

Oxygen and Carbon Isotope Trends in a Late Glacial-Holocene Pollen Site in Wyoming, U.S.A.

Le comportement des isotopes de l'oxygène et du carbone dans un site pollinique tardiglaciaire holocène du Wyoming, É.-U.

Verlauf von Sauerstoff- und Kohlenstoff-Isotopen in einem späteiszeitlichen Pollen-Gelände aus dem Holozän in Wyoming, U.S.A.

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Article abstract

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OXYGEN AND CARBON ISOTOPE TRENDS IN A LATE GLACIAL-HOLOCENE POLLEN SITE IN WYOMING, U.S.A.*

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ABSTRACT Stable isotope studies of North American Late Glacial and Holocene lake sediments are few. Previous studies of pollen sites in Indiana, South Dakota, and the Great Lakes area show low $\delta^{18}\text{O}$ values during deglaciation, rising to a Hypsithermal peak, and provide paleoenvironmental reconstructions similar to those obtained from pollen studies. Blacktail Pond, located in Douglas fir steppe in northern Yellowstone National Park, Wyoming, is one of the highest elevation lakes (2018 m) yet studied with both pollen and stable isotopes. Analyses of marls yield low oxygen and carbon isotope values at the base of the core probably due to meltwater influx at 12,500 to 14,000 BP. Tundra vegetation persisted for about an additional 1,500 years following the end of meltwater input. Later, more enriched isotope values fluctuate due to the high sensitivity of Blacktail Pond to evaporation and CO_2 exchange because of its shallow depth. These processes result in a covariance between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ related to the residence time of water in the pond; they exert a primary control on the isotopic composition of the Holocene marl. It may be possible to filter the data for residence time effects and extract additional paleoenvironmental information based on the offsets of isotopic data from the $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ trend for a particular pollen zone.

RÉSUMÉ Le comportement des isotopes de l'oxygène et du carbone dans un site pollinique tardiglaciaire holocène du Wyoming, É.-U. Les études sur les isotopes stables présents dans les sédiments lacustres tardiglaciaires et holocènes de l'Amérique du Nord sont peu nombreuses. Les quelques études antérieures dans des sites polliniques de l'Indiana, du South Dakota et de la région des Grands Lacs ont démontré que les valeurs de $\delta^{18}\text{O}$, faibles pendant la déglaciation, s'élevaient jusqu'à l'apogée au cours de l'hypsithermal. Ces études ont de plus permis d'effectuer des reconstitutions paléoenvironnementales semblables à celles que permettaient les études palynologiques. Le Blacktail Pond, situé en milieu de steppe arborée, au nord du parc national Yellowstone, au Wyoming, est un des lacs les plus élevés (2018 m) où l'on a effectué des études palynologiques et isotopiques. À la base des carottes, l'analyse des marnes fait ressortir de faibles valeurs des isotopes de l'oxygène et du carbone, probablement en raison d'une forte alimentation en eaux de fonte de 12 500 à 14 000 BP. La végétation de toundra a subsisté pendant les 1500 années qui ont suivi l'arrêt de la circulation des eaux de fonte. Plus tard, les valeurs plus élevées des isotopes ont varié à cause de la forte évaporation et des échanges de CO_2 qui affectaient plus particulièrement le Blacktail Pond, un lac de faible profondeur. Ces processus ont entraîné une covariance du $\delta^{13}\text{C}$ et du $\delta^{18}\text{O}$ reliée à la durée d'emmagasinement de l'eau dans le lac et ont exercé un contrôle direct sur la composition isotopique des marnes holocènes. On peut arriver à raffiner les données afin de tenir compte des conséquences de la durée d'emmagasinement de l'eau et de tirer des renseignements supplémentaires sur les paléoenvironnements d'une zone pollinique particulière à partir des écarts par rapport au comportement habituel des isotopes du $\delta^{13}\text{C}$ et du $\delta^{18}\text{O}$.

ZUSAMMENFASSUNG Verlauf von Sauerstoff- und Kohlenstoff-Isotopen in einem späteiszeitlichen Pollen-Gelände aus dem Holozän in Wyoming, U.S.A. Es gibt wenige Studien über dauerhafte Isotopen in den nordamerikanischen Seesedimenten der Späteiszeit und des Holozän. Frühere Studien von Pollen-Gebieten in Indiana, Süd-Dakota, und dem Gebiet der großen Seen weisen niedrige $\delta^{18}\text{O}$ Werte während der Enteisung auf, welche zu einem Höchstwert im Hypsithermal ansteigen und Rekonstruktionen der früheren Umweltbedingungen erlauben, welche denen, die durch Pollen-Analysen gewonnen werden, ähnlich sind. Der mitten in einer Douglas Tannen-Steppe im nördlichen Yellowstone Nationalpark, Wyoming, gelegene Blacktail-Pond, ist einer der höchstgelegenen Seen (2018 m), der bisher sowohl auf Pollen sowie auf dauerhafte Isotopen untersucht wurde. Die Analysen der Mergel ergeben niedrige Werte der Sauerstoffund Kohlenstoff-Isotope an der Basis der Bohrsprobe, möglicherweise wegen Schmelzwasserzufluß um 12,500 bis 14,400 v.u.Z. Tundra-Vegetation bestand weiter für etwa 1,500 zusätzliche Jahre nach dem Ende des Schmelzwasser-Zuflusses. Später schwankten die höheren Isotopen-Werte, da der Blacktail-Pond wegen seiner geringen Tiefe sehr empfindlich auf Verdunstung und CO_2 -Austausch reagierte. Diese Prozesse führen zu einer Kovarianz zwischen $\delta^{13}\text{C}$ und $\delta^{18}\text{O}$, welche in Verbindung zu der Dauer des Wasservorkommens in dem See gebracht wird; sie üben eine direkte Kontrolle auf die isotopische Zusammensetzung des Mergel aus dem Holozän aus. Es könnte möglich sein, die Ergebnisse in Bezug auf die Auswirkungen der Dauer des Wasservorkommens zu filtern und auch zusätzliche Informationen über die frühere Umwelt einer bestimmten Pollen-Zone herauszuziehen, mittels der Abweichungen der Isotopen-Data vom $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ Trend.

* Contribution du premier symposium de la CANQUA, sous la direction de René W. Barendregt

INTRODUCTION

Stable isotopes have been extensively used as a paleoenvironmental tool in studies of deep sea sediments, but relatively few studies have applied the technique to lake sediments to provide a record of terrestrial climate. This may be because the relatively small reservoir of water in lakes results in relatively large temporal changes in the temperature and isotopic composition of lake waters. One advantage lake studies have is that isotopic data can be combined with regional pollen studies to develop an integrated paleoenvironmental picture.

In principle the oxygen isotopic composition of lacustrine carbonates is dependent upon the temperature and the isotopic composition of the water. The temperature dependence of ^{18}O fractionation, about -0.2‰ per $^{\circ}\text{C}$ (EPSTEIN *et al.*, 1953), is overshadowed by the temperature dependence of the $\delta^{18}\text{O}$ of rainwater, about $+0.3$ to $+0.7\text{‰}$ per $^{\circ}\text{C}$ (YURTSSEVER and GAT, 1981; DANSGAARD, 1964). The carbon isotopic composition of the lacustrine carbonates should reflect

that of the dissolved inorganic carbon (DIC). Enrichment in ^{13}C occurs when photosynthesis removes ^{13}C -depleted dissolved carbon, or when dissolved CO_2 exchanges with atmospheric CO_2 . The DIC becomes depleted in ^{13}C through the oxidation of organic matter.

Surprisingly few studies have examined the oxygen and carbon isotope ratios of molluscs or marls in cores from North American lakes. STUIVER (1970), in what may have been the first comprehensive isotopic study of lacustrine paleoenvironments, examined cores from lakes in New York and Maine, as well as pollen sites from Indiana (Pretty Lake) and South Dakota (Pickeral Lake). In general, the $\delta^{18}\text{O}$ records from the various sites showed good agreement with each other (Fig. 1) and with ice core data from Greenland. Low $\delta^{18}\text{O}$ values (colder) before about 9000 to 10,000 BP were followed by higher values (warmer) culminating in a "Hypsithermal" interval. A later decrease in $\delta^{18}\text{O}$ was attributed to cooler temperatures during the Late Holocene.

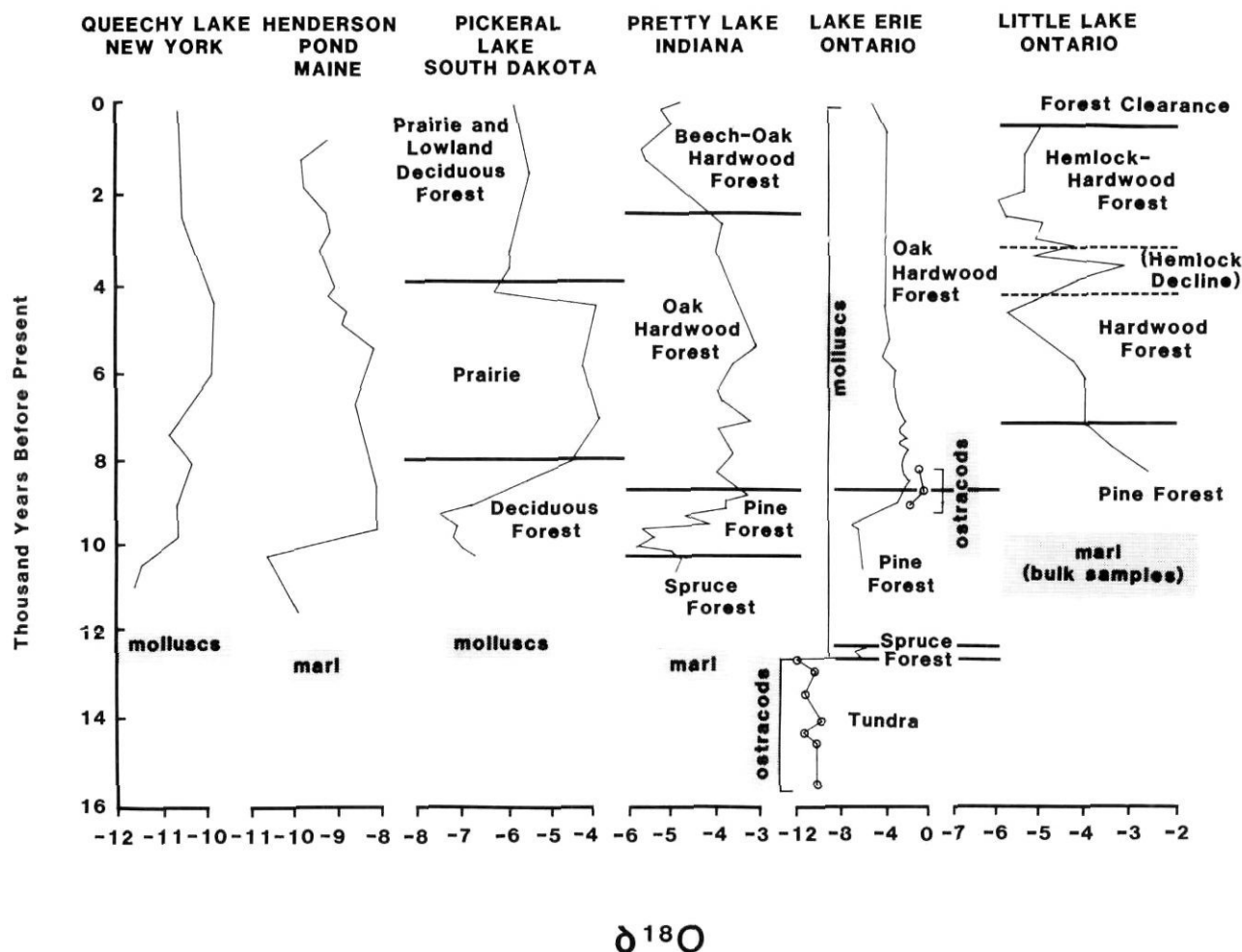


FIGURE 1. Summary of stable isotopic studies of North American lakes with pollen zonations where available (from STUIVER, 1970; FRITZ *et al.*, 1975; TURNER *et al.*, 1983). All isotopic values are in per mil vs. PDB.

Résumé des études sur les isotopes stables dans les lacs de l'Amérique du Nord avec les zones polliniques là où elles ont été établies (selon STUIVER, 1970; FRITZ *et al.*, 1975; TURNER *et al.*, 1983). Les valeurs isotopiques sont données en mille/PDB.

Stuiver's isotope intervals generally coincide with those of the pollen records (WATTS and BRIGHT, 1968; WILLIAMS, 1974). At Pickerel Lake, South Dakota, where the most direct comparison was made, light $\delta^{18}\text{O}$ values before 9000 BP correspond to deciduous forest. A 3‰ increase in $\delta^{18}\text{O}$ at 9000 BP, characterizing the hypsithermal event, is coincident with a change to almost ubiquitous blue-stem prairie. In the most recent section of core, a 2‰ decrease of $\delta^{18}\text{O}$ is accompanied by a shift to abundant oak and ash deciduous forests around the lake, with prairie upland.

In an isotopic study of molluscs and ostracods from Lake Erie, FRITZ *et al.* (1975) attributed low $\delta^{18}\text{O}$ values prior to about 13,000 BP to meltwater influx in addition to isotopically light precipitation (Fig. 1). In light of the absence of evidence of plant life, the high $\delta^{13}\text{C}$ values were attributed to exchange with atmospheric carbon dioxide. The $\delta^{18}\text{O}$ values increase at about 12,700 BP, reflecting either warming or increased evaporation. A decrease in $\delta^{13}\text{C}$ at this time suggests that the DIC $\delta^{13}\text{C}$ lowered due to oxidation of organic matter from more abundant aquatic vegetation. Forestation of the landscape during this period is indicated by a rise in spruce (*Picea*) pollen. Subsequent variations in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ could not be attributed to climate change, but instead were attributed to changes in the rate of evaporation, gas exchange, and residence time of water in the lake.

LERMAN (1974) presented isotopic results for lakes in Switzerland and Germany (Nielsen, unpublished). The isotopic records, after assignment of samples to the determined pollen zones, show very good agreement. Detailed comparison between pollen and $\delta^{18}\text{O}$ diagrams have been made by EICHER and SIEGENTHALER (1976) for two Swiss lakes and by EICHER *et al.* (1981) for Tourbière de Chirens in France. In general, isotopic curves for marls from these lakes correlate well with arboreal pollen vs. non-arboreal pollen curves from the same core, and with one another. Temperature changes interpreted from these $\delta^{18}\text{O}$ and pollen data are very similar, although some fluctuations in $\delta^{18}\text{O}$ are correlated not with the pollen record, but instead with other paleontological evidence for climatic change (*i.e.*, beetles), local hydrologic changes, or the Late Glacial ^{18}O depletion of ocean water. These studies also reveal a short lag between isotopic and vegetational changes.

The glacial to interglacial shift in $\delta^{18}\text{O}$ often serves as a datum to complement dates obtained with radiocarbon. Where sediments of glacial age are not recovered, isotopic paleoenvironmental studies must depend solely on ash and radiocarbon dates for the determination of the timing of climatic events. This can sometimes prove difficult, as in the study of FRITZ and KROUSE (1973). These authors measured the $\delta^{18}\text{O}$ values of Holocene ostracods and molluscs from Wabamun Lake, Alberta; these isotopic results were later compared with pollen data from the same lake (HOLLOWAY *et al.*, 1981; HICKMAN *et al.*, 1984). The most enriched isotopic values probably coincide with the occurrence of ditch-grass (*Ruppia*), which Hickman *et al.* interpreted as a hypsithermal, saline lake indicator. Dating of this event, however, has proved difficult because possible contamination by lignite has caused the

entire chronology of the core to be disputed, including the date for an ash layer previously identified as Mt. Mazama, which overlies the isotopic event.

TURNER *et al.* (1983) have suggested that the isotopic trends in lake marls may at times record the residence time of the water in the lake rather than the temperature. With long residence times, gas exchange and evaporation combine to enrich the DIC in ^{13}C and the water in ^{18}O . With short residence times, the water retains the original $\delta^{18}\text{O}$ of the meteoric water and the $\delta^{13}\text{C}$ of the soil gas CO_2 initially dissolved in the water entering the lake. These processes result in a covariance between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in carbonates precipitated in the lake, as summarized by Figure 2. Turner *et al.* cited the significant degree of covariance between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in the isotopic record of marls and molluscs from Little Lake, Ontario, to support the model, and used it to interpret radiocarbon dates. Such a trend was also noted by STUIVER (1970) for a core from Pretty Lake, Indiana. He considered the possibility that changes in evaporation rate were responsible for the trend, not unlike the residence time model of TURNER *et al.* (1983).

To provide a better understanding of climate change in western North America, we have combined stable isotopic and pollen records from a core taken from Blacktail Pond, Yellowstone Park, Wyoming. Although this pond is one of the lower elevation lakes (2018 m) used in palynological studies of Yellowstone Park, it is as far as we know the highest elevation lake to provide a detailed isotope stratigraphy.

The pollen record of late Quaternary climate change in Yellowstone Park is probably as complete as that of any

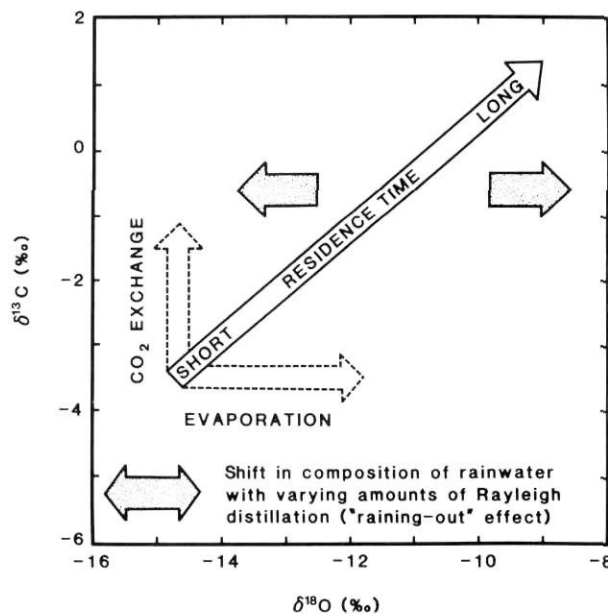


FIGURE 2. Effect of variations of residence time of lake water on isotopic composition of carbonates. Isotopic values vs. PDB.

Les conséquences de la durée d'emmagasinement de l'eau dans les lacs sur la composition isotopique des carbonates. Les valeurs isotopiques sont données en mille/PDB.

region in the Rocky Mountains (WADDINGTON and WRIGHT, 1974; BAKER, 1976, 1983; GENNETT and BAKER, 1986). These data suggest that during the Holocene, climatic change at various elevations in Yellowstone Park was not synchronous, and that the "Xerithermal" interval occurred at much earlier dates in areas of the park presently covered by subalpine forest. Previous to this study no late Quaternary paleoclimate data exist from this area other than pollen and plant macrofossil records.

STUDY AREA

Blacktail Pond is located in the northernmost section of Yellowstone Park in northwestern Wyoming, U.S.A. (Fig. 3). The valley was originally a meltwater channel during the Late Wisconsinan. The pond originated as a kettle and has been filling with sediment since Late Glacial times. At present it has an average depth of 3.5 m and a maximum depth of 8 m. Lacustrine sediments cover a large area around the modern Blacktail Pond, indicating higher lake levels in the past. Springs to the north and runoff supply water for the pond (Fig. 3; JONES *et al.*, 1984). The pond water is a calcium-bicarbonate type with a pH of 8.4. The lake supports a dense aquatic plant community which can condition the water for CaCO_3 precipitation. *Chara* sp. is the most abundant submerged plant and covers the substrate in much of the littoral zone. Calcareous deposits have been observed on *Chara* sp.; such deposits may be the source for the carbonate found in the marls analyzed in this study. However, no oögonia (female sex organs), common in *Chara*-derived carbonate, were observed in our samples.

Blacktail Pond is presently surrounded by sagebrush (*Artemisia*) steppe with mostly Douglas fir (*Pseudotsuga*) open forest immediately to the north, and a mixture of Douglas fir, pine (*Pinus* spp.), and spruce (*Picea* sp.) on the southern slope of the valley.

METHODS AND MATERIALS

The Blacktail Pond core was taken on a peat surface using a Livingstone piston corer. The base of the core consists of 5 cm of sand overlain by 310 cm of marl, and capped by 49 cm of peat (Fig. 4). An upper ash (236 cm) was identified as the Mt. Mazama (6700 BP), and a lower ash (450 cm) was dated at 14,360 BP (W-2780) in accordance with WADDINGTON and WRIGHT's (1974) date on a correlative ash at Cub Creek Pond. A date of 1320 BP (DIC-670) was obtained for wood from the 39-42 cm interval. Marl at the 360 to 370 cm interval was dated at 11,780 BP (DIC-669). It is possible that "old carbon" from calcareous tills caused the ^{14}C age of this level to be underestimated. Comparison of pollen sequences from within the park, however, indicates that the date is fairly accurate (GENNETT and BAKER, 1986).

Sediment was sampled at 5 or 10 cm intervals when possible. Pollen procedures are described in GENNETT and BAKER (1986). Procedures for isotopic analysis are as follows. Samples of one to several milligrams were roasted *in vacuo* at 380°C for one hour and reacted in "100%" phosphoric

acid at 50°C. The CO_2 evolved was analyzed for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ using a Finnigan MAT 251 Isotope Ratio Mass Spectrometer. All isotopic values are reported in δ -notation vs. the PDB standard. Replicate analyses were performed on 25% of the samples; the average precision (one standard deviation) is $\pm 0.22\text{‰}$ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ measurements. The high standard deviation for marl analyses reflects sample heterogeneity rather than analytical precision, which averages ± 0.09 and $\pm 0.13\text{‰}$ for carbon and oxygen isotopic analyses of carbonate standard.

Per cent carbonate was measured in duplicate by EDTA titration (TUREKIAN, 1956). For 75% of the samples, the average difference between duplicates was 2.2%. In the remaining 25%, agreement between duplicates was not as good, due either to interfering ions which altered the indicator

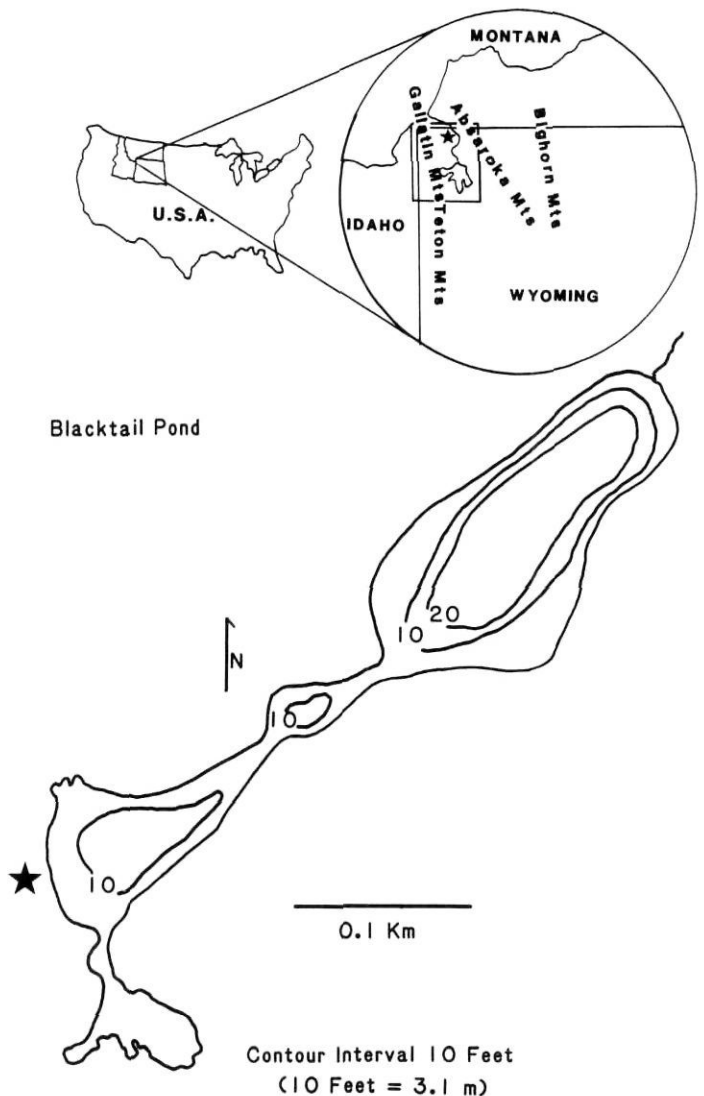


FIGURE 3. Location and bathymetric map of Blacktail Pond (from JONES *et al.*, 1984). Core location is shown by a star.

Carte de localisation et bathymétrie du Blacktail Pond, selon JONES *et al.*, 1984. L'étoile identifie l'emplacement de la carotte.

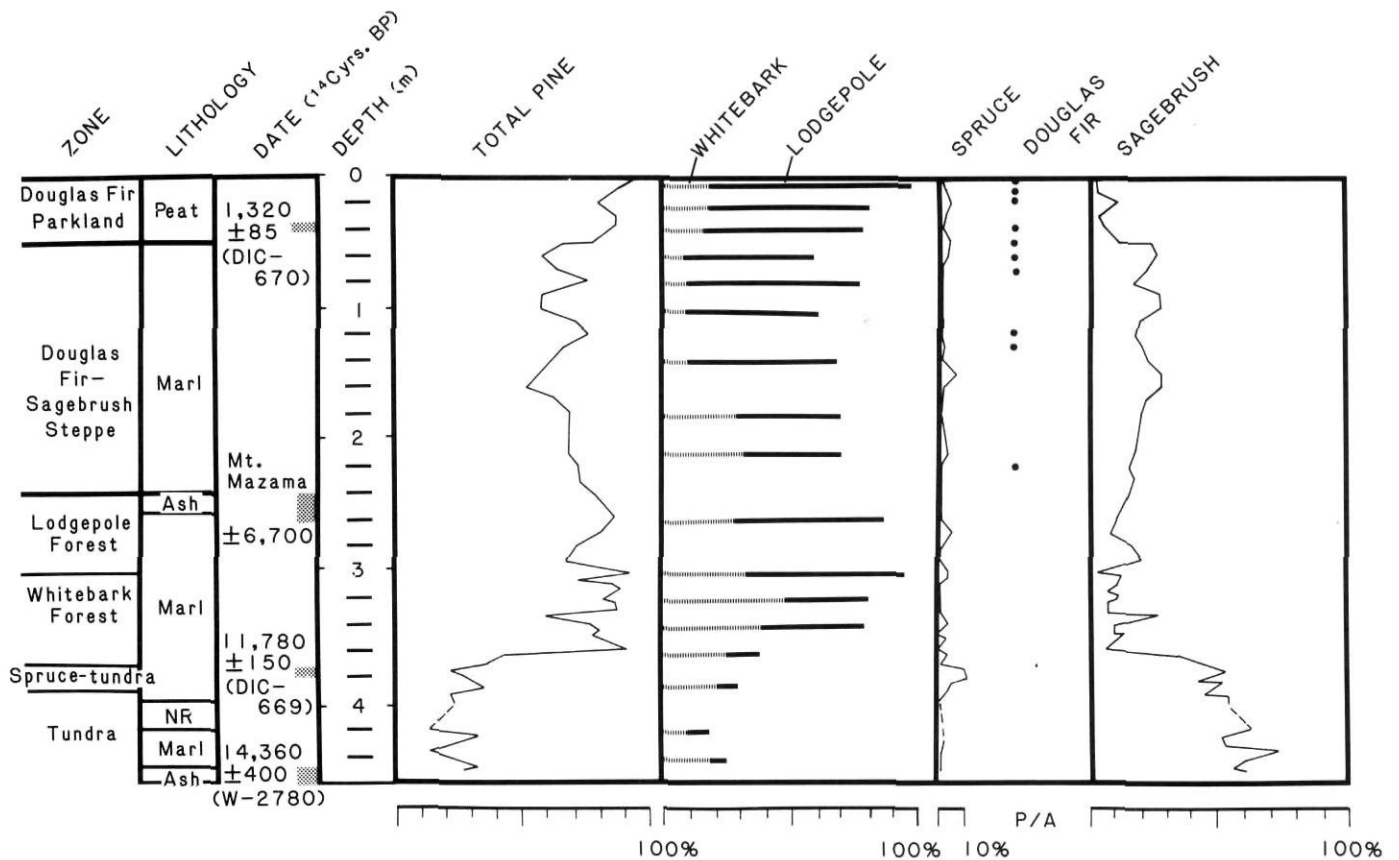


FIGURE 4. Simplified pollen percentage diagram for the Blacktail Pond core. Detailed diagrams are contained in GENNETT and BAKER (1986). NR = No recovery.

Diagramme pollinique en pourcentage simplifié de la carotte de Blacktail Pond. L'article de GENNETT et BAKER (1986) comprend les diagrammes détaillés. NR = perte de carotte.

color, or to sample heterogeneity. In these instances, additional replicates were performed, with a mean precision of ± 8.4 %. Using standard techniques, x-ray diffraction analyses were performed on more than half the samples to test for contamination of the calcitic marl by aragonitic shell fragments.

THE POLLEN RECORD

The pollen record at Blacktail Pond was described in detail by GENNETT and BAKER (1986; Fig. 4). Following deglaciation at around 14,600 BP, tundra vegetation covered the Blacktail Pond area. Subsequent spruce-tundra parkland was invaded by whitebark pine-spruce forest at about 11,800 BP. By 6700 BP, the forest was dominated by lodgepole pine. At that time Douglas fir and sagebrush steppe took the place of forest vegetation. Finally, lodgepole pine became more important in the record after 2000 BP.

Composite paleovegetational maps of Yellowstone Park (Fig. 5) were derived from pollen diagrams of Blacktail Pond (GENNETT and BAKER, 1986), Gardner's Hole (BAKER, 1983), Buckbean Fen (BAKER, 1976), and Cub Creek Pond (WADDINGTON and WRIGHT, 1974); they provide a qualitative estimate of extent of paleocommunities based on elevation. The maps indicate that around 11,000 years ago, forest above 2,500 m was composed of subalpine spruce and

fir, with lower elevations covered by mixed lodgepole and whitebark pine forests. By 7000 BP the whole park except for the lowest elevations was vegetated by lodgepole pine. Douglas fir and steppe expanded in the low elevation sites in the northern portion of the park contemporaneous with enlarged subalpine areas at high elevations by 5000 BP. DESPAIN's (1973) map of present-day vegetational zonation suggests further increase of spruce fir/subalpine areas principally at the expense of lodgepole pine areas.

ISOTOPIC RESULTS AND DISCUSSION

Isotopic compositions and per cent carbonate data are given in Figure 6. The oxygen isotopic compositions, ranging from - 16.9 to - 10.5 ‰, are among the most negative recorded for a lacustrine marl and reflect the temperate latitude and high altitude of the pond. The carbon isotopic compositions, ranging from - 2.7 to 2.5 ‰, are within the range of other North American marls analyzed by STUIVER (1970).

The isotopic values of the 430-449 cm interval are the lowest in the core. Probably at this time Blacktail Pond was larger and deeper and experienced less exchange with the atmosphere. Oxygen isotopic compositions near - 17 ‰ may reflect glacial meltwater input as well as cooler temperatures. The low carbonate content (40-60 %) during this period in-

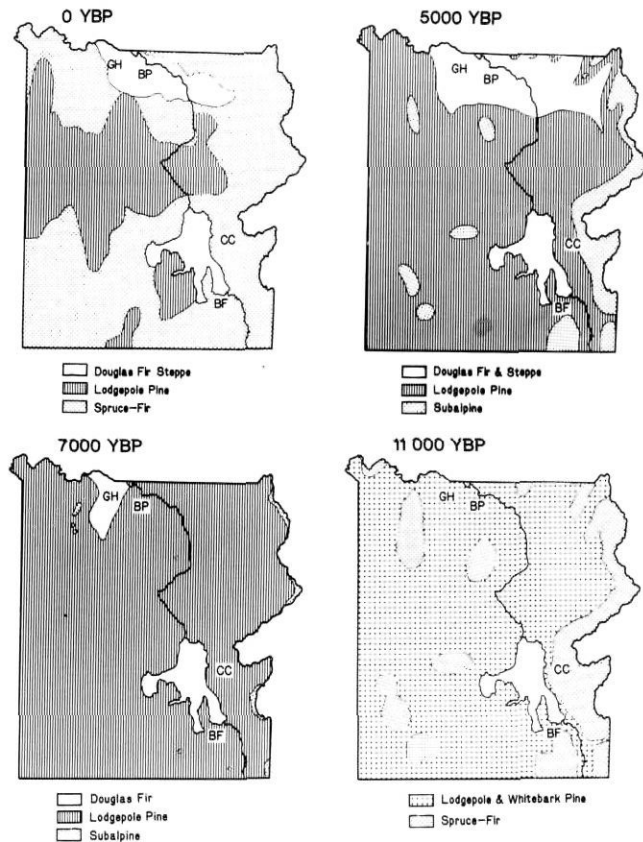


FIGURE 5. Composite paleovegetational maps of Yellowstone National Park, Wyoming, based on data from WADDINGTON and WRIGHT (1974), BAKER (1976, 1983) and GENNETT and BAKER (1986). BP, Blacktail Pond; CC, Cub Creek Pond; GH, Gardner's Hole; BF, Buckbean Fen. (0 YBP map after DESPAIN, 1973.)

Cartes de la paléo-végétation du parc national Yellowstone, au Wyoming, fondées sur les données de WADDINGTON et WRIGHT (1974), de BAKER (1976, 1983) et de GENNETT et BAKER (1986).

indicates low carbonate productivity in relation to non-carbonate clastic influx (Fig. 6), probably due to low temperatures and the influx of glacial meltwater. These low isotopic compositions and carbonate contents coincide with early tundra development. Later tundra samples share the low carbonate contents, indicating low productivity, but exhibit increases in $\delta^{18}\text{O}$ to about -13‰ , and in $\delta^{13}\text{C}$ to about -1‰ at 385 cm. This suggests a lag between carbonate productivity and paleoenvironmental change of about 1500 years.

In the upper part of the core, isotope data are highly variable. Carbonate content is quite high (around 80 %) indicating substantial productivity and/or low clastic input.

The isotopic record of Blacktail Pond is considerably more erratic than those of other North American lakes (STUIVER, 1970; FRITZ *et al.*, 1975; TURNER *et al.*, 1983; Fig. 1). This may be due to the pond's small size and high altitude. An extensive marshy area as well as mudcracked, marly areas surrounds the pond, and the shallow depth may have led to large scale evaporation during periods of drought, which would have resulted in increases in both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$, as shown in Figure 2. Drought may have also affected the proportion

of water derived from groundwater, snowmelt, and rainfall, and the elevation of these sources. In addition, subaerial exposure may have occurred, although we have observed no evidence for such an event.

The persistence of tundra after the cessation of significant influx of meltwater may have been caused by slow migration of trees into the Blacktail Pond Valley. The Blacktail Pond area was deglaciated before areas to the east (PIERCE, 1979), and meltwater drainage may have shifted to streams in that direction before the climate and/or soil became favorable for tree growth.

Plotted on a $\delta^{13}\text{C}$ vs. $\delta^{18}\text{O}$ scatter diagram (Fig. 7), the carbon and oxygen isotopic data can be seen to covary. This covariance was also observed in replicates. As mentioned earlier, TURNER *et al.* (1983) suggested that such a correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ can represent variations in the residence time of water in the lake. Furthermore, FRITZ and POPLAWSKI (1974) noted that the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of lake waters and DIC, respectively, from samples collected at different times of the year in a number of lakes, show this same correlation; the range in isotopic compositions is even larger than that observed in carbonates in this study.

Alternatively, the $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ correlation may result from admixture of an additional calcium carbonate source. Two likely possibilities are detrital carbonate and shell carbonate. Detrital carbonate, along with non-carbonate detritus, occur in glacial till in the Blacktail Pond area. We examined samples microscopically for detrital grains and found none except in the 460 cm sample, which contains a large portion of clastics and is probably glacial outwash. Because of the association with non-carbonate detritus, if detrital carbonate did influence the isotopic composition of the Holocene marls, there would be a correlation between carbonate content and isotopic composition. Correlation coefficients for $\delta^{18}\text{O}$ vs. percent carbonate and $\delta^{13}\text{C}$ vs. percent carbonate for Holocene samples are -0.16 and 0.07 ($N = 27$), respectively, indicating no relationship between isotopic composition and percent carbonate.

Isotopic analyses were performed on the gastropods *Physa* and *Gyraulus* from selected intervals to test the effect of "contamination" on the marl by fine-grained shell fragments. Table I shows that relative to the marl, the shells have about the same $\delta^{18}\text{O}$, but are depleted in ^{13}C by as much as 11 ‰. Thus the correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in the Blacktail Pond data could not result from the admixture of shell material to the marl, however, this admixture could produce the offset from the general isotopic trend as best represented by the Douglas fir zone samples (Fig. 7). As a test for shell material in the marl, samples were inspected for shell fragments with a microscope, and analyzed using x-ray diffraction (XRD). No shell fragments were observed microscopically with the exception of some small fragments in the 60 cm sample. Note that the isotopic compositions of the 60 cm sample do not differ substantially from those of the intervals above and below (Fig. 6). Because the marl is calcitic and the shells of gastropods like *Physa* and *Gyraulus* aragonitic, XRD can be used to detect minute shell fragments that might not be recognized under the microscope. We performed this analysis on more

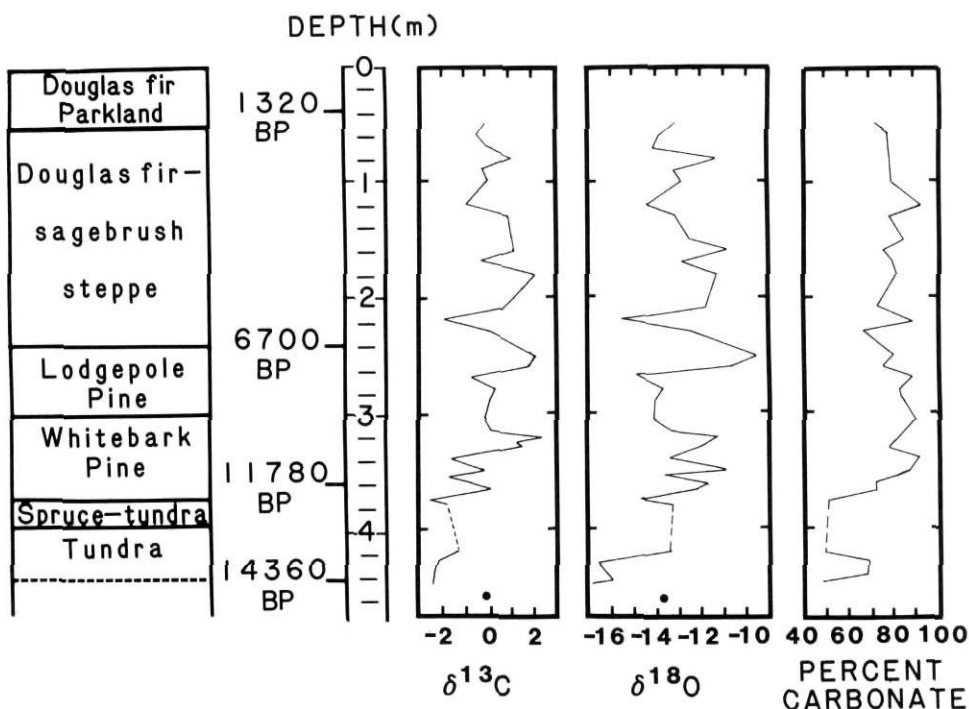


FIGURE 6. $\delta^{13}C$, $\delta^{18}O$, and percent carbonate curves for Blacktail Pond. The isotopic data for the outwash sample (460 cm) are represented by dots. All isotopic data are relative to PDB.

$\delta^{13}C$, $\delta^{18}O$ et les courbes du pourcentage de carbonate contenu dans les marnes du Blacktail Pond. Les données isotopiques de l'échantillon de matériaux fluvio-glaciaires (460 cm) sont représentées par des points. Les valeurs isotopiques sont données en mille/PDB.

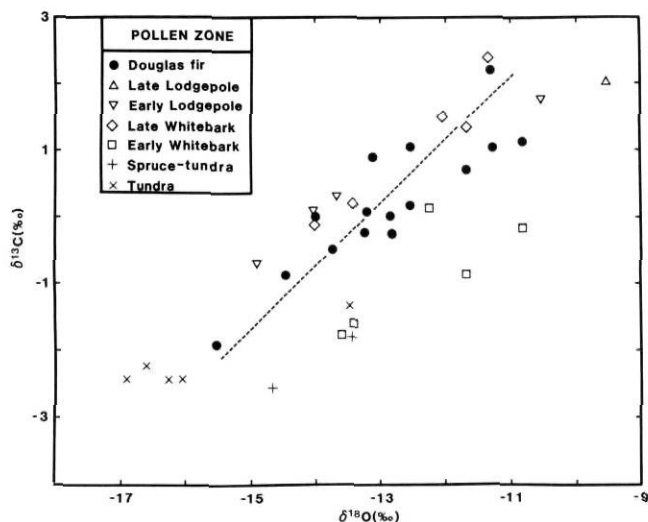


FIGURE 7. $\delta^{13}C$ versus $\delta^{18}O$ scatter diagram of the average compositions of Blacktail Ponds marls. All isotopic data are vs. PDB.

Diagramme de dispersion du $\delta^{13}C$ par rapport au $\delta^{18}O$ montrant la composition moyenne des marnes du Blacktail Pond. Les valeurs isotopiques sont données en mille/PDB.

than half of the marl samples and found no aragonite with the exception of the sample at 60 cm. In conclusion, we believe that the influence of detrital carbonate and shell material on the isotopic signature of the marls is insignificant, and that the $\delta^{13}C$ - $\delta^{18}O$ correlation in Holocene samples reflects variations in the residence time of water in the pond throughout its history.

Offsets in the $\delta^{13}C$ - $\delta^{18}O$ data trend ($\Delta^{13}C_{Df}$, $\Delta^{18}O_{Df}$), as observed in the Spruce-tundra and early Whitebark zone samples (Fig. 7), may reflect the influence of a second en-

TABLE I

Isotopic compositions of *Physa* and *Gyraulus* from the Blacktail Pond core (all values are relative to PDB)

Interval (cm)	Genus	$\delta^{13}C$ (‰)	$\delta^{18}O$ (‰)
190	<i>Physa</i>	-8.8	-13.0
290	<i>Gyraulus</i>	-5.0	-9.3
		-7.2	-10.3
330	<i>Physa</i>	-10.1	-13.0
		-8.8	-12.9
	<i>Gyraulus</i>	-8.2	-12.5
340	<i>Gyraulus</i>	-7.9	-12.2
350	<i>Physa</i>	-9.4	-12.6
	<i>Gyraulus</i>	-7.5	-12.8
435	<i>Gyraulus</i>	-7.4	-14.2
		-10.0	-13.5

vironmental factor. These offsets can be determined from the regression of the Douglas fir data:

$$\delta^{18}O_{Df} = -13.16 + 1.05 \delta^{13}C_{Df} \quad (r = 0.88, N = 15),$$

and the equations:

$$\Delta^{18}O_{Df} = \delta^{18}O_x - \delta^{18}O_{Df}$$

$$\text{and } \Delta^{13}C_{Df} = \delta^{13}C_x - \delta^{13}C_{Df}$$

where $\delta^{18}O_x$ and $\delta^{13}C_x$ are the isotopic compositions of the sample. As shown in Figure 8, this treatment of the data produces an interesting pattern. Tundra zone samples below 425 cm are offset toward depleted $\Delta^{18}O_{Df}$ (or enriched $\Delta^{13}C_{Df}$) values. Between 425 and 335 cm, which includes late Tundra, Spruce-tundra, and early Whitebark Pine zones, only positive $\Delta^{18}O_{Df}$ values are obtained. The $\Delta^{18}O_{Df}$ values become negative

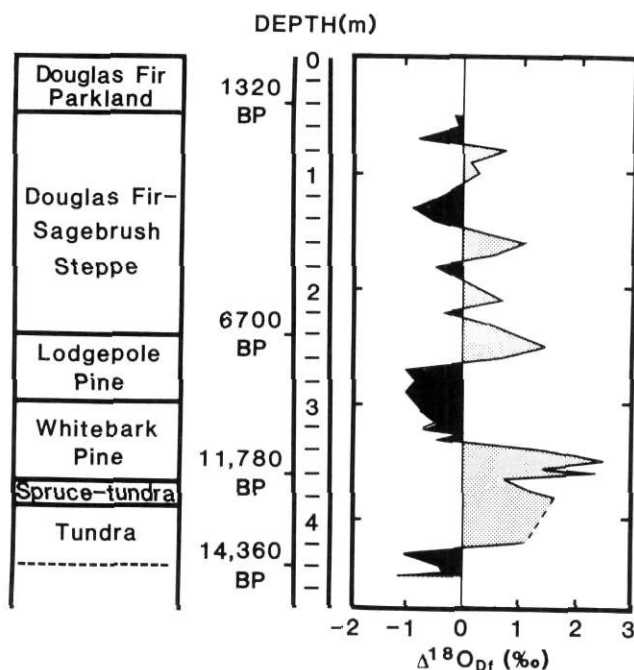


FIGURE 8. Deviation of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ data from the Douglas fir-steppe regression line ($\Delta^{18}\text{O}_{\text{Df}}$ and $\Delta^{13}\text{C}_{\text{Df}}$).

Déviations du $\delta^{18}\text{O}$ et du $\delta^{13}\text{C}$ de la droite de régression établie à partir des données de la zone pollinique à sapin Douglas ($\Delta^{18}\text{O}_{\text{Df}}$ et $\Delta^{13}\text{C}_{\text{Df}}$).

above 335 cm and remain negative during the late Whitebark and early Lodgepole zones, before shifting to positive during late Lodgepole at about 265 cm. Above 265 cm, the $\Delta^{18}\text{O}_{\text{Df}}$ values fluctuate between positive and negative with a higher frequency than in the lower half of the core.

But what factors could produce the observed $\Delta^{18}\text{O}_{\text{Df}}$ (and $\Delta^{13}\text{C}_{\text{Df}}$) pattern? As a first order approximation, $\delta^{18}\text{O}_{\text{Df}}$ and $\Delta^{13}\text{C}_{\text{Df}}$ will not be influenced by evaporation effects since an increase in the residence time of the water and a concomitant ^{13}C enrichment due to increased CO_2 exchange will be associated with evaporation-derived ^{18}O enrichment. Thus on a $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ plot, evaporation will cause variation along the residence time trend (Fig. 2), but not necessarily offsets from it. To produce offsets from that trend like those of the Spruce-tundra and early Whitebark Pine zone samples, one can call upon changes in the initial $\delta^{18}\text{O}$ of the water entering the lake, and/or changes in the $\delta^{13}\text{C}$ of the DIC in that water; that is, mechanisms which allow $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ to vary initially and independently.

The $\delta^{18}\text{O}$ of the precipitation may change due to temperature change or to greater raining out (distillation) of the meteoric water (DANSGAARD, 1964; YURTSEVER and GAT, 1981); the latter would be expected during drier periods. However, for these factors to cause an offset from the Douglas fir trend and not produce shifts in the $\delta^{18}\text{O}$ record, we must make the tenuous assumption that the residence time of pond water is not related to temperature or dryness. According to this interpretation, positive $\Delta^{18}\text{O}_{\text{Df}}$ values would result from moist or warm climate, while negative values would result from cool

or dry climate (again, evaporation effects are assumed to affect $\delta^{18}\text{O}$ but not $\Delta^{18}\text{O}_{\text{Df}}$). In the Blacktail Pond core, positive $\Delta^{18}\text{O}_{\text{Df}}$ values are coincident with the Spruce-tundra zone, indicative of moist climate and supporting our model (Fig. 8). Furthermore, the shifts to negative $\Delta^{18}\text{O}_{\text{Df}}$ values occurs at approximately the same time as a xeric period in the Northwest, including sites in Washington (BARNOSKY, 1984) and British Columbia (MATHEWES, 1985), and as evidenced by the arrival of *Pinus contorta* forest at Buckbean Fen in Yellowstone Park (BAKER, 1976). This relative aridity persisted until about 5000 BP at Buckbean Fen, but was over by about 7000 BP in the Pacific Northwest, somewhat synchronous with the shift to positive $\Delta^{18}\text{O}_{\text{Df}}$ values and the beginning of the Douglas fir zone. An interpretation of negative $\Delta^{18}\text{O}_{\text{Df}}$ values as indicators of dry climate would also be consistent with the occurrence of pioneer communities of lodgepole pine during the early Holocene at Blacktail Pond (Fig. 8) in response to increased fire frequency due to dryness (DAUBENMIRE, 1978).

The alternate view is that the offsets from the covariant $\delta^{18}\text{O}$ - $\delta^{13}\text{C}$ trend, as represented by the Douglas fir zone data, result from shifts in the $\delta^{13}\text{C}$ of the DIC in the water entering the pond. FRITZ *et al.* (1975) suggested that the carbon isotopic composition of DIC in lake carbonates can vary inversely with the abundance of organic matter available for decay. We find no evidence for a significant relationship between $\Delta^{13}\text{C}_{\text{Df}}$ and productivity, whether recorded as percent carbonate or as pollen concentration. However, it is not unlikely that if climate were moist during the 335-425 cm interval, more organic matter would be available for decay, resulting in the observed negative $\Delta^{13}\text{C}_{\text{Df}}$ values (Fig. 8). Conversely, organic matter for decay would be scarcer during dry periods, and thus $\Delta^{13}\text{C}_{\text{Df}}$ would be positive.

The approach of using the offset from the dominant $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ trend in the Douglas fir zone samples to filter the data for residence time effects is admittedly speculative. However, in cases where the lake may experience fluctuating volume and the water significant changes in residence times, our technique may permit extraction of useful paleoenvironmental information.

CONCLUSIONS

Low oxygen (-17‰) and carbon (-2.5‰) isotope values at the base of calcium carbonate-rich sediments at Blacktail Pond represent meltwater influx and low temperatures contemporaneous with tundra vegetation. Isotopic values indicate an end to significant meltwater influx at around 13,000 BP; however, tundra vegetation persisted until about 12,000 BP.

Isotopic values in the upper part of the core fluctuate greatly, and show a strong covariance between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ which we believe reflects the influence of variations in the residence time of pond water on isotopic composition. As residence time increases, $\delta^{18}\text{O}$ increases due to evaporation, while $\delta^{13}\text{C}$ increases due to increased dissolved CO_2 equilibration with atmospheric CO_2 . The fluctuating values at Blacktail Pond are probably due to variable influx from groundwater and snowmelt during short-term climatic fluctuations, combined

with the fact that the lake is sensitive to outside influences because of its small size.

It may be possible to use offsets in $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ trends to filter the isotopic data for residence time effects, and provide additional paleoenvironmental information.

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