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LAKE SEDIMENTS AS CONTINENTAL $\delta^{18}O$ RECORDS FROM THE GLACIAL/POST-GLACIAL TRANSITION

by

U. Siegenthaler, U. Eicher and H. Oeschger

(Physikalisches Institut, Universität Bern, Sidlerstrasse 5, CH-3012 Bern, Switzerland)

and W. Dansgaard

(Geophysical Isotope Laboratory, University of Copenhagen, Haraldsgade 6, DK-2200 Copenhagen, Denmark)

ABSTRACT

As in polar ice, ^{18}O variations of precipitation are recorded in carbonate sediments formed in lakes (lake marl). We have analyzed many late-glacial profiles from Europe. There are strong ^{18}O variations which coincide with well-known pollen zone boundaries and which indicate that abrupt, drastic climatic changes occurred in the late glacial period. These events are a major warming around 13 ka BP (pollen zone boundary Oldest Dryas/Bölling) and a marked cold phase between about 10.8 and 10 ka BP (Younger Dryas). Comparison of the $\delta^{18}O$ records of European lake sediments and of Greenland ice cores reveal a striking similarity which indicates that climatic changes in the late glacial and early postglacial were parallel in Greenland and in Europe. First results from North American lake-sediment profiles do not exhibit similar $\delta^{18}O$ variations. This pattern of climatic changes was probably caused by retreating and readvancing polar water in the high-latitude North Atlantic Ocean, as discussed by Ruddiman and McIntyre (1981).

INTRODUCTION

The polar ice sheets contain a deep-frozen record of $\delta^{18}O$ in precipitation and thus of climatic variations in ancient times. In temperate latitudes, such a direct storage of precipitation is available only to a limited extent, but $\delta^{18}O$ variations in precipitation from late-glacial time are recorded in lacustrine carbonate sediments. This has been demonstrated for a number of sites in Europe (e.g. Eicher and Siegenthaler 1976, Eicher and others 1981), and some main results are summarized below. A comparison of $\delta^{18}O$ results from European lake sediments and from Greenland ice cores exhibits striking similarities, so that a direct correlation between late-glacial deposits from central Europe and Greenland now appears feasible.

$\delta^{18}O$ variations of continental waters are recorded in lake carbonate sediments if these formed under conditions of isotopic equilibrium with the lake water. Lake marl is such a kind of sediment. It is precipitated in hard-water lakes when submerged aquatic plants withdraw dissolved CO_2 for assimilation. The resulting sediment contains a high percentage of carbonate, typically 70 to 90% in the sediment profiles considered here.

The profiles discussed here were obtained in central Europe in regions surrounding the Alps. They cover the time period between about 14 and 8 ka BP, and they were all analyzed for pollen composition so that indirect dating by means of pollen zone boundaries is possible.

Lake carbonates do not directly reflect $\delta^{18}O$ variations in precipitation. First, there is a temperature-dependent $\delta^{18}O$ shift between water and carbonate in isotopic equilibrium. Second, lake water is often enriched in $\delta^{18}O$ with respect to mean precipitation because of evaporation. Nevertheless, the variation of $\delta^{18}O$ in precipitation with mean air temperature is large enough to dominate these other effects. Since the isotopic fractionation between carbonate and water has a negative temperature coefficient, while $\delta^{18}O$ in precipitation is positively correlated with temperature, the variations recorded in lake marl may be somewhat smaller than originally occurred in the precipitation and can therefore be considered as the lower limit for the isotope variations in precipitation. A quantitative interpretation of the isotopic profiles on lake carbonates, for instance in terms of temperature, is, however, even more difficult than for polar ice (Eicher and Siegenthaler 1976).

$\delta^{18}O$ VARIATIONS IN LAKE MARL FROM THE LATE-GLACIAL AND EARLY POST-GLACIAL

Three sediment profiles will be discussed here. Gerzensee and Faulensee (both at 600 m a.s.l.) are two sites in Switzerland situated in the northern lower Alps, about 20 km distant from each other (Eicher and Siegenthaler 1976). Tourbière de Chirens (460 m a.s.l.) is a peat bog in the western Alps in France (Eicher and others 1981). Only at Gerzensee does a lake still exist, while at the other two sites, the former lakes have dried out. Pollen analysis was performed by M Welten on the Faulensee profile, by S Wegmüller on the profile Tourbière de Chirens and by U Eicher on the Gerzensee profile.

Figure 1(a) shows $\delta^{18}O$ expressed as relative deviation from the PeeDee belemnite PDB standard, plotted versus depth below the surface. In the left column, the pollen zones according to the system of Firbas are also indicated. Table I gives a list of these pollen zones, together with approximate ^{14}C ages for the pollen zone boundaries, as generally

TABLE I. POLLEN ZONES OF LATE-GLACIAL AND EARLY POST-GLACIAL IN CENTRAL AND NORTHERN EUROPE

Pollen zone		^{14}C age of zone boundaries (ka BP)
IV	Pre-Boreal (warm)	
		10
III	Younger Dryas (cold)	
		10.8
II	Alleröd (warm)	
Ib	Bölling (warm)	
		ca. 13
Ia	Oldest Dryas (cold)	

determined for central Europe (e.g. Welten 1972, Mangerud and others 1974).

The $\delta^{18}O$ profile exhibits clear variations which are associated with the history of the climate as known from the results of pollen analysis. At the boundary Oldest Dryas/Bölling (Ia/Ib) a sharp increase of 3°/oo occurs, and the beginning and end of the Younger Dryas cold phase are marked by rapid $\delta^{18}O$ shifts of about the same size. The changes of $\delta^{18}O$ in precipitation which are recorded here must each have occurred within a century or even less, considering the short sediment intervals in which they occurred and the time scale given in Table I. This suggests that some processes were operating at the end of the last glaciation which lead to abrupt and drastic climatic changes in Europe.

Besides these three rapid $\delta^{18}O$ shifts, slower variations are observed. A first maximum is reached immediately after the first rise, followed by a gradual decline from pollen zone Ib to II. This suggests that possibly the warmest phase of the last late-glacial interstadial (Bölling-Alleröd) was at its beginning. It should, however, be kept in mind that $\delta^{18}O$ in precipitation, and therefore in lake marl, is influenced not by temperature alone, but by other factors as well, so that such a statement cannot at present be made with certainty.

Figure 1(b) and (c) shows the $\delta^{18}O$ results for the lake marl profiles of Tourbière de Chirens and Faulensee. The same general patterns are observed as in Figure 1(a): three major changes at the pollen zone boundaries Oldest Dryas/Bölling (Ia/Ib), Alleröd/Younger Dryas (II/III) and Younger Dryas/Pre-Boreal (III/IV); a slow decrease with some minor fluctuations from Bölling to Alleröd; and after the transition from the Younger Dryas to the Pre-Boreal period, which marks the boundary between Pleistocene and Holocene, similar values as observed during the Bölling are found. The similarities between the three profiles also include minor fluctuations. Before the end of the Alleröd pollen zone, a $\delta^{18}O$ minimum is observed which has been found also in marl profiles from other sites. It is obviously a real feature and Eicher (1980) has termed it the Gerzensee fluctuation. At Gerzensee and Faulensee there is a small minimum after the rise at the transition Younger Dryas/Pre-Boreal, which is also observed in several other profiles not shown here. Thus we conclude that lake marl sediments form a reliable record of even minor variations of $\delta^{18}O$ in precipitation. The variations observed in marl may be different in amplitude (in general slightly smaller) than those in precipitation due to the temperature-dependent isotope fractionation

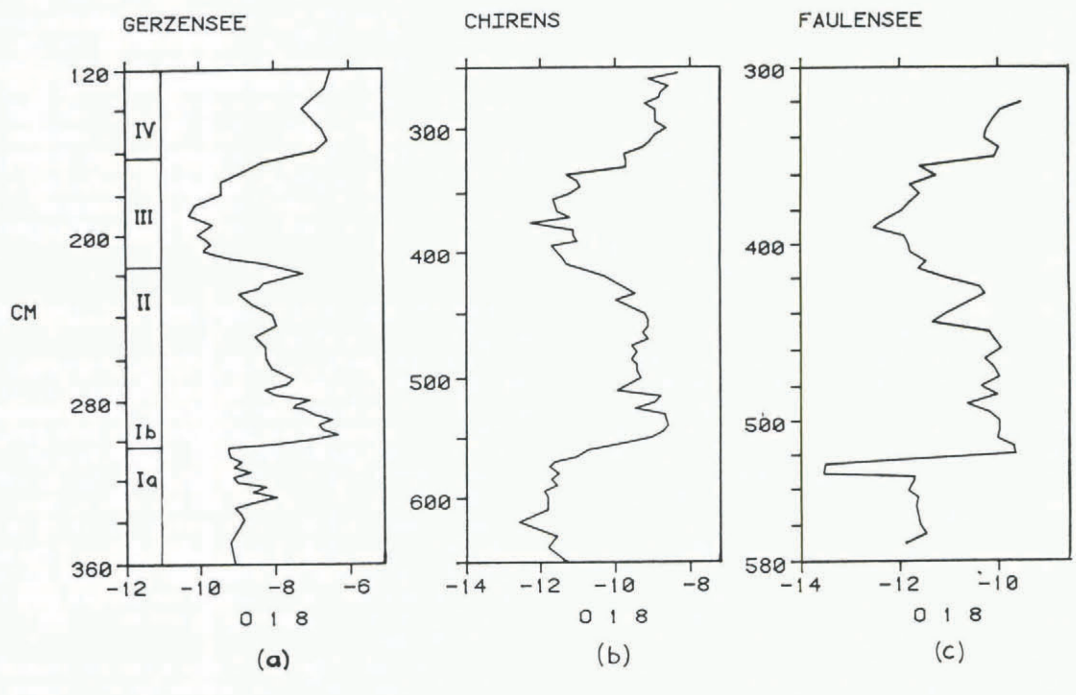

Fig.1. $\delta^{18}O$ profiles in lake marl sections from (a) Gerzensee, (b) Tourbière de Chirens, (c) Faulensee. Horizontal axes: $\delta^{18}O$°/oo (vs PeeDee belemnite (PDB) standard), vertical axes: depth in cm. Left column in Figure 1(a): pollen zones, cf. Table I.

between water and carbonate, but the relative pattern of changes is preserved.

The agreement between different profiles is not perfect; deviations can be observed from Figure 1. Thus there is a marked minimum in the Faulensee profile at about 530 cm depth which has no analogies at the two other sites (nor in profiles analyzed from elsewhere). Therefore, this probably reflects a local feature, for instance due to a transient change in the water or temperature regime of the lake. Only those features which show up in several independent profiles can be considered as representing large-scale climatic fluctuations.

The absolute $\delta^{18}O$ values of the three profiles differ by several parts per thousand. Since Gerzensee and Faulensee are at the same altitude and at a small distance from each other, geographical variations of $\delta^{18}O$ in precipitation cannot have been important for the difference between these two profiles. It must therefore have been caused by local influences such as ^{18}O enrichment of lake water, depending on the residence time of water in a lake. The water of Gerzensee is presently enriched by ca. 2.5°/oo by evaporation, compared to the water feeding it, and we have evidence also from ^{13}C and ^{14}C measurements that the water had a relatively long residence time in the lake already in late-glacial times (Eicher and others in press). This would explain why higher $\delta^{18}O$ values are observed in the profile from Gerzensee than in the two others.

The abrupt climatic variations reflected in the lake marl $\delta^{18}O$ profiles are well known also from other palaeoclimatic studies (e.g. Coope 1975, Watts 1980). An important question for the understanding of climatic change is whether these rapid changes were a hemispheric or even a global phenomenon, or whether they were restricted to Europe only. In order to find out if similar $\delta^{18}O$ variations in precipitation as recorded in lake sediments from central Europe also occurred in North America, Eicher and others (in preparation) have analyzed lake marl from the Great Lakes region. The results obtained so far do not show any indications of abrupt changes. This is consistent with the finding that there is no clear evidence in North America for climatic fluctuations in the late Wisconsin comparable to the Younger Dryas period (Wright 1977).

COMPARISON WITH $\delta^{18}O$ IN GREENLAND ICE CORES

The most direct information on past $\delta^{18}O$ variations in precipitation is stored in polar ice. Dansgaard and others (1982) presented a continuous $\delta^{18}O$ record of an ice core from Dye 3, Greenland, which had been obtained as the result of a joint American-Danish-Swiss project. Oeschger and others (in press) observed that the late-glacial part of the $\delta^{18}O$ record from the ice core of Dye 3, Greenland, correlates very well with the late-glacial $\delta^{18}O$ records from European lake marl profiles. In Figure 2, the record of Gerzensee is shown together with the section of the Dye 3 record between about 1 764 and 1 822 m depth. The similarity between the two curves is striking. Not only the three major shifts, corresponding to the pollen zone boundary Oldest Dryas/

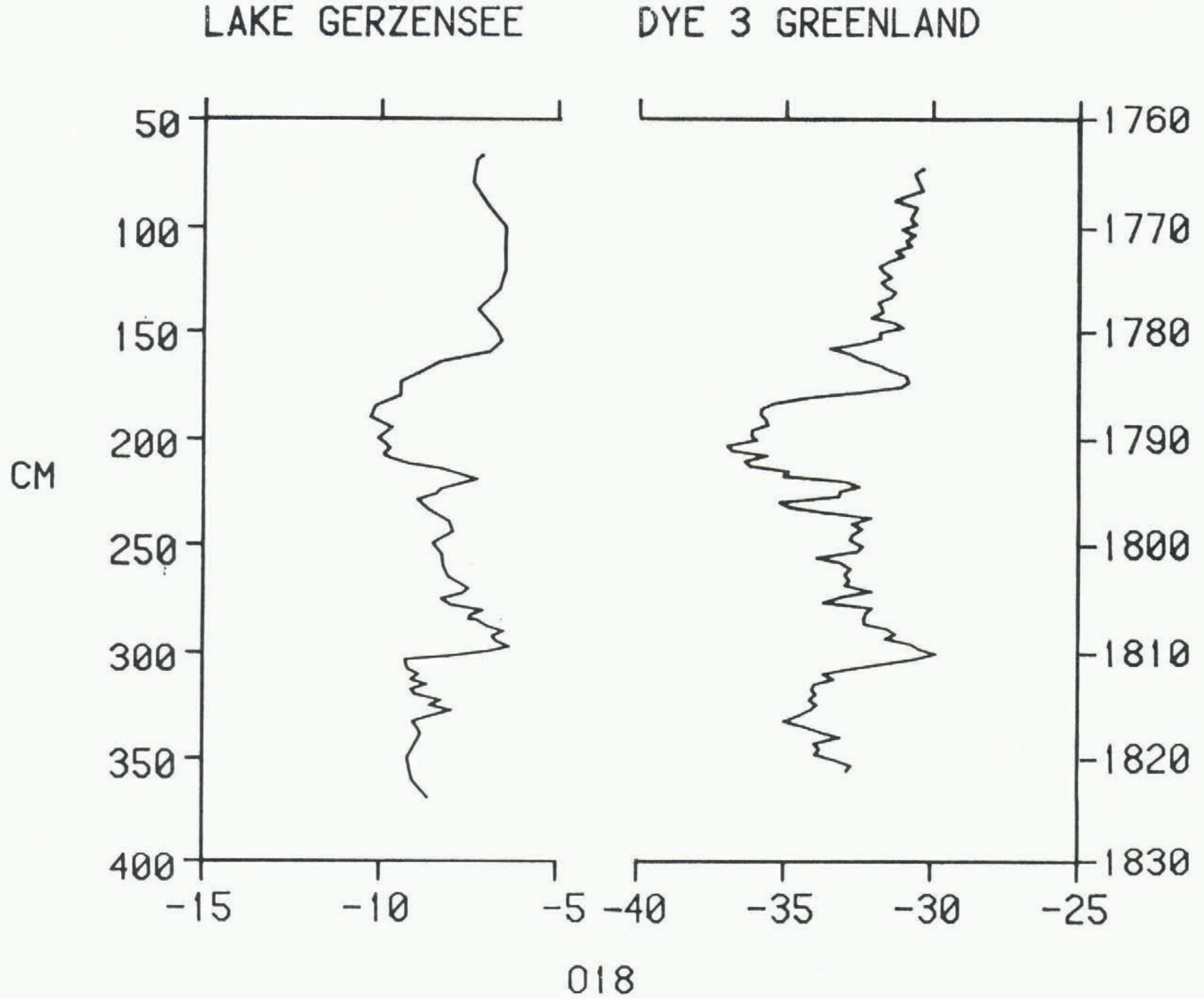

Fig.2. $\delta^{18}O$ profiles from Gerzensee lake marl (left) and from the Greenland ice core from Dye 3 (right) versus depth.

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Bölling and the beginning and end of the Younger Dryas, but also minor variations described above for the lake marl profiles, e.g. the gradual decrease after the first rise or the "Gerzensee fluctuation", are clearly represented in the Dye 3 ice core. In the isotope profile from Camp Century, Greenland, the same pattern as in the profile from Dye 3 is observed, although with a different baseline trend (cf. Dansgaard and others 1982). The excellent correlation indicates that there is a great probability that the two curves of Figure 2, which can be assumed to be representative of $\delta^{18}O$ variations in Greenland and central Europe, reproduce the same sequence of climatic events and therefore the same time interval. This means that the corresponding sections of the ice cores can be dated, based on the ^{14}C age scale determined for the late glacial in Europe (cf. Table I). The Dye 3 core section from 1 822 to 1 764 m depth shown in Figure 2 must correspond to the time span between about 14 and 8 ka BP.

POSSIBLE CAUSES FOR THE ABRUPT LATE-GLACIAL CLIMATIC EVENTS

We conclude from these findings that in the late-glacial and early post-glacial the climate in central Europe and in Greenland changed in a similar way, but that the changes observed here did not occur, or in an attenuated form only, in North America. The question arises as to what mechanism may have been responsible for the abrupt climatic changes that took place in the late glacial in Europe and in Greenland, that is in the north and east of the North Atlantic Ocean, and why analogous events cannot be observed in North America, that is in the west of the North Atlantic.

Probably the cause has to be found in changes of surface conditions in the ocean. Ruddiman and McIntyre (1981) have summarized convincing evidence that before ca. 13 ka BP, cold polar water had advanced as far south as about 40°N in the North Atlantic and that a rapid warming took place around 13 ka BP. A major cooling and readvance of the oceanic polar front took place between 11 and 10 ka BP, so that most of the high-latitude North Atlantic experienced almost full-glacial temperatures again. Duplessy and others (1981) have estimated that in the Bay of Biscay the warming around 13 ka BP and the cooling between 11 and 10 ka BP amounted to about 12°C (summer surface-water temperatures). Because of the prevailing west winds, mainly the continental regions east of the ocean were affected by changes of sea surface temperature, while North America, upstream of the North Atlantic Ocean, did not experience these changes so strongly. Palaeoclimatic studies have shown that the Younger Dryas cold period was most severe in north-west Europe (Watts 1980), which is consistent with the finding that polar waters readvanced to about 45°N in that period (Ruddiman and McIntyre 1981).

The presence of ice-rafted sand in deep-sea sediments indicates that in times of polar water advances icebergs must have been present in the North Atlantic north of about 40°N, which implies strong cooling of the sea surface due to ice melting (Ruddiman and McIntyre 1981). The icebergs may have been produced by surging of continental ice masses, but this mechanism requires further study. At any rate a strong coupling between sea ice, ocean temperatures and circulation and continental climate must have been involved in producing the climatic events of the transitional period from the glacial to the post-glacial.

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