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STABLE ISOTOPE GLACIOLOGY

BY

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WITH 11 FIGURES AND 3 TABLES
IN THE TEXT

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Abstract

The development of the ice core drilling technique has led to a broad variety of studies reaching far beyond glaciology itself. Radioactive and stable isotope analyses of polar ice cores are particular promising.

Under cold climatic conditions, the relative concentration (δ) of the heavy stable isotopes oxygen-18 and deuterium in precipitation mainly depends on the temperature of formation. This leads to a geographical δ-distribution and, at a given location, δ variations in phase with seasonal and climatic changes. Under favorable conditions informations may be obtained about present and past changes of climatic and ice flow parameters. However, isotopic homogenization in firn and ice delimits the application of the method, particularly on temperate glaciers and in low accumulation areas. Furthermore, in cases of unstable ice sheets, the climatic component in δ profiles is difficult to separate from the effect of surface altitude changes. The Mid Greenland ice sheet seems to render the most favorable conditions for stable isotope studies on deep ice cores.

The Greenland Ice Sheet Program, initiated in 1971, is an international joint effort to extract paleoclimatic and other geophysical informations contained in the ice sheet. Several drillings to 400–500 m depth will be performed in the coming years as precursors of drillings to bedrock in the last half of the 1970'ies.

Acknowledgements

Several people have contributed to the completion of this work. During many years of co-operation with U.S.A. Cold Regions Research and Engineering Laboratory, we have had pleasant and scientifically rewarding relations with B. Lyle Hansen, Chester C. Langway and John Rand. P. Theodorson, (University of Iceland), Claus U. Hammer and Kjeld Rasmussen were of great help in the G.I.S.P. field operations, and so were Bela Papp (Research Institute for Water Resources Development, Budapest) and Niels Reeh in our laboratory. During eight years, Birthe Truelsmark Poulsen gave an invaluable contribution in carrying out thousands of analyses with never resting care and cheerfulness. This work was successfully taken over by Tove Stougaard, Lis Amossen and Ellen Christiansen. H. Boye-Hansen and H. Dreyer Nielsen of the H. C. Ørsted Institute rendered excellent technical contributions.

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1. Introduction

Isotope Glaciology may be defined as the analysis and interpretation of radioactive and stable isotope variations in glaciers. This paper deals mainly with the heavy stable isotopes deuterium and oxygen-18 in the water molecule. The potentialities of stable isotope glaciology was first pointed out by DANSGAARD (1954) and EPSTEIN (1956). This discipline has in the past two decades become one of the most powerful tools in glaciology, particularly because the deep ice core drilling technique was developed in the same period.

The first known core drilling in glaciers was performed in Switzerland in 1842 (MILLER, 1954), but more than a century should pass until scientifically rewarding ice core drillings began with the efforts of Expéditions Polaires Francaises in Greenland in 1950–51 (HEUBERGER, 1954) and, particularly, with those of the Norwegian-British-Sweedish expedition to Queen Maud Land, 1949–52. The result was a, for that time, unique material for studying the firnification process (SCHYTT, 1958).

The modern deep drilling technique (HANSEN & LANGWAY, 1966; UEDA & GARFIELD, 1969) was developed mainly by B. LYLE HANSEN, whom the International Glaciological Society recently awarded the Seligman Crystal prize for his accomplishments. The many problems involved in the drilling technique are now recognized and overcome to the degree that "you just tell me the origin and the dimensions of the ice core you want, and you will get it" (B. LYLE HANSEN). In this paper we shall use the terms "shallow", "intermediate" and "deep" for drillings to depths until 100 m, 500 m and more than 500 m, respectively.

The modern development began when S.I.P.R.E. recovered a 411 m long ice core from Site 2, NW Greenland, in 1956. This was later followed up by deep drillings to bedrock at Camp Century (1966, 1387 m) and at Byrd Station (1968, 2164 m), and several drillings to intermediate depths at various sites in Antarctica, Greenland, Meighen and Devon Islands, and Iceland. Except for the latter, all of these drillings have been performed on cold glaciers.

The scope of ice core studies reaches far beyond glaciology itself, as it appears from the various promising aspects listed below. The use of stable isotopes will be treated in further detail later in this paper.

As to the applications of radioactive isotopes, reference is made to a recent review paper on isotope glaciology (DANSGAARD & OESCHGER, 1973). Non-isotopic studies on ice cores are referred to by exemplified references.

Ice Core studies

A. Glaciology

a. Accumulation rates – stable isotopes (section 4.3.1).

b. Stability of ice sheets – stable isotopes (section 6).

c. Ice flow patterns.

α. dating of deep strata – stable isotopes (section 4.3.2), Si^{32}, C^{14}, Pb^{210}, H^3 and Ar^{39},

β. verification of model calculations – stable and radioactive isotopes,

γ. deformation of bore holes indicates the horizontal velocity profiles and ice flow law parameters.

d. Temperature profiles – direct measurements. Indicate degree of stability (ROBIN, 1968), climatic changes, flow law parameters and dielectric absorption properties (RADOK *et al.*, 1968).

e. Metamorphosis of ice crystals – size and orientation can be studied as function of age, load, stress, temperature (SCHYTT, 1958; GOW, 1963).

B. Climatology

a. Past, present and possibly future temperature and accumulation changes – stable isotopes (sections 5 and 4.3.1)

b. Past storm activity — land and sea salts (MUROZUMI *et al.*, 1969; LANGWAY, 1970).

c. Past turbidity – dust (HAMILTON & O'KELLEY, 1971). Correlation with stable isotopes (HAMILTON & LANGWAY, 1967).

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a. Sequence of glaciations – stable isotopes (section 5.3.3).

b. Sub-bottom sediments and rocks (HANSEN & LANGWAY, 1966).

D. Volcanology

Volcanic activity – fallout of volcanic dust and ash and possible relation to climates (GOW & WILLIAMSON, 1971).

E. Atmospheric chemistry

a. Composition changes – composition of entrapped air bubbles (SCHOLANDER *et al.*, 1961; ALDER *et al.*, 1969).

b. Pollution – fallout of solid elements, such as lead (MUROZUMI *et al.*, 1969) and fission products (CROZAZ *et al.*, 1966). Organic matter from forest fires. Dust.

F. Meteorology

a. Circulation patterns – stable isotopes.
b. Exchange across the tropopause – radioactive isotopes, stratospheric dust.
c. Residense times in atmospheric reservoirs – radioactive isotopes, dust.

G. Cosmic physics

a. Changes in cosmic radiation flux – naturally produced radioactive isotopes, *e.g.* C^{14} and Si^{32}.
b. Cosmic dust – radioactive isotopes, particularly Mn^{53}.

H. Solar physics

Possible relation between stable isotopes in the ice and C^{14} in tree rings (DANSGAARD *et al.*, 1971).

I. C^{14} dating

Correction of the C^{14} scale beyond the range of the tree ring technique by absolute dating of climatic events – stable isotopes.

Although oceanography is not included in the list, it should be born in mind that isotope oceanography and isotope glaciology have one important goal in common, namely the establishment of long climatic records. Some of the difficulties inherent in the two approaches are similar (*e.g.* possible changes in accumulation rate and flow pattern), and so are the techniques: drilling of cores from the sea floor and the ice sheets, and oxygen-18 measurements on foraminifera (EMILIANI, 1966) and ice, respectively. Yet, the two methods are quite independent and, in fact, complementary, in so far as (i) the oceanographic method has its main application in the tropical and temperate oceans, the glaciological method in the polar regions; (ii) the deep sea cores may reveal the major climatic trends during millions of years, while the deep ice cores give a wealth of details spanning a shorter period. This is why the interplay of the two methods constitutes the most promising aspect in paleo-climatology today.

Thus, there are many reasons to recall A. P. CRARY's advise in his presidential address to the ISAGE meeting, Hanover, 1968:

"My suggestion for future glaciological studies is simple: add the thin dimension. Drill, drill and drill some more; know the ice-rock

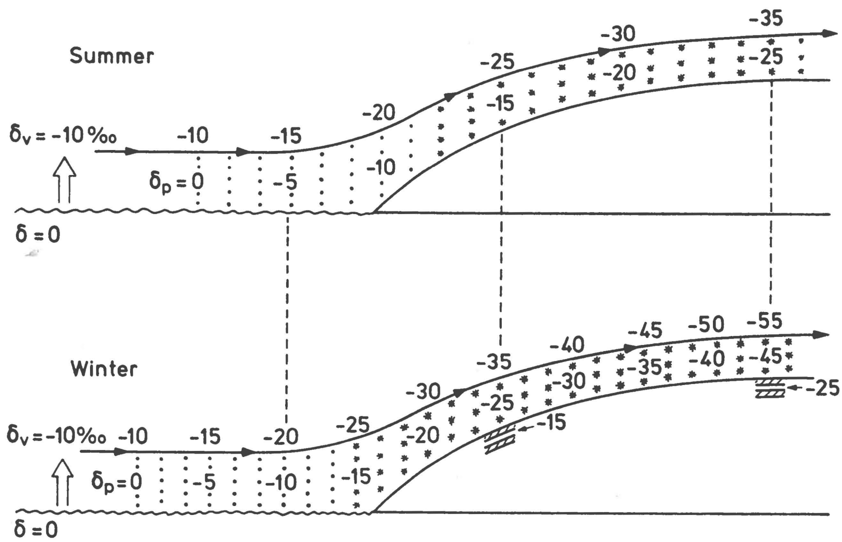


Fig. 1. *Upper part* (summer or warm climatic conditions): Simplified circulation model showing the oxygen isotope fractionation during the evaporation of ocean water (to the left) and the subsequent precipitation, when the air is gradually cooled off by travelling towards higher latitudes or ascending to higher altitudes over an ice sheet (to the right). Although the model is quantitatively unrealistic, it "explains" qualitatively the isotopic latitude and altitude effect at the late stages of the column (lower δ's at higher latitudes and/or altitudes) as being due mainly to preferential fallout of heavy components (Rayleigh condensation from limited amount of vapor).

Lower part (winter or cool climatic conditions): Same as above, except that a cooling, increasing with latitude, has changed the isotopic fall out pattern into lower δ's at any mid and high latitude locality. Snow of $\delta_p = -25\,^0/_{00}$ is assumed to be deposited on top of snow of $\delta_p = -15\,^0/_{00}$, deposited during the preceeding warm period. This explains qualitatively why δ profiles along ice cores reveal (i) seasonal δ-oscillations and (ii) long term climatic records.

interface as well as the surface is presently known. Study the internal ice so that we can learn and understand the history of accumulated snow and other material that is available to us as far back as the cores take us; drill on the continental divides, on the slopes, and on the shelves. And develop the radio echo sounding apparatus to supplement the drilling programmes and to correlate the interior ice horizons and ice-rock interface characteristics over vast areas of the continent".

In the following sections we shall outline the basis for the stable isotope method and discuss its potentialities and limitations within glaciology and other fields.

2. Basic relationships

2.1. The δ scale

In the commonly used δ scale, stable isotope data on natural waters are reported in terms of the ratio R between the concentrations of heavy and light isotopes (O^{18}/O^{16} or D/H). δ of a given sample is the relative difference between R_s in the sample and R_{st} in Standard Mean Ocean Water (SMOW). Unfortunately, SMOW is not a real water body, as the name indicates, but only the zero point of the δ-scale defined on the basis of the real National Bureau of Standards' water standard No. 1 (NBS-1) by

$$R_{SMOW} = 1.008\ R_{NBS-1} \text{ for } O^{18}/O^{16},$$
$$R_{SMOW} = 1.050\ R_{NBS-1} \text{ for } D/H$$

(CRAIG, 1961 a). Thus, by definition, $\delta(O^{18})$ for NBS-1 is

$$\delta = \frac{1 - 1.008}{1.008} \cdot 10^3 = -7.94\,‰.$$

Samples of NBS-1 and a number of secondary standards are available for calibration purposes at Section of Isotope Hydrology, International Atomic Energy Agency, Vienna.

The mass spectrometric technique for δ measurements has previously been outlined (DANSGAARD, 1969; NIEF, 1969). With minor improvements it gives an overall day-to-day reproducibility of $\pm 0.12\,‰$ on δ in routine oxygen isotope analyses, which is satisfactory in stable isotope glaciology.

2.2. Isotopic fractionation in the atmosphere

The main reason for fractionation of the three most important isotopic components of water (H_2O^{16}, H_2O^{18}, HDO) is that the vapor pressure of the heavier components is slightly lower (1 % for H_2O^{18}, 10 % for HDO) than that of the light component. Thus, in case of equilibrium, atmospheric water vapor contains 10 ‰ less O^{18}, and 100 ‰ less deuterium than mean ocean water. We denote this by writing $\delta_v(O^{18}) = -10\,‰$, $\delta_v(D) = -100\,‰$ (CRAIG, 1961 a), *cp.* left part of Fig. 1. If such vapor is separated from the ocean and cooled off, the first small amount of precipitation will get the same composition as the ocean water *i.e.* $\delta_p(O^{18}) = \delta_p(D) = 0\ ^0/_{00}$ (*cp.* Fig. 1), because the heavy components condensate with a 10, respectively 100 $^0/_{00}$, preference to the light component. For the same reason the remaining vapor is left a bit depleted in heavy isotopes, *i.e.* $\delta_v(O^{18}) < -10\ ^0/_{00}$, $\delta_v(D) < -100\ ^0/_{00}$. Further cooling leads to further depletion, both of the vapor and of the condensate given off at later stages of the process. In Fig. 1, δ_p has been put equal to $\delta_v + 10\ ^0/_{00}$ at any stage.

However, evaporation is not an equilibrium process and, furthermore, the atmospheric air is never completely separated from water bodies in nature. Therefore, kinetic effects separating the O^{18} and D components, as well as exchange processes across liquid-vapor boundaries are important for the isotopic fractionation during the evaporation of ocean water and of liquid drops falling from the clouds (DANSGAARD, 1964; CRAIG & GORDON, 1965). Exchange between ocean and atmosphere also influences the fractionation in the first stages of the condensation process. However, in the last stages of the condensation process, which provide the snow that feeds the glaciers, we may neglect both the kinetic effect (because evaporation from solids causes no isotopic fractionation) and the exchange (*e.g.* the ice fields add only small amounts of (isotopically light) vapor to precipitating air masses). The simplified fractionation model in Fig. 1 is therefore realistic enough to qualitatively explain the spacial and temporal isotopic variations observed in the polar regions:

Geographical δ-variations:

Latitude effect (lower δ's at higher latitudes) due to preferential removal of heavy components from precipitating clouds moving toward higher latitudes.

Altitude/inland effect (lower δ's at higher altitudes and/or further inland) for similar reasons.

Temporal δ-variations:

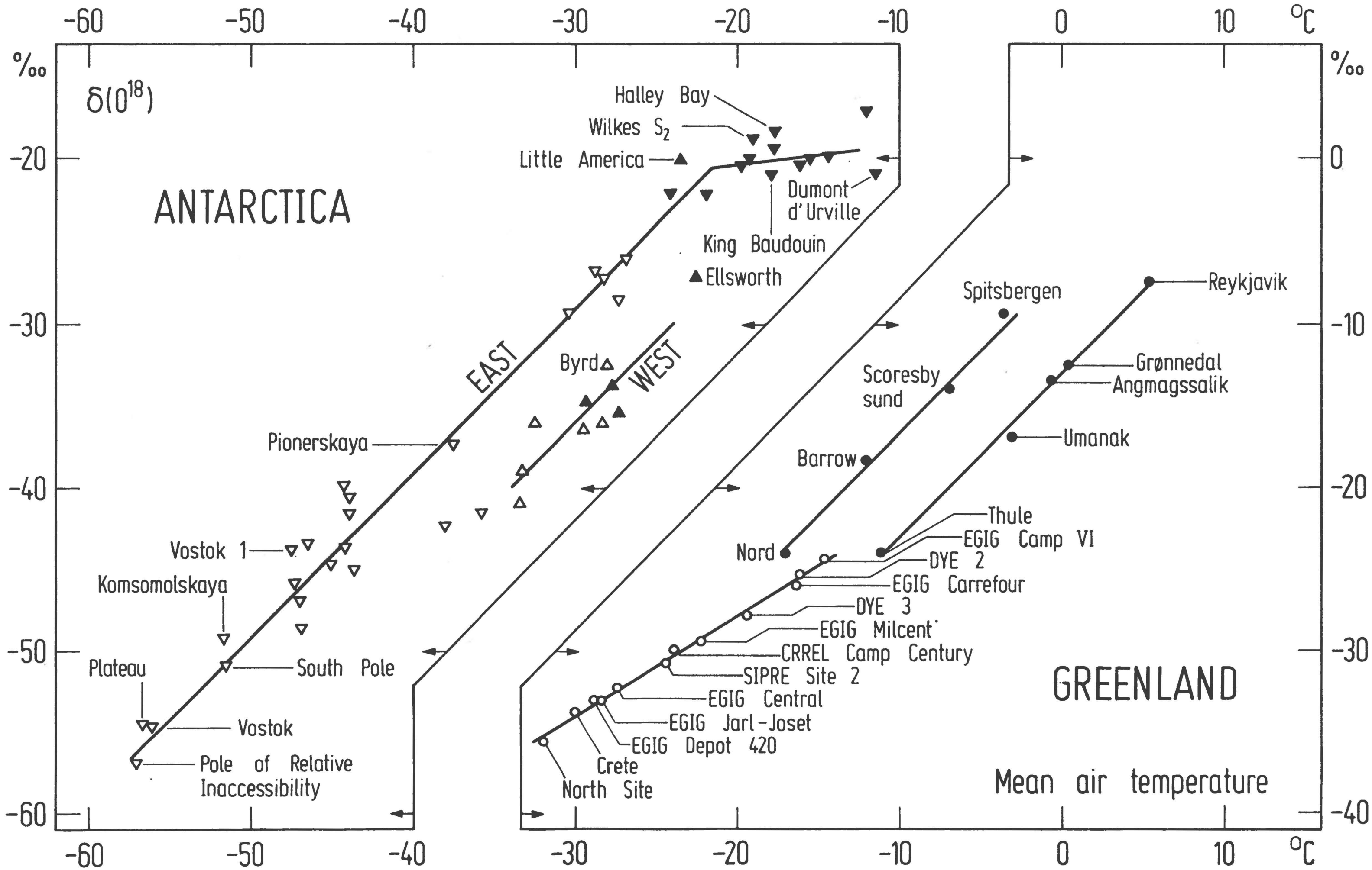
Seasonal effect (lower δ's in winter than in summer) in polar and continental regions, because of deeper winter cooling in such areas than in the low latitude source area of the atmospheric vapor.

Paleo-climatic effect (lower δ's in cold than in warm periods at a given location) for similar reasons.

Due to the lack of kinetic effects in the formation and deposition of snow, $\delta(D)$ and $\delta(O^{18})$ are linearly related in snow and ice at high latitudes and/or altitudes. At present,

$$\delta D = 8.0\ \delta(O^{18}) + 10\ ^0/_{00}$$

(CRAIG, 1961b; DANSGAARD, 1964). However, one cannot rule out the possibility that the linear relationship has been slightly different in periods of entirely different climatic conditions. If so, the deviations might reflect changes in the atmospheric circulation pattern. Therefore, a combined deuterium $-O^{18}$ study on series of samples from well defined climatic periods might be rewarding, yet difficult because it would deal with a second order effect. However, the first order effects, *i.e.* geographical and temporal δ-variations, can be studied exhaustively using one of the heavy isotopes. In the rest of this paper, δ means $\delta(O^{18})$, but all data can be read in terms of $\delta(D)$ using the linear relationship given above.



‰
δ(O¹⁸)
ANTARCTICA
GREENLAND
EAST
WEST
Halley Bay
Wilkes S₂
Little America
Dumont d'Urville
King Baudouin
Ellsworth
Byrd
Pionerskaya
Vostok 1
Komsomolskaya
Plateau
South Pole
Vostok
Pole of Relative Inaccessibility
Reykjavik
Spitsbergen
Grønnedal
Angmagssalik
Scoresby sund
Umanak
Barrow
Nord
Thule
EGIG Camp VI
DYE 2
EGIG Carrefour
DYE 3
EGIG Milcent
CRREL Camp Century
SIPRE Site 2
EGIG Central
EGIG Jarl-Joset
EGIG Depot 420
Crete
North Site
Mean air temperature
°C
-60 -50 -40 -30 -20 -10 0 10
0 -10 -20 -30 -40

Fig. 2.

3. Geographical δ distribution

The geographical variance of the mean δ of annual precipitation is influenced by many parameters (DANSGAARD, 1964). In cold regions, the dominating parameter is $T_c - T_c^\circ$, *i.e.* the difference between the condensation temperature, T_c, at a given location and T_c° at the first stage of the precipitation process. None of these temperatures are generally known, and therefore one has to tentatively replace $T_c - T_c^\circ$ by $T_m - T_m^\circ$, *i.e.* the mean air temperature difference at ground level, or even by T_m itself, assuming the temperature T_m° at the early stage of the condensation to be essentially the same for all stations.

Fig. 2 shows the mean annual δ in precipitation plotted against the mean annual air temperature T_m at ground level for high latitude stations, including several high altitude stations (> 1000 m, open symbols) on the Greenland (Fig. 3) and the Antarctic ice sheets (Fig. 4). In the right part of Fig. 2 each of the steep lines represents the fractionation during the isobaric cooling of precipitating air masses traveling over the oceans from the source area of the vapor. The position of the steep lines in the diagram depends on the temperature of condensation, T_c°, at the beginning of the process. The curves suggest higher values at T_c° for air masses in South and West Greenland than for those in the far North and in Northeast Greenland. This suggests that the latter air masses take up at least part

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Fig. 2. Mean δ of precipitation plotted against mean air temperature at ground level. Notice different δ scales for the right and left part of the figure.

Circles (triangles): Northern (respectively southern) hemisphere stations.

Filled (open) signatures: Low (respectively higher than 1000 m) altitude stations.

The origins of the δ-data are given in parantheses below: Halley Bay and most of the low altitude stations in the Northern Hemisphere (I.A.E.A., 1969); most of the high altitude Greenland stations (DANSGAARD *et al.*, 1969a); King Baudouin (GONFIANTINI, 1965); Ellsworth and Wilkes (EPSTEIN *et al.*, 1963); Byrd (JOHNSEN *et al.*, 1972); Little America (DANSGAARD *et al.*, 1973); South Pole (EPSTEIN *et al.*, 1965); Pole of Relative Inaccessibility (PICCIOTTO *et al.*, 1968); Plateau, Vostok, Komsomolskaya and Pionerskaya (LORIUS *et al.*, 1968a); most of the filled triangles (LORIUS *et al.*, 1969). The equation on p. 10 has been used to transfer LORIUS' deuterium data into $\delta(O^{18})$ values.

of their vapor from the Arctic oceans. The slope of the steep lines, 1.0 ‰ per °C, is in essential agreement with thermodynamic calculations on the later stages of the process (DANSGAARD, 1964).

The isotopic latitude effect appears to be approximately −1 ‰ per degree latitude along the coasts of North Greenland, less in South Greenland. Of course, the latitude effect highly depends on the patterns of winds and sea currents in the area. For example, δ for Isfjord, Spitsbergen, at latitude 78° N is higher than any of the δ's for the Greenland stations, of which only Nord is located at higher latitude (81° N).

When an air mass reaches the coasts of Greenland, further cooling takes place under moist-adiabatic conditions. Due to the expansion of the air, the decrease of the mixing ratio (gram vapor per kg of air) per degree of cooling is now less than under isobaric conditions, and so is the isotopic fractionation per degree centigrade. The slope of the high altitude line in the right part of Fig. 2 (0.62 ‰ per °C) corresponds to the fractionation during moist-adiabatic cooling of an air mass (DANSGAARD, 1964). This agreement would seem to be accidental, if high altitude snow is formed mainly from high strata of the air mass, as suggested by the high tritium concentrations found in high altitude snow in Greenland (AEGERTER *et al.*, 1969; VERGNAUD *et al.*, 1973).

The data from Antarctica, mainly measured by LORIUS *et al.* (1968a, 1969) and shown to the left in Fig. 2, are more scattered. Some of the scatter may be due to improper sample collection technique, several of the samples being collected by unexperienced people. Thus, recent data (MERLIVAT & LORIUS, 1973) on carefully collected samples from a confined area in E Antarctica (Terre Adélie) suggest a closer δ to temperature relationship (slope 0.76 ‰ per °C) than the older data from the same area and other parts of the continent. Other reasons for high scattering could be low accumulation rates and high storminess, which make wind erosion, snow drift etc. more effective in disturbing the isotopic distribution pattern. For example, in the coastal area of Terre Adélie, where the accumulation varies unsystematically inland between 2 and 60 g $\cdot$cm$^{-2}\cdot$yr^{-1}, LORIUS (1963, Fig. 44) found an anticorrelation between δ and accumulation rate, which might be explained by local redistribution of the light winter snow by drift. No such anticorrelation exists in Greenland to judge from data at stations with accumulation rates ranging from 16 to 86 g$\cdot$cm$^{-2}\cdot$yr^{-1}. The importance of drifting snow in Antarctic low accumulation areas appears from LOEWE's (1954) estimate of some $25\cdot10^6$ tons of snow per year and per km of the coast line drifting out from the East Antarctic continent at Port Martin. If representative, this figure suggests that 50 % of the precipitation in Terre Adélie is removed by drift (ASTAPENKO, 1964). Since the drift is most effective

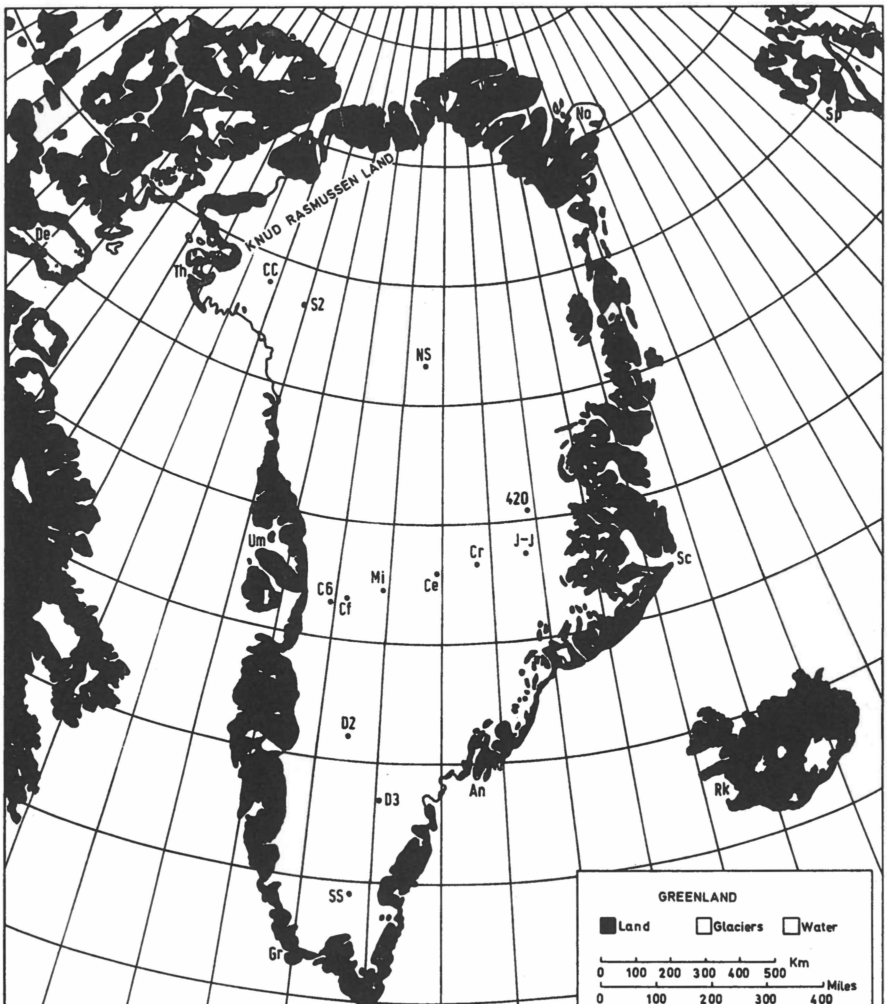

Fig. 3. Map of Greenland with most of the Northern Hemisphere stations mentioned in the text or in Fig. 2. Alphabetically, they are: An = Angmagssalik, CC = Camp Century, Ce = Central, Cf = Carrefour, Cr = Crete, C6 = Camp VI, De = Devon Island, D2 = Dye 2, D3 = Dye 3, Gr = Grønnedal, J-J = Jarl-Joset, Mi = Milcent, No = Nord, NS = North Site, Rk = Reykjavik, Sc = Scoresbysund, Sp = Spitsbergen, SS = South Site, S2 = Site 2, Th = Thule, Um = Umanak, 420 = Depot.

in removing the light winter snow, it must affect the isotopic distribution over the continent, *cp.* Epstein *et al.*, (1963), who found considerably lower δ's in blowing snow at Little America than in any layer in a pit.

Additional evidence for irregular δ distribution inland from the Antarctic coasts appears from the δ profile along the ice core through the Ross Ice Shelf at Little America (DANSGAARD *et al.*, 1973). The δ's vary from -20 ‰ near the surface to -35 ‰ close to the ice/sea water interface. Unless the grounding line of the Ross Ice Shelf is retreating with a dramatic velocity, the δ's decreasing downward in the core reflect a substantial isotopic inland effect of no less than -4 ‰ per 100 km, in case the ice shelf is in essential mass balance as suggested by the results by CRARY *et al.* (1962). -35 ‰ in the ice originating near the present grounding line is lower than the δ value at Byrd Station 600 km further inland and at 1500 m higher altitude. A possible explanation is that warm and dry foehn and catabatic winds cause evaporation of part of the deposits in the grounding line area. This process would be most effective in the warm season, and would leave mainly the isotopically light winter deposits in this area. Support for this idea is to be found near the grounding line of the Amery Ice Shelf, where even areas of ablation have been noticed (BUDD, 1966, Fig. 7).

Hence, there are many reasons why the δ's and the mean surface temperatures are poorly correlated in Antarctica. Nevertheless, the high altitude stations in the left part of Fig. 2 show some degree of systematism. The West Antarctic stations (triangles with peaks upwards) are significantly lower than the East Antarctic stations. The two lines drawn to give the best possible fit to the high altitude stations have the slope 1.0 ‰ per °C, which corresponds fairly well to the relationship between δ of falling snow and its temperature of formation in Antarctic clouds at Station King Baudouin (PICCIOTTO *et al.*, 1960). Furthermore, the slope equals that of the two steep lines in the Greenland section of Fig. 2, which should be expected for the generally flat interior of Antarctica, if most of the precipitation is assumed to be formed by isobaric cooling of air masses.

The isotopic altitude effect on precipitation, defined as the change of δ per 100 m increasing altitude, often includes both a latitude effect and an inland effect (decreasing δ's inland from the coast at unchanged altitude).

In temperate regions the altitude effect is of the order of -0.2 ‰ per 100 m (DANSGAARD, 1961). However, details of a rough, mountainous topography cannot be expected to be reflected in δ-variations in precipitation, because the snow falling in valleys and on intermediate mountain peaks originates from essentially the same stage of the condensation column (AMBACH *et al.*, 1968). For the same reason, recent snow in the accumulation zone of local glaciers does not exhibit any simpel or pronounced relationship between altitude and δ. Nevertheless, if the δ of the precipitation varies appreciably with the season (*cf.* section 4), the summer melting (and run off), increasing downwards from the summit

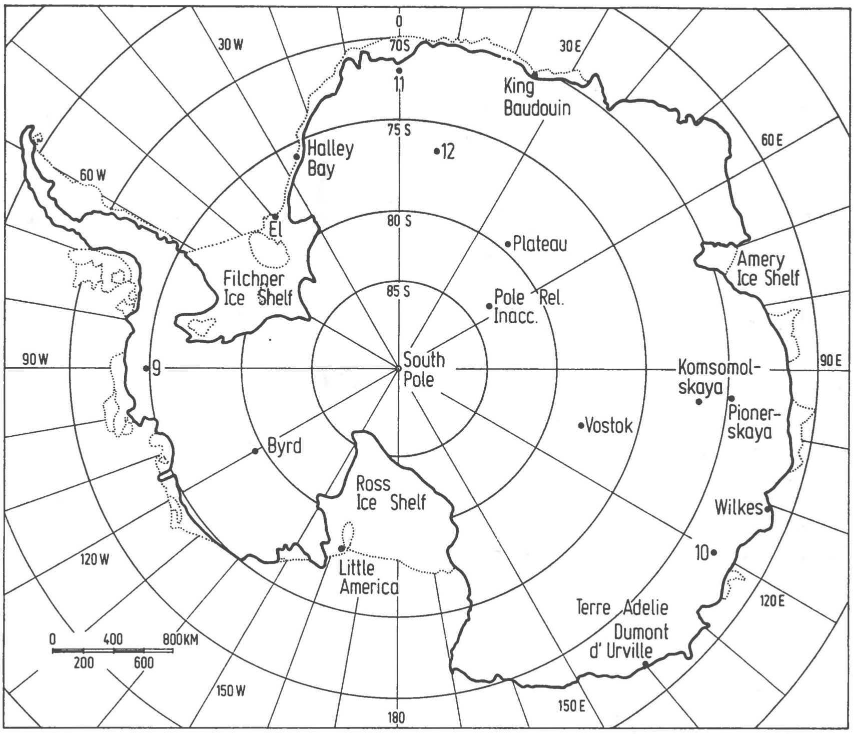

Fig. 4. Map of Antarctica with most of the Southern Hemisphere stations mentioned in the text or in Fig. 2. Stations 9 through 11 are those listed in Tables 2 and 3. El = Ellsworth.

of the glacier, may cause an isotopic altitude effect in the net-accumulated material (AMBACH *et al.*, 1968). In turn, this will show up as generally decreasing δ's of the surface ice from the firn limit to the terminus of the glacier, revealing the inner flow pattern of the ice, *e.g.* in Saskatchewan Glacier (EPSTEIN & SHARP, 1959a), Blue Glacier (SHARP *et al.*, 1960) and Kesselwandferner (AMBACH *et al.*, 1968). However, the picture is often blurred by snow drift, and in regions with no or little seasonal variations in δ, *e.g.* Iceland and Western Norway, the effect is hardly pronounced enough to reveal the flow pattern.

In Greenland, the combined altitude/inland effect appears by multiplying the slope of the high altitude Greenland line in Fig. 2 (0.62 $^0/_{00}$/°C) by the lapse rate (−1°C/100 m), *i.e.* −0.62 $^0/_{00}$/100 m. This has been used as a tool for determining the altitude of deposition of icebergs and marginal glacier ice (DANSGAARD, 1961), assuming essentially unchanged temperature pattern since the time of deposition.

The East Antarctic stations at altitudes below 1000 m (filled triangles

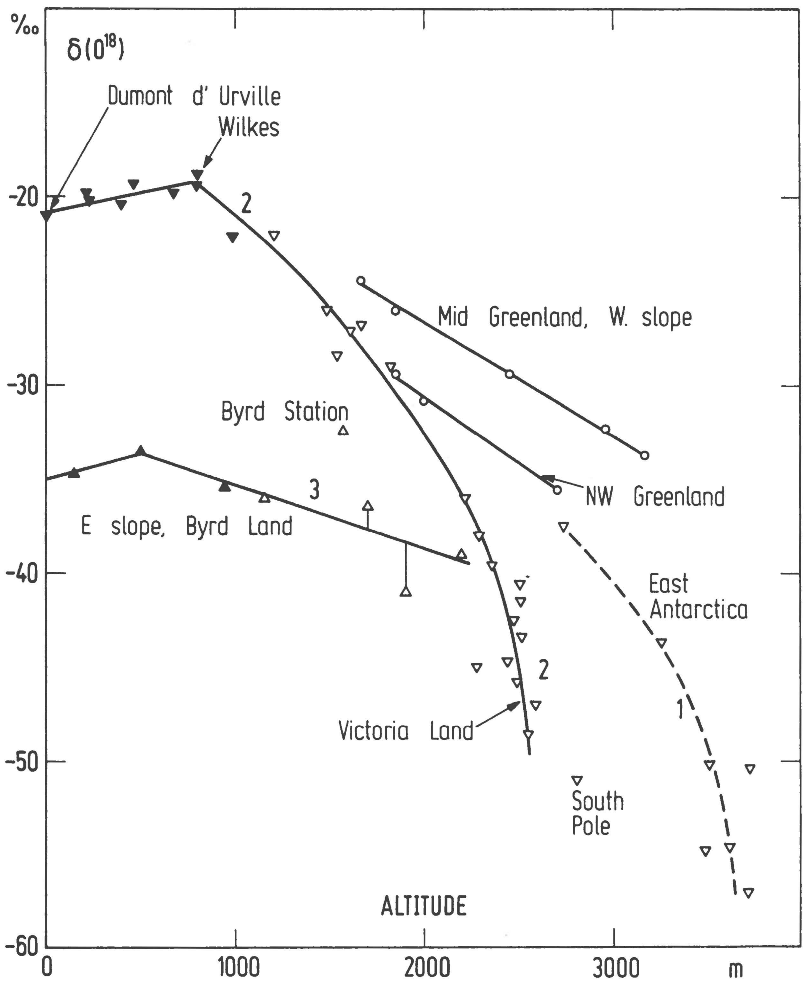

Fig. 5. Mean δ of firn plotted against altitude of deposition. At altitudes lower than 1000 m, the isotopic altitude effect seems to be positive in Antarctica.

pointing downward in Fig. 2) have δ's that are almost independent of temperature, showing little or no altitude effect as first pointed out by LORIUS *et al.*, 1969). Frequent re-distribution of the deposits by drift in the coastal regions may also be responsible for this. In fact, the data from the Terre Adélie coast (LORIUS *et al.*, 1969) rather suggest an inverse isotopic altitude effect with slightly higher δ's at higher altitudes (*cp.* Fig. 5). Their data from the Eastern slope of Marie Byrd Land are less conclusive, but one cannot rule out a similar situation there. Finally, since the δ value at Byrd Station (-32 ‰) is higher than that of the low strata in the Little America ice core ($-35\,^0/_{00}$), which cannot have been deposited as far as Byrd Station, an inverse altitude effect must also exist on at least part of the Western slope of Marie Byrd Land. Only a systematic survey can clarify this complex situation and show, if the non-thermodynamic processes mentioned above can explain the phenomenon.

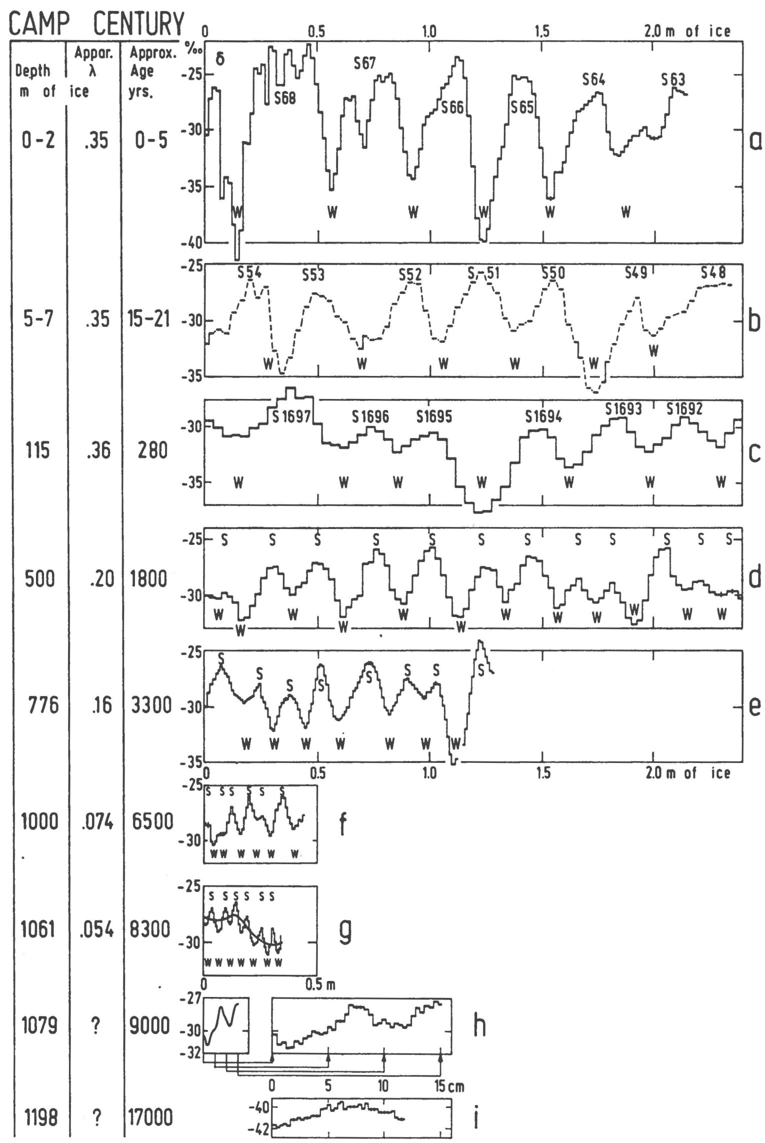

Fig. 6. (From JOHNSEN *et al.*, 1972). The δ oscillations in firn and in ice cores from the depths below surface indicated to the left of the figure at Camp Century. S and W indicate interpretations of summer and winter layers, respectively. As the ice sinks towards the bottom, thickness, λ, of the annual layers is reduced due to plastic deformation. Within a few years (a and b) short-term δ variations are obliterated by mass exchange in the porous firn. After some decades (b and c) the seasonal δ amplitude is reduced to about 2 ‰. Further reduction takes place only by molecular diffusion in the solid ice and becomes effective only when, after thousands of years, the thinning of the layers has increased the δ gradients considerably (d to i).

4. Seasonal effect

The δ of precipitation varies seasonally in the temperate and polar zones. The summer to winter differences range from very little at temperate ocean islands to 10 ‰ at high polar stations (DANSGAARD, 1964, p. 466; I. A. E. A., 1969, 1970), and even more on high polar glaciers, *cp.* Fig. 6a and top of Fig. 7. Under favorable conditions the seasonal δ variations are preserved in glacier ice (first shown by EPSTEIN & SHARP, 1959b), which leads to some important applications described in section 4.3. First, however, we shall discuss some processes, that tend to obliterate the isotopic stratification of snow and ice after deposition.

4.1. Smoothing of δ variations during firnification

On temperate glaciers, the seasonal oscillations in the snow pack are rapidly reduced during recrystallization in presence of percolating melt water. Simultaneously, isotopic fractionation occurs leaving the solid phase enriched relative to the liquid phase (ARNASON, 1969; BUASON, 1972). δ profiles from Icelandic, Norwegian and Austrian glaciers contain only slight or no seasonal variations (ARNASON, 1969; DEUTSCH *et al.*, 1966).

On polar glaciers, the disturbances due to percolation are small or negligible. However, other smoothing effects must be quite active in the porous firn, since δ variations in the precipitation are always smoothed considerably, and sometimes obliterated shortly after deposition. Thus, Fig. 6 (from JOHNSEN *et al.*, 1972), shows a reduction of the seasonal amplitude at Camp Century by a factor of 4, at the same time as δ oscillations with frequencies higher than 1 yr^{-1} are completely obliterated. Fig. 7 shows various degrees of smoothing at 3 other Greenland stations, all close to the ice divide in North, Mid and South Greenland, respectively. It appears from Fig. 7a that the existence of pronounced seasonal δ variations in the upper firn is no guarantee for preservation during firnification. In low accumulation areas, such as most of Antarctica, many seasonal δ oscillations are simply missing due to drift or lack of winter (summer) snow, or they are completely smoothed within a few years, *cp.* EPSTEIN *et al.*, (1963); PICCIOTTO *et al.*, (1968); LORIUS *et al.*, (1968a). Some of our own results are compiled in Fig. 8, showing the gradual reduction of the summer to winter δ difference during the first part of the firnification process.

According to JOHNSEN (1973) there are two important reasons for the isotopic homogenization in polar firn, both connected to recrystallization of the grains via the vapor phase. Firstly, storms and barometric pressure changes cause vertical air movements, particularly in the upper firn, where mass exchange between the strata is further accentuated by high temperature gradients. This might account for the fast smoothing at Byrd and Little America V (Fig. 8) that have much higher storminess than the Greenland stations. Secondly, diffusion in the vapor phase causes considerable interstratificial mass exchange down to "the critical depth", d_c, where the density has reached the "critical value", ϱ_c, approximately 0.55 gr/cm³ (BENSON, 1962; ANDERSON & BENSON, 1962). The total vertical diffusion length at this level seems to be only slightly dependent on the temperature, which may be explained by the temperature influencing d_c and the rate of diffusion in such a way that the effects essentially cancel out (JOHNSEN, 1973). The combined mass exchange between firn layers down to d_c may be expressed by the mean vertical displacement, L_c, of the material (relative to its original layer) caused by convective air movements and by diffusion. A given harmonic δ-oscillation of "wave length" λ_0 will be essentially obliterated, if L_c exceeds $\frac{1}{3}\lambda_0$. Hence, assuming the seasonal δ-variations to be harmonic, which implies the precipitation be uniformly distributed throughout the year, the smoothing effect down to the critical depth depends mainly on the storminess and on the rate of accumulation. If the winter precipitation

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Fig. 7. Seasonal $\delta(O^{18})$ oscillations in the upper firn at 3 stations close to the ice divide in North, Mid and South Greenland. The sampling frequency was approximately 8 samples per annual layer (16 in the top 2 meters of ice equivalent). δ-values higher than average are set off in black. Tentative interpretations are given to the left of the curves. At Dye 3, the dating was supported by a β-activity profile (C. HAMMER, to be published). High frequency δ oscillations ($> 1\ yr^{-1}$) disappear within a few months at North Site, a few years at Crete, and essentially within a few decades at Dye 3, where impermeable layers of refrozen melt water delay smoothing by vapor diffusion. In the first few years after deposition, the amplitudes of the seasonal δ-oscillations are highest at North Site. However, due to the low accumulation rate, they are also most rapidly reduced at this station. Complete obliteration begins at a depth corresponding to 6.5 m of ice equivalent. This is not the case at Crete, much less at Dye 3, where the seasonal δ-variations are known to survive the entire firnification process (*cp.* Fig. 8). At the "critical depth", corresponding to density 0.55 gr/cm³, or some 3 m of ice equivalent, the isotopic homogenization is essentially suspended for a long period of time (*cp.* Fig. 6). The mean accumulation rates appear to be 16.5, 27 and 53 cm of ice per year, respectively.

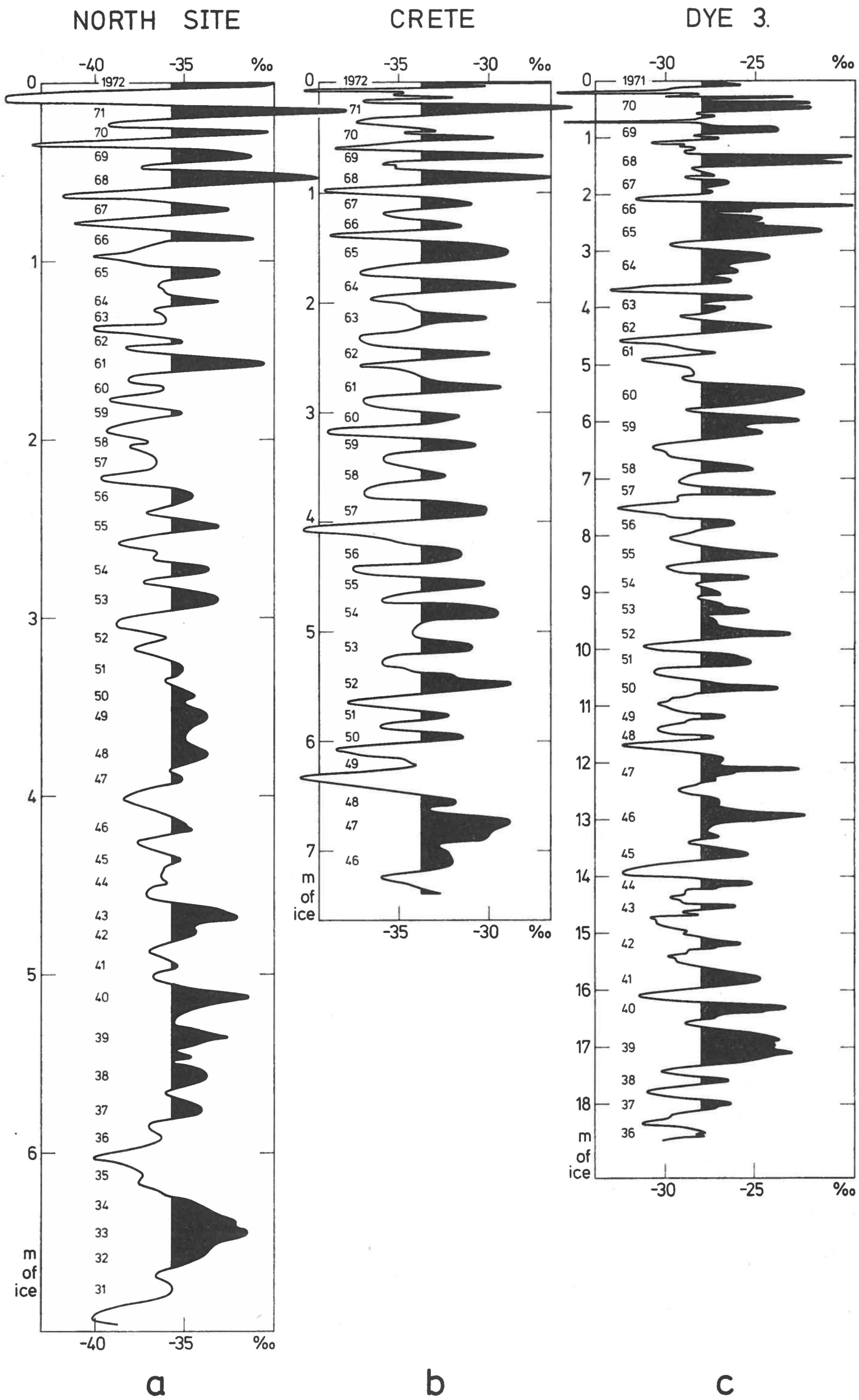


Fig. 7.

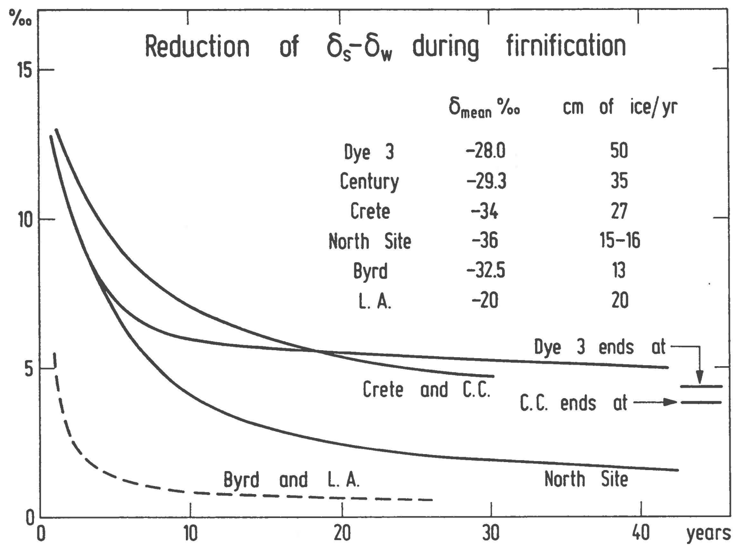


Fig. 8. Measured reduction of summer to winter difference, $\delta_S - \delta_W$, in accumulated snow as a function of time since deposition. Seasonal δ variations, and thereby age and accumulation rates, can be determined as long as $\delta_S - \delta_W$ remains higher than 2 ‰. If this is the case, when the firnification is completed, the stable isotopes can be used for age and accumulation measurements thousands of years backwards in time.

is considerably less than the summer precipitation, the smoothing is much more effective or, in other words, obliteration occurs for values of L_c considerably less than $\frac{1}{3}\lambda_0$. Similarly, individual years with unusually low accumulation will "disappear" in the δ profile much faster than normal years.

Unfortunately, seasonal δ variations have been measured down to, or below, the critical depth only at a few stations. These are listed in Table 1 together with other stations (indicated by stars), where δ oscillations have been measured to depths less than d_c. In column 4 and 5, a plus sign means that seasonal δ cycles are detectable below the critical depth, respectively only a few years. At the station 73°25′ S, 14°10′ W (samples collected by OLAV LIESTØLL, Norsk Polarinstitutt), a δ amplitude of approx. 2 ‰ remained at 9 m depth, and the oscillations are therefore estimated to be preserved. On the other hand, only a 1 ‰ δ amplitude remained at 0.5 m depth on the South Pole (EPSTEIN *et al.*, 1965, Fig. 2), and it is therefore assumed to disappear before the critical density is reached.

Columns 2 and 4 in Table 1 suggest that an accumulation rate of approx. 26 cm of ice equivalent per year, corresponding to 24 $gr \cdot cm^{-2} \cdot yr^{-1}$,

Table 1

1	2	3	4	5	6
Station	Accumulation rate cm ice·yr^{-1}	Mean temp. °C	Seasonal δ variations detectable: below d_c	a few years	Reference
Carrefour	62	− 18	+	+	AMBACH & DANSGAARD (1968)
Dye 3	53	− 20	+	+	Fig. 7c.
King Baudouin	42	− 15	+	+	GONFIANTINI (1965)
Site 2	40	− 24	+	+	EPSTEIN & SHARP (1959 b)
73°25′ S, 14°10′W*	37	− 25	(+)	+	LIESTØL (1970)
Camp Century	35	− 24	+	+	Fig. 6
Crete	27	− 30	+	+	Fig. 7 b
Pionerskaya*	25	− 37	−	−	LORIUS *et al.* (1968 a)
Little America V	24	− 23	−	−	EPSTEIN *et al.* (1963)
NW Devon Island	22	− 23	−	+	PATERSON *et al.* (1973)
North Site	16	− 31	−	+	Fig. 7 a
Wilkes S 2*	14	− 19	−	+	EPSTEIN *et al.* (1963)
Byrd Station	13	− 29	−	−	JOHNSEN *et al.* (1972)
South Pole*	9	− 51	(−)	+	EPSTEIN *et al.* (1965)
Komsomolskaya	5	− 51	−	−	LORIUS *et al.* (1968a)
Pole of Relative Inaccessibility	3	− 57	−	−	PICCIOTTO *et al.* (1968)
Vostok*	2 − 4	− 57	−	−	LORIUS *et al.* (1968a)

is critical for the survival of seasonal δ variations throughout the first stage of the firnification process. Once the critical density ϱ_c is reached, further smoothing takes place mainly by self-diffusion in solid ice, which is a very slow process. The δ oscillations found at the density level ϱ_c are therefore only exposed to little further smoothing during the rest of the firnification process and, in fact, during several thousands of years till the annual layers have reached a thickness of a few cm, *cp.* section 4.2 and Fig. 6.

Column 5, Table 1, shows on the other hand that the rate of smoothing in the top layers of the firn does not depend exclusively on the accumulation rate. The existence of unusually regular δ oscillations at North Site, Greenland, (Fig. 7 a) in spite of only 16 cm of accumulation points

Table 2. *Distance y_o above bottom, to which harmonic δ oscillations of period n years can be detected.*
Layer thickness $n\lambda$ and its age t_m at y_o. Age of ice at y = 10 m. y for 75,000 yrs old layers.

1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
					n = 1 yr			n = 10 yrs			n = 100 yrs			n = 1000 yrs			n = 10^4 yrs				
St. No.	Site of Drilling	H m of ice	T_o °C	λ_o cm of ice	y_o m ice	$n\lambda$ cm ice	t_m yrs $\times 10^3$	y_o m ice	$n\lambda$ cm ice	t_m yrs $\times 10^3$	y_o m ice	$n\lambda$ cm ice	t_m yrs $\times 10^3$	y_o m ice	$n\lambda$ cm ice	t_m yrs $\times 10^3$	y_o m ice	$n\lambda$ cm ice	t_m yrs $\times 10^3$	Age 10 m above bott. yrs $\times 10^3$	y m for age 75.000 yrs
1	Camp Century	1370	– 24	35	270	4	8.3	101	6	19	40	10	43	16	16	100	–	–	240	150	38
2	Jarl-Joset	2600	– 28.5	29.4	550	4	16	220	7	34	81	13	74	33	23	160	13	35	380	400	44
3	Crete	3200	– 30	27	650	4	22	280	9	42	106	14	88	42	25	190	17	42	430	700	118
4	Central	3150	– 27.6	38	560	4	16	210	7	34	81	14	70	33	25	150	13	37	360	460	74
5	Milcent	2350	– 22.3	55.6	570	3	13	220	6	31	87	11	65	34	19	140	–	–	–	400	75
6	Dye 3	1900	– 19	53	325	4	8.6	122	8	19	48	14	42	19	21	95	9	40	200	180	24
7	South Site	1800	– 20	65	245	4	7.8	93	7	17	37	10	37	15	18	87	–	–	180	120	17
8	North Site	2550	– 32	17	–	–	–	258	7	50	95	10	110	37	18	240	16	32	520	800	152
9	75° S, 90° W	1900	– 25	50																	
10	68.4° S, 120° E	2300	– 35	50																	
11	72.4° S, 0° E	1000	– 25	30																	
12	77° S, 10° E	2800	– 50	5	–	–	–	653	8	95	268	19	190	100	28	410	41	56	850	3000	844
13	South Pole	2600	– 50	7	–	–	–	445	7	82	174	14	170	70	24	350	28	40	780	2000	500
14	Pole Rel. Inaccess.	2700	– 57	3	–	–	–	822	7	120	325	16	260	130	25	530	52	56	1100	5000	1233
15	Vostok	3700	– 56	4	–	–	–	955	8	140	400	23	270	160	42	530	63	69	1100	5700	1710
				2	–	–	–	757	3	330	313	7	650								2488

at the storminess as an important parameter, because this location is unusually quiet to judge from the extremely loose consistence of the surface snow (people sank 30 cm into the snow). Very low winter accumulation is undoubtedly a contributary reason for the lack of seasonal δ variations at several Antarctic stations.

4.2. Smoothing of seasonal δ variations in ice

At great depths the remains of the seasonal δ oscillations are gradually obliterated by diffusion in the solid ice (*cp.* Fig. 6). Based on (i) JOHNSEN's equation for the temperature profile (1973), (ii) RAMSEIER's equation for the constant for self diffusion in ice (1967) and (iii) PHILBERT & FEDERER's generalization (1971) of DANSGAARD & JOHNSEN's flow model (1969), a numerical integration can be performed (JOHNSEN, 1973) to give λ and the total diffusion length L at any depth in any ice core, if the surface temperature T_0, the accumulation rate λ_0 and the total thickness H of the ice sheet are known. Applying this method on the Camp Century ice core shows that, after firnification, a further 10 % reduction of the δ amplitude takes 5,000 years, and a 50 % reduction takes 8,000 years. After 10,000 years only 10 % is left. This is in fairly good agreement with the data in Fig. 6.

Table 2, column 8 shows for some Greenland and Antarctic stations calculated approximate "life times" t_m of seasonal δ oscillations, defined as the time needed to reduce the δ amplitude left after firnification by a factor of 4. Column 7 shows the layer thickness by the time, when the oscillations essentially disappear by diffusion.

At Milcent (far from the crest) the decreasing accumulation and temperature, as well as the increasing thickness upstream have been accounted for by a two-dimensional numerical integration. Lack of information prevents this at stations Nos 9, 10 and 11. At Nos 8 and 12 through 15 the seasonal δ oscillations do not survive the firnification process.

4.3. Applications

4.3.1. Accumulation rates. The distance between summer maxima in a detailed δ profile in the upper firn indicates the annual net-accumulation. In high accumulation areas, δ oscillations with frequencies higher than 1 yr^{-1} impede the interpretation of summer maxima (*cp.* Fig. 7c), and cross checks should be looked for by classical stratigraphic methods (density variations, textural observations), or by measuring β-activity profiles for identification of reference horizons (PICCIOTTO & WILGAIN, 1963; CROZAZ *et al.*, 1966; AMBACH *et al.*, 1967; AMBACH &

Dansgaard, 1968). In low accumulation areas in Antarctica the stable isotope method often fails completely, as mentioned on p. 21.

At depths that are not small relative to the thickness of the ice sheet, the measured annual layer thickness must be corrected for vertical strain since the time of deposition. In simple cases, *i.e.* no melting, and uniform vertical strain rate right down to the bottom, the correction factor to be applied down to considerable depths is simply H/y, H being the total thickness of the ice sheet, y the present distance of the layer from the bottom. However, the total strain since deposition also depends upon the entire temperature history of the layer in question. The correction factor to be applied on ice layers deposited during the glaciation or in the early post-glacial period, must therefore account for the influence of the low temperatures in the past.

The data available suggest the conclusion, that recent accumulation rates can be determined by stable isotopes in most of the dry and percolation zones of the Greenland ice sheet, because only a small part in northeast has accumulation rates considerably lower than the 15 $gr \cdot cm^{-2} \cdot yr^{-1}$ at North Site (Mock, 1967). In contrast, twice this accumulation rate seems to be needed in most of Antarctica in order to secure a reliable interpretation of seasonal δ variations even in the uppermost firn. Accumulation rates higher than 30 $gr \cdot cm^{-2} \cdot yr^{-1}$ occur in Antarctica only in relatively small, coastal areas (Bull, 1971). On temperate glaciers the method may be applicable under particularly favorable conditions. However, this possibility remains to be checked.

4.3.2. Dating of ice cores is possible by using (i) the classical stratigraphic methods (back to some 50 years), (ii) the radioactive isotopes H^3 and Pb^{210} (100 years), Si^{32} and Ar^{39} (1000 years), and C^{14} (20,000 years), (iii) flow model calculations (Dansgaard & Johnsen, 1969), and by (iv) counting δ summer maxima, or (v) counting maxima of long-term δ oscillations (Dansgaard *et al.*, 1971).

The accuracy of the methods (ii) and (iii) is seldom better than $\pm$ 10 %. The fourth method renders a better accuracy. It is applicable, if the summer to winter δ difference left after firnification averages more than 2 ‰. The mean thickness λ of annual layers, measured in reasonably spaced increments of the core, gives the relationship $\lambda = f(y)$ between λ and the distance y above the bottom. Age in years (t) can be calculated as

$$t = \tau \int_{H}^{H-y} \frac{dy}{f(y)}$$

τ being 1 year.

Absolute dating implies a detailed continuous δ record downward from surface (8 samples per annual layer). At present, the Dye 3 core is dated this way back to 1230 A.D. with an estimated accuracy of $\pm$ 5 years. The uncertainty is due to occasionally dubious interpretations. Exact dating may be obtained by correlating a dust concentration profile with dust veil indices (LAMB, 1970).

Table 2, column 8, suggests that method (iv) is applicable in extended areas in Greenland several thousand years backward in time; and in Mid Greenland, where the higest thickness of the ice is found, even prior to the termination of the last glaciation. This latter possibility implies however that the accumulation by that time was high enough to secure preservation of seasonal δ oscillations. If so, deep ice cores from Mid Greenland offer unique possibilities, not only for absolute dating of late glacial climatic events, and thereby correction of the C^{14} scale beyond the range of the dendrochronological technique, but also for measurements of accumulation rates in glacial times.

5. Climatic effect

As mentioned in section 3, the δ value of a given snow sample depends on many parameters, of which the most dominating one is the difference between the condensation temperature, T_c, in the precipitating cloud and the temperature, T_c°, at the first stage of the condensation process. Thus, a general 1° C cooling all over the Earth would hardly have any noticeable effect on the δ value of snow at a given locality, in so far as $T_c-T_c^\circ$ would remain unchanged. The reason why, nevertheless, δ is a climatic indicator, is that climatic changes, being most pronounced at high latitudes, do cause a change in $T_c-T_c^\circ$.

In Fig. 9 the two curves to the outer left and right show 10 years running mean air temperatures as observed at sea level in Upernavik (NW Greenland) and Angmagssalik (SE Greenland) since the beginning of this century. Due to the imperfect control of these less accessible stations, particularly before World War II, the temperature records are hardly fully reliable in all detail. Nevertheless it is reasonable to conclude that (i) a considerable warming took place at both stations during the 1920'ies, (ii) the temperature optimum occurred shortly after 1930, and (iii) already 20 years later, the temperatures in Upernavik had dropped to close to the mean value for the whole period, whereas the warm conditions lasted more than a decade longer in Angmagssalik.

In between the two outer curves, Fig. 9 shows six δ-records from stations on the Greenland ice sheet, as indicated in Fig. 3. The δ records from Jarl-Joset and Central were measured on firn cores hand augered by E.G.I.G. (Dansgaard *et al.*, 1969a). Like the observed temperatures, the δ's can of course not be expected to vary completely in parallel from South to North. Nevertheless, the 4 records reaching back to before 1920 show a considerable increase in δ during the 1920'ies (at Jarl-Joset the δ maximum occurred simultaneous with a secondary temperature maximum in Angmagssalik around 1944), followed by a decrease during the recent decades. The lower part of the Camp Century curve in Fig. 9 is not considered completely reliable because the firn samples in question had to be collected (kindly by Dr. S. Leung of C.R.R.E.L.) under non-ideal conditions in the inclined trench at Camp Century.

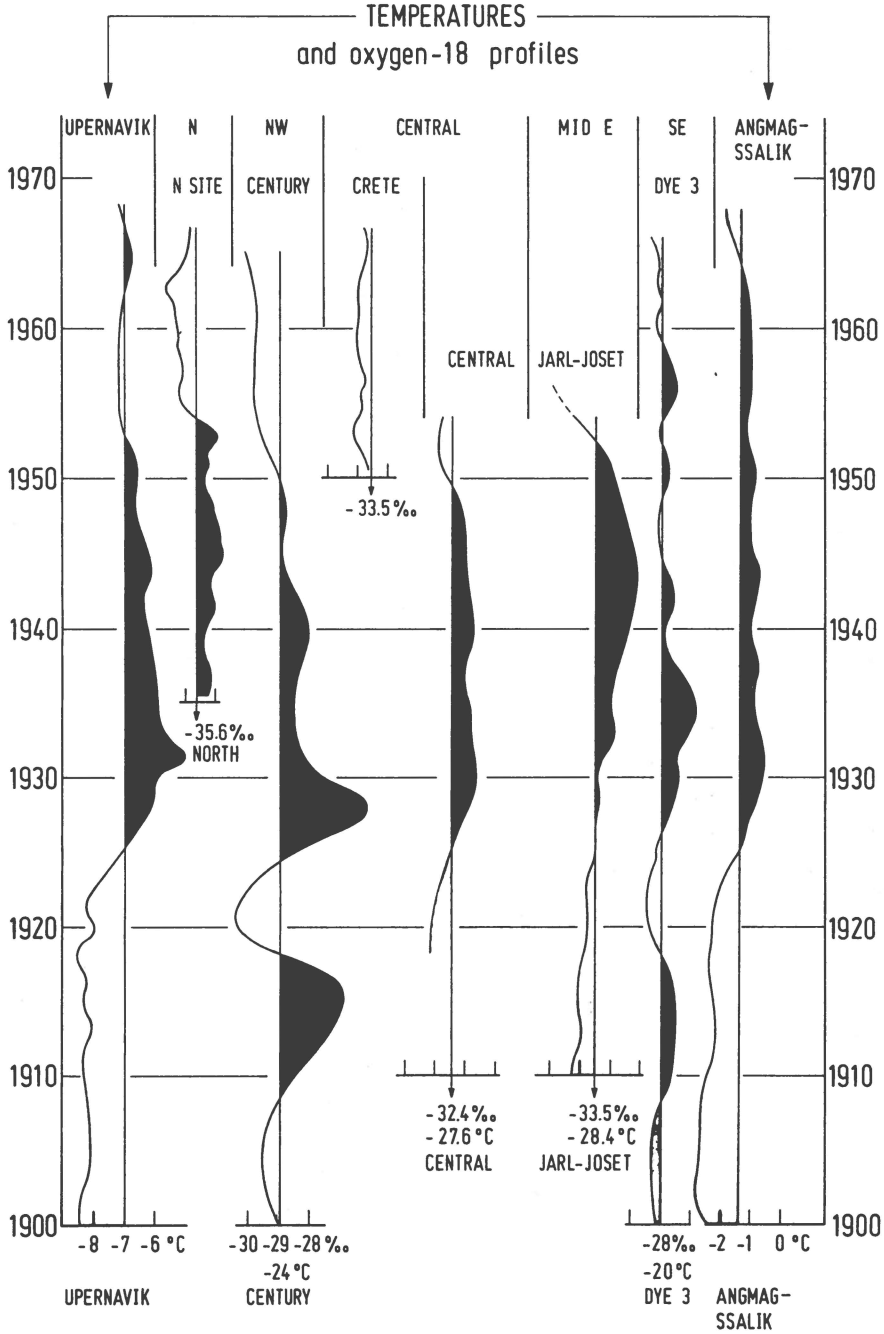
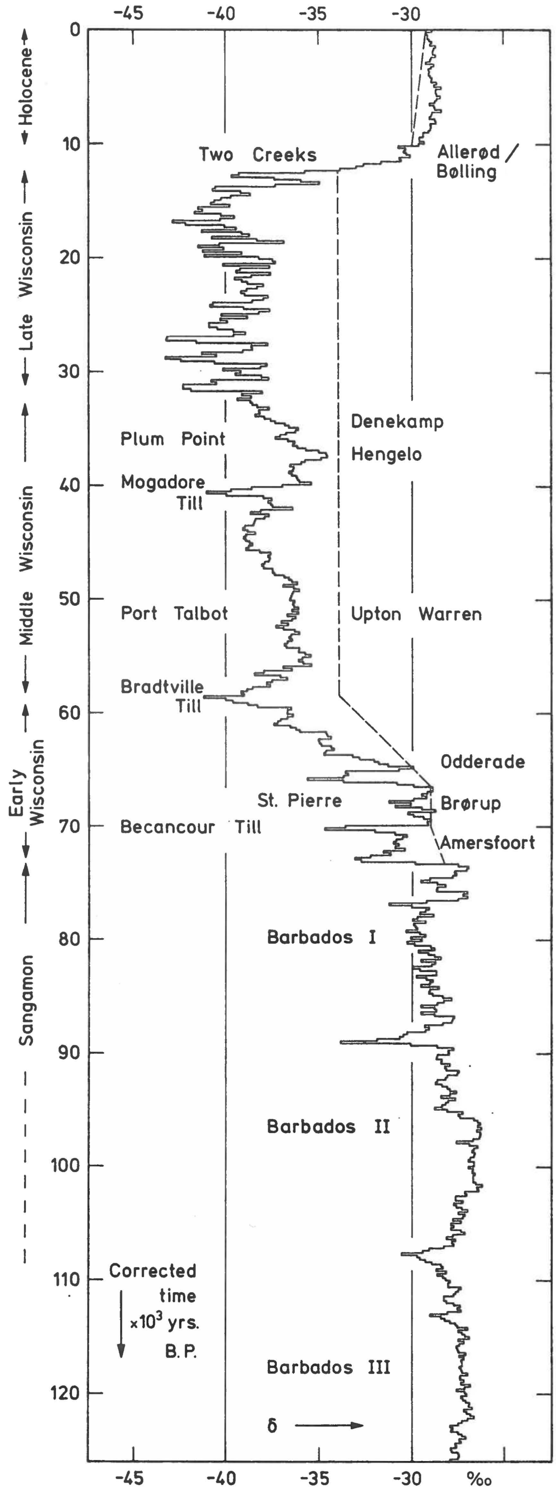


Fig. 9. Observed 10 years running mean temperatures at Upernavik (outer left) and Angmagssalik (outer right) compared with 10 years running mean δ values at 6 Greenland ice sheet stations. The lower part of the Camp Century δ-record may not be fully reliable, *cp.* text on p. 30.



-45
-40
-35
-30
0
10
20
30
40
50
60
70
80
90
100
110
120
Holocene
Late Wisconsin
Middle Wisconsin
Early Wisconsin
Sangamon
Two Creeks
Allerød / Bølling
Plum Point
Denekamp
Hengelo
Mogadore Till
Port Talbot
Upton Warren
Bradtville Till
Odderade
St. Pierre
Brørup
Becancour Till
Amersfoort
Barbados I
Barbados II
Barbados III
Corrected time
$\times 10^3$ yrs. B.P.
δ
‰

Fig. 10

The entire Camp Century δ-record from the 1387 m long ice core is shown in Fig. 10, plotted on a time scale that was originally evaluated by ice flow considerations (Dansgaard & Johnsen, 1969), and later provided with minor corrections that make the time scale independent on ice flow parameters (Dansgaard *et al.*, 1971). The record is undoubtedly continuous and reaches back to long before the onset of the last glaciation. It has been provided with tentative interpretations in European and American terminology, and seems to reveal all of the climatic events within the last 80,000 years known from other independant studies and in hitherto unseen detail. In addition, other known and unknown events prior to the Wisconsin glaciation appear in the δ-record. Thus, at approx. 90,000 yrs B.P. in the adopted time scale, a shift in δ suggests an almost instantaneous climatic cooling to full glacial severity, followed by a climatic recovery within a thousand years (Dansgaard *et al.*, 1972b). However, the significance of such apparently catastrophic events, as well as the time scale applied, and the climatic and glaciological implications of the more general trends of the Camp Century curve all call for further investigations.

As pointed out in the next section, deep ice cores from the Northwestern part of the Greenland ice sheet are not expected to reveal purely paleoclimatic records. The δ values shown in Fig. 10 may need considerable correction for past changes of the ice flow pattern, particularly in the glaciation period. On the other hand, these changes do not remove the climatic influence on the δ curve.

5.1. Corrections to long-term δ profiles

The interpretation of δ profiles in terms of temperature changes is uncertain for several reasons as repeatedly pointed out since 1969. Thus,

> "(i) the deeper strata originated further inland, where perhaps slightly different climatic conditions existed; (ii) the isotopic composition of seawater, which provides the moisture for the precipitation, changed; (iii) the ratio of summer to winter precipitation possibly changed; (iv) the main meteorological wind patterns changed; (v) the flow pattern of the ice in the accumulation area possibly changed; and (vi) the thickness of the ice sheet changed.

Fig. 10. (From Dansgaard *et al.*, 1971) δ-record through the Greenland ice sheet at Camp Century (down to 17 m above the bottom) plotted on a time scale (in units of 10^3 years before present) that is based on the assumption of observed persistant δ oscillations having a constant period of 2400 years. The curve is provided with tentative interpretations in European and American terminology. Correction for the influence of changing surface altitudes is suggested by the dashed curve, which is assumed to roughly reveal the δ trends in case of constant climatic conditions.

The changes in the isotopic composition of sea water are less important in changing ice δ's than changes in the temperature of precipitation formation (EMILIANI, 1966). The extremely low δ's shown during the Wisconsin/Würm do not necessarily correspond to temperatures that were many degrees centigrade lower than the present—if the thickness of the ice sheet were considerably greater than it is today, this would, in itself, cause lower surface temperatures and lower δ's of precipitation at a given geographical location.

For the same reason, the δ's during the climatic optimum between 4500 and 7000 years B.P. would probably have been even higher had the ice sheet not still been thicker that at present. However, a major increase (or decrease) of the surface altitude could hardly occur without a general cooling (or warming) of the climate. Both of these effects influence the δ's in the same direction, but it is not yet possible to distinguish between the individual contributions to an observed change in δ's" (DANSGAARD *et al.*, 1969b).

As to the latter point, LORIUS *et al.*, (1968b) have suggested that the total gas content in the ice depends on the altitude of deposition. An altitude effect on the gas content, of the order of 1 % change or -0.08 volume % per 100 m may be expected, assuming the gas content to be proportional to the mean atmospheric pressure at the site of formation. This is not much compared to the scatter of available data on individual blocks of ice obtained by other authors (SCHOLANDER *et al.*, 1961; MATSUO & MIYAKE, 1966). Furthermore, one has to assume that the pores in the firn are closed off into separate air bubbles at the same stage of the firnification (density approximately 0.82 $gr \cdot cm^{-3}$), independent of all firnification parameters. The gas content in ice is rather sensitive to changes of the close-off density ϱ', in fact proportional to $\frac{1}{\varrho'} - \frac{1}{0.92}$. Thus a 2 % decrease of ϱ' from glacial to post-glacial conditions is more than enough to explain the increase in gas content during the climatic transition indicated by recent analyses of the Camp Century core, *cp.* RAYNAUD & LORIUS (1973). Unfortunately, no data were given on the number of bubbles per cm^3 of ice, which should remain constant during the shift in case of unchanged firnification. However, a comprehensive study should be performed to clarify the potentialities of the gas content method. At present, one can only estimate a correction to the δ of a given ice layer to account for the difference between its altitude of deposition and that of the surface at the drilling site.

In the first place, one may assume that the present ice flow and meteorological condition have been valid for the entire period considered. Even with this assumption, correction requires detailed knowledge about the accumulation and ice flow patterns in the drainage area upstream from the drilling site. In the Camp Century case, the present accumulation pattern and the surface and bottom topographies are sufficiently well known (MOCK, 1968; GUDMANDSEN, 1970), but the divergence, or the strains, of the ice flow in the drainage area need to be further investigated, before the δ curve can be corrected with reasonable accuracy for lower temperatures of deposition for the lower strata than for the upper strata. As to deep strata, the correction is probably much less than the 6 ‰ δ-difference between Camp Century and North Site (*cp.* Figs 3 and 9), because these strata have, for tens of thousands of years, moved very close to the bottom, where the velocities are much lower than at surface. Furthermore, most of the ice flowing from far inland toward the Thule peninsula is drained off into the Melville Bugt and the Humbolt Gletscher. Consequently, in case of steady state even the deep strata in the Camp Century area, is most likely of fairly local origin. And anyway, the correction would be continuously increasing with the age of the strata, which means that the interpretation of relatively short-term δ variations as being due to climatic changes would not be invalidated. The situation is still more complicated at Byrd Station, where not even the accumulation and topography in the drainage area are sufficiently well known.

Going a step further, we now cancel the reservation of steady state and include possible changes of the shape of the ice sheet that have taken place since the time of deposition of the strata. Of course, this must be somewhat speculative, because one has to envisage flow patterns that do not exist any more. If we begin with the deep layers in the Camp Century ice core originating from the Sangamon/Eem interglacial, their very existence, as well as the lack of evidence of melting at the bottom, indicate that the bottom temperature was below the pressure melting point. Therefore, the considerations presented in the preceeding paragraph still apply for the lowest part of the ice core. In other words, we see no reason for significant correction of the pre-glacial δ values.

Before we proceed to the part of the δ record that covers the glaciation period, a few ice-dynamic remarks might be useful. In general terms the climatic deterioration after the onset of the glaciation 73,000 years B.P. (Fig. 10) made it possible for the Greenland ice sheet to grow in both lateral extent and in thickness. When full glacial severity was encountered at 59,000 B.P., the ice sheet probably covered all available land area as in the late stages of the glaciation, indicated by isostatic rebound of the coastal areas all around Greenland (WEIDICK, 1972). The altitude difference H between the crest and the margin of an ice sheet in mass balance can be deduced from ice-dynamic considerations (HAEFELI, 1961; PATERSON, 1972):

$$H = k \cdot \sqrt{L}$$

L being a characteristic dimension of the base and k a constant that depends only slightly on the temperature and accumulation. This important equation shows that H is mainly determined by the horizontal extension of the ice sheet, which of course grows in times of cooling. On the other hand, the sea around Greenland limits the extent of the base, which may vary slightly with the sea level. The degree of cooling needed to make the ice sheet reach this limit is unknown, but it was most probably attained early in the glaciation. Therefore, the shape of the ice sheet has most likely remained essentially unchanged during most of the glaciation. Within this same period we may therefore apply the steady state considerations presented in the preceeding paragraphs.

During the workshop meeting on Temperature and Isotopic Profiles in Ice Sheets in Cambridge, January 1973 (ROBIN, 1973a), we discussed the consequences of possible fusion of the Greenland and the American ice sheets during the glaciation. The above considerations are also valid for such ice ridge as soon as it reached its maximum extent. Uplift data from Ellesmere Island (WALCOTT, 1972) and NW Greenland (WEIDICK, 1971) suggest that the ice ridge built up to a width of some 700–800 km between Knud Rasmussen Land and the growing Innuitian ice sheet over The Queen Elizabeth Islands. In view of the small distance between Northwest Greenland and Ellesmere Island (~ 50 km) a fusion of the two ice sheets could have taken place early in the glaciation followed up by the build-up of a ridge to a steady state with altitudes reaching 2500–3000 m to judge from the 800 km wide Mid Greenland ice sheet today.

In the following we assume that the width and, thereby, the thickness of the ridge was only slightly influenced by sea level changes, as long as the sea level was low. On the other hand, at the end of the glaciation, the ridge must have disintegrated extremely fast, when the rising sea, assisted by the rapidly warming climate, undermined the glacier in the deep channel between Greenland and Ellesmere Island. This appears from the almost instantaneous rise of the δ curve at 12,500 B.P. in the time scale used in Fig. 10. This event left the Northwestern part of the Greenland ice sheet higher than today, but still shrinking till it reached the present stage.

The development of the glaciological conditions described above, suggests the following tentative conclusions as to the Camp Century δ record: During the build-up of the ridge, the altitude of deposition of the layers from that period increased by some 800 m, probably with a halt during the Brørup Interstadial. The corresponding δ change in the period 73,000 to 59,000 B.P. was some 6 ‰. The remaining 6 ‰ drop in δ within the same period (Fig. 10) reveals the climatic cooling. In the period 59,000 to 12,600 B.P. we apply, as a first approximation, a correction of 6 ‰, assuming (i) essentially steady state of the ridge, and (ii) sites of deposition close to its ice divide. The latter assumption is reasonable, because during the rapid thinning that followed only the ice in the central part of the ridge would be left on the Thule peninsula. As a consequence of the suggested correction, the interstadial interpreted as Plum Point in Fig. 10 appears to have reached close to present day climatic conditions, which is supported by the Devon ice cap δ profile (PATERSON *et al.*, 1973). We desist from correcting the Camp Century δ curve during the dramatic disintegration of the ridge, though part of the smooth increase of the δ's above -34 ‰ undoubtly reveals the postglacial thinning of the ice sheet in Northwest Greenland. However, another part (of the order of 1 ‰) of the total increase must be ascribed to the deeper layers coming from further inland. As described on p. 35, lack of data prevents us from specifying the post-glacial δ corrections further than shown in Fig. 10.

Obviously, several other interpretations and corrections can be applied to the Camp Century record. In view of the complex glaciological development outlined above, the climatic component of the δ-record can hardly be separated completely. However, if a purely climatic δ record could be obtained from a more favorable location on the Greenland ice sheet, the Camp Century record could be corrected for the climatic component and would thereafter render extremely valuable informations about past changes of the Northwest Greenland Ice Sheet.

Looking at the rest of the Greenland ice sheet, we find the most favorable ice flow conditions on the crest, where the deposits sink toward the bottom without any horizontal velocity. In other words, if the position of the crest has remained unchanged, any layer in a vertical ice core has been deposited at the same geographical position. WEERTMAN (1973) has recently shown that even considerable changes of the accumulation rate and distribution do not cause appreciable changes in the position or altitude of the crest.

Thus, we only have to consider possible secular altitude changes due to growth or shrinkage of the ice sheet in horizontal direction. If the length of the western slope was 750 km during the glaciation, in stead of the present 500 km, we must expect that H (the difference between the elevations of the crest and the margin) for the Mid Greenland ice sheet was $\sqrt{750/500} = 1.23$ times higher, *i.e.* 600 m higher than today. However, at present the western ice margin in Mid Greenland is located at an altitude of approximately 500 m above sea level, whereas during the glaciation the margin was close to the then sea level (some 100 m lower than the present sea level). Hence, it seems that the crest has had essentially the same altitude (relative to the sea level) during the climatic extremes in glacial and interglacial periods.

This makes Station Crete an ideal location for deep drilling, as far as the ice flow is concerned, which does not necessarily mean that it is also the best suited drilling site from a climatological point of view, because the meteorological pattern there may be more complicated than on the slopes. No matter where a deep drilling site is chosen, the accumulation and ice flow pattern in the drainage area should be investigated with great care.

As to Antarctica various parts of the ice sheet may react differently to changing climatic conditions. The extent of the continent is so large that it may be questionable, if changes of the Antarctic ice sheet are due to climatic changes, or vice versa. In times of high negative mass budget, enormous exhaust of glacier ice covering large areas of the Antarctic Ocean may reduce the albedo sufficiently to initiate a global climatic change. At least, the shape of the West Antarctic ice sheet suggests that it is presently off balance (HUGHES, 1972). Consequently, it is questionable to which degree the Byrd Station δ profile reflects variable climatic conditions. The δ trend within the last 4000 years suggests a lowering surface altitude, and the high δ values close to the bottom suggest much lower altitudes in West Antarctica prior to the Wisconsin/Würm glaciation than today (JOHNSEN *et al.*, 1972).

5.2. Smoothing of long-term δ variations

Up till now, no isotope profiles have been obtained through temperate glaciers. Smoothing by exchange between firn and percolating melt water may show to be completely destructive. On low latitude glaciers at extremely high altitudes (Himalaya, Chile) the conditions may be favorable enough for preservation of long term δ variations. At low latitudes, however, climatic changes are not very pronounced, yet probably the more important in global climatic connection.

The smoothing by diffusion in polar firn, mentioned in section 4.1. is negligible as regards long-term δ variations due to climatic changes: If the amplitude of seasonal δ-oscillations is reduced by a factor F during the first, important stage of the firnification, a climatic δ cycle with period n years will be smoothed by a factor of the order of $(F)^{1/n^2}$ (JOHNSEN, 1973). Even for F as low as 0.01, this is in practice equal to 1 already for $n = 5$. Therefore, under otherwise favorable conditions long term climatic changes may be revealed by δ records through polar glaciers even in areas, where no seasonal δ oscillations survive the firnification process.

At greater depths, a harmonic δ oscillation of period n years will be smoothed by diffusion in the solid ice according to equation (20) in JOHNSEN (1973) if λ is replaced by $n \cdot \lambda$. Table 2, columns 10 and 11 suggest for $n = 10$, that a possible climatic δ variation with a period corresponding to the sun spot cycle is detectable back to $t_m = 17{,}000$ yrs in South Greenland and to 50,000 yrs in North Greenland. In East Antarctica t_m may be more than 100,000 yrs. At the level of disappearance the layer thickness $n\,\lambda$ is 5 to 9 cm for $n = 10$.

Columns 7–8, 13–14, 16–17 and 19–20 show that if the period, n yrs, of the considered δ cycle is increased by an order of magnitude, the "life time" t_m of the δ cycle increases at all stations by a factor close to 2.3, and $n\,\lambda$ by approximately 1.7. The significance of y_0 (m above the bottom) is that below this level the n years δ cycle cannot be expected to be detectable although in some cases past temperatures below the present one have given better conditions for preservation than assumed in the calculation.

5.3. Applications

Obviously, the cold ice sheets contain climatic informations spanning very long periods of time. It may be rational to distinguish between (i) short-term climatic changes with periods up to a few thousand years and (ii) long-term climatic changes. In the first category, we find all the climatic oscillations that are of interest for short-term prognostic purposes, in addition to all the variations that describe the main features of the present interglacial period. The second category is important for the study of the glacial chronology in the Pleistocene, including the problem of the duration of the present interglacial period.

5.3.1. Long-term dating. The 80 and 180 years δ cycles found in the upper 800 years δ record from Camp Century (Johnsen *et al.*, 1970; Dansgaard *et al.*, 1972a) need to be checked; as to persistency over much longer intervals of time and, as to global significance, on ice cores from other locations in Greenland, Antarctica and, if possible, other glaciers. Based on the detailed δ record from Dye 3, spanning some 740 years (Fig. 11, p. 47), Hibler *et al.*, (1973) have found evidence for the 80 years cycle in South Greenland.

Table 2, columns 11, 14 and 17, shows that, if global, these δ cycles may be found back to the beginning of the Wisconsin/Würm, in some cases even much further, both in Greenland and Antarctica. If these cycles reflect solar activity variations, as suggested by an anticorrelation between δ and C^{14} in treerings (Dansgaard *et al.*, 1971), they may be used for dating purposes just like the seasonal δ variations (section 4.3.2) and far beyond the range of the latter (*cp.* columns 8 and 14 in Table 2).

5.3.2. Climatic prognoses for the coming decades or centuries have become an urgent need. Long-term planning of industrial, agricultural and communication developments have always rested upon the assumption of essential stability of the environments. However, the growing understanding of the existence and importance of natural climatic changes, in addition to the potentialities of modern man to influence the environment, has emphasized the need for clarifying the complex processes that create our climate, the interplay of numerous feed-backs, self-adjusting or amplifying processes that is our dynamic environment. Great efforts are devoted to the task of establishing a comprehensive model that accounts for all the atmospheric, continental, oceanic and extra-terrestrial parameters that influence the climatic balace. A better knowledge about the natural climatic changes during the past centuries or millenia is a basic condition for verifying the validity of any model, including its ability to predict the consequences of man's impact on the climate. This is why the very registration of past climatic events, for example by the isotope method used on ice cores, is more important than ever. If such studies show that the climatic conditions in considerable parts of the world are subjected to regular, persistant oscillations, this must be accounted for in the model. It might even be possible to make a short-cut by predicting future changes simply as a continuation of those of the past, without complete understanding of the complex mechanism behind. The confidence limit of such predictions could be checked by using part of the known data to predict the rest. Such prognoses of future climates would of course imply negligible influence of future human activity but, on the other hand, the model needed to calculate this inflence as a correction would be simpler than a complete model.

The first attempt to use this technique on isotope data was presented by JOHNSEN *et al.*, 1970 and DANSGAARD *et al.*, 1972a. Another attempt based on a different technique and a different material (the Dye 3 ice core) will be presented by HIBLER *et al.*, (1973). Some results of these investigations are briefly outlined at the end of section 7.1., p. 48.

5.3.3. Glacial chronology. According to EMILIANI (1966) the mean frequency of the occurrence of glaciations is one per 40,000 years. Other claim much lower frequencies (WOLLIN *et al.*, 1971). Thus $n = 10^4$ years is the longest period of interest to the study of glacial sequences. Such long δ cycles may be preserved for half a million years in Greenland, and for one million years in East Antarctica (Table 2, column 20). Strata of these ages are to be found nearly 20 m above the bottom in Mid Greenland and some 40 m above the bottom in East Antarctica (column 18). If the potential radioactive dating methods based on Kr^{81} and Mn^{53} (DANSGAARD & OESCHGER, 1973) are firmly developed, the study of these very old layers would give an independent climatic record, spanning a great part of the Pleistocene and invaluable as a check on, and supplement to the ocean floor records, *cp.* p. 7. This is important the more so as deep disagreement still exists among paleo-climatologists, not only on the general problem of the character and frequency of pleistocene glaciations, but also on the specific problem of the duration of the Sangamon/Eemian interglacial that preceeded the Wisconsin glaciation.

The general problem raises the question of how old ice one can expect to find deep in the large ice fields. The answer depends highly on the temperatures at the bedrock now and in the past, which is another unsolved problem (BUDD *et al.*, 1971). Assuming that the bedrock temperatures have always been below the pressure melting point, we find the ages of ice 10 m above the bottom given in Table 2, p. 26, column 21. Apparently, one might get close to 1 million years in Mid and North Greenland, and reach several million years in East Antarctica. Such old ice might give interesting informations about chemical compositions etc., but the diffusion delimits the range of the stable isotope method to the ages given in column 20.

The specific problem (the duration of the Sangamon/Eem) raises another question, namely how much pre-Wisconsin ice is available at the various locations. In the Camp Century ice core, the layer indicating the Sangamon/Wisconsin transition is located some 38 m above the bottom, or only 21 m above the silty ice, the nature of which is not clearly understood. This is one reason why the age of the clear/silty ice interface, estimated at 127,000 years, is encumbered with considerable uncertainty. Table 2, column 22, shows that considerably more pre-Wisconsin ice (more than 100 m) exists in other parts of Greenland, and a great deal more ($\sim$ 1000 m) in Central East Antarctica, still with the assumption of no bottom melting now or ever since the build-up of the ice sheets in the distant past.

It seems most likely that ice cores from Mid Greenland and East Antarctica contain sufficient information to determine the duration of the Sangamon/Eem interglacial and to give an independant check on the validity of a great part of EMILIANI's glacial chronology. If this is proven correct, there are good reasons for emphasizing the unusually long duration (10,000 years) of the present interglacial period (EMILIANI, 1972). In turn, this would stress the necessity that man desists from further impact on the global environment, because such impact might lead to uncontrolable disturbances of a possibly labile climatic balance.

6. Conclusions

As listed in the introduction, there are numerous reasons to believe that continued and intensified ice core drilling will show to be scientifically rewarding to a higher extent than any other technique in geophysical Polar research. As to the application of stable isotopes in ice cores, the perspectives may be summarized as follows:

Accumulation rates during many thousands of years can be determined over most of the Greenland ice sheet and in certain marginal areas of Antarctica by using seasonal δ variations. Over the rest of the two large ice sheets accumulation rates may be determined by considering long term δ variations (or radioactive dating). Such measurements are essential for estimating the mass balance now and in the past. Dating of deep ice cores is possible by the same techniques.

Stable isotope profiles through ice sheets reveal past climatic conditions, if surface altitude changes and redistribution of deposits can be neglected or corrected for. In Greenland such records may reach half a million years backwards in time, in Antarctica several million years in areas where the bottom temperature has never reached the pressure melting point. Such records would be extremely valuable (i) for atmosphere model calculations, because detailed long-term climatic records do not yet exist, (ii) for determining a possible phase difference between climatic changes in the Northern and Southern Hemisphere, (iii) for checking the possibility of predicting the climatic development in the next decades or centuries, (iv) for checking the validity of the climatic records obtained from deep sea cores, (v) for estimating the probable duration of the present interglacial, and (vi) for estimating the corrections to the C^{14} scale beyond the range of the dendrochronological technique.

In case of unstable ice sheets δ profiles reflect a combination of surface altitude and climatic changes, each of which can be determined, if the other can be corrected for.

The most promising aspects of ice core drilling are offered in the dry and percolation zones of Polar ice sheets. In view of the high cost, any deep drilling should be preceeded by drilling to intermediate (400–500 m) and/or shallow depths (50–100 m) at the intended deep drilling site and at nearby alternative sites.

A 100 m core represents approximately 80 m of ice equivalent or between 125 and 4000 years of accumulation at the stations listed in Table 2. In most cases this is enough not only for checking the preservation of seasonal δ variations, but also to give an idea of to which degree the long-term δ variations reflect the known climatic changes within the recent centuries. A number of 100 m cores from localities well spread over the Greenland ice sheet, combined with complete maps of the bottom and surface topography, would be an excellent basis for a choise of intermediate and deep drilling sites.

If the main purpose with deep drilling is a search for climatic records, the ice core should of course contain δ variations that are easy to interprete in terms of climatic variations, *i.e.* δ variations due to changing surface altitudes, or to secular changes in the summer to winter precipitation ratio, should be negligible or possible to correct for. Only direct checks by intermediate or shallow drilling can show, if any part of the Antarctic ice sheet or shelves meets this requirement.

Furthermore, the ice core should be easy to date, either by seasonal δ variations or by ice flow considerations. This requirement counts in favor of high accumulation areas close to the crest or the main ice divides. On the other hand, in such areas the origin or direction of precipitating air masses may shift, when the climate changes, complicating the meteorological pattern. This is probably of minor importance in Greenland, to judge from the close relationship between δ and temperature today, *cp*. Fig. 2, p. 12, but, again, drilling to intermediate or maybe even shallow depths can clarify this point.

Finally, it should be emphasized that the climatic informations contained in the two large ice sheets can only be revealed by several deep ice cores from locations that are well distributed in the various climatic regimes. This may raise the need for two deep drillings far apart on the same flow line. However, it is difficult to see any other argument for it. Thus, determining the position of a given layer inland from the margin is much easier by radio-echo sounding. But it is important that a deep drilling site far from the crest be chosen in consideration of a simple flow pattern upslope; thorough strain measurements and several shallow drillings in the drainage area are necessary for correction of the δ-profile for the isotopic altitude/inland effect.

The combination of generally high accumulation and simple flow pattern, stability, extent and accessibility makes most of Greenland well suited for climatic isotope studies on ice cores.

If the main purpose of a deep drilling is to get information about flow patterns, the suitable areas are characterized by a high altitude/inland effect. Furthermore, the core must be datable, because the δ profile has to be corrected for secular climatic changes. This technique will in some cases give informations about past surface altitude changes.

Table 3

Station		Close to crest or main ice divide	Smooth bottom topography in drainage area	Simple flow pattern	Simple meteorological pattern	Cores datable by seasonal δ variations		Climatic records beyond		Climatic details beyond 120,000 yrs. B.P.
No.	Name					back to 7000 yrs. B.P.	beyond 10,000 yrs. B.P.	80,000 yrs. B.P.	120,000 yrs. B.P.	
1	Camp Century	−	+	−	+	+	−	+	−	−
2	Jarl-Joset	(+)	+	+	+	+	+	+	+	+
3	Crete	+	+	+	(+)	+	+	+	+	+
4	Central	(+)	+	+	(+)	+	+	+	+	+
5	Milcent	−	+	+	+	+	+	+	+	−
6	Dye 3	+	−	−	+	+	−	+	+	−
7	South Site	+	−	−	+	+	−	+	−	−
8	North Site	+	+	(+)	+	−	−	+	+	+
9	75° S, 90°W	−	−	−	+	+	−	−	−	−
10	68.4° S, 120° E	−	−	−	+	+	−	+	(+)	(+)
11	72.4° S, 0° E	−	−	−	+	+	−	+	−	−
12	77° S, 10° E	+	+	(+)	−	−	−	(+)	(+)	(+)
13	South Pole	−	−	−	−	−	−	(+)	(+)	(+)
14	Pole Rel. Inaccess.	+	+	+	−	−	−	(+)	(+)	(+)
15	Vostok	+	+	+	−	−	−	(+)	(+)	(+)
16	Ross Ice Shelf	−	−	−	+	−	−	−	−	−
17	Byrd	−	−	−	−	−	−	−	−	−
18	Little America	−	−	−	+	−	−	−	−	−

In Table 3 are given a number of expected features for the stations listed in Table 2 and a few more in Antarctica. As long as the Dye 3 core is the only intermediate or shallow core recovered from the stations 2 through 14, any of them may, for individual reasons, be regarded as potential deep drilling sites. The stations 9, 10 and 11 are all locates in those high accumulation margin areas in Antarctica (Fig. 4, p. 17) that seem to offer the best possibilities for preservation of seasonal δ variations. However, the flow and accumulation conditions upslope from these station are too complicated and poorly known to justify calculations as those shown in Table 2. It should be kept in mind that a δ profile through fastly moving marginal ice reflects the past conditions far inland.

At this stage it would be hazardous to try to put up a priority list for deep drillings. The only reasonable proceedure is to use the shallow and intermediate drilling techniques at stations 2 through 14, and/or alternative stations, and let the results guide further action, both as regards when and where.

7. Appendix: The Greenland Ice Sheet Program (G.I.S.P.)

The previous sections, and particularly Table 3, show that only parts of the Greenland ice sheet and, maybe, a few isolated Canadian glaciers have ice flow and meteorological regimes that, a priori, look promising as to simple interpretation of a deep δ profile in terms of datable climatic changes. Furthermore, the Greenland ice sheet reaches from the Arctic Ocean to close to the main track of the North Atlantic cyclones governing the climate of a great and, society-wise, important part of the temperate belt of the Northern Hemisphere. Quite naturally, the Greenland ice sheet has therefore, after some years of intermission, regained its position as one of the foci of interest for deep drilling projects.

G.I.S.P. was initiated in 1971 as a long-term joint effort with scientific and technical contributions from U.S.A. Cold Regions Research and Engineering Laboratory (C.R.R.E.L., responsible for logistics, core drilling, physical and chemical core studies, surface strain measurements), University of Copenhagen (Geophysical Isotope Laboratory, responsible for stable isotope analyses, Si^{32} dating, dust concentration and β-activity measurements), Technical University of Denmark (Laboratory for Electromagnetic Field Theory — radio-echo sounding), University of Bern (Physikalisches Institut — radio-carbon dating of ice in situ), and University of Iceland (Science Institute — tritium dating of firn strata). Financial support has been given by the U.S. and Swiss National Science Foundations, the Ministry of Greenland and by the Universities involved.

The first step was to establish a solid basis for choosing the most favorable sites for deep drilling. An important requirement to sites far from the crest is that the sub-surface upslope must be relatively plain. Fortunately, radio-echo sounding from aircrafts, perfected by GUDMANDSEN and his co-workers, has shown to be a valuable tool for mapping the topographies of the surface (ROBIN, 1973b) and the bedrock underneath the ice sheet (ROBIN *et al.*, 1969). A systematic survey has been worked on since 1968 with impressive results (GUDMANDSEN, 1973). It shows a very smooth bottom topography along the EGIG track all the way from Camp VI to Jarl-Joset. Just East of the latter station subsurface mountains complicate the ice flow pattern. However, Jarl-Joset and all E.G.I.G. stations West of it seems to be well suited for a deep drilling operation as far as the ice flow pattern is concerned, the more so as great efforts have already been devoted to clarify the strain pattern (E.G.I.G.).

In the following, we shall briefly outline some other G.I.S.P. activities that have taken place up till now, or are planned for the years to come.

7.1. Some results from the field season 1971

The first practical result of the G.I.S.P. drilling activity was the recovery of an ice core to intermediate depth (377 m) at the Dye 3 Station (65°11′ N, 43°50′ W), where the total ice thickness is approximately 1900 m, according to radio-echo sounding measurements (GUDMANDSEN, 1970). The choice of Dye 3 was made for economical and logistic rather than for glaciological reasons. The station is located some 30 km East of the ice divide, and the bottom topography is rough. However, this latter point is not essential, since the ice core spans, less than 25 $^0/_0$ of the total ice thickness.

The ice core was split along the axis and shared between C.R.R.E.L. and our laboratory. Some 6000 δ-measurements on small increments (each c. $^1/_8$ year) in a continuous sequence show approximately 740 seasonal δ oscillations. Hence, the ice core reaches back to c. 1230 A.D. Furthermore, the accumulation rate in the Dye 3 area has remained surprisingly constant, *cp.* the left part of Fig. 11 showing an annual accumulation record, in which oscillations shorter than 30 yrs have been essentially removed by applying a digital low-pass filter (DANSGAARD *et al.*, 1971). The most pronounced deviation from the overall mean value, 48 $gr \cdot cm^{-2} \cdot yr^{-1}$ apparently occurred around 1400 A.D. It is interesting to note that in this same period Iceland was ravaged upon frequent, heavy land slides (THORODDSEN, 1916), and that in Denmark farming was possible only on the high-lying lands ("højbrug"), the low-lying ones being frequently flooded. Was all this due to increased cyclonic activity over the North Atlantic Ocean?

The mid section of Fig. 11 shows the corresponding δ record. Some degree of correlation exists between the accumulation and the δ curves, which might be explained by varying mean position of the Polar front. As expected the δ's vary less than at Camp Century (right section) due to the more oceanic character of the climate in South Greenland. However, the two δ curves exhibit several mutual features, *cp.* for example the generally sub-normal δ's during the "little ice age" in the 16'th and 17'th century. In some periods the two δ curves are out of phase (*cp.* most of the 18'th century and the time around 1300 A.D. and just before 1500 A.D.). This is probably due to the fact that no exact dating of the

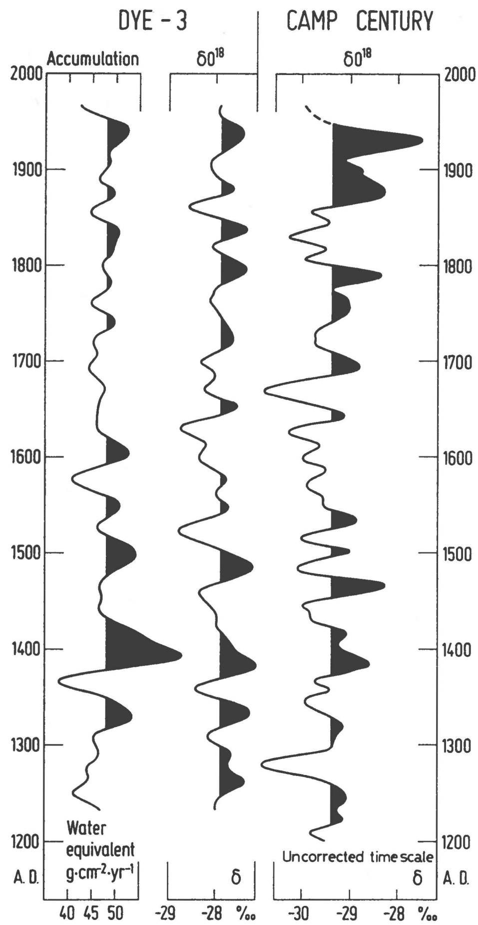


Fig. 11. Digital-filter smoothed records of annual accumulation at Dye 3 (left) and $\delta(O^{18})$ at Dye 3 and Camp Century. The Dye 3 data (including the time scale) are based on a continuous series of 740 seasonal δ oscillations. The Camp Century time scale is based on the assumption of unchanged accumulation rate. This might be the reason why the two δ curves are occasionally out of phase. However, both of them show the climatic optimum around 1930, and generally sub-normal δ-values during "the little ice age". Some correlation (coefficient 0.6) appears between the accumulation and δ curves from Dye 3, sub-normal accumulation occurring in cold periods.

Camp Century ice core has been performed by counting seasonal δ variations.

The strain rate measurements, radio-carbon measurements, as well as dust and other physical and chemical analyses will be reported elsewhere, *e.g.* silicon-32 dating measurements by CLAUSEN (1973). A preliminary time series analysis of the isotopic and stratigraphic features will soon be published by Hibler *et al.*, (1973). In essential agreement with the previous one (Dansgaard *et al.*, 1972a) the tentative prognosis suggests that the cooling within the recent decades may soon be replaced by a slight warming up till approximately 1990. Thereafter, the tentative prognosis suggests minor fluctuations with cold peaks around 2005 and 2060 A.D. The warm peaks foreseen for the coming century will hardly bring back the extremely warm conditions of the 1930'ies. However, once again it should be emphasized that longer records from different climatic regions and other studies are needed.

7.2. The field season 1972

In spring 1972 analysis of firn strata on the Devon Island ice cap indicated a fast smoothing of seasonal δ variations present in the upper firn, in spite of a considerable accumulation rate ($20\ \mathrm{gr \cdot cm^{-2} \cdot yr^{-1}}$). This observation stressed the need for investigating the isotopic homogenization at the intended intermediate drilling sites in Greenland, Crete and North Site. Handaugered cores were recovered in June 1972. The results, shown in Fig. 7, p. 23, suggest that the seasonal δ variations probably survive the firnification process at Crete. If so, it further supports the first priority of Crete as a deep drilling site. See text to Fig. 7.

7.3. The field season 1973

In the summer of 1973 an intermediate drilling at the E.G.I.G. station Milcent (295 km from the ice divide) resulted in the recovery of a 398 m long ice core that spans the last c. 780 years. 7000 samples for O^{18} analyses were cut from the core in a continuous sequence. Detailed sampling from pits and hand augered cores collected in the vicinity of Milcent is intended to reveal possible local effects upon the isotopic stratification of the deposits, as well as the internal mixing of the layers. Near surface samples spanning the last 27 years were collected from various locations for tritium and chemical analyses.

In addition, valuable experience was gained in relation to two devises for shallow drilling (to 50–100 m) recently developed at the University of Bern and C.R.R.E.L.. When perfected, they will be extremely useful tools for survey purposes, because they will be easy to move, install and use without heavy logistic support.

References

AEGERTER, S., OESCHGER, H., RENAUD, A., & SCHUMACHER, E., 1969. Tritium studies on firn samples collected by E.G.I.G. in Greenland. *Meddelelser om Grønland*, Bd. **177**, No. 2, pp. 76–88.

ALDER, B., GEISS, J., GROEGLER, N., & RENAUD, A., 1969. Gas composition in ice samples collected by E.G.I.G. in Greenland. E.G.I.G. 1957–60. No. 5, No. 3 (ed. A. RENAUD). In *Meddelelser om Grønland*, Bd. **177**, No. 2, pp. 100–107.

AMBACH, W., & DANSGAARD, W., 1968. Fallout and climate studies on firn cores from Carrefour, Greenland. *Earth and Planetary Science Letters*, Vol. **8**, pp. 311–16.

AMBACH, W., DANSGAARD, W., EISNER, H. & MØLLER, J., 1968. The altitude effect on the isotopic composition of precipitation in the Alps. *Tellus*, Vol. **20**, pp. 595–600.

AMBACH, W., EISNER, W., & PRANTL, F. A., 1967. Investigation on fission products in the accumulation area of an Alpine glacier. I.U.G.G. General Assembly at Bern, Sept. Oct. 1967. International Association of Scientific Hydrology, Commision on Snow and Ice, *Publication*. No. **79**, pp. 117–125.

ANDERSON, D. L., & BENSON, C., 1962. The densification and diagenesis of snow In Ice and Snow (ed. KINGERY, W. D.), *Proceedings of a conference held at The Massachusetts Institute of Technology*, February 12–16, 1962. pp. 391–411.

ARNASON B., 1969. The exchange of hydrogen isotopes between ice and water in temperate glaciers. *Earth and Planetary Science Letters*, Vol. **6**, pp. 423–30.

ASTAPENKO, P. D., 1964. *Atmospheric Processes in the high latitudes of the Southern Hemisphere*. Translated from Russian, Jerusalem, 1964. Distributed by Oldbourne Press, London, 1964.

BENSON, C., 1962. Stratigraphic studies in the snow and firn of the Greenland ice sheet. *S.I.P.R.E. Research Report*, No. **70**, pp. 1–93.

BUASON, T., 1972. Equation of isotope fractionation between ice and water in a melting snow column with continuous rain and percolation, *Journal of Glaciology*, Vol. **11**, pp. 387–405.

BUDD, W., 1966. The dynamics of the Amery Ice Shelf. *Journal of Glaciology*, Vol. **6**, pp. 335–58.

BUDD, W., JENSSEN, D., & RADOK, U., 1971. Derived physical characteristics of the Antarctic ice sheet. Mark I. University of Melbourne, Meteorological Department, *Publication*, No. 18, 178 p.

BULL, C., 1971. Snow accumulation in Antarctica. In *Research in the Antarctic* (ed. L. O. QUAM), A.A.A.S. *Publication* No. **93**, pp. 367–421.

CLAUSEN, H. B., 1973. Dating of polar ice by silicon-32. *Journal of Glaciology*, Vol. **12**, No. 66.

CRAIG, H., 1961 a. Standards for reporting concentrations of deuterium and oxygen-18 in natural waters. *Science*, Vol. **133**, pp. 1833.

— 1961 b. Isotopic variations in meteoric waters. *Science*, Vol. **133**, pp. 1702–3.

Craig, H., & Gordon, L., 1965. Deuterium and oxygen-18 variations in the ocean and marine atmosphere. In *Stable Isotopes in Oceanographic Studies and Paleotemperatures* (ed. E. Tongiorgi). Laboratorio Geologia Nucleare. Pisa, pp. 9–130.

Crary, A. P., 1968. Presidential address to the ISAGE meeting in Hanover, N.H. In *Proceeding of International Symposium on Antarctic Glaciological Exploration* (ed. A. J. Gow, C. Keeler, C. C. Langway and W. F. Weeks), Publication No. 86 of the International Association of Scientific Hydrology and Scientific Committee on Antarctic Research, 1970.

Crary, A. P., Robinson, E. S., Beunet, H. F., & Boyd, W. W., Jr., 1962. Glaciological regime of the Ross Ice Shelf, *Journal of Geophysical Research*, Vol. **67**, pp. 2791–2807.

Crozaz, G., Langway, C. C. Jr., & Picciotto, E., 1966. Artificial radioactivity reference horizons in Greenland firn. *Earth and Planetary Science Letters*, Vol. 1, pp. 42–48.

Dansgaard, W., 1954. The oxygen-18 abundance in fresh water. *Geochimica et Cosmochima Acta*, Vol. **6**, pp. 241–60.

— 1961. The isotopic composition of natural waters, *Meddelelser om Grønland*, Bd. **165**, No. 2, 120 p.

— 1964. Stable isotopes in precipitation, *Tellus*. Vol. **16**, pp. 436–68.

— 1969. Oxygen -18 analysis of water. In *Meddelser om Grønland*, Bd. **177**, No. 2, pp. 33–36.

Dansgaard, W., & Johnsen, S. J., 1969. A flow model and a time scale for the ice core from Camp Century, Greenland, *Journal of Glaciology*, Vol. 8, pp. 215–23.

Dansgaard, W., & Oeschger, H., 1973. Isotope Glaciology, to be published in *Geopedia*, Pergamon Press, Oxford, U.K.

Dansgaard, W., Merlivat, L., & Roth, E., 1969a. Stable isotope measurements on firn cores. *Meddelelser om Grønland*, Bd. **177**, No. 2, pp. 62–76.

Dansgaard, W., Johnsen, S. J., Møller, J., & Langway, C. C. Jr., 1969b. One thousand centuries of climatic record from Camp Century on the Greenland ice sheet. *Science*, **166**, pp. 377–81.

Dansgaard, W., Johnsen, S. J., Clausen, H. B., & Langway, C. C. Jr., 1971. Climatic record revealed by Camp Century ice core. In *Late Cenozoic Glacial Ages* (edit. K. K. Turekian), pp. 37–56, Yale University Press, New Haven, Conn., 1971.

— 1972a. Climatic oscillations depicted and predicted by stable isotope analyses of a Greenland ice core. In *Proceedings of the First Int. Conf. on Port and Ocean Engineering under Arctic Conditions*, Vol. **1**, pp. 17–22.

— 1972b. Speculations about the next glaciation. In The Holocene Interglacial—how and when will it end? (ed. A. L. Washburn). *Quarternary Research*, Vol. **2**, No. 3, pp. 296–8.

— 1973. Oxygen isotope profile through the Ross Ice Shelf. Manuscript in preparation.

Deutsch, S., Ambach, W., & Eisner, E., 1966. Oxygen isotope study of snow and firn on an Alpine glacier. *Earth and Planetary Science Letters*, Vol. **1**, pp. 197–201.

Emiliani, C., 1966. Isotopic paleotemperatures, *Science*, Vol. **154**, pp. 851–57.

— 1972. Quarternary paleotemperatures and the duration of the high-temperature intervals. *Science*, Vol. **178**, pp. 398–401.

Epstein, S., 1956. Variations in the O-18/O-16 ratios of fresh water and ice. National Academy of Science, Nuclear Science Series, *Report* No. **19**, pp. 20–25.

Epstein, S., & Sharp, R. P., 1959a. Oxygen isotope variations in the Malaspina and Saskatchewan Glaciers, *Journal of Geology*, Vol. **67**, pp. 88–102.

— 1959b. Oxygen isotope studies. *Transactions of American Geophysical Union*, Vol. **40**, pp. 81–84.

EPSTEIN, S., SHARP, R. P., & GODDARD, I., 1963. Oxygen-isotope ratios in Antarctic snow, firn and ice. *Journal of Geology*, Vol. **71**, pp. 698–720.

EPSTEIN, S., SHARP, R. P., & GOW, A. J., 1965. Six-year record of oxygen and hydrogen isotope variations in South Pole firn, *Journal of Geophysical Research*, Vol. **70**, pp. 1809–14.

GONFIANTINI, R., 1965. Some results on oxygen isotope stratigraphy in the deep drilling at King Baudouin Station, Antarctica, *Journal of Geophysical Research*, Vol. **70**, pp. 1815–19.

GOW, A. J., 1963. The inner structure of the Ross Ice Shelf at Little America V, Antarctica, as revealed by deep core drilling. I.U.G.G. General Assembly at Berkeley, Int. Association of Scientific Hydrology, *Publication* No. **61**, pp. 272–284.

GOW, A. J., & WILLIAMSON, T., 1971. Volcanic ash in the Antarctic ice sheet and its possible climatic implications, *Earth and Planetary Science Letters*, Vol. **13**, pp. 210–18.

GUDMANDSEN, P., 1970. Notes on radar sounding of the Greenland ice sheet. *Proceedings of International Meeting on Radio-glaciology*, Technical University of Denmark, Lyngby, May 1970, pp. 124–133.

— 1973. Radioglaciology. Soundings at proposed drill sites, Report D 185, Laboratory of Electromagnetic Theory, Technical University of Denmark, August 1973.

HAEFELI, R., 1961. Contribution to the movement and the form of ice sheets in the Arctic and Antarctic. *Journal of Glaciology*, Vol. **3**, pp. 1133–51.

HAMILTON, W. L., & LANGWAY, C. C., 1967. A correlation of microparticle concentrations with oxygen isotope ratios in 700 year old Greenland ice. *Earth and Planetary Science Letters*, Vol. **3**, pp. 363–6.

HAMILTON, W. L., & O'KELLEY, M. E., 1971. Investigations of Particulate Matter. In *Antarctic Snow and Ice Studies II*, Antarctic Research Series (ed. A. P. CRARY), Vol. **16**, pp. 355–62.

HANSEN, B. L., & LANGWAY, C. C. Jr., 1966. Deep core drilling in ice and core analyses at Camp Century, Greenland, 1961–66. *U.S. Antarctica Journal*, Vol. **1**, pp. 207–208.

HEUBERGER, J. C., 1954. *Greenland, Glaciologie*, Vol. I. Forages sur l'Inlandis. Hermann and Cie, Paris.

HIBLER, B., *et al.*, 1973. Isotopic and stratigraphic time series analysis of an ice core. Manuscript in preparation.

HUGHES, T., 1972. Is the West Antarctic ice sheet disintegrating? *Journal of Geophysical Research*, in press.

I.A.E.A. INTERNATIONAL ATOMIC ENERGY AGENCY, 1969, 1970. Environmental isotope data, World survey of isotope concentration in precipitation, No. 1, 1953–63. No. 2: 1964–65. Vienna.

JOHNSEN, S. J., 1973. Isotopic homogenization in firn and ice strata. Manuscript in preparation.

JOHNSEN, S. J., DANSGAARD, W., CLAUSEN, H. B., & LANGWAY, C. C. Jr., 1970. Climatic oscillations 1200–2000 A.D., *Nature*, Vol. **227**, pp. 482–483.

— 1972. Oxygen isotope profiles through the Antarctic and Greenland ice sheets, *Nature*, Vol. **235**, pp. 429–34, and Vol. **236**, p. 249.

LAMB, H. H., 1970. Volcanic dust in the atmosphere; with a chronology and assessment of its meteorological significance. *Philosophical Transactions of the Royal Society of London*. Vol. **266 B**, pp. 425–539.

LANGWAY, C. C. Jr., 1970. Stratigraphic analysis of a deep ice core from Greenland. The Geological Society of America, *Special paper* **125**, 186 p..

LIESTØL, O., 1970. Private communication.

ˀLOEWE, F., 1954. Études de glaciologique en Terre Adélie, 1951–52. Expéditions Polaires Francaises, No. **9**. Hermann, Paris.

LORIUS, C., 1963. Le deuterium, possibilités d'application aux problèmes recherche concernant la neige, le névé et la glace dans l'Antarctique. *Comité National Francais des Recherches Antarctique*, No. **8**, pp. 1–102.

LORIUS, C., LAMBERT, G., HAGEMANN, R., MERLIVAT, L., & RAVOIR, J., 1968a. Dating of firn layers in Antarctica: application to the determination of the rate of snow accumulation. In *Proc. of Int. Symp. on Antarctic Glaciological Exploration*, Hanover, N.H., Sept. 1968. Publ. No. 86 of Int. Acc. Sci. Hydrol. and Sci. Comm. on Antarctic Res., 1970, pp. 3–14.

LORIUS, C., RAYNAUD, D., & DOLLE, L., 1968b. Densité de la glace et étude des gaz en profondeur dans un glacier Antarctique, *Tellus*, Vol. **20**, pp. 449–59.

LORIUS, C., MERLIVAT, L., & HAGERMANN, R., 1969. Variation in the mean deuterium content of precipitations in Antarctica, *Journal of Geophysical Research*, Vol. **74**, pp. 7027–31.

MATSUO, S., & MIYAKE, Y., 1966. Gas composition in ice samples from Antarctica. *Journal of Geophysical Research*, Vol. **71**, pp. 5235–41.

MERLIVAT, L., & LORIUS, C., 1973. Manuscript in preparation. Isotopic temperature effect demonstrated by Fig. 7 *in* C. LORIUS: Les calottes glaciaires, témoins de l'environment. *La Recherche*, Vol. **4**, No. 34, pp. 457–472, 1973.

MILLER, M. M., 1954. Juneau Icefield Research Project, Alaska, 1950. American Geographical Society, *J.I.R.P. Report* No. **7**.

MOCK, S. J., 1967. Calculated patterns of accumulation on the Greenland ice sheet, in *Journal of Glaciology*, Vol. **6**, No. 48, pp. 795–803.

— 1968. Snow accumulation studies on the Thule Peninsula, Greenland. *Journal of Glaciology*, Vol. **7**, pp. 59–76.

MUROZUMI, M., CHOW, TSAIHWA, J., & PATTERSON, C., 1969. Chemical concentrations of pollutant lead aerosols, terrestrial dusts and sea salts in Greenland and Antarctic snow strata. *Geochimica Cosmochimica Acta*, Vol. **33**, pp. 1247–94.

NIEF, G. 1969. Determination de la concentration en deuterium des echantillons. E.G.I.G. 1957–60, Vol. **5**, No. 3 (ed. A. RENAUD). In *Meddelelser om Grønland*, Bd. **177**, No. 2, pp. 36–42.

PATERSON, W. B. S., 1972. Laurentide ice sheet: Estimated volumes during late Wisconsin. *Reviews of Geophysics and Space Physics*, Vol. **10**, pp. 885–917.

PATERSON, W. B. S., *et al.*, 1973. Oxygen isotope profile through the Devon Island ice cap. Manuscript in preparation.

PHILBERT, K., & FEDERER, B., 1971. On the temperature profile and the age profile in the central part of cold ice sheets. *Journal of Glaciology*, Vol. **10**, No. 58, pp. 3–14.

PICCIOTTO, E., & WILGAIN, S. E., 1963. Fission products in Antarctic snow: a reference level for measuring accumulation. *Journal of Geophysical Research*, Vol. **68**, pp. 5965–72.

PICCIOTTO, E., DE MAIRE, X., & FRIEDMAN, I., 1960. Isotopic composition and temperature of formation of Antarctic snows. *Nature*, Vol. **187**, pp. 857–9.

PICCIOTTO, E., CAMERON, R., CROZAZ, G., DEUTSCH, S., & WILGAIN, S., 1968. Determination of the rate of snow accumulation at the Pole of Relative Inaccessibility, Eastern Antarctica: A comparison of glaciological and isotopic methods. *Journal of Glaciology*, Vol. **7**, pp. 273–287.

RADOK, U., JENSSEN, U., & BUDD, W., 1968. Steady state temperature profiles in ice sheets. In *Proceeding of International Symposium on Antarctic Glaciological*

Exploration, Hanover, N.H., Sept. 1968, Publication No. **86** of Int. Acc. Sci. Hydrol. and Sci. Comm. on Antarctic Res. 1970, pp. 151–165.

RAMSEIER, O. R., 1967. Self-diffusion in ice monocrystals, U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, U.S.A. *Research Report* No. **232**.

RAYNAUD, D., & LORIUS, C., 1973. Climatic implications of total gas content in ice at Camp Century, *Nature*, Vol. **243**, pp. 283–4.

ROBIN, G. DE Q., 1968. Stability of ice sheets as deduced from deep temperature gradients, in *Proceeding of International Symposium on Antarctic Exploration*, Hanover, N.H., Sept. 1968, International Association of Scientific Hydrology and Scientific Commision on Antarctic Research, Publication No. **86**, 1970, pp. 141–151.

— 1973a. Temperature and isotopic profiles in ice sheets. Workshop held at Scott Polar Research Institute, Cambridge, England, 1'st meeting, January 1973.

— 1973b. Private communication.

ROBIN, G. DE Q., EVANS, S., & BAILEY, J. T., 1969. Interpretation of radio echo sounding in polar ice sheets. *Philosophical Transactions of the Royal Society of London*, Vol. **265**, pp. 437–505.

SHARP, R. P., EPSTEIN, S., & VIDZIUNAS, I., 1960. Oxygen isotope variations in the Blue Glacier, *Journal of Geophysical Research*, Vol. **65**, pp. 4043–59.

SCHOLANDER, P. F., HEMMINGSEN, E. A., COACHMAN, L. K., & NUTT, D. C., 1961. Composition of Gas Bubbles in Greenland Icebergs, *Journal of Glaciology*, Vol. **3**, pp. 813–22.

SCHYTT, V., 1958. Norwegian-Brittish-Swedish Antarctic Expedition, 1949–52. *Scientific Results* (Oslo, Norsk Polarinstitutt), Vol. **4**, A, B, C, pp. 1–152.

THORODDSEN, TH., 1916. Arferdi A Islandi I thusund Ar. Hid Islenzka Fræðafelag, Copenhagen 1916.

UEDA, H. T., & GARFIELD, D. E., 1969. Core drilling through the Antarctic Ice Sheet. U.S. Army Cold Region and Research Laboratory, *Technical Report* No. **231**.

VERGNAUD, J. P., BOTTER, R., LORIUS, C., MERLIVAT, L., RAVOIR, J., & ROTH, E., 1973. Tritium and deuterium content of snow in Greenland. In press.

WALCOTT, R. I., 1972. Late Quaternary vertical movements in North America: Quantitative evidence of glacio-isostatic rebound. *Reviews of Geophysics and Space Physics*, Vol. **10**, pp. 849–84.

WEERTMAN, J., 1973. Position of ice divides and ice centers on ice sheets. *Journal of Glaciology*, in press.

WEIDICK, A. 1971. Quarternary map of Greenland, Geological Survey of Greenland, Copenhagen.

— 1972. Holocene shore-lines and glacial stages in Greenland—an attempt at correlation. Geological Survey of Greenland. *Report* No. **41**, 139 p.

WOLLIN, G., ERICSON, D. B., & EWING, M., 1971. Late Pleistocene climates recorded in Atlantic and Pacific deep-sea sediments. In *Late Cenozoic Glacial Ages* (ed. K. K. TUREKIAN), pp. 199–214, Yale University Press, New Haven, Conn.

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