12300 to 12000 yr BP (see also Gaillard and Moulin, 1989) for the observed regressive phase in vegetation development. Consequently, this phase is definitely older than 12000–11800 yr BP and hence cannot be attributed to the north European Older Dryas chronozone (OD *sensu* Mangerud *et al.*, 1974). Counts on sediment thin-sections of the laminated Soppensee record (Lotter, 1991a; Fig. 7) indicate that this decrease in birch pollen may have lasted for a maximum of ca. 100 varves. Therefore, its duration is too short to be reliably radiocarbon dated (see also Welten, 1982; Björck, 1984). However, this event might well be synchronous with the climatic oscillations observed in Scandinavia (e.g. Björck et al., 1988), Great Britain and Ireland (e.g. Watts, 1980; Atkinson et al., 1987), and France (e.g. Ponel, 1989; Ponel and Coope, 1990), which took place between ca. 12500 and 12000 yr BP.

In the $\delta^{18}O$ records this climatic oscillation is often indicated in an indistinct way. In the Aegelsee core AE-3 there is one unusually low value at the corresponding depth (835 cm). Inspection of the carbonate content and $\delta^{13}C$ shows some irregularities in this sample, perhaps due to higher minerogenic input. At Faulenseemoos and Rotsee there is one, and at Gerzensee there are two relatively low $\delta^{18}O$ values at the corresponding levels, with a normal carbonate content and $\delta^{13}C$. Inspection of other lacustrine sediment cores (see Eicher, 1979, 1980; Eicher and Siegenthaler, 1976; Eicher et al., 1981; Kaiser and Eicher, 1987) shows that in many, but not all, profiles one or a few lower $\delta^{18}O$ values occur within the regional PAZ CHb-3, at the increase of the pine curve, suggesting a short period of climatic deterioration. However, the example from Rotsee (Fig. 6) illustrates the importance of a high-resolution sampling of the pollen and isotope record in order to detect this oscillation and to verify that these low values are not outliers.

As this oscillation is of very short duration and occurs at a time of a highly dynamic environment, its detection in the sedimentary record may be further obscured by bioturbation, thus smoothing its occurrence. Both sampling resolution and bioturbation may explain the different local picture. The two cores from Aegelsee, taken only ca. 50 m apart, are a striking example of two possible palynostratigraphical reflections of the same event at the same locality: a fact which makes life for a pollen analyst challenging but not easy!

With the available palynological data it is not possible to decide whether this oscillation in the pollen record was caused by lower summer temperatures (Iversen, 1973) or by increased continentality and a drier climate (Kolstrup 1979, 1982; Ammann and Tobolski, 1983; Björck et al., 1988; Gaillard and Moulin 1989). The $\delta^{18}O$ minimum observed in several profiles suggests that the temperature was colder.

Gerzensee fluctuation

During regional PAZ CHb-4a, corresponding to the Allerød biozone (II *sensu* Firbas), a second increase of birch associated with a decrease in pine often can be observed, usually starting just before the occurrence of the LST, i.e. shortly before 11 000 yr BP. This feature is more or less consistent on the Swiss Plateau. In the stable isotope record a decrease in the $\delta^{18}O$ values at the corresponding depths was first described from Gerzensee (Eicher, 1980) and can be observed at many sites. Judging from the stable-isotope record, this oscillation might have been of the same magnitude as the Aegelsee oscillation. The absence of a NAP phase and of any minerogenic lithological evidence, however, indicates that its impact on the vegetation may have been smaller. This might have been due to different reasons, acting singularly or together. First, as oxygen isotopes record large-scale climate changes rather than local ones, this oscillation may have had a smaller impact in central Europe and the Alps. Second, established pine forests may not be as sensitive to climatic change as a pioneer birch forest. Interspecific competition may lead to a favouring of birch during a short period, without any NAP phase. The pine forest would subsequently need more time to regain its predominance.

Younger Dryas

The Younger Dryas as a biozone is well characterised in both the pollen and stable-isotope records of the Swiss Plateau. Concerning the duration and the amplitude, this climatic oscillation is the most important during the whole late-glacial period. At altitudes between 400 and 1000 m a.s.l. on the Swiss Plateau this climatic deterioration did not, however, result in complete deforestation.

Preliminary varve counts at Soppensee (Fig. 7) suggest a duration of the Younger Dryas biozone in the range of 600–800 varves (Lotter, 1991a, b). However, chronological studies on laminated sediments from maar lakes of the German Eifel mountains suggest a duration of 400 years (Zolitschka, 1990) whereas counts at the Polish lake Gościąż suggest up to 1400 years for the same biozone (Rozanski et al., 1992). Welten (1944) counted ca. 700 varves for this biozone at Faulenseemoos; nevertheless, this number has to be evaluated critically, as the techniques at that time were not very refined. Until we have more varve counts from different cores at Soppensee and other sites we can only speculate about the absolute duration of this biozone.

Shift 2 in the oxygen-isotope record usually takes place before the major changes in the pollen record, i.e. the increase of NAP. This suggests a lag phase between the onset of the climatic change and the reaction of the vegetation since oxygen isotopes reflect the climatic change without a time-lag. The end of the Younger Dryas biozone, in the oxygen-isotope record represented by shift 3, is hardly affected by any time-lag. These patterns contradict our expectations as one would expect a fast reaction of vegetation to a climatic deterioration and a time-lag in the recovery phase (see also Wright, 1984; Ammann, 1989c).

A duration in the range of 50 yr has been suggested (Dansgaard et al., 1989) for shift 3 of the $\delta^{18}O$ curve. If we accept that pollen and stable isotopes react more or less synchronously at the end of the Younger Dryas biozone, the annually laminated Soppensee record would yield a comparable duration for this transition.

The regional *Pinus*–Poaceae-NAP PAZ (CHb-4b) can often be subdivided into two parts. The first part is characterised by a decrease and subsequent increase of pine percentages. This phenomenon also can be observed in pollen diagrams from above the timberline (e.g. Küttel, 1974, 1977; Wegmüller and Lotter, 1990). Furthermore, slightly higher values of juniper and of grasses occur during the first part, whereas the second part is often characterised by higher *Filipendula* and birch values (Fig. 8). This bipartition of the Younger Dryas biozone (III *sensu* Firbas, 1949) can be observed in many profiles of the Swiss alpine foreland. The sequence splitting results (Table 5) suggest that statistically significant divisions occur within the Younger Dryas at Aegelsee AE-1, Gerzensee and Faulenseemoos but not at Rotsee or Aegelsee AE-3.

The bipartition of the Younger Dryas, evidenced in many pollen records, is also present in the oxygen-isotope record of many lake deposits on the Swiss Plateau. The section between shifts 2 and 3 (Fig. 9) is often divided into two phases by an intermediate $\delta^{18}O$ minimum. The transition from subzones CHb-4b$_1$ to CHb-4b$_2$ often takes place during or shortly after the decline of the $\delta^{18}O$ values to their minimum. All sites (Table 6) show a significant relationship between pollen stratigraphy and oxygen isotopes and depth, as assessed by Monte Carlo permutation tests of the first dominant RDA axis (99 permutations). However, when oxygen-isotope composition is the only external predictor, there is a significant relationship between pollen and oxygen-isotope stratigraphy at Aegelsee AE-1 and Faulenseemoos only, suggesting that depth rather than oxygen-isotope stratigraphy may be a better predictor of the pollen stratigraphy. This hypothesis was tested by partial RDA, where the effects of depth are partially removed statistically, with depth as a co-variable (ter Braak, 1988). There is still a statistically significant relationship at Aegelsee AE-1 and Faulenseemoos between the pollen stratigraphy, as summarised by the first two or three DCA axes, and the oxygen-isotope stratigraphy when the effects of depth are partially removed. Possible explanations for the absence of this relationship at the other sites could again be the sampling resolution and/or smoothing of the signal by bioturbation.

Based on pollen data, Markgraf (1969) and Welten (1972) suggested that the second part of the Younger Dryas was drier than the first, Poaceae-rich part. Glacier data from the Central Alps (Kerschner, 1980) seem to corroborate this. At first sight our pollen data, especially the increase in *Filipendula* (assuming it is *F. ulmaria*) values during the second part of the

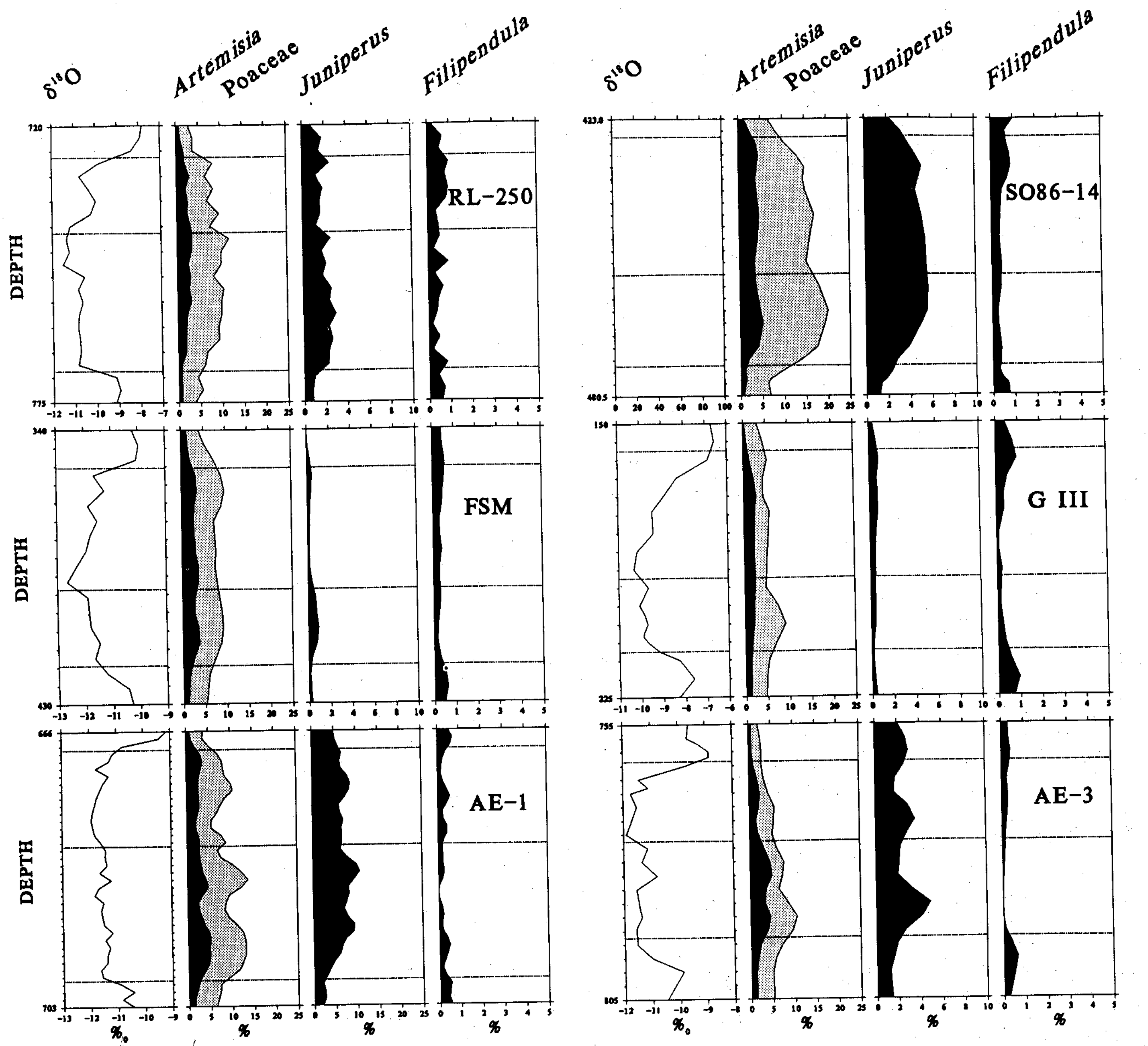

Figure 8 Oxygen isotopes and selected pollen curves for the Younger Dryas biozone. Upper and lower dashed lines mark the boundaries of regional PAZ CHb-4b (i.e. Younger Dryas biozone) whereas the middle dashed line indicates the subdivision of CHb-4b into CHb-4b$_1$, and 4b$_2$. The pollen values are smoothed by a three-sample running mean.

Younger Dryas biozone seem to contradict this interpretation of a drier or more continental second part. Moreover, there are no unambiguous indications for a wetter first part that are consistently present in diagrams from all sites. However, if we assume that the second part of the Younger Dryas biozone was drier, a possible explanation for the increased values of *Filipendula* and *Betula* during PAZ CHb-4b$_2$ could be a lowering of lake levels. This would result in more wet or moist soils immediately around the lake available for the colonisation of local plants such as *Filipendula* and *Betula*. Increased input of minerogenic sediment during this subzone, CHb-4b$_2$ (e.g. in AE-3 or RL-250, Figs 3 and 6), would then be explained not only by soil erosion but also by shore erosion.

Preboreal oscillation

During the Preboreal, shortly after shift 3 in the $\delta^{18}O$ curve, another small oscillation in the oxygen-isotope record often can be observed at ca. 9500 yr BP (P in Fig. 9). Becker *et al.* (1991) reported a δD minimum occurring around 9350 yr BP measured on tree rings from southern Germany. It might be possible that this represents the same event seen in the lake $\delta^{18}O$ records, although the date seems to be too young compared with the observed Preboreal oscillation.

In the pollen record this oscillation, like the Gerzensee oscillation, is not very prominent. It often occurs during a phase of increasing *Betula* values while the *Pinus* curve is decreasing. Both had already started well before this feature in the stable-isotope record. Besides the weak pollen evidence for this oscillation it is difficult to correlate it with other Preboreal oscillations as it occurs at ca. 9500 yr BP, i.e. during another phase of constant radiocarbon age (Becker and Kromer, 1986; Becker et al., 1991). For the Alpine region the Preboreal 'Piottino' oscillation has been described (Zoller, 1960). However, it is not useful to continue using this term as it has been shown that this oscillation is part of the Younger Dryas (Küttel, 1977). Nevertheless, there are many indications for Preboreal climate oscillations in the Alps as well as in

Table 5 Results of the sequence splitting applied to the first four principal components of each sequence. The PCA axis that is split, the position of the split, the likelihood ratio Λ_t, and the number of samples (n) in the section in which the statistical test is conducted are listed

Site	PCA axis	Split (cm)	Λ_t	n
Aegelsee AE-1	1	668.5	7.09	41[a]
	1	699.5	9.62	37[a]
	1	683.5	7.54	31[a]
	2	674.5	5.02	41
	2	682.5	6.45	31[a]
	2	703.5	2.70	23
	2	696.5	2.26	21
	3	693.5	23.92	41[a]
Aegelsee AE-3	2	778.5	2.84	18
	3	788.5	4.15	18
Gerzensee G-III	1	192.5	7.36	9[a]
Faulenseemoos	2	392.5	4.78	13[a]
	4	372.5	4.65	13
Rotsee RL-250	1	736.25	3.97	19
	2	763.75	3.75	19

[a]Significant splits at $p \leq 0.05/n$

central and northwestern Europe, which are termed e.g. 'Youngest Dryas' (Behre, 1978; Schneider, 1985b), or 'Búði' (Ingolfsson, 1991). The absence of any palynological evidence for this short climatic oscillation recorded in the oxygen-isotopes may be due to the absence of any clear vegetational ecotone on the Swiss Plateau at that time.

Teleconnection of oxygen-isotope records

Only recently has it become possible to obtain reliable (hard-water error free) ^{14}C datings and stable isotope measurements on the same core by means of AMS techniques. Such studies were carried out at Zürichsee (Lister, 1988) and at Rotsee (Lotter and Zbinden, 1989). Furthermore, high-resolution AMS dating results revealed problems in connection with constant radiocarbon ages that encompass the $\delta^{18}O$ shifts towards positive values (Ammann and Lotter, 1989; Zbinden et al., 1989). Shift 1 (Fig. 9) lies within the 12 700 yr BP plateau, whereas shift 3 occurs in the 10 000 yr BP plateau. Shift 2, however, has been dated at 10 700 to 10 800 yr BP. Because of these plateau phases it is not possible to estimate any meaningful rates of biotic change during these periods of climatic warming on the basis of radiocarbon dates (see Ammann, 1989b; Lotter et al., 1992a).

Several attempts have been made to correlate continental, deep-sea, and ice-core records (e.g. Broecker et al., 1988; Lehman and Keigwin, 1992). It has been shown that the major fluctuations (shifts 1–3 in Fig. 9) and the shape of the $\delta^{18}O$ curve from European lacustrine sediments can be correlated with the Dye 3 $\delta^{18}O$ profile of the Greenland ice sheet (Oeschger et al., 1984; Siegenthaler et al., 1984). Both curves seem to represent not only the same time interval but also the same climatic fluctuations. The palynologically dated sedimentary oxygen-isotope record therefore could be used to date the major shifts in the Greenland ice cores as well as the marine cores.

It is not only possible to correlate the major shifts between the continental and the Greenland oxygen-isotope records but also some of the minor shifts. Pollen and oxygen-isotope analyses have shown that the Aegelsee fluctuation (A in Fig. 9), the Gerzensee fluctuation (G in Fig. 9), and the Preboreal fluctuation (P in Fig. 9) are synchronous, at least within the investigated area of the Swiss Plateau. The similarity of the $\delta^{18}O$ profiles in Swiss lake sediments and in Greenland ice is a strong indication that they reflect large-scale climatic variations in the North Atlantic region.

The results of the sequence slotting, in terms of psi, a standardised measure of correspondence between pairs of sequences (see Gordon, 1980; Birks and Gordon, 1985) are tabulated in Table 3. Although little is known about the statistical significance of different values of psi, values less than 1.0 can be tentatively interpreted as indicating a very good fit between the sequences, values in excess of 2.0 can be regarded as suggesting a poor comparison, and values between 1.0 and 2.0 imply a fairly good correspondence (Gordon, 1980). The psi-values for the constrained slotting are consistently equal to or larger than the corresponding values for unconstrained slottings. The very low psi-values for many of the unconstrained slottings appear, however, to be a result of excessive blocking of samples, arising because there is very little difference between several consecutive samples within a sequence (see Birks and Gordon, 1985, p. 139). The slottings suggested by the constrained procedure have slightly higher psi values but show much less blocking than the unconstrained slottings. We therefore regard the results of constrained slottings as being more reliable and robust than the unconstrained results. In general the psi values for constrained slotting are low (< 1.5), suggesting a good correspondence between the oxygen-isotope sequences. Exceptions are between Dye 3 and Aegelsee (AE-3 psi = 2.041; AE-1 psi = 1.725).

If we concentrate on the Younger Dryas, i.e. the part of the oxygen-isotope curves between shifts 2 and 3, we notice a discrepancy between the continental records on the one hand and the Dye 3 or Camp Century records (Patterson and Hammer, 1987) on the other: after shift 2 most of the continental records show a first phase with approximately constant low values before they decrease further to reach their minimum values, whereas the Greenland record seems to decrease more or less steadily to its lowest values. The increase to shift 3 again seems comparable: both systems show a short phase of decrease in the slope. Possible explanations for this phenomenon might be sought for in the differences between sedimentation in lacustrine environments and glacier accumulation. Higher sediment accumulation rates may stretch the $\delta^{18}O$ signal in the sediment, as seems to be the case for the first part of the Younger Dryas.

Conclusions

Our results permit the following conclusions.

1. The palynostratigraphy of the investigated sites can be subdivided into regional pollen assemblage zones. A regional zonation including four major PAZ, which often can be subdivided further, is proposed for the late-glacial and early Holocence of the central part of the Swiss Plateau.
2. Pollen analysis reveals two distinct phases of late-glacial climatic oscillation after the initial reforestation. The first occurred before 12 000 yr BP and had a short duration of

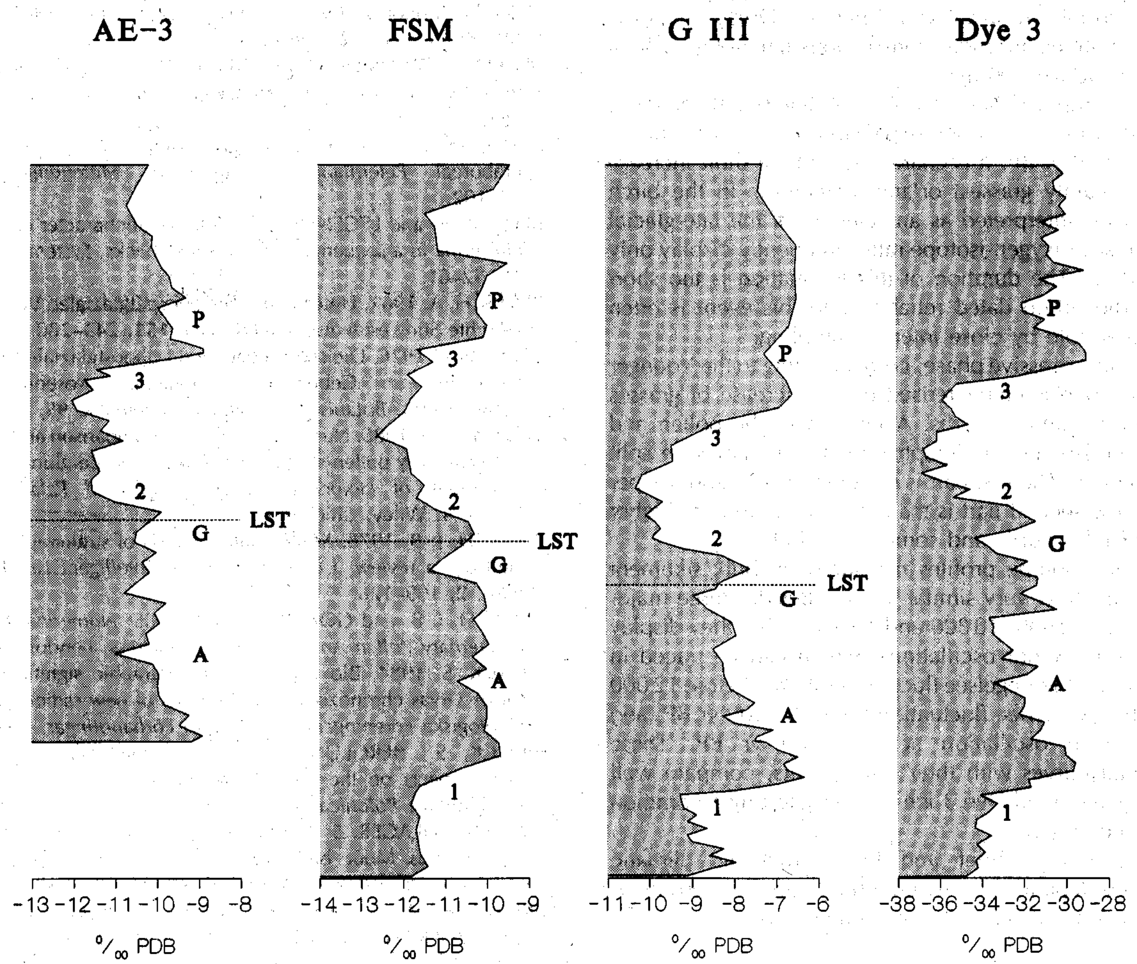

Figure 9 Comparison of oxygen-isotope records from Swiss lakes Aegelsee (AE-3), Faulenseemoos (FSM), and Gerzensee (G-III) with the Greenland Dye 3 record (Dansgaard et al., 1982). LST marks the position of the Laacher See Tephra (11 000 yr BP). Letters and numbers mark the position of synchronous events (for details see text).

Table 6 Results of redundancy analysis and partial redundancy analysis permutation tests for the significance of axis 1 when oxygen isotopes and depth are predictor variables, when oxygen is the only predictor, and when oxygen isotopes are the predictor variable and depth is a covariable

Site	Predictor variable: $\delta^{18}O$ and depth	Predictor variable: $\delta^{18}O$ Covariable: depth	Predictor variable: $\delta^{18}O$	Number of response variables (DCA axes)
Aegelsee AE-1	0.01[a]	0.01[a]	0.02[a]	2
Aegelsee AE-3	0.01[a]	0.16	0.20	1
Gerzensee G-III	0.01[a]	0.46	0.57	1
Faulenseemoos	0.01[a]	0.01[a]	0.01[a]	3
Rotsee RL-250	0.01[a]	0.21	0.08	2

[a]Significant at $p < 0.05$

the order of ca. 100 yr. The second is more prominent and is correlated with the Younger Dryas biozone. Both oscillations are more conspicuous palynologically at elevations above 600 m.

3. The first regressive phase, termed the Aegelsee fluctuation, occurred at the time of the expansion of pine, i.e. shortly before 12 000 yr BP. It is characterised by a short increase in NAP (mainly grasses) or/and a decrease in the birch percentages, interpreted as an opening of the late-glacial birch forests. Oxygen-isotope ratios frequently display only a brief drop. The duration of this fluctuation is too short to be radiocarbon dated reliably, and this event is often only discernible by close interval sampling.
4. The second regressive phase, corresponding to the Younger Dryas biozone is characterised by an increase of grasses and light-demanding herbs. According to the pollen and the oxygen-isotope stratigraphy it can sometimes be split into two parts. The first part is richer in grasses and juniper whereas the second part is mainly characterised by higher values of *Filipendula* and sometimes birch.
5. The oxygen-isotope profiles measured on bulk sediment samples all show very similar curves. Besides three major shifts (at ca. 12 600, 10 800 and 10 000 yr BP) they display at least three minor oscillations, which can be traced in most profiles: the Aegelsee fluctuation shortly before 12 000 yr BP, the Gerzensee fluctuation before 11 000 yr BP, and the Preboreal oscillation at ca. 9500 yr BP. These continental curves with their minor shifts, compare well with the Greenland Dye 3 ice core, suggesting a common climatic control.
6. The detection of brief and highly dynamic climatic oscillations depends heavily on a high sampling resolution as well as on a high sediment-accumulation rate.

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