GEOLOGIE EN MIJNBOUW VOLUME 48 (3), p. 349-361 1969

ON THE WÜRM – FLANDRIAN BOUNDARY IN DEEP-SEA CORES

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SUMMARY

Due to the present drainage pattern the North Atlantic receives much more weathered products than the other oceans do. Dissolutions and oxidations at the bottom of the North Atlantic seem to change in time, being climatologically controlled. The dissolution of carbonates in North Atlantic has been intensified during ice ages mainly in the zone of Polar Bottom Water.

During the Main (Upper) Würm the Arctic Ocean was not ice-covered because of the absence of a salinity discontinuity (halocline). It is believed that an intense formation of bottom water occurred there and that this water then flushed the North Atlantic bottom and caused the increased dissolution of carbonates.

At the transition Alleröd/Younger Dryas (Würm/Flandrian boundary) the Arctic Ocean was covered by a pack-ice. The developed halocline stopped the sink water formation. Then the outflow of cool bottom water to the Norwegian Sea dropped suddenly. The lack of this bottom water caused an increased carbonate deposition in the North Atlantic approximately 11,000 B.P., and somewhat later on, a decrease of it occurred in the Indian and Pacific Oceans (see fig. 1).

The isotopic paleotemperature method has been found invalid. High percentages of warm-water-indicating foraminifera suggest that the North Atlantic gradually became warmer from the Alleröd up to the climatic optimum about 6000 B.P.

The precipitation in the arid lands around the Gulf of Aden during the Late Würm and Flandrian are given in Fig. 6. The highest precipitation during aforementioned times occurred during the Atlantic subage and the lowest one during the Preboreal subage.

It is suggested that the Red Sea was dry during maximum of the Main (Upper) Würm Ice age. A marked change in the vertical circulation happened about 8000 B.P.

The Eastern Mediterranean and the Cariaco Trench changed into a (quasi)stagnant phase approximately during or at the end of the Alleröd subage.

The climatic deterioration during the Younger Dryas stadial is explained as caused by the albedo change which occurred when the Arctic Ocean went into the pack-ice phase. The duration of this stadial is about 600 yrs. Such a short interval is hard to trace in abyssal cores. The last large change in deep-sea cores has probably happened at the end of the Alleröd subage. These two points suggest that the upper boundary of the Würm age is to be placed at the Earlier Dryas – Alleröd boundary.

The present age is here called Flandrian since Holocene is a term of higher rank (epoch). I also suggest that Pleistocene should be considered as the only epoch in the Quaternary period. "Holocene" could be discarded.

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INTRODUCTION

This paper deals with the history of the oceans during the Late Würm and Early Flandrian. The stratigraphy of the sediments from the Eastern Mediterranean, from the Gulf of Aden and the Cariaco Trench are discussed.

The upper boundary of the Würm age is placed at the Earlier Dryas-Alleröd boundary. Flandrian is introduced as the name of the present age. Then each age during the Pleistocene period has a name of its own. "Holocene" can be discarded.

GEOCHEMICAL AND GEOECONOMICAL ASPECTS ON CARBONATE DEPOSITION

If the chemical composition of sea water should remain constant for a certain time, it is necessary that the supply of each element from weathering be equal to the withdrawal of it from sea water. The accumulation of $CaCO_3$ at the sea floor can thus be estimated from analyses of river waters.

The land area drained to the Atlantic is nearly

twice as large as that drained into the Indian Ocean and the Pacific together. Therefore, the Atlantic, particularly the North Atlantic, receives much more weathered products from rivers than the other oceans do. Thus, the North Atlantic receives 70% of all Ca, South Atlantic 8%, Indian Ocean 10%, and the Pacific 12% (O l a u s o n, 1967). However, water exchanges between the oceans more or less effectively smooth out differences caused by the present drainage pattern which concern both the chemical composition of the sea water and the accumulation rates of each element. The present accumulation of $CaCO_3$ in per cent of the total deposition ($\frac{1}{2}22.10^{13}$ g $CaCO_3$) is: North Atlantic = 33, South Atlantic = 20, Indian Ocean = 20, and the Pacific = 25 (*l.op. cit.*). Consequently, the North Atlantic loses about half of the received amount of calcium from rivers and this amount is deposited in other oceans. However, the net outflow of Ca changes in time, being climatologically controlled.

There is no surface outflow from the North Atlantic. Therefore, the net outtransport occurs in the cool water sphere. The calcium is removed from the surface water by organisms, chiefly foraminifera and coccolithophorids. The settling time compared to the time during which the calcareous remains are in contact with the bottom water is negligible. Hence, the dissolution of the carbonates can be supposed to occur chiefly at the bottom. This dissolution can be described by the following equations:

$$CaCO_3 + H_2O + CO_2 \rightarrow Ca^{2+} + 2\,HCO_3^-$$

$$CaCO_3 + CH_2O + O_2 \rightarrow Ca^{2+} + 2\,HCO_3^-$$

where CH_2O stands for organic matter (soft organic remains in various degrees of decomposition).

Calcareous matter and large, soft organic remains can reach the bottom due to their own weight but smaller organic particles, colloidal organic matter and organic matter in true solution seem to reach larger depth chiefly by descending water masses. The car-

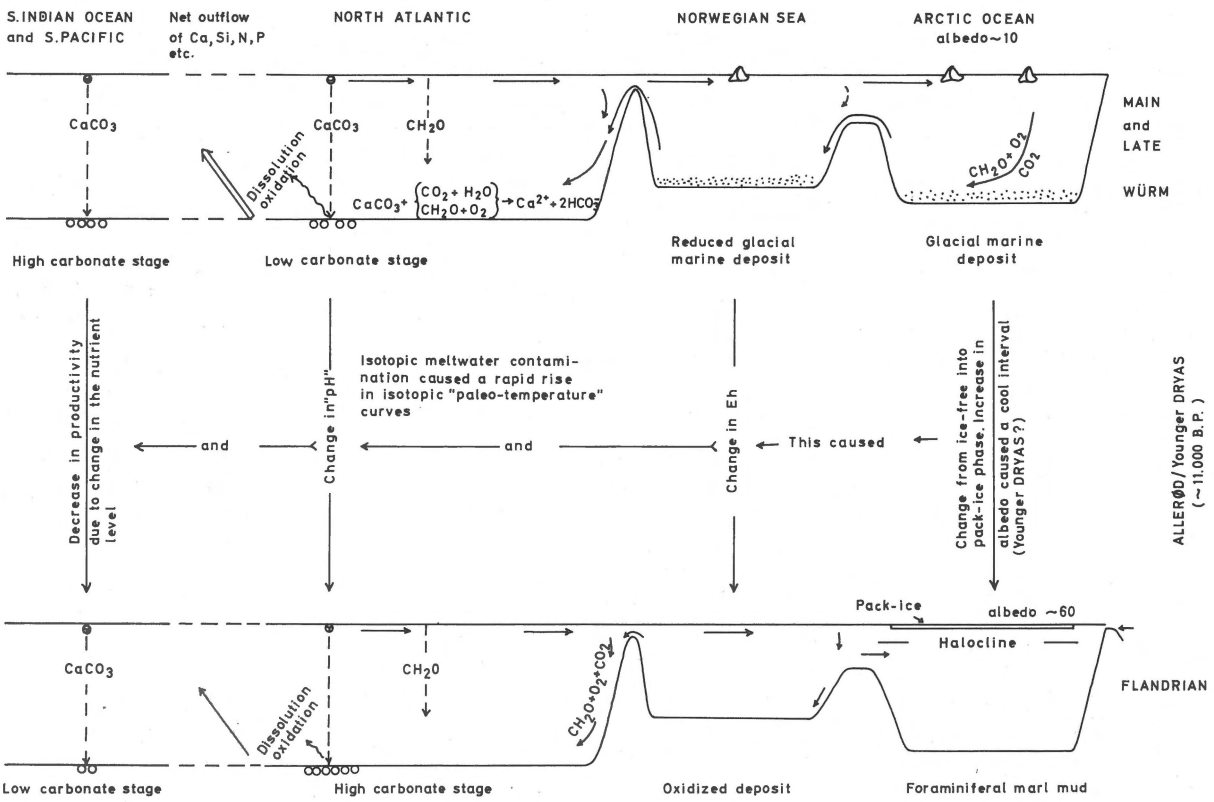


Fig. 1

The vertical circulation of the oceans during the Main and Late Würm and the Flandrian. The change in the hydrography of the Arctic Ocean at the end of the Alleröd subage is believed to have started a chain reaction illustrated in the figure.

bonate content of sediments, the dissolution rate at the bottom as well as the pH of the environment depend on the supply (in moles per time unit) of $CaCO_3$ (chiefly vertical supply), CH_2O (both vertical and lateral supply), CO_2 and O_2 (both by descending water masses). These processes are illustrated in Fig. 1.

In areas flushed by North Atlantic sink water oxygen is usually present in excess. If we assume the proportion of $CaCO_3$ and CH_2O settled down from the overlying surface water as rather constant, changes in the dissolution of $CaCO_3$, and thus changes in the carbonate content of the final sediment, are due to changes in the intensity of the vertical circulation.

The organic matter which settles down both from the surface water and from sink water masses which flush the bottom, contains not only CH_2O but also elements as N, P, Si etc. After oxidation of the compounds these elements are released back to the water and then transported out to other oceans (see Fig. 1).

The areas of sink water formation in the Northernmost Atlantic are at either side of southern Greenland and between Greenland and Spitsbergen. Since the carbonate content of North Atlantic cores is lower in glacial stages than in interglacial ones and the reverse situation occurs in cores from the Indian and Pacific Oceans, we must look for a new source area operating during ice ages. Further, the intensified dissolution of carbonates in North Atlantic occurred chiefly in the zone of Polar bottom water. The additional area of bottom water formation may be the Arctic Ocean.

THE ARCTIC OCEAN AND ITS RÔLE IN THE VERTICAL CIRCULATION OF THE OCEANS

There have been considerable discussions about the distribution of ice in the Arctic Ocean during earlier (sub)ages (e.g. by E w i n g and D o n n, 1956, 1958). It seems to me that the aforementioned discussions have not dealt with the main problem, i.e. why we have at present a permanent ice cover there, and why it must have been absent during some earlier times.

In an Arctic Ocean with a vertically uniform salinity, ice cannot be formed until the entire water column has been cooled to freezing-point. The salinity stratification in the present Arctic makes the situation different: the surface temperature will rapidly reach freezing-point because the salinity stratification prevents downward convection below about 30 m. The salinity stratification is for 50% due to the river discharge, for 46% due to the inflow of Pacific water with a salinity of 32‰, and for about 4% due to precipitation.

During an ice age large amounts of water were transferred from the sea and large ice sheets covered the northernmost America, Europe and Asia. During ice advance phases the amount of water discharged into the Polar Sea from rivers must have diminished considerably compared to their present quantities, probably to near nil. The sill depth of the Bering Strait is about 38 m (H o p k i n s, 1967, p. 29). When the sea-level dropped, the inflow from the Pacific decreased gradually and became nil when the sill was left dry. The lack of both fresh water sources will create a vertically uniform salinity of the Arctic Ocean. In such a water cooling will start a vertical circulation reaching down to the bottom (Fig. 1). Then an open Arctic Ocean will prevail until the salinity stratification is developed again. This occurs when the melt waters and river waters are discharged into the Polar Sea in rather high amount and/or the inflow from the Pacific becomes important.

The rise in sea-level above – 38 m has probably occurred during the Alleröd subage (11,900-10,800 B.P.). Particularly during the Alleröd large amount of meltwater was discharged into the Polar Sea. These together with a climatic improvement could have created difficulties to form sink water with sufficient density to reach large depths. A salinity discontinuity (halocline) was established not only during the summer but for the whole year. Then a permanent pack-ice could have been formed.

In the sediment cores from the Alpha Rise glacial marine sediment alternate with foraminiferal marl mud (H e r m a n, in press). The former is interpreted as accumulated under open Arctic condition while the foraminiferal marl mud is formed during times when a permanent pack-ice existed. The last change into pack-ice stage in the Arctic Ocean is by O l a u s s o n and J o n a s s o n (in press) concluded to have happened at the transition between the Alleröd and Younger Dryas. This is suggested by the following indications:

a) largest supply of fresh water to the Arctic Ocean during the Alleröd (and Bölling),

b) the inflow from the Pacific began during the Alleröd subage,

c) if the pteropods (Fig. 3) entered the Arctic Ocean from the Pacific, which is probable, the lithologic change happened during or after the Alleröd subage,
d) the change into pack-ice stage must have reduced the global temperature by some 0.5°C; such a drop occurred during the Younger Dryas, a lesser one at the Bölling/Earlier Dryas transition,
e) the increase in carbonate deposition in the North Atlantic occurred approximately 11,000 B.P., and
f) the age of the boundary Alleröd/Younger Dryas is generally given as 10,800 B.P.

The surface circulation in the North Atlantic is given in Fig. 2. It is possible that the Gulf Stream Water Mass entered the Norwegian Sea and the Arctic Ocean during (part) of the Würm, constituting a compensation current for the outflowing Arctic Bottom Water Mass.

Further details about the Arctic Ocean during the Würm and Early Flandrian are given by O l a u s s o n and J o n a s s o n (in press).

THE NORTH ATLANTIC AT THE ALLERÖD–YOUNGER DRYAS BOUNDARY

Judging from the above it seems likely that the Arctic Ocean was the most important contributor of the cool water containing CO_2 and CH_2O. In Fig. 1 the vertical circulation of the oceans are schematically given both for the Würm and for the Flandrian (see also Fig. 2).

It appears from this figure that the change from ice-free into pack-ice stage stopped the sink water formation in the Arctic Ocean. Then the outflow of cool bottom water into the Norwegian Sea dropped suddenly. The supply of organic matter to the Norwegian Sea floor diminished and probably changed the colour of its deposits. The overflow to the Atlantic diminished too. This led to the following chain reaction:

(i) a decreased supply of CO_2, O_2, and CH_2O to the bottom of the North Atlantic,
(ii) which caused a decrease of the oxidation of CH_2O (or $C_{106}N_{16}P$),
(iii) which together with the decreased supply of CO_2, caused a decrease in the dissolution of calcite, apatite etc. and, thus, an increased accumulation of both there,
(iv) which resulted in a decrease of the amount of Ca, P, Si, N etc. which were transported out from the North Atlantic.
(v) which caused a decrease of the rate of organic production in the other oceans due to a decreased amount of nutrients available there,
(vi) which caused a decrease of the rate of accumulation of calcium carbonate and the number of entire foraminifera (O l a u s s o n, 1961a, 1967, 1969) and a decrease in the deposition of phosphate, silica etc. This change in the Indian and Pacific Oceans might be expected to have happened, say, 1000 yrs later than the aforementioned changes in the Atlantic. A date of the faunal change in an Andaman Sea core at 9000 B.P. (F r e r i c h s, 1968) is in good agreement with my theory.

At the beginning of an ice age the reverse happened.

The exchange across the Equator is large only in the Atlantic Ocean. Therefore, the increased outflow of nutrients from the Atlantic during glacials chiefly or exclusively influenced the area around or south of the South Equatorial Divergence in the Indian and Pacific Oceans (O l a u s s o n 1969).

ON THE SURFACE TEMPERATURE OF THE NORTH ATLANTIC AT THE WÜRM – FLANDRIAN BOUNDARY

Two paleotemperature methods, their validities and results are to be discussed, namely the isotopic and foraminiferal analyses (see also O l a u s s o n, 1965, 1969).

The relationship between 1. temperature (T), 2. the per mil difference between the O^{18}/O^{16} ratio in the sample and in a fixed standard (δ'), and 3. the per mil difference between the water of ancient composition applicable to the tests being examined and the present day marine water of average composition (δ) is

$$T = 16.5 - 4.3\,(\delta' - \delta) + 0.14\,(\delta' - \delta)^2$$

Only δ' can be determined; in order to get an estimate of T, the numerical value of δ must also be estimated or guessed at. The question is then whether it is easier to estimate δ than T. The answer is not necessary: "yes".

Due to the glacial storage effect there has been a change in the average value of δ. According to E m i-

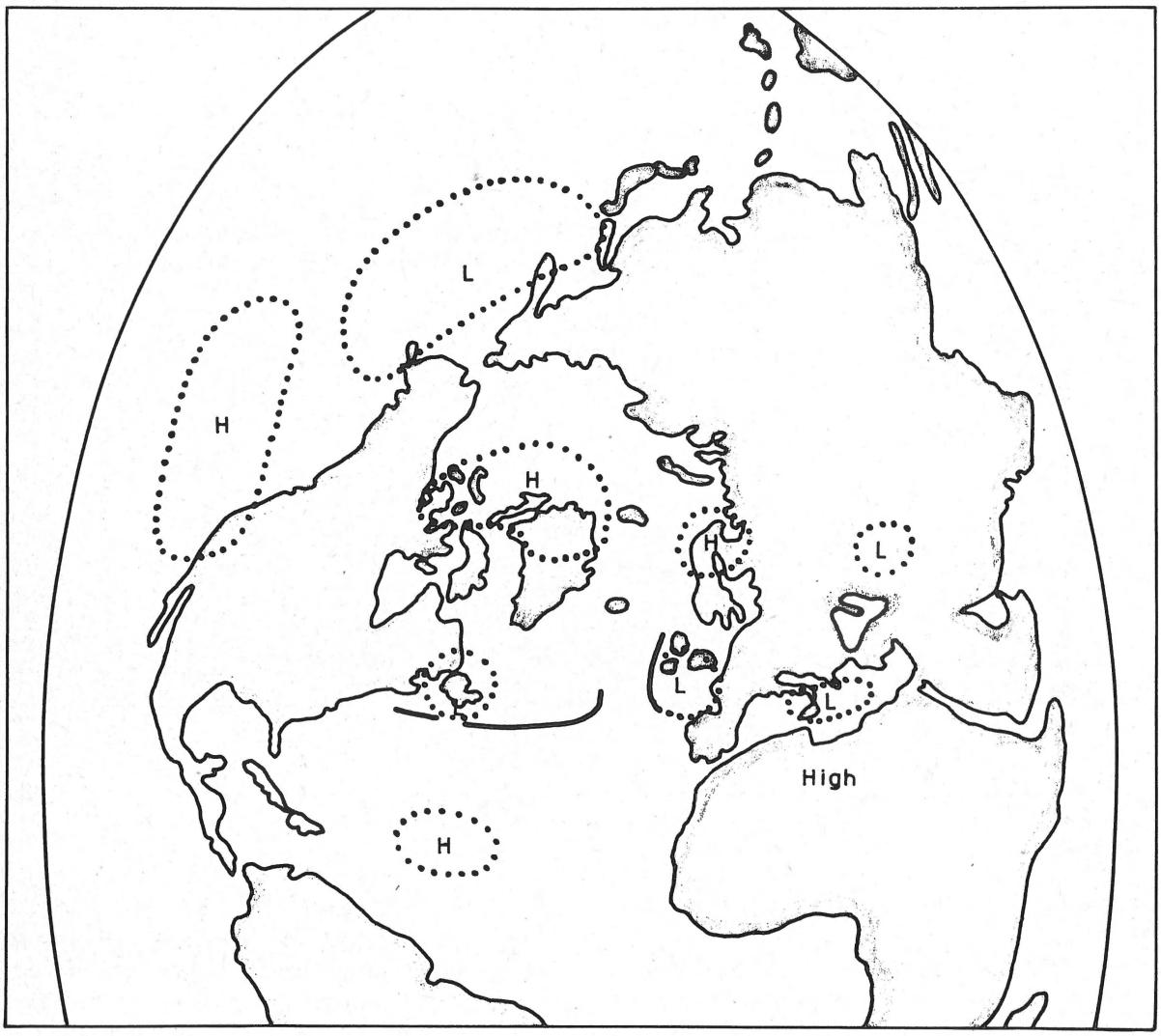


Fig. 2

The mean surface pressure distribution in January about the maximum of the Würm according to L a m b (1961) and tentative boundaries between the Gulf Stream Water Mass and the Subarctic Water Mass. The later are based on the coiling directions of *Globigerina pachyderma* in the Lamont cores A 157-11, A 157-13, R 5-36 and SP 8-4 (Ericson in W e y l (1968)), and in the Swedish cores 280-299.

l i a n i (1955a) the amplitude of this change was 0.7 per mil, i.e. equal to a temperature change of 3°C. A new estimate was published ten years later (O l a u s s o n, 1965). I found an amplitude which was about 3 times as large. Later on other scholars have accepted my results (*e.g.* S h a c k l e t o n, 1967; W e y l, 1968). The result of this calculation was that the temperature change in the subtropical and tropical part of the world oceans during the Pleistocene has not been 6-10°C as believed before but 0-3°C. For the Pacific the amplitude seems to be close to 0 (O l a u s s o n, 1965), *i.e.* a corrected temperature curve should more or less be a straight line. As to the North Atlantic cores their curves were believed to be influenced by isotopic meltwater contamination during deglacial times.

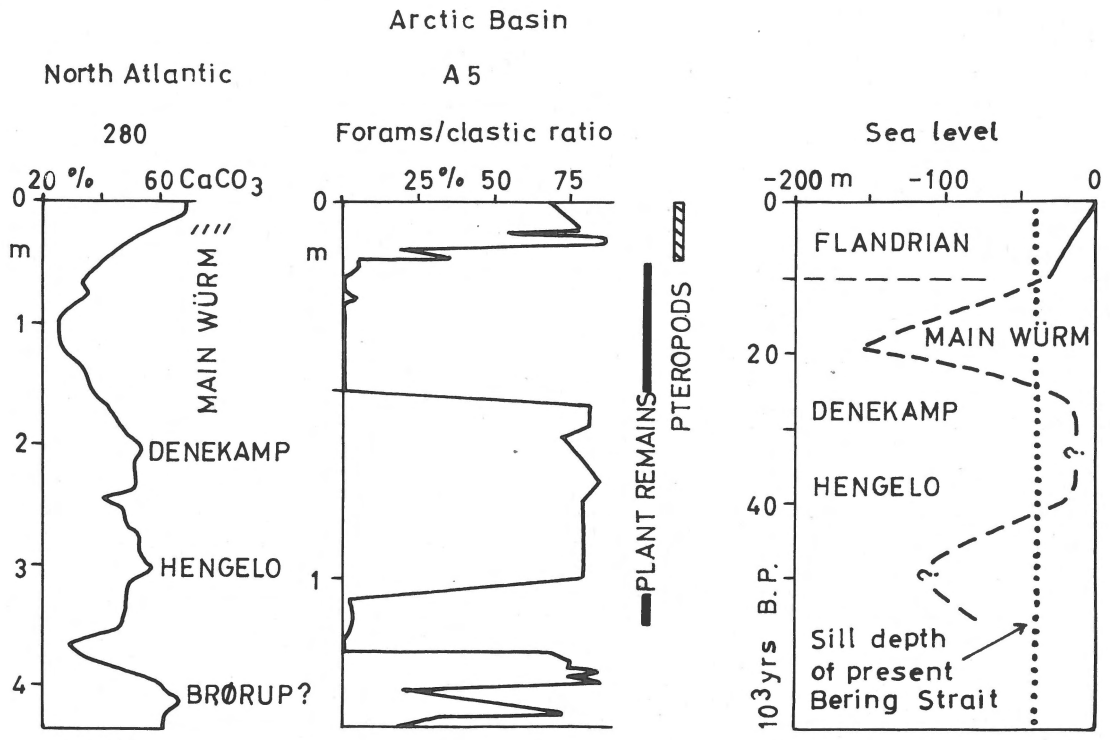


Fig. 3

The distribution of $CaCO_3$ in the North Atlantic core 280 (O l a u s s o n 1960), the ratio foraminifera/clastic grains and the distribution of plant detritus and pteropods in the Artic Basin core D.st. A 5 (H e r m a n, in press) and the dates of sea level stand (mainly after S h e p a r d and C u r r a y, 1967).

E m i l i a n i (1966) thinks that a strict boundary condition for possible glacial-interglacial variations in the average oxygen composition of the ocean could be estimated from analyses of benthonic foraminifera in Pacific cores. As stated by him, I did not consider this possibility and the reason herefore was that it seemed to me quite unprobable. As appears from Fig. 4 the O^{18}/O^{16} ratios varies irregularly. Thus, these are amplitudes within the Holstein interglacial of 0.8 per mil. The delta value can also be higher in samples from interglacial stages than in glacial ones. The amplitude between glacial and interglacial groups is 0.16 per mil (according to Emiliani, *op. cit.* it is 0.5 ± 0.2). The large dispersion found in Fig. 4 suggests that such a mean value has no meaning in this connection. The isotopic composition of benthonic foraminifera in Pacific core samples depends on several factors. Many of them cannot be assumed to be constant. Emiliani's attempt to minimize the error in isotopic paleotemperature analysis is thus not supported by the evidence.

It is necessary to recognize three different problems connected with the Würm-Flandrian boundary in North Atlantic cores according to O l a u s s o n (1965):

a) as stated in this paper, the change in carbonate accumulation at the bottom does not necessary indicate any change in the conditions in the surface water above; it can reflect only changes which happened elsewhere,

b) the change in frequency of foraminifera can be due to change in the surface conditions and/or change in the bottom water conditions,

c) the possibility of an isotopic meltwater contamination of this ocean which could give a too high isotopic "temperature" during deglacial times.

This discussion was by E m i l i a n i (1966, footnote 5) referred "to noise alone". Aforementioned

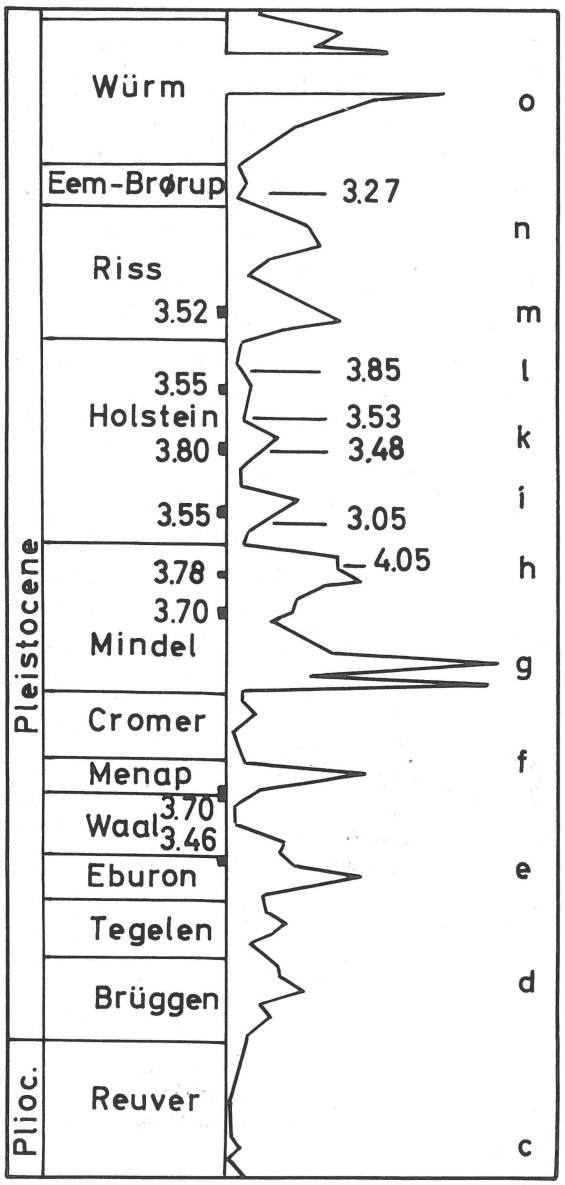


Fig. 4

The figure shows the δO^{18} values of benthonic foraminifera of the Pacific cores 58 (the figure in the column of the time-stratigraphic units) and 60 (the figure to the right of the curve) in relation to the climatic curve of core 58. Stratigraphic data by Olausson (1961b, 1967), and isotopic data by Emiliani (1955a).

footnote is here left aside since Emiliani does not visualize the difference between reactions within the surface water and those happening at the bottom.

30 out of 60 foraminiferal tests which settle down annually per cm^2 in the North Atlantic are at present dissolved at the bottom. Due to selective dissolution a change in the dissolving capacity of the bottom and interstitial waters will also alter the percental distribution of "warm" and "cool" species (Olausson, 1969). It is therefore very difficult to say something from a change in thanatocoenosis (*l. op. cit.*). The interpretation of it is a complicated geochemical-paleooceanographical problem. However, an increase in the frequency of thin-walled species might indicate an increase in the "pH" at the bottom. An increase of a thick-walled species in or after a rise in carbonate accumulation must mean something else.

In connection with the last rise in carbonate content, approximately 11,000 B.P., there is a rapid increase in the frequency of the thin-walled *Globigerinoides sacculifera* in North Atlantic cores (Fig. 5). It marks, according to my opinion, only a decrease in the dissolution at the bottom (Olausson, 1965, 1969). Later on an increase for the thick-walled *Globorotalia menardii-tumida* group is noticed (Fig. 5). Since last group is considered as "warm" species, their appearance indicate probably a slight increase in the water temperature. This increase happened some thousands years after the carbonate increase.

Judging from Fig. 5 we can state that the increase in carbonate and isotopic plaeotemperature occurred about 11,000 B.P. and reached interglacial values some 10,000 B.P. Since that time there is an increase in warm-water indicating foraminifera. If we assume a constant rate of deposition during the Flandrian age, the maximum of the "warm" group could have occurred during the climatic optimum (about 6000 B.P.; see further Olausson, 1965). Thus, if we question the isotopic paleotemperature method, there is a good agreement between the climatic data obtained from continental sequences and deep-sea cores. If the isotopic method is correct, it means that the Atlantic Ocean reached climatic optimum 4000 yrs earlier than the surrounding continents. This seems impossible to me.

The problem about the O^{18}/O^{16} ratio in the North Atlantic during the Würm/Flandrian transition must be discussed further (see also Olausson, 1965, p. 233-238).

Due to the extraction of $50\text{-}55 \cdot 10^6$ km^3 O^{18}-poor water the world ocean was isotopically heavier at the maximum of the Würm ice age than during the Flandrian or Eem. The increase of the O^{18}/O^{16} ratio was

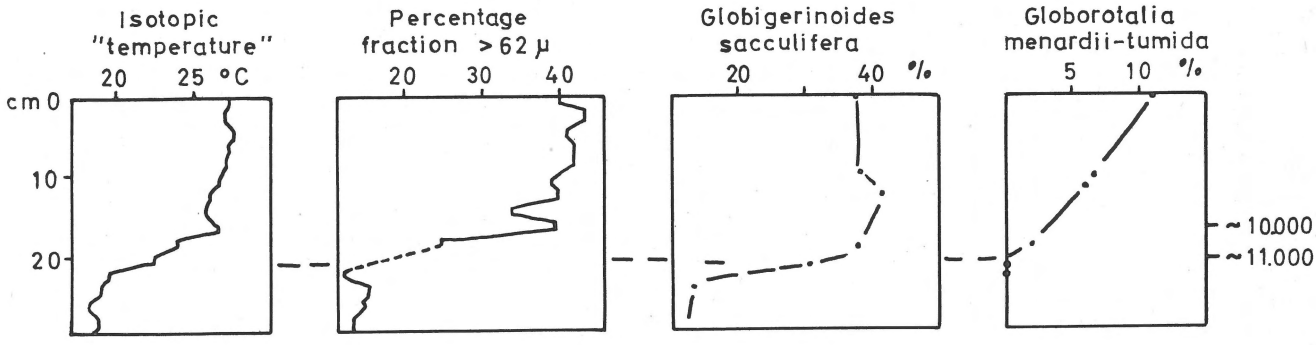


Fig. 5

Core 235 A from 3°12′N, 28°28′W. Percentages of the fraction larger than 62 μ, isotopic temperatures obtained from *Globigerinoides sacculifera*, the frequency of the same species and of the *Globorotalia menardii-tumide* group. Data by Emiliani (1955a), Olausson (1965) and A. Andersson.

somewhere between 1.27 (a sea-level stand at 1.40 m below present mean sea level) and 1.4‰ (a sea-level stand at −154 m) if the average δ value of the ice sheets was −33‰ (Olausson, 1965).

During the deglaciation more than 75% of all meltwater was discharged into the North Atlantic and its adjacent seas. It means a volume of nearly $40 \cdot 10^6$ km^3. This water was spread out on its surface lowering both the salinity and the O^{18}/O^{16} ratio. The decrease of both in the world oceans began in the North Atlantic. It is then the question as to whether the mixing in the North Atlantic and between it and the other oceans was rapid enough or not, to prevent a too low a O^{18}/O^{16} ratio in the surface water of the North Atlantic.

At an early deglacial stage the input of water into the Gulf of Mexico may have been at about 5.10^3 km^3 annually above the present normal rate (Fairbridge, 1967, p. 471). Aforementioned figure is equal to a water layer of 3 m thick over the whole Gulf. Foraminifera from such a water would give too high isotopic "temperature". Later on the Laurentian ice sheet was drained eastwards. During intense deglaciation phases the North Atlantic received annually up to about 10,000 km^3 of meltwater (up to 18,000 km^3 according to Fairbridge, 1967). This water was discharged into the ocean particularly during summers. Due to the low O^{18}/O^{16} ratio of meltwater, say 30‰, its contamination effect was very large. If uniformly distributed all over the North Atlantic and adjacent seas, this amount could have lowered the O^{18}/O^{16} ratio by 1‰ of the uppermost 4 m of the surface water. The outflow of water from the North Atlantic takes place only at large depths. The surface water must therefore sink down to large depths before it can leave the North Atlantic. According to the more traditional concept a density increase by cooling during winters is fundamental to the formation of deep-water. The way the North Atlantic gets rid of the meltwater is thus not a simple one. A temporarily stopped or diminished rate of sink water formation – e.g. due to a meltwater supply[2]) and an on the whole increasing

[2]) Deffeyes estimates (Wey l, 1968) that complete blocking of the Panama water vapor flux for 600 yrs would remove the salinity contrast between the Atlantic and Pacific. The meltwater addition to the North Atlantic may have been up to 3 times (6 times according to Fairbridge, op. cit.) larger than Deffeyes' figure for the size of aforementioned flux. If the Panama water flux is responsible for the salinity in the North Atlantic being about 1.5(1.5–2)‰ higher than at corresponding latitudes in the Pacific, then, during deglacial phases, the salinity of the North Atlantic surface water would have been at least as low (or lower) than that of the North Pacific.

The decrease of the average salinity during deglacial times, about 1.5‰, as an average for the world ocean, began in the North Atlantic and spread to the other oceans. This decrease in salinity (density) has probably retarded the vertical circulation in the North Atlantic and may have contributed to the concentration of the isotopically light meltwater in its surface water and to the rapid increase of carbonate accumulation at its floor.

temperature – together with convection within the surface water may make the whole surface water mass markedly contaminated. A stopped ventilation during, say, 50 years might have been enough to create a water layer, about 200 m thick, over the whole North Atlantic with a O^{18}/O^{16} ratio up to 1 per mil lower than the "average", i.e. showing up to 5°C too high an isotopic paleotemperature and lowering the salinity by some 1.5 per mil. It remains to be explained by those who deny my meltwater contamination theory how all living foraminifera could have been prevented from being influenced by meltwater masses.

The contamination, of course, has varied in space and has been particularly large near the outflow of glacial water. In the Eastern Mediterranean the meltwater contamination (from the Black Sea area) was at least 2.6‰ during the Riss deglaciation (O l a u s s o n, 1965), which is also admitted by E m i l i a n i (1955b). For the southern North Atlantic 0.5-1 per mil is suggested. The contamination may have been less pronounced within the North Atlantic anticyclonic area, from the periphery of which Core 280 was raised (*cf.* E m i l i a n i 1966, p. 115).

During deglacial phases the sink water from the northernmost Atlantic must have had a low O^{18}/O^{16} ratio. The isotopic "temperature" of benthonic foraminifera given by E m i l i a n i (1955a, Table 14) for Cores 234 and 246 are too high, because the correction applied by Emiliani is too small. The bottom water in the Early Flandrian was therefore much more contaminated by meltwater than Emiliani believes. Assuming a constant bottom temperature through the Flandrian, a correction of –1.3‰ instead of –0.3 per mil is justified for Core 246, sample 5-15 cm; for Core 234 an additional contamination of 0.5-1‰ for that time is likely.

The rapid rise in the isotopic curves at the Würm-Flandrian transition is thus interpreted as due to meltwater contamination. Such a contamination is probable for every large deglacial phase. This statement suggests that the isotopic curve of North Atlantic cores cannot be used to distinguish interglacials from interstadials with intense deglaciation (O l a u s s o n, 1967, p. 261). We have to apply a larger correction figure for every deglacial section of the isotopic paleotemperature curve than suggested by Emiliani. A corrected curve for the Würm-Flandrian would have a sinusoidal trend with a maximum around the climatic optimum and with a glacial-interglacial amplitude of 0-3°C for Central and Southern North Atlantic.

As pointed out earlier (O l a u s s o n, 1965, p. 233), the O^{18}/O^{16} ratio of the surface water in the Pacific might have been larger in the Eem than in the Early Würm interstadials, the opposite case of that of the North Atlantic. This too is an illustration of the fact that the mixing rate of the oceans is finite.

GULF OF ADEN

A yearly fluctuation in the supply of eolian matter to the Gulf of Aden causes microstratification with an average thickness of 1 mm per microvarve. The varve stratigraphy arrived at is connected to the C^{14}-time scale.

It is shown (O l a u s s o n and O l s s o n, in press) that there is a reverse correlation between the content of eolian matter and the phosphate content. The phosphate curve could thus mirror the amount of precipitation in adjacent arid lands during the Late Pleistocene.

There is a marked drop in the precipitation during or after the Alleröd subage. Preboreal was the driest subage during the Flandrian and a time of great wind erosion. A marked increase of the precipitation occurred during the Atlantic subage. The first millenium of the Subboreal subage was comparatively dry whereafter (~ 1800 B.C.) a slight increase of precipitation occurred. A new decrease in rainfall gradually took place between some 750 and 500 B.C. Further details can be found in Fig. 6.

RED SEA

During a glacial age the sea-level may have dropped down to or below the sill depth (125 m). No surface inflow could then have compensated for the evaporation loss within the Red Sea. Supposing an evaporation rate during glacials of 2 m per yr, most of the present Red Sea area (about 90%) was dry after, say, 500 yrs, and after, say, 1100 yrs of isolation only some deep holes contained brine as long as connate water was added from below. Since a sealevel lowering down to about –154 m is probable (S h e p a r d and C u r r a y, 1967; D o n n *et.al.*, 1962; R o b i n, 1964) and a lowering and rise of 2 cm per yr seems to be plausible the time available for evaporation is about 3 times as large as necessary for a complete

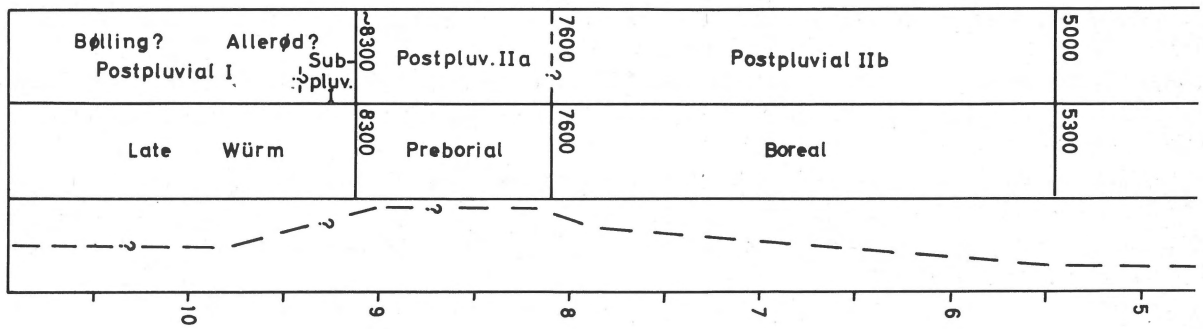


Fig. 6

Chronology of Core 162 from the Gulf of Aden. A curve of the precipitation in the areas near the Gulf is given. The same curve can also illustrate the size of the supply of wind-transported matter.

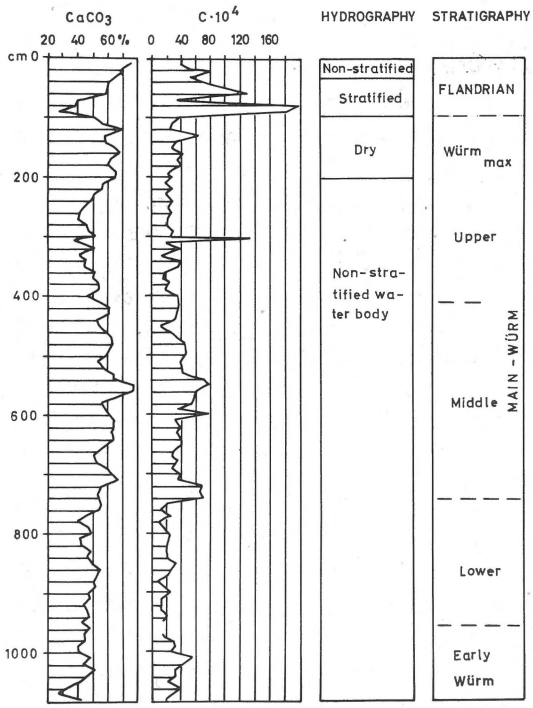


Fig. 7

Chemistry and stratigraphy of the Red Sea core 165. The hard layer sequence corresponds to the dry period in the hydrographical column.

evaporation. A hard layer sequence (sabkha or dissolution breccia) is found over large areas of the Red Sea floor and can hardly be attributed to other processes than an evaporation. It seems to have been formed between 16,000 and 22,000 B.P. The maximum extent of the ice sheets is supposed to have occurred around 19,000 B.P.

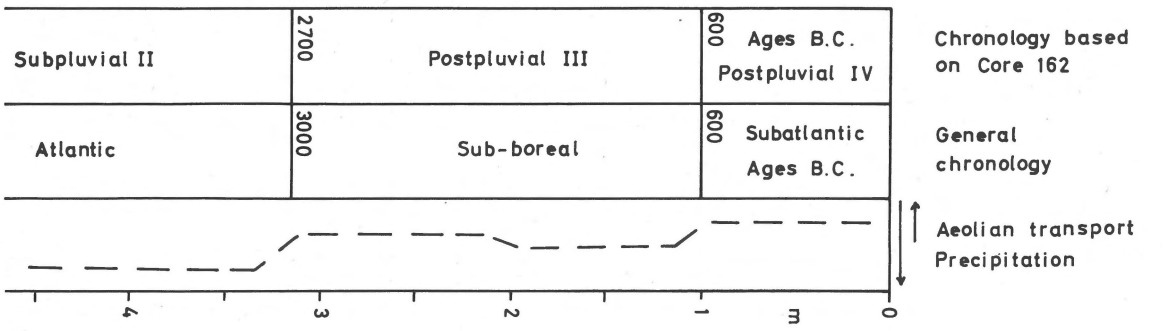
The limy crust may have been crushed by the wave action during the transgression. The rock salts precipitated below about 1000 m depth have been dissolved. This dissolution has increased the salinity of the water in the median valley which may have reached up to about 30‰ while the water above may have been less dense. A stagnant phase began. This stagnation lasted to approximately 8000 B.P. (Fig. 7). Then normal conditions entered. The brines found in the central part is interpreted by the author as a rest of the brine which filled most of the median valley in the Late Würm – Early Flandrian.

Further details could be found in other papers (O l a u s s o n, 1966, 1969).

EASTERN MEDITERRANEAN

The Eastern Mediterranean cores consist of alternating layers of oxidized and reduced deposits suggestive of cycles in the oxygen content at the bottom. A renewal of the bottom water is possible if the density of the surface water becomes larger than that of the bottom water. During the time from a thermal maximum to a succeeding glacial maximum the density of the surface water is increased due to a temperature lowering (3-4°C) and a globally increasing salinity (up to about 1.5 per mil in the world oceans). This increase results in a good ventilation of the basin.

However, during deglacials and early interglacials as well as in the Early Flandrian (transgression phases) the increase in temperature and excess supply of fresh and less saline waters (meltwater from the



Black Sea, precipitation, and Atlantic waters) have tended to increase the density difference between the bottom and surface waters in various degrees, causing stagnant conditions. A temperature rise from 8 to 12°C must be compensated by a salinity increase by 0.8‰ if the density of the surface water is to remain constant; since, instead, the salinity decreased, the ventilation was halted. The oxygen can be used up within some 1000 yrs and black, sapropelitic mud accumulates.

The Post-Würmian stagnant phase of the Eastern Mediterranean began approximately at the Alleröd – Younger Dryas boundary (see further O l a u s s o n, 1961a, 1966). The euxinic phase lasted about 4000 yrs. Fig. 8 shows the distribution of sapropelitic mud in Eastern Mediterranean cores and its correlation with climatic curves from other regions.

CARIACO TRENCH

The Cariaco Trench is at present anerobic beneath 400 m. According to H e e z e n *et.al.* (1959) the stagnation of the trench occurred 11,200 yrs B.P. Thus, the warming up of the basin created the anaerobic conditions which began during or at the end of the Alleröd subage.

THE YOUNGER DRYAS AND THE WÜRM–FLANDRIAN BOUNDARY

It is suggested above that the change into a pack-ice phase of the Arctic Ocean caused a climatic deterioration which can be the Younger Dryas. A decrease of the total heat absorbed by the Earth by about 0.5% is probable. If also the Norwegian Sea as a result of this cooling was temporarily ice-covered (O l a u s s o n and J o n a s s o n, in press) the effect have been somewhat larger.

The Arctic Ocean receives radiation from the sun

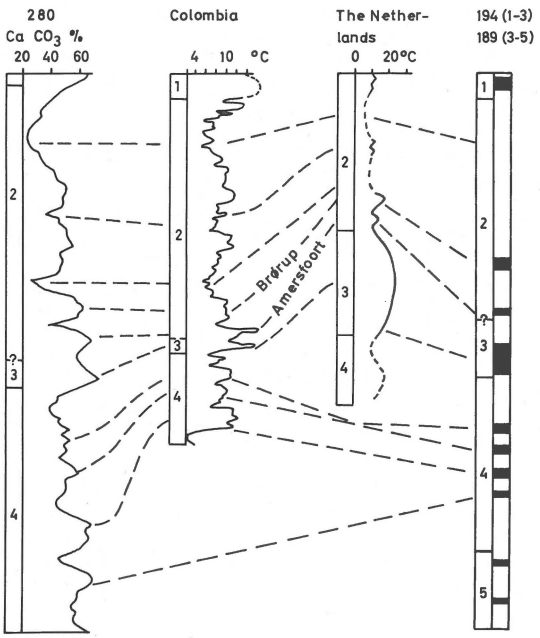


Fig. 8

The figure consists of a) the carbonate curve of Core 280 (O l a u s s o n, 1960) from 34°57′N, 44°16′W; b) the curve of annual average temperature for the Upper Pleistocene in the Sabana de Bogotá (v a n d e r H a m m e n, 1961); c) the estimate of mean temperature in July, The Netherlands (Z a g w i j n, 1963); d) the distribution of sapropelitic muds in the Eastern Mediterranean cores 194, stages 1-3, and 189, stages 4-5 (O l a u s s o n, 1961a, 1965).
Stage 1 = Flandrian, 2 = Würm, 3 = Eem, 4 = Riss and 5 = Holstein.

during half of the year (northern summer). Thus the result will be cooler summers while during winters the temperature may have been the same during the Younger Dryas as during the Alleröd time. It seems also possible that the temperature decline chiefly or only occurred in the northern hemisphere. The winters at the southern hemisphere can theoretically have been slightly cooler if the exchange across the tropical zone was large enough; however, summers need not to have been cooler.

Butzer (1958) states that a moderate increase of precipitation was apparent in the Near East during the Younger Dryas, accompanied by lower temperature effective to at least as far south as latitude 33°N. The core from the Gulf of Aden (Fig. 6) seems rather suggest a decrease than an increase of precipitation in adjacent arid lands during this subage. Coetzee (1964) studied sediments of Sacred Lake on the north-east flank of Mount Kenya (37°32'E, 0°03'N; altituda of 2440 m). Her analyses indicate a rising tree-line during the Younger Dryas without any temperature decline. Van Zinderen Bakker (1964) has published a pollen diagram of a peat profile from the same region. The diagram is similar to Coetzee's with some small differences concerning the proportion between *Graminaea* and *Compositae* in the Younger Dryas – Early Atlantic substages, probably indicating only very local changes at or near the swamp studied. Van der Hammen and Gonzalez (1965) have studied a sequence from Colombian Eastern Cordillera (ca 6°N) and in this the Younger Dryas is represented by a sandy deposit. The change in vegetation is small and occurs gradually from the Alleröd (Bölling) subage up to the end of the Preboreal. In another pollen diagram the Younger Dryas section shows the same vegetation as during the Flandrian while the change in the pollen rains occurred during Bölling-Alleröd subages (Gonzalez *et.al.* 1966, Section VL-V). Consequently, a climatic deterioration during the Younger Dryas is not yet proved for the equatorial belt of the Earth.

As to the southern hemisphere the data about the Late Würm-Holocene climate are sparse. A parallel in the Flandrian climatic history between the northern and southern hemisphere was drawn by von Post (1946) but the synchrony attached to the climatic zones was only a surmise. C^{14}-dated profiles are the only facts on which we can base a discussion about this problem and their number is insufficient.

A pollen diagram of La Misión, Tierra del Fuego, shows a marked increase of rain forest species and four *Nothofagus* species in a section which, according to an extrapolated C^{14}-time scale, may be Younger Dryas in age (Auer, 1965). A similar pollen flora occurs also in the Boreal stage. However, it is very difficult to deduce what changes in the diagram are due to changes in the depositional environments and/or fluctuations of the water table of the adjacent land area, and what are due to climatic changes.

In summary it can be said that the climatic deterioration during the Younger Dryas is in evidence at the northern hemisphere as far south as latitude 33°N. South of this latitude the climate during the Younger Dryas could have been similar to that during the Alleröd or somewhat warmer.

If we try to translate a decrease in the heat absorbed into a semiglobal temperature drop during the summer half year, a decrease by 2°C is likely. The drop must have been larger at higher latitudes than at lower latitudes. According to various authors the mean July temperature in the North Europe dropped 3-5°C during the Younger Dryas (see *e.g.* van der Hammen *et.al.*, 1967, Fig. 8). This is in good agreement with my theory of the cause of the climatic change at the Alleröd-Younger Dryas boundary.

It is necessary that the climate during the Late Würm-Early Flandrian at lower latitudes and in the southern hemisphere is clarified by further studies before the boundary Würm-Flandrian is defined. In this paper several indications are given about large changes in deep-sea sediments during or at the end of the Alleröd subage. It is also stated that so short an interval as Younger Dryas is hard to trace in oxidized deep-sea deposits. If we consider the Younger Dryas climatic deterioration as only a semi-global event which more or less is recognized only in continental and shelf deposits from the northern hemisphere, than a global Würm-Flandrian boundary can be tentatively placed at the transition Earlier Dryas/Alleröd.

ACKNOWLEDGEMENT

This study has been supported by the Swedish Natural Research Council.

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