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EVIDENCE FOR AN ABRUPT CHANGE IN CLIMATE CLOSE TO 11,000 YEARS AGO*

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ABSTRACT. Evidence from a number of geographically isolated systems suggests that the warming of world-wide climate which occurred at the close of Wisconsin glacial times was extremely abrupt. Surface ocean temperatures in the Atlantic Ocean increased by several degrees centigrade and deep-sea sedimentation rates sharply decreased in the equatorial Atlantic. The bottom waters of the Cariaco Trench, a restricted basin in the Caribbean, abruptly stagnated. The pluvial lakes in the Great Basin rapidly shrank from their maximum volume to nearly their present size. Silty clay which was flushed into the Gulf of Mexico during the glacial period was suddenly retained in the alluvial valley and delta of the Mississippi River and the glacially-dammed Great Lakes drainage was returned to the northern outlet. As indicated by pollen profiles a pronounced warm period in northwestern Europe also occurred at this time. In each case the radiocarbon age determinations suggest that the changes occurred in less than 1000 years close to 11,000 years ago.

INTRODUCTION

The problem of reconstructing the variations in climate which characterize the Pleistocene has long challenged the geologist. Although the collection, correlation and interpretation of data has progressed steadily during the past 100 years, the recent development of geochemical methods for both time and temperature measurement have greatly accelerated the progress in this field. The C^{14} method of age determination in particular has provided not only absolute estimates of time over the past 40,000 years but has also allowed precise correlations to be made between events in widely separated geographic areas and in vastly different depositional environments.

This paper summarizes the evidence for an abrupt world-wide change in climate close to 11,000 years ago marking the end of the Wisconsin glacial period. That this climatic change occurred suddenly was first recognized by Ewing as a result of the study of deep-sea cores (Ewing and Donn, 1956). Subsequently radiocarbon measurements on these cores were presented by Ericson, et al. (1956). The study has since been extended to other climate-controlled systems.

The evidence to be summarized is largely correlated on the basis of radiocarbon measurements. These include not only those reported in "Lamont Natural Radiocarbon Measurements" (Broecker, et al., 1956, and Broecker and Kulp, 1957) but also the results published by other laboratories. Since the period of interest spans less than 2000 years, careful attention must be given not only to the laboratory errors, which range between 100 and 500 years, but also to the possibility of systematic errors for samples of different material types and growth environments. An attempt will be made to evaluate these factors as the discussion proceeds.

The systems studied include deep-sea sediments from the Atlantic Ocean and adjacent seas, deposits from the pluvial lake area of the western U. S. A., sediments from the Great Lakes and their associated drainage networks and

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the pollen sequences of northwestern Europe. Each of these systems shows a strong dependence on climate.

EVIDENCE FROM DEEP-SEA SEDIMENTS

Surface temperatures.—The first and most important of these systems is the ocean. Its vast extent and fairly rapid internal mixing rate enable it to record world-wide climate changes, being less influenced by local events than any other system. This broad picture is recorded in a number of different ways. Three of these have been investigated in the course of this study: 1) variations in surface ocean temperature, 2) changes in the rate of deep-sea sedimentation, 3) stagnation of restricted basins.

A record of surface ocean temperature is provided by the tests of planktonic foraminifera which make up the majority of the coarse fraction ($>74\mu$) of the carbonate material in globigerina ooze. This record can be read by either the micropaleontological method used by Ericson and Wollin (1956) or the oxygen-isotope method used by Emiliani (1955). The first method is based on the temperature dependence of the relative abundance of certain species of planktonic foraminifera (Schott, 1935). By observation of the relative abundance of these species an estimate of the relative surface water temperature for the time corresponding to any depth in the core can be made. The second method relies on the temperature dependence of the oxygen-isotope fractionation that occurs during the formation of carbonate. The isotopic composition of the oxygen from carbonate samples can be measured with a precision equivalent to 1°C (Urey, 1951). The uncertainty in the absolute temperature computed from these ratios is somewhat larger than 1°C because the oxygen-isotope ratio for the sea water from which the carbonate was precipitated must be estimated from other considerations. For example, Emiliani (1955) has pointed out that the preferential storage of O^{16} in glacial ice could have caused a significant change in the ratio for seawater during Wisconsin time (equivalent to at least 2°C change in temperature).

As shown in figure 1 the two methods have shown excellent agreement where they have both been applied to the same core. Both methods show a rather sharp increase from the relatively cool temperatures of Wisconsin time to the warm temperatures of the present. The magnitude of this change as given by the paleotemperature method is 6° to 10°C (Emiliani, 1955), depending on the assumptions made about the glacial storage effect.

Since the span of time required for the temperature change is critical to the arguments which follow, it is important to consider what processes might alter the true shape of the curve. The curve might conceivably be sharpened by erosion or removal by slumping. Such effects should be apparent as sudden changes in lithology or as time discontinuities. Careful inspection has not revealed either of these effects associated with the temperature transition in the cores of interest. On the other hand, as pointed out by Emiliani (1957), post-depositional mixing of the sediment by bottom organisms would tend to spread out the change. Such activity is known to occur but is difficult to evaluate quantitatively. It may be concluded that the temperature transition is at least

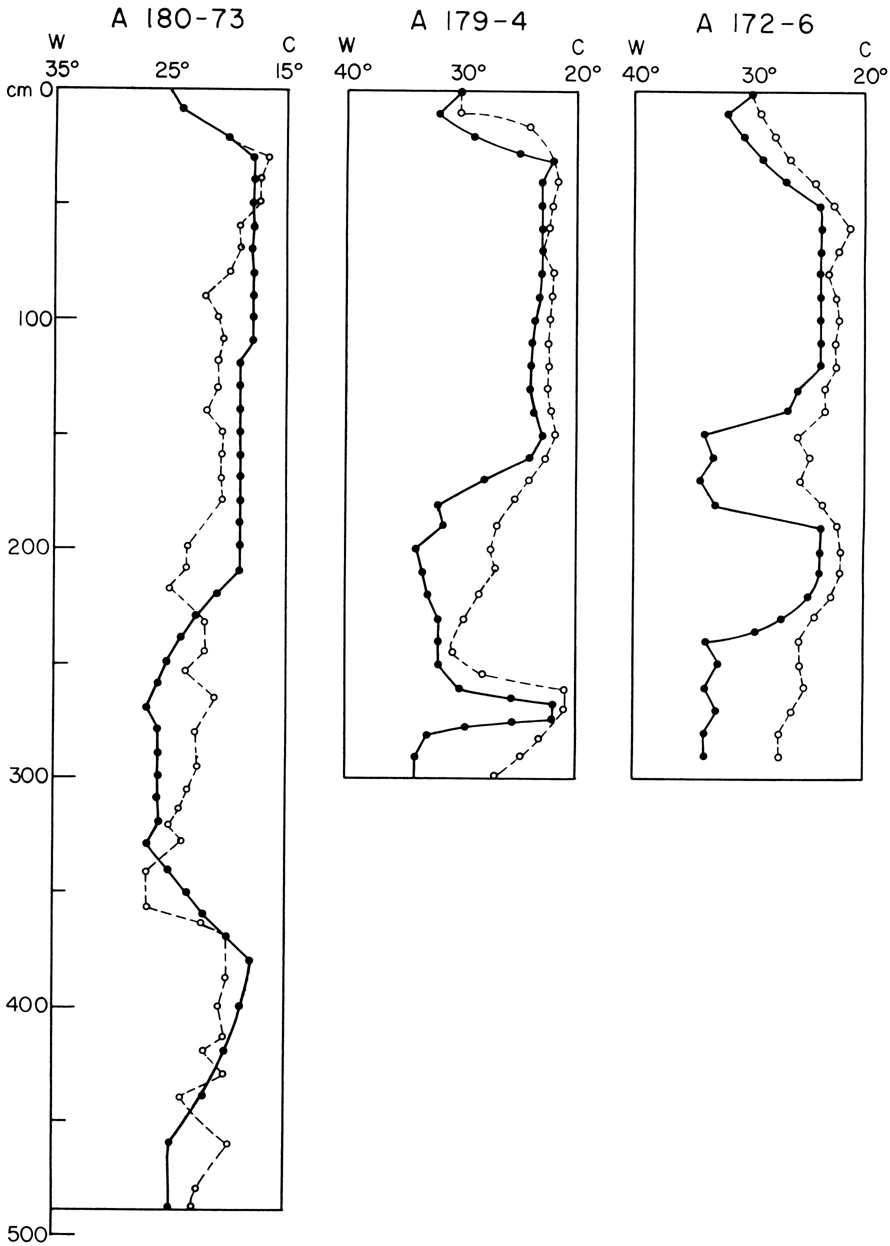


Fig. 1. Comparison of surface ocean temperature curves based on foraminifera abundance (Ericson and Wollin, 1956a and 1956b; solid curves, warm-cold scale) with those based on oxygen isotope measurements (Emiliani, 1955; dotted curves, degrees *centigrade* scale).

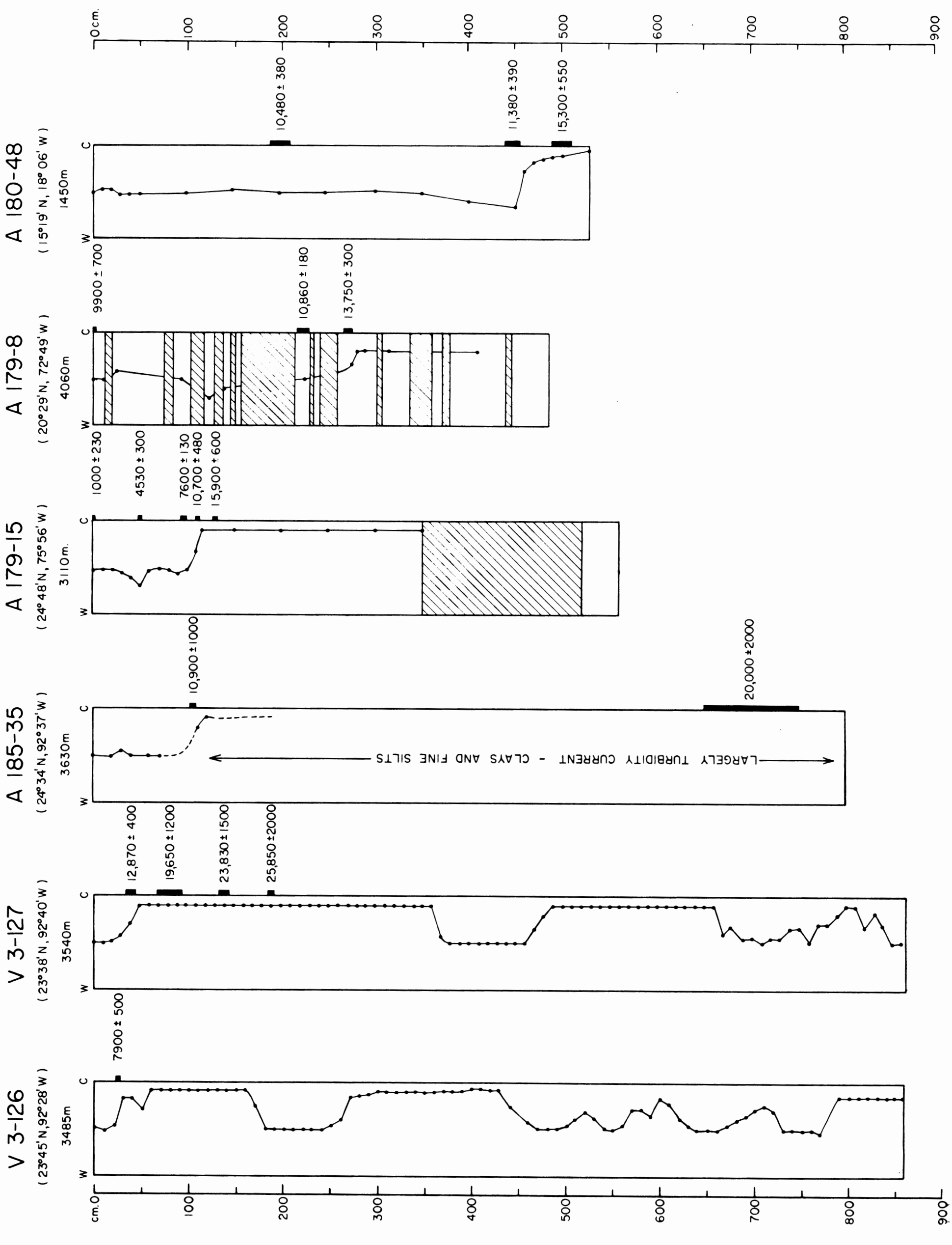
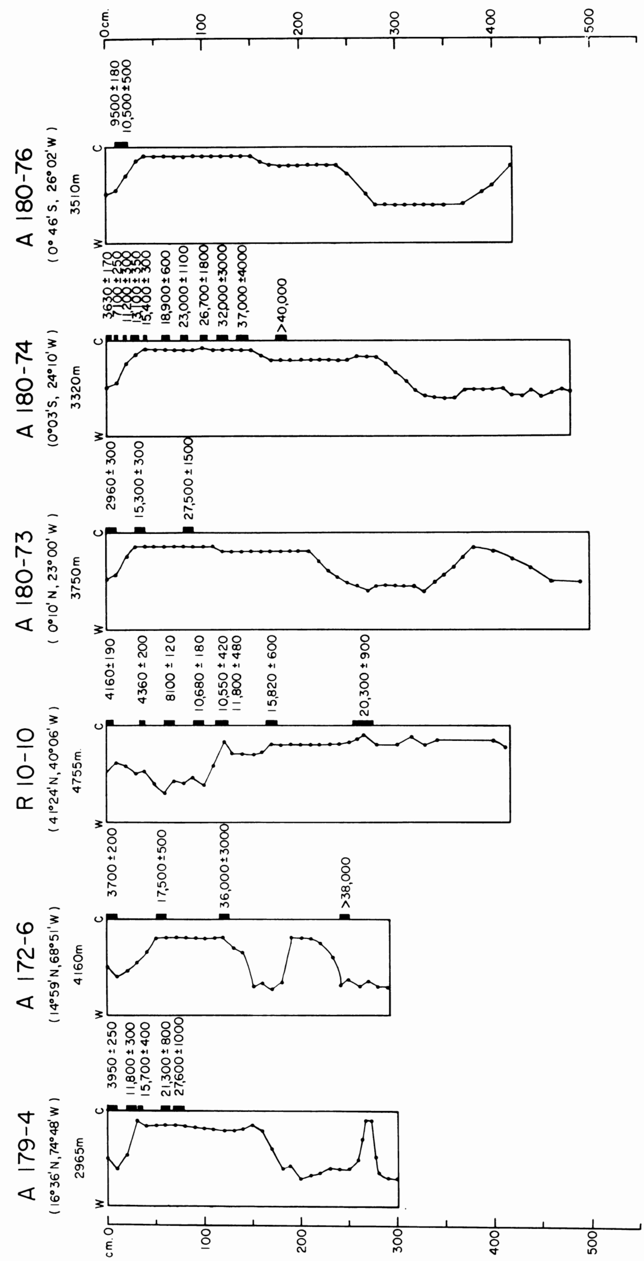


Fig. 2. Foraminifera temperature curves and radiocarbon ages for deep sea cores. The climate curves were constructed by Ericson (Ericson and Wollin, 1956a and b; Ewing,



et al., 1958). The radiocarbon ages on cores A 179-4, A 172-6 and A 180-73 were obtained by Rubin and Suess (1955 and 1956). The dates on the other cores were obtained at the Lamont Geological Observatory (Broecker, et al., 1956; Broecker and Kulp, 1957).

as sharp as indicated by the temperature curves.[1]

Since the carbonate material in the cores of interest consists mainly of foraminiferal tests and coccoliths, both of which grow in the surface water, C^{14} age determinations can be made by comparing the C^{14}/C^{12} ratio in the core material with that in dissolved carbonate in present day surface ocean water. The difference is attributed to radioactive decay. There are two major potential sources of error in such determinations: 1) post-depositional exchange, 2) presence of reworked carbonate material. The fact that the oxygen isotope record is preserved and that C^{14} ages of $>30{,}000$ years have been observed on core materials suggest that the amount of exchange is small and should be negligible in the age range of interest. Suess (1956) pointed out that comparison of C^{14} ages on the coarse and fine fraction should give an indication as to the presence of reworked carbonate. In one case he found that the fine fraction showed an appreciably greater age. Two such comparisons by Broecker and Kulp (1957) showed a greater age for the fine fraction in one case but no significant difference in the other. For the age range of interest reworked carbonate appears to be the only important source of systematic error.[2] Thus it may be concluded that the ages obtained are either equal to or *greater than* the true age of the core material.

The micropaleontologic temperature curves obtained by Ericson and Wollin (1956) and the radiocarbon dates obtained by Rubin and Suess (1955, 1956), Broecker, et al. (1956), Broecker and Kulp (1957) are shown in figure 2. With the exception of cores A 179-8 and A 185-35, which contain numerous turbidity current deposits and core A 180-48 which shows an abnormally high sedimentation rate, the cores may be taken to represent normal deep-sea sedimentation. A more complete discussion of the lithology of these cores has been published elsewhere (Ericson, et al., 1956; Ericson and Wollin, 1956a, and 1956b; Ewing, et al., 1958).

Table 1 summarizes the data. The best estimate of the age of the midpoint of the temperature transition in each core is given in column 7. Columns 5 and 6 give respectively the estimates for the maximum age for the beginning of the temperature transition and the minimum age for the end. The difference between these two ages provides an upper limit on the duration of the transition interval (column 8).

[1] Emiliani (1957) concludes that the temperature changes shown in several of the cores used in this study (R 10-10, A 180-48, A 179-15) are anomalously sharp and that the more gradual changes suggested by some of the cores (A 172-6, A 180-73, A 180-74) give a better picture of the actual shape of the temperature transition. Since the latter are slow deposition cores mixing by bottom borers would be more effective in spreading out the time record. Unless erosion removed a portion of the rapid deposition cores at precisely the time when the temperature was changing it is difficult to postulate a mechanism for sharpening the curves. Neither the lithologic or age data gives any indication of loss of the transition period sediment by erosion. For these reasons more weight is given to the cores showing sharp transitions.

[2] Emiliani (1957) has suggested that mixing by bottom borers can make the age at a given depth in a core too old. As evidence he uses the finite ages of the tops of several cores. However, if the mixing process has proceeded at a more or less constant rate, the amount of younger material mixed downward should roughly compensate for the amount of older material mixed upward in samples collected from below the depth of mixing. In this case ages at the depths of the temperature break should be unaffected by mixing. Also, the existence of rather sharp color changes in many of the equatorial cores suggests that the mixing produced by bottom borers is more limited than Emiliani suggests.

The majority of the midpoint ages lie close to 11,000 years. Those which are significantly older probably reflect the influence of reworked carbonate. Although the midpoint ages close to 11,000 years may also be somewhat too old, the excellent agreement between cores of widely different lithology and geographic locations suggests that this effect is small (<1000 years).

The maximum time for the transition is in all cases less than 6000 years. Since the manner in which the limits were determined, the finite spacing of the temperature measurements, and the possibility of homogenization by bottom borers all tend to increase the magnitude of this range, the smaller times (~2000 years) obtained for cores R 10-10 and A 180-48 and the very small time (~200 years) for V 12-97 are probably significant.[3] No minimum estimate can be made from the data in hand.

The conclusion is drawn that a large portion of the Atlantic Ocean underwent a temperature increase of 6° to 10°C during a period of less than 2000 years and that the midpoint of this change was within 300 years of 11,000 years ago.

Sedimentation rate.—A study of the variation in sedimentation rate with time in core A 180-74 from the mid-equatorial Atlantic by Broecker, et al. (1958) indicates that both the carbonate and the clay fraction in this core underwent a significant decrease in sedimentation rate close to 11,000 years ago. Cumulative plots of weight of clay and of weight of carbonate as a function of time before present as determined by radiocarbon dates (Broecker, et al., 1956; Broecker and Kulp, 1957) is shown in figure 3. From the change in slope close to 11,000 years ago it can be shown that the clay rate decreased by a factor of 3.7 and the carbonate rate by a factor of 2.1. Although this represents only one locality and does not permit an extrapolation to other parts of the Atlantic, the data supports the hypothesis that an abrupt change in climate occurred close to 11,000 years ago.

Ventilation of restricted basins.—The depletion of oxygen from the bottom water in restricted marine basins is often recorded in the sediments by the preservation of organic materials, the extermination of the benthic fauna, and the presence of H_2S gas. Such a situation exists in the upper 5 to 10 meters of sediments in the Cariaco Trench off Venezuela. Although the stagnation of the deep water in the trench appears to have been continuous during the period of deposition of this sediment, the presence of a layer of oxidized clay beneath the organic rich ooze demonstrates aeration at some time in the past. A radiocarbon date on organic material from the base of the anaerobic layer indicates that the abrupt stagnation of the trench occurred within 300 years of 11,000 years ago (Heezen, et al., 1958 and 1959).

EVIDENCE FROM FRESH-WATER DEPOSITS

Deposits formed by fresh-water drainage systems also record the abrupt climatic change seen in deep-sea deposits. Two such systems will be considered:

[3] Emiliani (1957) concludes that the transition occurred gradually over a period of about 8000 years. He considers cores R 10-10 and A 179-15 which show rapid transitions unreliable because of the large changes in sedimentation rate suggested by the ages. Ericson (personal communication) who did the lithologic studies of these cores, considers them both reliable in the depth range of interest. Core A 179-15 shows no evidence of abnormal sedimentation above a depth of 300 cm. Even if abnormalities were present it is difficult to construct a sequence of events which would yield the observed temperature curves and radiocarbon dates if a gradual temperature change over a period of 8000 years did occur.

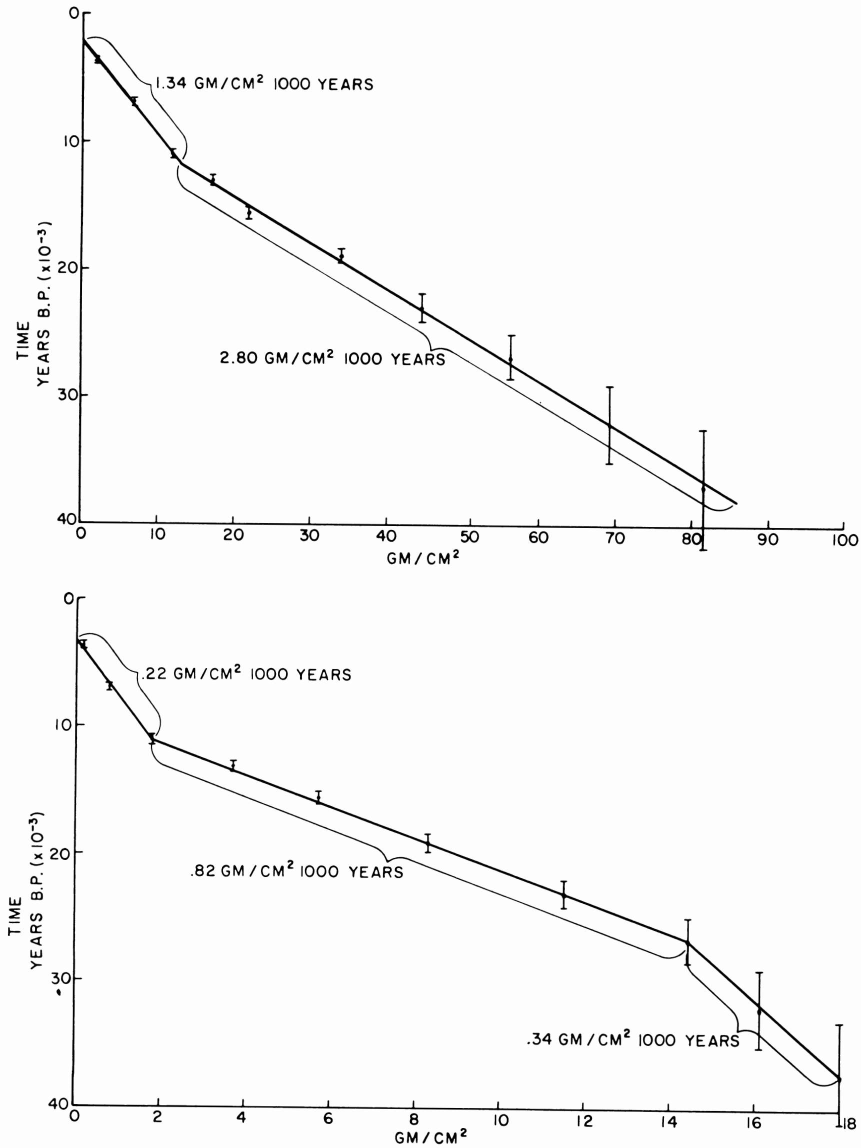


Fig. 3. Cumulative weight of $CaCO_3$ (upper diagram) and of clay (lower diagram) as a function of time in the past (Broecker, et al., 1958). The points represent radiocarbon ages with their respective errors. The slope of the lines connecting points gives the sedimentation rate at a given time. For both components a sharp change in rate occurs close to 11,000 years ago.

1) the system of lakes in the Great Basin of the western United States and 2) the main midcontinent drainage system consisting of the Great Lakes, the Mississippi River and the St. Lawrence River.

Pluvial lakes.—The Great Basin lakes are of particular interest since they are direct indicators of climate. Since these lakes have no outlet, their size is controlled primarily by a combination of rainfall and evaporation rates. As shown by Broecker and Orr (1958) the lakes respond rapidly to climatic change so that their level at a given time provides an excellent index of climate. Radiocarbon dates for Searles Lake by Libby (1955) (see Flint and Gale, 1958) and for lakes Lahontan and Bonneville by Broecker and Orr (1958) provide reasonably detailed data on the level of these lakes as a function of time.

Since these curves are based primarily on C^{14} dates obtained on material formed in fresh water lakes, the problem of estimating the initial C^{14} concentration in such materials must be considered. As shown by Deevey, et al. (1954) in some cases the concentration is significantly lower than in terrestrial plants or surface ocean water. Broecker and Orr (1958) used the present C^{14}/C^{12} ratio in the dissolved carbonate of Pyramid Lake as the control value. Broecker and Walton (1959) on the basis of numerous C^{14} measurements on currently forming materials from Great Basin fresh-water systems concluded that the C^{14}/C^{12} ratio in these lakes was within 5 percent of the present Pyramid Lake value during high water periods. This corresponds to a 400-year age uncertainty.

A second possibility of systematic error, especially for carbonate samples, is the incorporation of C^{14} by the sample after deposition. This could occur through transport and redeposition by rain or ground water or by exchange with the atmosphere. Experiments by Broecker and Orr (1958) suggest that contamination by atmospheric exchange is negligible in nearly all cases. Only the dry climate of the Great Basin and the internal consistency of the data can be used as evidence against the importance of redeposition. The ages given by Broecker and Orr on carbonate materials are probably good to 400 years but in some cases may represent only minimum estimates.

The available radiocarbon data, summarized in tables 2 and 3, strongly suggest an abrupt decrease in level from near maximum to near desiccation close to 11,000 years ago. The evidence may be summarized as follows. Except for the samples from the Winnemucca Cave area, the tufa samples from the highest level of Lake Lahontan yielded ages between 11,000 and 12,000 years B.P. averaging 11,600 years B.P. Samples from the Winnemucca Cave area suggest a second maximum close to 10,000 years ago. Although the Bonneville samples show evidence of high lake levels at both of these times no distinct separation into two groups exists. The Searles Lake data indicate that the now-dry lake contained water prior to 11,000 years ago. Data from each of the three lakes clearly demonstrates the return of arid conditions close to 11,000 years ago. Searles Lake became completely desiccated. Lake Lahontan fell below the level of Fishbone Cave and Leonard Rock Shelter and Lake Bonneville fell below the level of Danger Cave and deposited a sequence of carbonates on the salt flats.

The evidence for a rapid rise following the low levels of these lakes is problematic. In both the Lahontan and the Bonneville basins C^{14} data suggest a maximum at about 10,000 years ago. No evidence for this maximum appears to be present in the Searles Lake record, however. More work is needed to settle this question.

The events recorded in the deposits of the Great Basin lakes are summarized in figure 4. A period of continuous relatively high-water levels be-

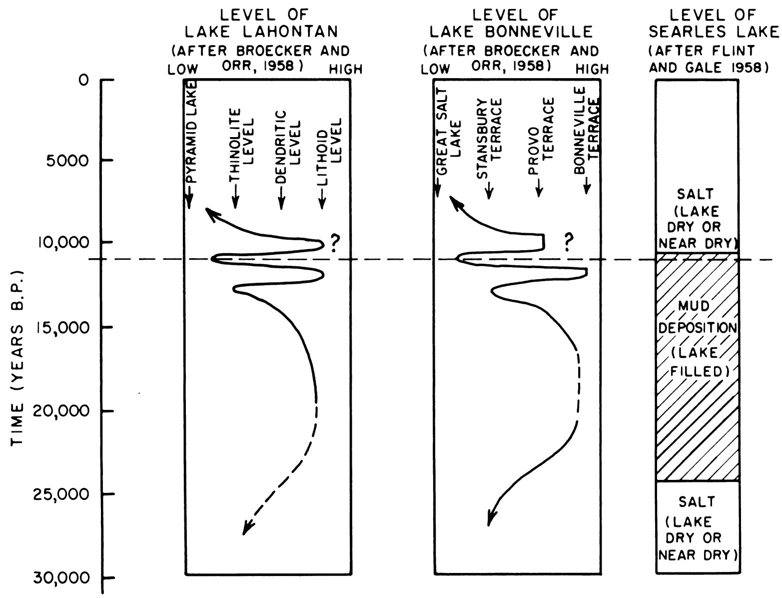


Fig. 4. The height of the major Great Basin lakes as a function of time as interpreted from radiocarbon ages.

ginning about 25,000 years was culminated by a sharp drop in lake level close to 11,000 years ago. This drop may have been followed by a brief high-water period ending about 9500 years ago. Since that time the lakes have had levels similar to those observed at present.

Midcontinent drainage system.—The midcontinent drainage system records climate in a somewhat more indirect manner. The sediments in the alluvial valley of the Mississippi River, as well as those on the Sigsbee Abyssal Plain of the Gulf of Mexico, record a marked change in the operation of the drainage system. Ewing, et al. (1958) have shown on the basis of sediment core studies that great thicknesses of gray silty clay were deposited on the abyssal plain during late Wisconsin time. They showed that this silty clay emanated from the mouth of the Mississippi River and was transported by turbidity currents. Radiocarbon dates (Broecker and Kulp, 1957) on globigerina ooze from immediately above the silty clay sequence from one of the cores (A 185-35) indicates that the supply of silty clay was cut off abruptly close to 11,000 years ago.

The evidence used by Ewing, et al. (1958) to establish the origin of the silty clay deposits is summarized in an idealized diagram in figure 5. Cores (see, for example, V 3-127, fig. 2) on knolls rising above the abyssal plain

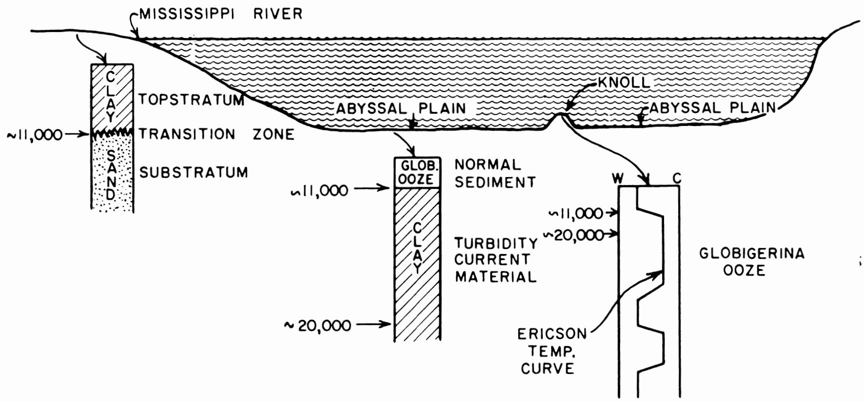


Fig. 5. Idealized diagram showing relationships of sediments on the floor of the Gulf of Mexico to those in the alluvial valley of the Mississippi River. The Gulf sediments are described by Ewing, et al. (1958) and the alluvial valley sediments by Fisk and McFarlan (1955).

are composed entirely of foraminiferal ooze, show normal climatic curves and have, as shown by radiocarbon dates, sedimentation rates similar to those in other normal deep sea oozes. Since the cores taken on the abyssal plain surrounding the knolls all show a thick silty clay sequence, Ewing, et al. conclude that the sediment must have been carried in along the bottom rather than having fallen from the surface.

Further evidence for an abrupt change in the operation of the Mississippi River system is found in the sediments on the alluvial plain in Louisiana. Fisk (1955) has classified the sediments filling the channel cut by the river during the last period of sea level lowering into two categories; the substratum consisting of sand and gravel deposits and the top stratum consisting of silt and clay. A number of radiocarbon dates (Broecker, et al., 1956; Broecker and Kulp, 1957; Brannon, et al., 1957) on samples taken near this transition zone indicate that it has an age close to 11,000 years. The results are summarized in table 4. It seems reasonable to believe that whereas much of the clay and silt load of the Mississippi River was swept beyond its channel and beyond the continental shelf into the Gulf of Mexico during the late Wisconsin, the characteristics of the river changed close to 11,000 years ago in such a manner that these clays and silts were deposited in the present alluvial valley and in the modern delta. In the absence of other evidence it is difficult to say whether this change was brought about by a change in sea level, discharge, or quantity and character of the detrital load.

The level of Lake Michigan provides an indirect index of climate. High levels are thought to correspond to times when glaciers covered the Straits of Mackinac causing the lake to drain to the south. On the other hand, when the ice receded to a position north of the glacially depressed straits drainage to the east (perhaps via the St. Lawrence) allowed the lake level to fall. Thus levels of Lake Michigan at or below its present level indicate periods of relatively warm climate.

Evidence for low water intervals in Lake Michigan close to 11,000 years ago is provided by both the Two Creeks Forest dated at 11,300 years B.P. (see table 5) and the peat bed exposed at South Haven dated at 11,200±600 (Crane, 1956). 10.860±350 (Rubin and Suess. 1955), and 10,790±200 (Preston, et al.. 1955). The use of this data to support a warming in climate close to 11.000 years ago is complicated by the Valders readvance. If the age averages of 11.300 for Two Creeks and 10.900 for South Haven are accepted at face value and if the Two Creeks Forest was overrun by Valders ice shortly after growth, then two low levels for Lake Michigan. one preceding and one following the Valders. are required. The consequent high advance and retreat rates of the Valders ice can be avoided by assuming that the Two Creeks Forest and South Haven peat are both pre-Valders in age. This would require that either or both of the radiocarbon averages are slightly in error or that a significant lag occurred between the growth and destruction of the Two Creeks Forest. Again, as in the case of the Great Basin, more work is needed to resolve the multiple events which occurred in the time interval between 10.000 and 12.000 years ago.

Radiocarbon ages on marine shell from the St. Lawrence Valley provide additional evidence for a rapid ice retreat. In order for these glacially depressed lowlands to be invaded by marine waters the ice must have retreated considerably from its maximum extent (see Flint, 1956). The results are as follows: Hull, Ontario, Y 215, 10,630 ± 330, Ottawa, Ontario, Y 216, 10,850 ± 330, and Montreal, Quebec, 11,370 ± 360 years (Preston, et al., 1955). Again the weighted average lies slightly below 11,000 years.

Leverett and Taylor (1915), however, correlated the Champlain Sea with the Lake Algonquin stage of Lake Michigan and more recently Flint (1956) correlated it with the Lake Chippewa stage. In either case the Champlain Sea invasion would be a much more recent event (4000 to 5000 years B.P.). A radiocarbon date of 9430±250 on a peat sample from sediments post-dating the Champlain Sea (Terasmae, 1959) verifies the antiquity of the shell ruling out the explanation given by Flint (1956) that the shell contained considerable carbonate derived from ancient limestones.

EVIDENCE FROM POLLEN PROFILES

Another index of climate is the type of plant and animal remains found in terrestrial deposits. Of these the pollen profiles of northwestern Europe have been most thoroughly studied and are best understood. Also, a large number of radiocarbon measurements are available allowing a fairly detailed chronology to be established (see table 6). A portion of this profile is reproduced in figure 6. The outstanding feature of all the diagrams is the pronounced warm period referred to as *Alleröd*. It is preceded and followed by colder climates, the Older Dryas and Younger Dryas respectively. The available radiocarbon data are summarized in table 6. From these average ages for the various pollen zones and for the transition can be computed. These averages are shown in figure 6. It is clear that the warm *Alleröd* climates prevailed during the period of increase in surface ocean temperature. A problem exists in that there is not a unidirectional change in climate from cold to warm following the Older

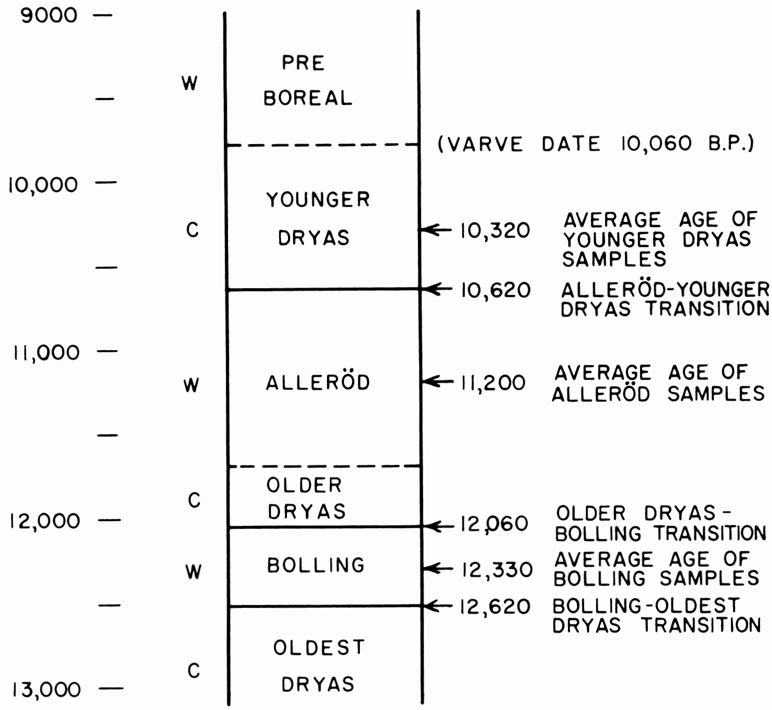


Fig. 6. Sequence of climate periods in the standard European pollen profile (based on diagram given by Barendsen, et al., 1957). The average ages for the periods and transitions are based on the data summarized in table 6. Warm periods are indicated by W and cold by C. No radiocarbon data are available for the dotted transitions.

Dryas for the cold climate is re-established during the period from approximately 10,650 to approximately 9800 years ago. The significance of this oscillation is not clear at present. It is interesting to note that an identical oscillation appears in the Great Basin and Great Lakes chronologies. It is entirely possible that the sharp change in oceanic conditions correlates with the post Younger Dryas rather than the pre Younger Dryas warm period as has been implied above.

CONCLUSIONS

From the evidence listed above it is clear that a major fluctuation in climate occurred close to 11,000 years ago. The primary observation that both surface ocean temperatures and deep sea sedimentation rates were abruptly altered at this time is supplemented by evidence from more local systems. The level of the Great Basin lakes fell from the highest terraces to a position close to that observed at present. The silt and clay load of the Mississippi River was suddenly retained in the alluvial valley and delta. A rapid ice retreat opened the northern drainage system of the Great Lakes and terrestrial temperatures rose to nearly interglacial levels in northwestern Europe. In each case the transition is the most obvious feature of the entire record.

Correlation of these systems on a finer scale is difficult in that oscillations preceding and following the major change in the case of the European pollen profiles, Great Basin lakes and Great Lakes region have not been observed in the record left in the Mississippi alluvial deposits or deep ocean deposits. More work is needed before a satisfactory explanation can be given for these differences.

Certainly if the abruptness of such a major change in world climate can be firmly established it will be of the utmost importance that acceptable theories are able to account for it. It is interesting to note that the recent "Theory of Ice Ages" proposed by Ewing and Donn (1956 and 1958) was stimulated by the observation that the change in climate which occurred at the close of the Wisconsin glacial period was extremely abrupt.

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TABLE 1

Estimate of time interval during which the increase in surface ocean water temperature occurred

1 Core	2 Minimum depth of end of change (cm)	3 Maximum depth of start of change (cm)	4 Depth of midpoint (cm)	5 Maximum age of start of change (yr)	6 Minimum age of end of change (yr)	7 Age of midpoint (yr)	8 Time span for change from Wisconsin to Recent temperature (yr)	9 Laboratory
				North Atlantic				
P 10-10	100	120	110	11,650	10,500	10,920	<1850	L
A 179-15	100	120	110	13,600	8200	10,450	<5400	L
A 179-8*	225	280	270	15,200	10,500	13,500	<4700	L
				Equatorial Atlantic				
A 180-74	10	35	18	14,600	7500	10,750	<7100	L
A 180-73	10	30	18	15,000	—	11,300	—	W
A 180-76	10	35	20	—	—	10,750	—	L
A 180-48*	450	480	457	13,500	11,600	12,100	<1900	L
				Caribbean				
A 172-6*	10	40	35	19,000	—	—	—	W
A 179-4*	15	30	24	14,200	—	11,000	—	W
				Gulf of Mexico				
A 185-35*	—	—	105	—	—	10,900	—	L
V 3-126	20	60	40	—	—	10,900	—	L
V 3-127	30	50	40	13,900	—	12,900	—	L
				Cariaco Trench				
V 12-97	445	455	450	10,950	10,750	10,850	<200	L

* Cores which might contain a substantial amount of reworked carbonate material. Ages quoted may hence be considered maximal.

W—Washington (Rubin and Suess, 1955, 1956).

L—Lamont (Broecker et al., 1956; Broecker and Kulp, 1957; Olson and Broecker, 1959).

TABLE 2

Materials from deposits formed during high water periods of the Great Basin lakes

Height Above 1890 Lake Level (feet)	Location	Material	Sample No.	Age (years)
	Pyramid Lake Basin			
525	Lake Winnemucca Cave Area, Nev.	tufa	L 289-G*	9700±200
525	Lake Winnemucca Cave Area, Nev.	tufa	L 356-G*	10,000±220
411	Lake Winnemucca Cave Area, Nev.	tufa	L 356-H*	9700±200
560	Lake Winnemucca Cave Area, Nev.	tufa	L 364-AA*	9500±200
570	Anaho Island Pyramid Lake, Nev.	tufa	L 289-N*	11,800±200
520	Anaho Island Pyramid Lake, Nev.	tufa	L 289-M*	11,700±200
390	Anaho Island Pyramid Lake, Nev.	tufa	L 289-L*	11,570±250
560	Mullen Pass Pyramid Lake, Nev.	tufa	L 289-I*	11,250±350
250	Fishbone Cave Lake Winnemucca, Nev.	tufa	L 289-C*	11,700±500
570	Anaho Island Pyramid Lake, Nev.	tufa	W 442**	12,050±400
	Great Salt Lake Basin			
660	South end Oquirrh Mtns., Utah	tufa	L 363-D*	11,000±600
660	South end Oquirrh Mtns., Utah	tufa	L 435-G*	11,300±250
660	South end Oquirrh Mtns., Utah	tufa	L 435-C*	10,600±300
>660	South end Oquirrh Mtns., Utah	tufa	W 409**	11,300±300
600	Pleasant View Salient, Ogden, Utah	tufa	W 456**	11,650±450
400	Weber River Delta Ogden, Utah	marl	W 382**	12,960±350†
400	Weber River Delta Ogden, Utah	peat (from W 382)	W 326**	9925±300†
400	Weber River Delta Ogden, Utah	peat (repeat W 326)	W 440**	10,260±300†
	Searles Lake Basin			
—	Searles dry lake, California	organic material from top of lake muds	C 894***	10,494±560
—	Searles dry lake, California	carbonate from top of lake muds	Y 574-a****	11,810±140
—	Searles dry lake, California	organic material from top of lake muds	Y 574-b****	10,700±130

* Broecker and Orr (1958)

** Rubin and Alexander (1958)

*** Libby (1955)

**** Flint and Gale (1958)

† A new set of samples collected from this locality by one of the authors (Wallace S. Broecker) yielded ages of 9500±200 for the organic fraction and 9700±200 for the carbonate fraction. The high age obtained by the Washington laboratory for the original marl has been checked and confirmed (Rubin, personal communication) strongly suggesting that this sample contained reworked detrital carbonate. Since these peats are overlain by lake sediments, these dates provide excellent evidence for a post 11,000 year B.P. high lake stand.

TABLE 3

Materials from deposits formed during low water periods of the Great Basin lakes

Height Above 1890 Lake Level (feet)	Location	Material	Sample No.	Age (years)
		Pyramid Lake Basin		
180	Fishbone Cave Lake Winnemucca, Nev.	twigs overlying lake sediments	L 245*	11,200 ± 250
—	Leonard Rock Shelter Humboldt Lake, Nev.	guano overlying lake sediments	C 599**	11,200 ± 570
90	Needles Area Pyramid Lake, Nev.	outermost tufa layer on dome	L 364-CE*	8500 ± 200
		Great Salt Lake Basin		
50	Danger Cave, Bonneville Salt Flats, Utah	sheep dung overlying beach gravels	M 118****	11,000 ± 700
50	Danger Cave, Bonneville Salt Flats, Utah	wood overlying lake deposits	C 610**	11,150 ± 570
50	Danger Cave, Bonneville Salt Flats, Utah	twigs associated with M 118	M 119****	10,400 ± 700
35	Hooper, Utah	plant stems in growth position	W 386***	9730 ± 350
17	Bonneville Salt Flats Knolls, Utah	carbonate from one foot beneath surface	L 485*	11,150 ± 450
		Searles Lake Basin		
—	Searles dry lake California	organic material from mud within upper salt	RC-36*****	10,270 ± 450
—	Searles dry lake California	organic material from mud within upper salt	RC-50*****	11,400 ± 600
—	Searles dry lake California	carbonate from RC-50	RC-50c*****	9900 ± 500
—	Searles dry lake California	trona from base of upper salt	W-248***	8550 ± 250

* Broecker and Orr (1958)

** Libby (1955)

*** Rubin and Alexander (1958)

**** Crane (1956)

***** Flint and Gale (1958)

TABLE 4

Radiocarbon dates defining the transition from sand to clay deposition in the Mississippi Alluvial Valley

Location	Sample Depth (feet)	Depth to Transition Zone (feet)	Sample No.	Age (years)
La Fourche Parish, La.	95	150	HOR-73***	8800±180
Duck Lake Area, La.	125	140	L-175B*	9750±550
Dulac Area, La.	150	210	L-291X**	10,700±150
Grand Isle Area, La.	215	245	L-291G**	11,050±300
Grand Isle Area, La.	240	245	L-291H**	10,530±350
Bayou Pigeon Area, La.	220	160	HOR-126***	13,650±300

* Broecker, et al., 1956
** Broecker and Kulp, 1957
*** Brannon, et al., 1957

TABLE 5

Radiocarbon dates on Two Creeks forest samples

Laboratory	No. of Samples	Average Age (years)
Chicago	4	11,400±350 (Libby, 1955)
Yale	1	11,130±350 (Preston et al., 1955)
Washington	2	11,375±150 (Suess, 1954)
Michigan	1	10,550±450 (Crane, 1956)
	Weighted Average	11,300±130

TABLE 6

Radiocarbon dates for northeastern Europe pollen profiles

	Sample No.	Age (years)
Younger Dryas		
	GRO 419	10,345±275
	K 111	10,300±350
	Y 139-3	11,350±150*
Younger Dryas/Alleröd Boundary		
	K 101	10,890±240
	GRO 454	10,995±250
	W 82	10,260±200
	W 86	10,510±180
	K 102	10,500±400
	Y 157A	10,560±200
	H 105-87	11,500±300*

TABLE 6 (Continued)

Alleröd		
	K 103	11,060 ± 480
	K 113	10,930 ± 380
	K 104	10,990 ± 240
	K 110	10,770 ± 360
	K 105	11,800 ± 410
	K 106	11,880 ± 340
	K 112	11,700 ± 360
	K 107	11,160 ± 320
	C 337	11,044 ± 500
	GRO 909	11,025 ± 120
	GRO 907-937	10,650 ± 120
	GRO 920	10,660 ± 90
	GRO 647	11,230 ± 400
	GRO 925	11,065 ± 120
	GRO 948-933	11,700 ± 90
	GRO 947	11,470 ± 90
	GRO 921	11,560 ± 100
	GRO 410	11,200 ± 320
	GRO 413	11,560 ± 260
	H 1-8	11,900 ± 500
	H 1-48	11,800 ± 300
	H 21-18	11,550 ± 280
	H 18-11	11,930 ± 290
	H 75-68	11,450 ± 180
	Y 139-1	12,500 ± 180*
	Y 139-2	10,880 ± 160
	Y 442	11,220 ± 350
Alleröd/Older Dryas		
Older Dryas		
Older Dryas/Bolling		
	GRO 926	11,825 ± 120
	H 77-54	12,300 ± 260
Bolling		
	GRO 927	12,355 ± 170
	GRO 1104	12,300 ± 100
Bolling/Oldest Dryas		
	GRO 928	12,200 ± 100
	GRO 935	12,380 ± 130
	H 88-74	13,250 ± 280
	H 106-89	12,700 ± 320

H—Hiedelberg (Munnich, 1957)

GRO—Groningen (DeVries, et al., 1958)

K—Copenhagen (Anderson, et al., 1953)

W—Washington (Rubin and Suess, 1955 and 1956)

Y—Yale (Barendsen, et al., 1957)

* Not included in average in figure 6.