NORTH ATLANTIC CLIMATIC OSCILLATIONS REVEALED BY DEEP GREENLAND ICE CORES

W. Dansgaard, S.J. Johnsen, H.B. Clausen, D. Dahl-Jensen, N. Gundestrup and C.U. Hammer

Geophysical Isotope Laboratory, University of Copenhagen, Haraldsgade 6 DK-22000 Copenhagen, Denmark

H. Oeschger

Physikalisches Institut, Universitat Bern, Sidlerstrasse 5, CH-3012 Bern, Switzerland

Abstract. Five long-term oxygen isotope (δ) records along ice cores are discussed, in particular two from the Greenland ice sheet that exhibit persistent δ oscillations with a quasi-periodicity of ca. 2550 years. A detailed study of the δ cycles in the Wisconsin glaciation show that they cannot be ascribed to discontinuities in the cores, nor to ice-dynamic instabilities in the ice sheet. In the Holocene, the δ cycles are less pronounced, but they are concurrent with the fluctuating glacier extention elsewhere, which substantiates their climatic significance. An anti-correlation with ^{14}C concentration in atmospheric CO_2, and with ^{10}Be deposition rates on the ice sheets, suggests a connection between climate and solar processes, but a conclusion on this point must await clarification of the terrestrial circulation and mixing processes, and the relationship between the solar outputs of radiation and particulate matter.

Introduction

The American-Danish-Swiss joint effort, Greenland Ice Sheet Program (GISP, 1971-81), emerged from the successful deep ice core drilling by U.S.A. C.R.R.E.L. at Camp Century, NW Greenland, in 1966 (Hansen and Langway, 1966), and from C.C. Langway's (1970) studies on the ice core from the nearby Site 2 that demonstrated the great ice sheets as rich sources of information on past environmental conditions.

The main objectives of GISP (1976) were to extract this paleoenvironmental information by drilling and studying new ice cores and to survey the Greenland ice sheet with a view to ice flow modelling.

Prior to 1978, most of the ice sheet was surveyed by airborne radio-echo sounding equipment (Gudmandsen, 1976). Three ice cores were drilled to 400 m depth by a thermal drill (Ueda and Garfield, 1969), and a number of cores up to 110 m were drilled by other techniques (Rufli et al., 1976; Johnsen et al., 1980) at the locations shown in Fig. 1. The GISP field activities culminated in 1981, when a new deep drill (Gundestrup et al., 1983) penetrated the ice sheet and reached bedrock 2037 m below surface at Dye 3, South Greenland. In parallel with the field activities, extensive stratigraphic, chemical and physical studies were performed on the ice cores, and ice flow models were developed. The initial results were presented at a GISP symposium in May 1982 (Langway et al., 1983).

According to the simple Rayleigh condensation model, the isotopic composition (oxygen-18 or deuterium concentration in the δ scale) of a given polar snow fall depends on several parameters, of which the most important one is cooling of the precipitating air mass, since the last substantial uptake of water vapor from the ocean. As the seasonal or longer term climatic conditions in the source area of the vapor are considerably more stable than those at high latitudes, the isotopic composition of the polar snow is strongly influenced by the temperature at the site and time of snow deposition (Picciotto et al., 1960; Dansgaard, 1964). With some reservations (Johnsen et al., 1972; Johnsen, 1977) $\delta^{18}O$ profiles along polar ice cores therefore exhibit seasonal oscillations (Epstein and Benson, 1959; Hammer et al., 1978) and changes due to general climatic temperature variations (see e.g. Dansgaard et al., 1973; 1975).

However, the δ profiles may also be influenced by varying summer to winter precipitation ratio; by surface elevation changes, which cause surface temperature variations that are not necessarily connected to general climatic changes; and by changing sea-ice cover, i.e. changing distance to the open ocean, which may affect the degree of cooling of precipitating air

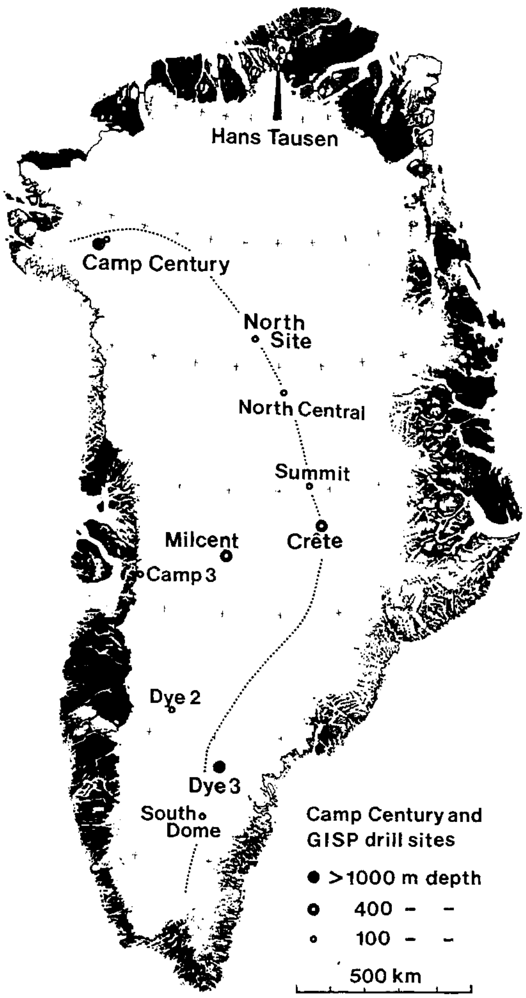

Fig. 1. GISP drill sites. The dotted curve shows the main ice divide that encloses the West greenland discharge area.

masses, but not necessarily the mean air temperature on the ice sheet, at least not to the same degree. In general, it should be kept in mind that the isotopic composition of a given ice core increment may give a hint as to the climatic conditions that prevailed during the relatively short periods of snowfall when the material was deposited, but it tells nothing about the much longer periods of dry weather conditions inbetween.

The concentration of solid particles (mainly continental dust) in Holocene ice in Greenland is of the order of 0.05 mg/kg of ice, which is 6 times higher than in W. Antarctica, but one to two orders of magnitude lower than in Greenland ice from late Wisconsin (Cragin et al., 1977; Thompson, 1977; Petit et al., 1981; Hammer et al., 1983). An important reason for the extremely high Greenland dust contents around the glacial maximum is that the strong meridional temperature gradients (CLIMAP, 1976) and the consequent high storminess at mid-latitudes (Ruddiman and Glober, 1975) created rich sources of aerosols in large, dry areas covered by glacial outwash south of the North American and Eurasian ice sheets (Schultz and Frye, 1968). But, the high content of calcium rich carbonates in late Wisconsin ice suggests that a new source of alkaline aerosol was formed, when the sinking sea level left vast high- and mid-latitude areas of the continental shelves exposed to scouring (Cragin et al., 1977; Hammer et al., 1983).

The latter source was cut off, when the sea level began to rise, but the former areas expanded during the disintegration of the ice sheets and, according to Ruddiman and McIntyre (1981), they stayed dry and cold up to 13 ka B.P. (1 ka (kiloannum) B.P. = 1000 years before present). In the same period (20 to 13 ka B.P.) an intensified supply of icebergs from the disintegrating ice sheets cooled off the North Atlantic surface water (Denton and Hughes, 1981), thus reducing the moisture content of maritime air masses and, hence the precipitation at high latitudes (Ruddiman and McIntyre, 1981). For any given annual fall-out of dust, lower annual accumulation on the Greenland ice sheet is, in itself, a contributory cause for higher dust concentration in the ice.

Even since the Wisconsin glaciation terminated, the deposition of dust in Greenland has varied seasonally with a maximum preferably in spring, probably because the stationary anti-cyclone and the associated high tropospheric winds bring dust from spring-dry areas at mid latitudes. The oscillating deposition of dust provides an alternative means for annual layer identification and, thereby, for absolute dating of ice cores (Hammer et al., 1978).

Also the acidity of the deposited snow varies with the season, which can be used for annual layer identification in periods of no or low volcanic activity (Hammer, 1980). Large and violent volcanic eruptions in the northern hemisphere leave a high acidity signal in the ice due to washout of, particularly, sulfuric acid and hydrochloric acid from the atmosphere (Hammer et al., 1980). However, the seasonal acidity variations, and even the volcanic signals are suppressed in ice deposited in Greenland during most of the Wisconsin glaciation. This ice is alkaline due to the then high alkaline aerosol load that obviously neutralized the atmospheric acids, at least at high northern latitudes (Hammer et al., 1980).

Absolute dating of Holocene ice in Greenland by the stratigraphic methods outlined above is laborious, but very accurate and, in principle, simple. These methods are being applied on the new Dye 3 core, which has so far been absolute dated back to ca. 4 ka B.P., but data on the earlier part of the Holocene are not yet available. As to the Camp Century ice core, annual layer thicknesses measured on essentially all suitable Holocene increments defined the input parameters of a simple steady state ice flow model (Dansgaard and Johnsen, 1969), which was subsequently used to date the younger part of the Camp Century record (Hammer et al., 1978). The dating accuracy was estimated at a few percent back to 8.5 ka B.P., but beyond 10 ka B.P. the steady state assumption behind this time scale is probably unrealistic.

The early attempts to date the Wisconsin part of the Camp Century record by various techniques were discussed by Dansgaard et al. (1982), who outlined a new tentative Camp Century time scale by correlating the major trends in isotope profiles along ice cores and deep sea cores. The new time scale lower glacial than post-glacial accumu-

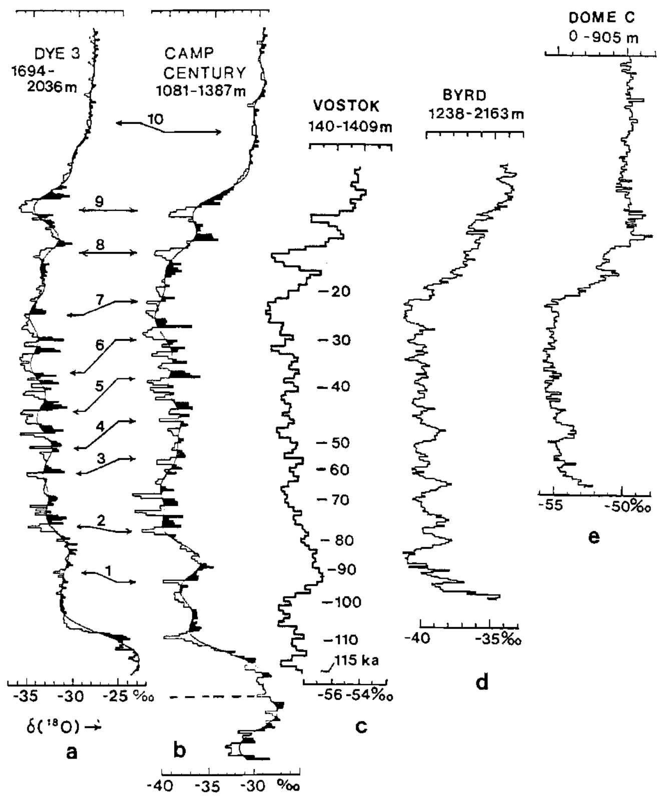

Fig. 2. Continuous $\delta(^{18}O)$ records along five deep ice cores, plotted on linear depth scales and spanning the depth intervals indicated in the upper part of the figure. Each of the samples represent increments of 1 meter in the two Greenland records (a and b), 10 meters in the Vostok record (c); 4 meters in the Byrd Station record (d); and generally 3-4 meters in the Dome C record (e). The arrows between (a) and (b) indicate common high frequency features, used to transfer the new Camp Century time scale to the Dye 3 core, cp. Fig. 3. The Vostok time scale shown to the right of curve c was calculated by Gordienko et al. (1982) on the basis of annual layer measurements by Wilson and Hendy (1981). It does not deviate considerably from the new Camp Century time scale shown in Fig. 3, if the δ maximum around 90 ka B.P. is interpreted as stage 5a.

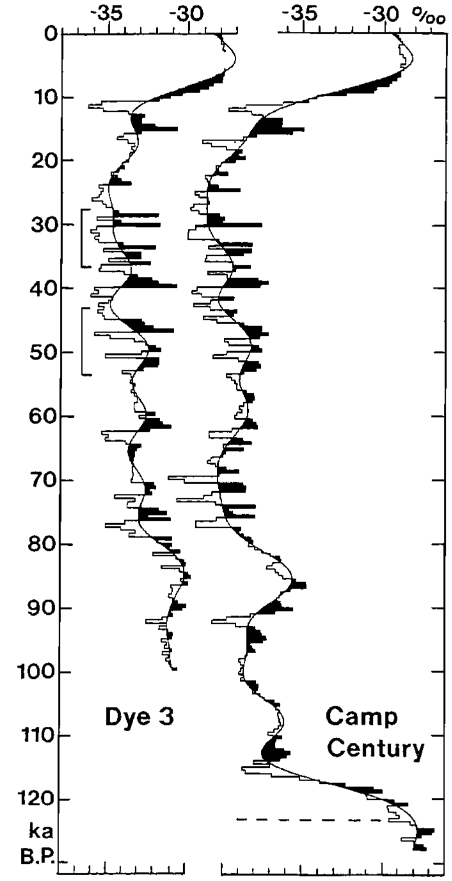

Fig. 3. $\delta(^{18}O)$ profiles along the Dye 3 (0 to 1982 m depth) and the Camp Century (0 to 1370 m depth) ice cores plotted on a common linear time scale as outlined in the text.

lation, and sliding of the ice over the bedrock or at least fast deformation of the ice close to the bedrock. Its accuracy is hardly better than ± 5 ka for ice older than 20 ka, even if the suggested correlation proves to be correct. In turn, the two Greenland δ records (plotted on linear depth scales in Fig. 2a and 2b) were correlated by their common high frequency features indicated by the arrows, and the new Camp Century time scale could then be transferred to the Dye 3 record. In Fig. 3 the datable parts of the two data sets are plotted as functions of time. The smoothed curves are obtained by low pass filtering technique (with sharp cut off of all cycles shorter than 20 steps (10 ka), and less than 10% leakage), combined with a maximum entropy prediction filter (Andersen, 1974) that allows extention of the smoothed curves through the entire range of the raw data.

The new time scale puts all Camp Century δ values higher than those in the Holocene maximum into the 130-120 ka B.P. time interval (Eemian interglacial, or Emiliani (1966) Stage 5e). This is consistent with McIntyre et al.'s (1972) evidence that in this very period subtropical Atlantic

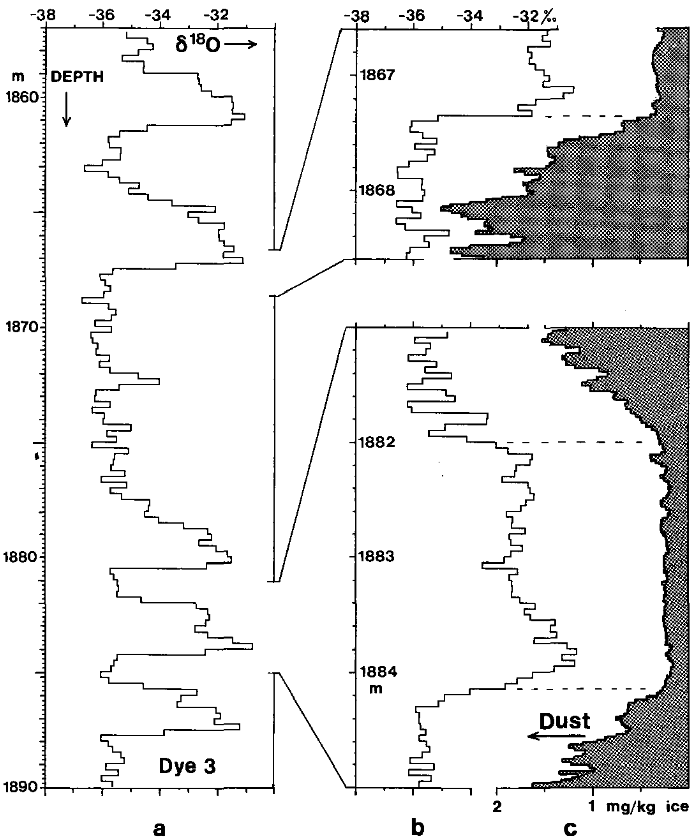

Fig. 4. a: Details of the δ oscillations marked to the left of the Dye 3 record in Fig. 3. b: Further details of some of the δ shifts. c: Concentration of insoluble microparticles. Notice, increasing values towards the left.

water masses advanced to more than 50°N in the Atlantic ocean for the first and only time in the last 225 ka. Accordingly, the deepest parts of the Dye 3 (Fig. 2a) and Devon Island (Paterson et al., 1977) ice cores with more than 2^{0}/oo higher δ's than in the Holocene were probably deposited in the Eemian interglacial.

Obviously, the Pleistocene to Holocene shift in δ is 7^{0}/oo (per mille) at Dye 3, but 11 to 12^{0}/oo at Camp Century. These shifts were discussed by Dansgaard et al. (1983), who suggested that the termination of the Wisconsin glaciation is marked by a δ-increase of generally 6 ± 1^{0}/oo in ice cores from the main parts of the ice sheets in Greenland and Antarctica, cp. Fig. 2c: Vostok, East Antarctica (Gordienko et al., 1983); Fig. 2d: Byrd Station, West Antartica (Johnsen et al., 1972); and Fig. 2e: Dome C, East Antarctica (Lorius et al., 1979). Furthermore, they suggested that the higher δ-shifts found around the Baffin Bay (11-12^{0}/oo at Camp Century; 10^{0}/oo on Agassiz ice cap, Ellesmere Island (Fisher et al., 1983); 15^{0}/oo on Barnes ice cap, Baffin Island (Hooke, 1976) may reflect that the disappearance of the North American ice sheet made the climatic change more drastic in this area than in Southeast Greenland or in Antarctica.

Abrupt Environmental Changes in the Wisconsin

The mid and late Wisconsin from 77 to 10 ka B.P. is characterized by large amplitude oscillations of $\delta^{18}O$ in both of the records in Fig. 3. In the Camp Century record they are present even before 90 ka B.P., yet of lower amplitude. Although the transfer of the new Camp Century time scale to the Dye 3 record was based on the assumption that 9 characteristic features in the two records could be correlated (cp. Fig. 2a and 2b), the synchroneity between the high frequency oscillations is considered real, firstly because the said procedure results in a correlation coefficient of no less than 0.88 (P > 99.98 % for estimated 10 degrees of freedom) between the filter-smoothed curves in Fig. 3; secondly, because the shape and the sequence of the δ peaks and valleys are often so similar in the two records that it cannot be due to a coincidence, cp. for example, the transition from a relatively quiet δ record below arrow No. 2 to the region of vigorous oscillations above it, beginning with a characteristic bipartite δ maximum; and the sequence of a broad peak followed by three sharp ones in the interval between arrows No. 5 and 6, which, in addition, adjoins relatively broad intervals of generally extremely low δ's; cp. also the shape of the δ maxima between arrows No. 8 and 9.

In a large part of the Dye 3 record, the δ's have a tendency to alternate between two levels of approximately -35.5^{0}/oo and -32^{0}/oo. Fig. 4a shows a detailed version of the δ oscillations in the Dye 3 core from 1857 to 1890 m depth, corresponding to times of deposition from approximately 28 to 37 ka B.P., according to the time scale in Fig. 3. Obviously, the peaks are saw tooth shaped with considerably more abrupt δ shifts towards "warmer" values. All of the δ peaks are provided with a secondary peak or, at least, a "shoulder" on the younger side.

The abruptness of the δ shifts may be connected to fast changes of the mean latitude of the polar front in analogy with the event 13 ka B.P. described by Ruddiman and McIntyre (1981). Once started by some kind of external forcing (varying insolation or volcanic activity?), the effect of a displacement of the polar front may be amplified by feed-back mechanisms, like the one suggested by Flohn (1982): "...in the case of a cooling ...meridional temperature gradients will become greater, latitudes of the subtropical anticyclones become lower, intensity of cell winds (trade) increases, equatorial sea surface temperature becomes cold, atmospheric content of CO_2 and H_2O becomes lower, thus resulting in further cooling". Although the importance of the greenhouse effect is not yet clarified, it should be mentioned here that CO_2 measurements on the Dye 3 core show that the CO_2 concentration in the atmosphere varied considerably and in phase with the δ oscillations in the Wisconsin glaciation (Stauffer et al., 1983).

Continuity of the Time Scale

On the face of it, a possible explanation for the abrupt alternations would seem to be that, somehow, two homogeneous ice masses of different isotopic compositions were folded repeatedly into each other, which would invalidate any continuous time scale along the core. However, in view of the completely different ice flow conditions in Southeast and Northwest Greenland, the close correlation between the Dye 3 and Camp Century δ records is strong evidence against the folding hypothesis.

But, this argument does not exclude that the δ shifts indicate "holes" in the time scale. Such "holes" might be caused either by essential cease of Greenland accumulation, or by negative accumulation (due to wind erosion, as observed today in some areas close to Vostok, East Antarctica; cp. Barkov, 1975) through long periods of time till the onset of different climatic conditions, including resumed accumulation of ice of different δ. In both cases, the other environmental characteristics of the ice (concentrations of dust, sea salts etc.) must be expected to shift exactly at the same depths as does the isotopic composition, though not necessarily with the same abruptness, because the various isotopes and impurities have migrated differently since the time of deposition, depending on their diffusion coefficients in the ice. Hence, the continuity problem may be solved by comparing high resolution δ and impurity profiles across the δ shifts:

The sharp δ shift at 1867.35 m depth, and the entire δ oscillation between 1881.0 and 1885.0 m depth are shown in great detail (5 cm samples) in Fig. 4b along with a high resolution dust concentration profile in Fig. 4c (2.5 cm samples). Both of them are typical for the other oscillations studies in detail so far. The dust concentrations (in mg dust per kg ice) were measured in the field by a light scattering technique (Hammer et al., 1983) applied on melted ice soon after the recovery of the ice core increments in question. The technique involved some degree of "cross-talk" between adjoining samples, and it may therefore be more realistic to consider the 5 cm resolution in the δ profile as valid for the dust concentration profile as well.

In general, high dust concentrations correspond to low δ values, and vice versa, in ice from mid and late Wisconsin (Hammer et al., 1983). This is, in part, due to colder climates being dominated by higher storminess at mid latitudes, but changes of the general atmospheric circulation pattern, and of the extent and location of the source areas of the dust may also be important.

However, Fig. 4b and 4c show the following features: (i) by the end of the periods of extremely low δ's the dust concentration begins to decrease (notice the scale at the bottom of Fig. 4c) long before the δ shift; (ii) it reaches the new level before or at the same time as the δ shift is completed; (iii) the dust concentration stays low till after the onset of the less well defined δ shift back to extremely low values, in fact till δ has dropped below $-34^{0}/00$; and (iv) the total change in dust concentration extends over a longer core increment (more than 0.6 m) than does the corresponding δ shift (less than 0.3 m).

These features indicate that the δ shifts are not due to discontinuities in the time scale, because, as mentioned above, such discontinuities would cause δ and dust concentration shifts at exactly the same depths, and with <u>smaller</u> δ than dust concentration gradients, since the diffusion coefficient of insoluble microparticles in the 0.1-2 μm size range (Hammer et al., 1983) is negligible, and at least much lower than that of $H_2{}^{18}O$.

Periodicity and Persistency of δ Fluctuations

The duration of a given δ oscillation cannot be determined until new techniques are developed to measure annual stratification in Wisconsin ice in Greenland. The new tentative time scale was established by considering the long term δ trends only, and should therefore not be used to estimate the duration of, for example, the δ oscillations shown in Fig. 4, much less the duration of the δ shifts.

However, the Camp Century record contains 18 oscillations from 77 to 30 ka B.P. in the new time scale (Fig. 3), i.e. one per 2600 yrs. This leads back to the assumed 2400 yr periodicity in climate that was previously used (Dansgaard et al., 1971) to correct Dansgaard and Johnsen's (1969) initial Camp Century time scale (the reason why the result differed considerably from the time scale in Fig. 3 is probably that the initial one deviated too much from absolute chronology to serve as a basis for the correction procedure).

Using data from a deep sea core back to 133 ka B.P., Pisias et al. (1973) demonstrated a similar periodicity in ocean surface summer temperatures west of Ireland (53°N, 22°W). Fourier spectral analysis of palaeotemperatures derived from palaeoecological transfer functions (Imbrie and Kipp, 1971) "shows a very significant peak representing 2600 years". In interglacial times, the sea surface temperature oscillations are less pronounced than under glacial conditions, and this feature is also present in the ice cores. It appears from the Camp Century record in Fig. 3 that the δ amplitudes are modulated by the δ level: The highest amplitudes occur at extremely low δ values, whereas the δ cycles are hardly recognizable in the high δ range in the Holocene (the Dye 3 record is not yet completed).

Nevertheless, as shown by Dansgaard et al. (1971) and Fisher (1982) the Holocene Camp

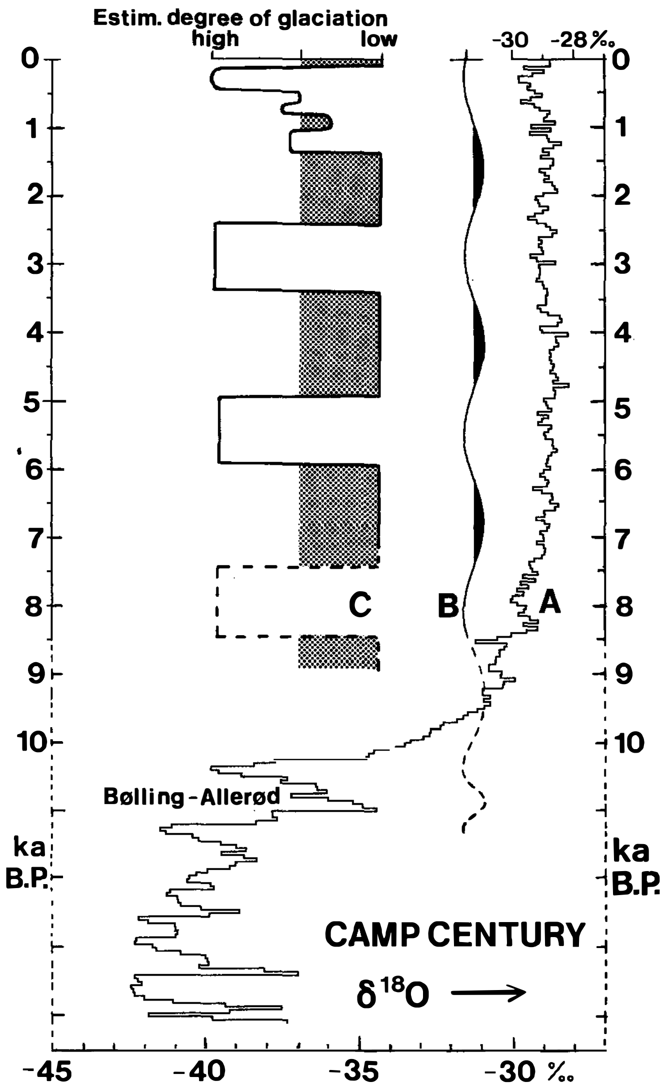

Fig. 5. A: The Holocene and late glacial part of the Camp Century δ record, plotted on a linear time scale that is close to absolute chronology back to 8.5 ka B.P. B: Band pass filtered version calculated for the last 8,500 yrs, and suggested by the dashed curve beyond this range, where the time scale is less accurate due to poorly known strain and accumulation rate. C: Estimated degree of worldwide glaciation (from Denton and Karlen, 1973).

Century δ record does contain an approximately 2500 yr cyclic component. The Holocene part of the Camp Century record is shown in the upper part of Fig. 5A. It has been dated by stratigraphic methods (Hammer et al., 1978) to an accuracy that is probably better than 2% back to 8.5 ka B.P. Fig. 6 shows a logarithmic power density spectrum of this part of the record calculated by the maximum entropy method (Ulrych and Bishop, 1975; Johnsen and Andersen, 1977). The 7200 yrs peak contains 50% of the power, but this is essentially a reflection of the long-term δ trend depicting the post-glacial climatic optimum in the middle of the Holocene. Apart from this trend, the 2550 yr oscillation is dominating with 17% of the power. A band-pass (2200 to 2800 yrs periods) filtered version of the raw data is shown with an amplification factor of 2 in Fig. 5B.

In times of transition between glacial and interglacial conditions the situation is less clear, i.e. because the time scale is most difficult to verify during drastic changes of the glaciological environment. Nevertheless, when comparing the curves in Fig. 5A and 5B, it is tempting to extend the latter one beyond the well-dated time interval as shown by the dashed curve, i.e. to add at least 1.5 more δ oscillations with maxima at approximately "9.3" and "10.9" ka B.P. on the dashed time scale. As shown by Siegenthaler et al. (1983), the latter maximum most likely corresponds to the late glacial Bolling-Allerod interstadial known from Western Europe (Iversen, 1954; Nilsson, 1968; Eicher and Siegenthaler, 1976; Watts, 1980) and the North Atlantic ocean (Ruddiman and McIntyre, 1981).

Assuming a constant δ periodicity of 2550 yrs and following the procedure of Denton and Karlen (1973), addition of two full periods to the oldest well-defined maximum in Fig. 5B (6.8 ka B.P.) puts an age of 11.9 ka B.P. on the middle of Bolling-Allerod. Furthermore, addition of a quarter of a period to the latter maximum gives 13.2 ka B.P. for the abrupt δ shift from extremely low preBolling δ values, which is in agreement with ^{14}C dating of the onset of Bolling elsewhere (Wegmuller and Welten, 1973) and of a radical change of the circulation in the North Atlantic ocean - atmosphere system, including a drastic northward displacement of the polar front 13 ka B.P. (Ruddiman and McIntyre, 1981). This suggests that the dashed time scale is ca. 2 ka off around the "11" ka B.P. mark, which is quite possible in view of the drastic changes of the ice flow conditions (accumulation, ablation, mechani-

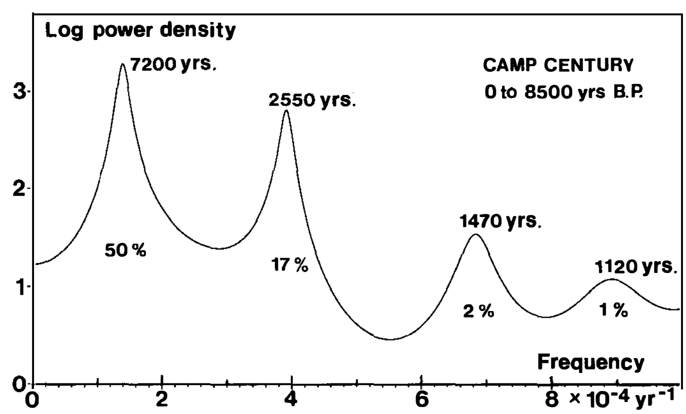

Fig. 6. Logarithmic power density spectrum calculated on Camp Century δ values back to 8,500 yrs by the maximum entropy method.

cal properties of the ice, cp. Hammer et al., 1978; Shoji and Langway, 1983; Gundestrup and Hansen, 1983) during the period of transition.

Hence, there are good reasons to believe that δ of Greenland snow oscillates persistently with quasi-periodicity of approximately 2550 yrs, and that the amplitude of the δ cycles is modulated by the degree of glaciation (and perhaps the latitude), being small under warm conditions and high under cold conditions.

Climatic Significance of the δ Cycles

The amplitude of the δ cycles in Greenland and on the Canadian Islands (Paterson et al., 1977); Fisher, 1982) was an order of magnitude higher in mid and late Wisconsin, than in the Holocene. The corresponding shifts in surface temperatures may also have been much higher, though it is difficult to tell how much, cf. the remarks in the 5'th paragraph of section 1 about δ profiles being influenced by factors that are not directly related to the local mean temperature of the troposphere.

One of these factors is surface elevation change that might happen in case of fast run-off of huge amounts of ice (surge) as a result of internal instability in the ice flow system, in analogy with observed surges of minor glaciers (Post, 1960; 1969; Dolgushin and Osipova, 1975). On the face of it, the abrupt ca. $5^{0}/_{00}$ δ increases in mid and late Wisconsin (Fig. 3) might be due to surges associated with several hundred meter lowerings of the surface elevation and, consequently, several degrees centigrade warmings of the surface.

On the other hand, as the ice flow conditions in Southeast and Northwest Greenland are quite different, it is not likely that the entire ice sheet could surge at one and the same time as a result of an internal instability alone. Therefore, general surges would have to be initiated by some external impact. For example, a substantial rise in sea level, caused by a great surge in Antarctica, would be extremely dangerous to the stability of at least the marginal zones of an expanded Greenland ice sheet, if they were resting on sub-sea level areas and fringed by ice shelves like most of West Antarctica today. The West Antarctica ice sheet is considered most vulnerable to surging, because it is grounded far below sea level (Wilson, 1964; Mercer, 1978). Therefore, it is remarkable that the abrupt mid-Wisconsin δ changes are not considerably higher in the Byrd Station profile (Fig. 2d), than in any other δ profile, including those along the Dome C and the Vostok (Fig. 2c and 2e) ice cores. This does not support the idea of surges as a cause for the fast δ increases.

Furthermore, it is difficult to envisage how the South Greenland ice sheet, which rests on high elevation bedrock, could surge to a degree that lowered the surface elevation close to the ice divide by many hundred meters. Even if the western edge of the ice sheet were located 170 km west of its present position, i.e. at the submerged moraine ridges found 50 km off-shore, the surface elevation at the ice divide would not be significantly higher than at present, according to 3-dimensional ice sheet model calculations (Reeh, 1982; 1983). A possible surge might disintegrate the marginal zone of an ice sheet advanced that far, but the edge would retreat to higher elevations, and the surge would be stopped by the blocking effect of the mountains. Far inland, including the area where the Dye 3 ice was formed, the stabilized ice sheet would be left with essentially unchanged surface elevation.

Finally, if the repeated δ shifts to higher values were to be explained by surges, the ice sheet must have rebuilt itself in between, i.e. the reverse δ trends would have to be ascribed to increasing surface elevation. However, build-up of an ice sheet is a very slow process (Weertman, 1964; Oerlemans, 1981) and although the δ decreases following the five peaks in Fig. 4a are more smooth than the δ increases, the five peaks terminate by abrupt drops in δ that correspond to considerable thickening of the central part of the ice sheet within a time frame of the order of a century. This implies a much higher accumulation rate than today, which is unrealistic under full glacial conditions.

Hence, surges are implausible as an explanation for the high frequency δ oscillations in the Wisconsin, that are rather to be explained by general climatic changes at high northern latitudes. This is the case with their successors at the end of the glaciation (Siegenthaler et al., 1983), and in the Holocene as well, judging from the concurrence of the curve in Fig. 5B with Denton and Karlen's (1973) independent estimate on the degree of glaciation (or rather deglaciation - notice the reversed scale on top of Fig. 5C), which is based on mainly geological evidence of worldwide (before 7 ka B.P. mainly Swedish) Holocene glacier fluctuations. According to Fig. 5C, considerable glacier expansion took place in periods of approximately 1000 years duration around 5500, 3000 and 500 years ago.

For obvious reasons, Denton and Karlen's estimate is most accurate and detailed in the latter period that spanned most of our own millennium, and culminated in the "Little Ice Age". It is interesting that the glacier advance lasted more than 1000 years, and was interrupted by halts and minor retreats, whereas the main retreat began around A.D. 1890 and was nearly completed 50 years later. In other words, the best documented glaciation cycle in Fig. 5C is saw-tooth shaped like the mid Wisconsin δ oscillations in Fig. 4a (the shape of the Holocene δ oscillations is doubtful, because the signal to noise ratio is small).

According to the conclusion in section 4, the abrupt temperature rise in the 1920's may thus be the latest member of a very long series of similar events occurring once every ca. 2550 years to an extent that is modulated by the degree of glaciation and dependent on the latitude: The 1920-30 temperature rise was 4°C in North Greenland, 2°C in South Greenland (Dansgaard et al., 1975), but only half a degree as an average over the northern hemisphere (Mitchell, 1963). Data from the southern hemisphere are sparse, but recent estimates (Hansen et al., 1981) suggest an 0.3°C warming through the last 100 years, interrupted by a minor cooling in the 1920's.

This suggests that short-term climatic changes in the southern hemisphere are not concurrent with, or at least that they are less abrupt than those in the northern hemisphere, probably due to the different land-sea distribution in the two sub-polar belts. There is also evidence that the post glacial climatic optimum occurred already between 11 and 8 ka B.P. in the southern hemisphere (see Lorius et al., 1979, and references therein).

It is still doubtful, if the southern hemisphere climate has a 2550 yr cycle. The Vostok record (Fig. 2c) is not yet detailed enough to show it, but the Byrd and Dome C records (Fig. 2d and 2e) both contain several oscillations that might reflect damped versions of this cycle.

Why?

The coupled atmosphere-ocean-cryosphere system may have a built-in feed-back mechanism that favors a 2550 yr cycle in the high latitude climate, but climate models are not yet developed to a stage that allows any conclusion on this point.

Another possibility is that some periodic external forcing is responsible for the climatic cycle, for example varying volcanic activity that modulates the load of aerosols in the stratosphere and, thereby, the solar radiation flux in the troposphere. However, although 30% of the climatic variability in the last 1400 yrs may be ascribed to varying volcanic activity (Hammer et al., 1980), there is no evidence that this activity is periodic. The only cyclic variations of the solar radiation flux in the atmosphere known at present are the Milankovich effects (cp. e.g. Berger, 1978), but they are too slow to be of interest in this context.

Since the solar radiation is the only important input of energy to the climatic system, it is most obvious to seek an explanation in solar processes (Eddy, 1977). Unfortunately, we know much less about the solar radiation output than about the emission of solar particulate matter in the past. The latter has modulated the production rate of cosmogenic isotopes (reduced production rates in times of high sunspot activity; deVries, 1958; Stuiver, 1961; Stuiver and Quay, 1981). There is evidence that the ^{14}C concentration in atmospheric CO_2, has varied in a non-random way in the past (see e.g. Neftel et al., 1981), and that its Holocene variations are in antiphase with climatic temperature changes (Suess, 1968; Sonett and Suess, 1983), including the 2550 yr cycle (Fisher, 1982).

Furthermore, the deposition rate of another cosmogenic isotope, ^{10}Be, on the ice sheets also seems to be in antiphase with climatic changes: Raisbeck et al. (1981) found 50% elevated ^{10}Be concentrations in Dome C ice from the 17'th century, i.e. more than can be explained by the estimated 25% lower accumulation rate in this cold period (Lorius et al., 1979). Beer et al. (1983a) have shown a similar feature in 17'th century ice from Milcent (70.4°N, 44.6°W), Greenland, where no significant accumulation changes occurred (Reeh et al., 1978). Since this period coincides with the Maunder sunspot minimum, increased ^{10}Be production rate is at least a contributory reason for the high concentrations. The ^{10}Be data so far obtained from the Dye 3 ice cores show a close anticorrelation with the δ's in late Wisconsin to early Holocene (Oeschger et al., 1983). The Younger Dryas value is 3 times higher than those from Bolling-Allerod and Preborial time. A similar ^{10}Be to δ anticorrelation has most recently (Beer et al., 1983b) been demonstrated to hold for the drastic δ oscillations in Wisconsin ice.

This is additional evidence that low solar activity in terms of particle emission is associated with cold climatic conditions. Hence, "whatever might cause changes in the ^{14}C level of the atmosphere CO_2 might well have an influence upon the global climate" (Suess, 1980).

But, as long as the long-term relationship between the solar outputs of radiation and particles is an open question (as to the short-term, i.e. a few years, relationship, see Eddy et al., 1982) and as long as the influence of shifting atmospheric circulation and mixing patterns is unknown, correlation of past cosmogenic isotope deposition rates with climatic changes contains several elements of speculation.

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