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GLACIOLOGICAL RECONNAISSANCE OF AN ICE CORE DRILLING SITE, PENNY ICE CAP, BAFFIN ISLAND

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Abstract. A site situated close to the main divide of the Penny Ice Cap, Baffin Island was occupied in 1979 for the purpose of determining the suitability of this ice cap for providing proxy climatic data and other environmental time series for a span of 10^4 a. A 20 m core was extracted and analysed for stable oxygen isotopes, tritium concentration, pH, electrolytic conductivity, major ion concentrations, and particulate concentration. An adjacent dedicated shallow core was analysed for pollen content to determine if a significant seasonal variation in the pollen rain existed. From these measurements, and from the observations made on the stratigraphic character of the core, the mean net accumulation rate over the approximately 30 year period covered by the core is found to be about 0.43 m water equivalent per year. This is in agreement with a single value determined 26 years earlier at a nearby site (Ward and Baird, 1954). The mean annual temperature in the bore hole was found to be close to -14.4°C, possibly some 2–5 deg warmer than the expected mean annual surface air temperature at the site. This difference is due to the expulsion of latent heat upon freezing of melt water at depth in the snow-pack which gives rise to the many ice layers observed in the core. The percentage thickness of ice layers per year may be correlated with summer temperatures.

Total ice depths were measured using a 620 MHz radar echo-sounder. In the vicinity of the divide, over an area of 1 km^2, the ice depths vary from about 460 to 515 m. These values compare favourably with values determined from an airborne radar depth-sounding flight carried out over the ice cap by a joint U.S.–Danish mission operating out of Søndre Strømfjord, Greenland. The data suggest that the ice-cap divide would be a worthwhile location to deep core drill with an expected useful coverage of at least the Holocene period.

Résumé. *Reconnaissance glaciologique d'un site de forage dans la glace sur la calotte glaciaire de Penny dans l'île de Baffin.* Un site proche de la chaîne principale de la calotte glaciaire de Penny dans l'île de Baffin a été occupé en 1979 pour étudier la possibilité d'utiliser cette calotte glaciaire pour obtenir des données climatiques proches et d'autres séries dans le temps sur l'environnement sur un laps de temps de 10^4 ans. Une carotte de 20 m a été extraite et analysée pour les isotopes stables de l'oxygène, la concentration en tritium, le pH, la conductivité électrolytique, la concentration en ions majeurs et la teneur en particules. Une carotte adjacente a subi des analyses polliniques pour déterminer s'il y a une variation saisonnière significative des retombées de pollen. De ces mesures et des observations faites sur le caractère stratigraphique de l'échantillon on a conclu que l'accumulation moyenne annuelle sur la période d'environ 30 ans couverte par la carotte est d'environ 0,43 m d'équivalent en eau par an. Ceci est cohérent avec une valeur singulière déterminée 26 ans auparavant en un site voisin (Ward et Baird, 1954). La température moyenne annuelle dans le forage était proche de $-14,4$°C probablement quelque 2,5 deg plus chaude que la température moyenne annuelle de l'air attendue en surface à ce site. Cette différence est due à l'expulsion de chaleur latente de regel d'eau de fusion en profondeur dans le manteau neigeux qui donne la plus grande part des niveaux de glace observés dans la carotte. La proportion de l'épaisseur des niveaux de glace créée chaque année peut être corrélée avec les températures d'été. Les épaisseurs totales de glace ont été mesurées avec une sonde à écho radar à 620 MHz. Dans le voisinage de la chaîne centrale sur une surface de 1 km^2 les épaisseurs de glace varient d'environ 460 à 515 m. Ces valeurs se comparent correctement aux valeurs déterminées par un sondage radar aérien effectué sur l'ensemble de la calotte glaciaire par une mission commune U.S.–danoise opérant depuis le Søndre Strømfjord, Groenland. Les données suggèrent que la chaîne centrale de la calotte glaciaire pourrait être un bon emplacement pour un forage profond avec l'espoir de l'utiliser pour couvrir au moins la période de l'holocène.

Zusammenfassung. *Glaziologische Erkundung einer Kernbohrstelle auf dem Penny Ice Cap, Baffin Island.* Eine Stelle nahe der Haupteisscheide des Penny Ice Cap, Baffin Island wurde 1979 mit dem Ziel besetzt, die Eignung dieser Eiskappe zur Gewinnung repräsentativer Klimadaten und anderer umweltrelevanter Zeitserien über die letzten 10^4 Jahre zu bestimmen. Ein 20 m langer Bohrkern wurde gezogen und auf stabile Sauerstoffisotope, Tritium-Konzentration, pH-Wert, elektrolytische Leitfähigkeit, grössere Ionen-Konzentrationen und besondere Konzentrationen untersucht. Ein dicht danebenliegender, kurzer Bohrkern wurde auf seinen Pollengehalt untersucht, um zu erfahren, ob eine signifikante jahreszeitliche Schwankung im Pollenregen besteht. Aus diesen Messungen und aus den Beobachtungen über den stratigraphischen Aufbau des Bohrkerns wurde die mittlere Nettoakkumulation über die etwa 30 Jahre, die der Bohrkern überbrückt, zu *c.* 0,43 m Wasser pro Jahr bestimmt. Dies stimmit mit einem Einzelwert überein, der vor 26 Jahren an einer nahe benachbarten Stelle gewonnen wurde (Ward und Baird, 1954). Die mittlere Jahrestemperatur in dem Bohrloch ergab sich nahe an -14.4°C, mögicherweise etwa 2–5 deg wärmer als die erwartete mittlere Jahrestemperatur der Luft an der Bohrstelle. Diese Differenz ist mit dem Austritt latenter Wärme beim Gefrieren von Schmelzwasser in der Tiefe der Schneedecke zu erklären; dabei entstehen die vielen Eisschichten, die im Bohrkern zu finden waren. Die prozentuale Dicke der Eisschichten pro Jahr dürfte mit den Sommertemperaturen korreliert sein. Die Gesamtdicke des Eises wurde mit einem 620 MHz-Radarecholot gemessen. In der Nachbarschaft der eisscheide, in einem Gebiet von 1 km^2, schwanken die Eisdicken Zwischen etwa 460 und 515 m. Diese Werte passen sehr gut zu jenen, die bei einem Radarechoflug über der Eiskappe, ausgeführt im Rahmen einer gemeinsamen Unternehmung der USA und Dänemarks von Søndre-Strøm-Fjord in Grönland aus, ermittelt wurden. Die Daten lassen vermuten, dass die Eisscheide ein geeigneter Ort für eine tiefe Kernbohrung wäre, deren Datenausbeute zumindest das Holozän überdecken würde.

INTRODUCTION

Previous known glaciological research on the Penny Ice Cap, south-east Baffin Island (Fig. 1) was carried out in 1953 by an Arctic Institute of North America expedition (Baird and others, 1953; Ward and Baird, 1954) and by Weber and Andrieux (1970). The former publication reports on measurements confined to the far south-east sector of the ice cap, about 25 km distant from and possibly lower than the site herein described. The latter publication gives ice depth data to within 11 km to the north-west of the present site.

Four low domes, each reaching to about 1990 m above sea-level, exist in the east-central part of the ice cap (Fig. 2). The site, occupied in April-May 1979, lies between the two most northerly domes, but nearer the southern and higher one of the two. (The intention was to occupy this latter dome, but in landing by aircraft, the surface relief was lacking in contrast and a misidentification of position was made. Because of limited surface mobility, it was decided to establish the reconnaissance camp at the landing site about 500 m to the east of the main ice-cap divide).

The present site reconnaissance was carried out to determine if useful paleo-environmental information might be extracted from the ice cap and where a suitable deep drill site might exist. Of prime interest was the potential of the core for yielding proxy climatic and other atmospheric data, and, if positive, what useful time span was likely to be covered. Such data would be useful input to previous climatological modelling studies for Baffin Island (Barry and Fogarasi, 1968; Andrews and Barry, 1972; Andrews and others, 1980), as well as Holocene glacier fluctuations and palynological studies over the same interval (Miller, 1973).

Climatic data from this area, in conjunction with similar data from southern Greenland (Dansgaard and others, 1975) would be extremely useful in long-term studies related to the behaviour of the Icelandic

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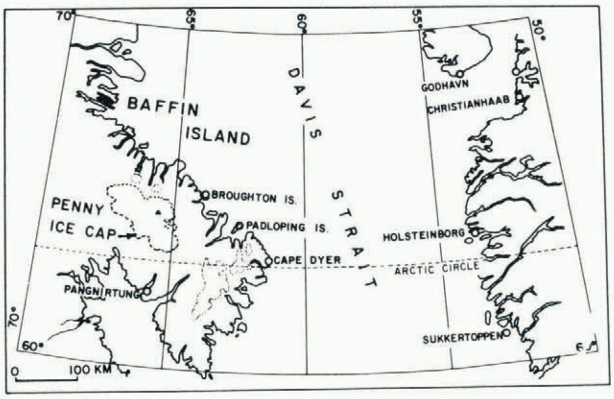

Fig.1. Map showing south-eastern Baffin Island and the location of Penny Ice Cap. Broughton Island and Cape Dyer are two sources of instrumental climatic data in the area.

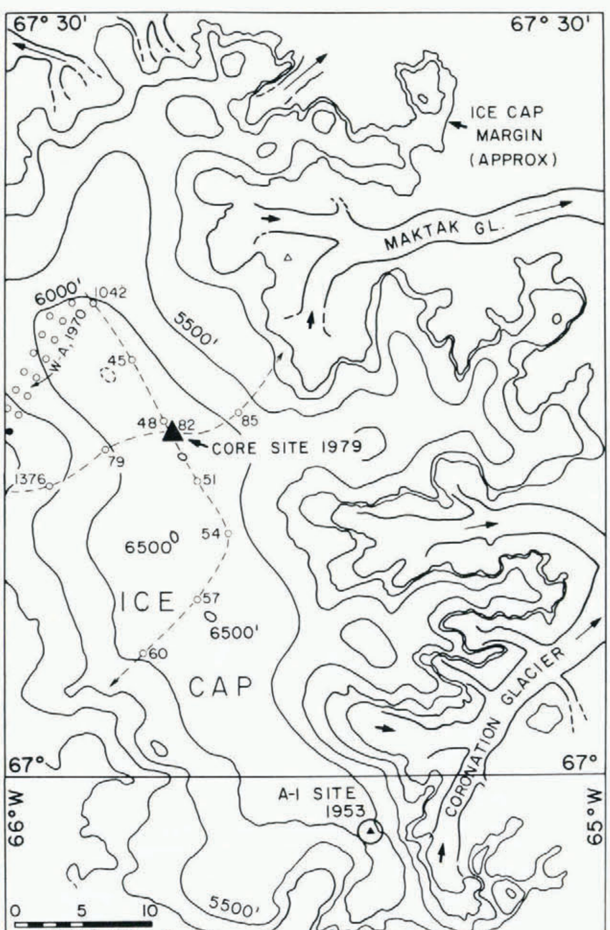

Fig.2. East-central section of the Penny Ice Cap showing the location of the 1979 field camp and core site. The highest part of the ice cap is defined by four broad domes rising above the 6500 ft (1980 m) contour. Site A-1, to the south-east was occupied in 1953 by an Arctic Institute of North America Party. The traverse and pole positions of Weber and Andrieux (1970) are also marked: W-A,1970. Two flight tracks of the airborne radar operation are shown with identifying numbers (see profiles in Fig. 17).

Low. This major climatic feature is of extreme importance for understanding the climate of the North Atlantic area and adjacent regions.

From available field evidence (Dyke and others, 1982) it may be assumed that the present location of the Penny Ice Cap divide is not significantly different from its location at any time over the last 10^4 years. From knowledge of the position of the ice margin during most of that time, it is unlikely that the ice thickness near the divide has changed by more than a few per cent over that time. This allows a relatively straightforward interpretation of an ice core that might be retrieved by drilling through the divide.

SITE SURVEY

A site, about 500 m east of the divide of the ice cap, at lat. 67°14'N., long. 65°43'W., altitude 1975 m (Fig. 2) was occupied from 24 April to 14 May 1979. During this time, a continuous 20 m firn and ice core was recovered for oxygen-isotope and other analyses, a 6 m core was obtained for pollen studies, an accumulation pole array was established, and some ice depth soundings were carried out. Ice temperatures were monitored in the 20 m bore hole. During this time, some limited meteorological measurements were also recorded.

The camp was used as a reference point for an airborne radar ice thickness survey of the complete ice cap, carried out on 9 May. Results of this survey will be published separately.

Site characteristics

The site is located in the percolation zone (Paterson, 1981) and most years, except the very coldest, are characterized by moderate to high ice-layer formation. Spring and winter snow is quite heavily covered by sastrugi formed by the intense storms which occur on the ice cap.

Core retrieval and processing

Core drilling was accomplished with a standard SIPRE corer. The core was immediately placed into plastic tubing and stored in a transit case set in a snow pit covered by an igloo. The core was later air freighted to Ottawa. Apart from a short section below the base of the pit, which was poorly packaged and labelled, the core is considered good and will be referred to in the text as complete. A separate 6 m core was retrieved for pollen analysis, which has now been completed (a paper on this is in preparation by S. Short and G. Holdsworth). This study showed that significant seasonal variations in pollen rain do exist at this site but that in order to obtain a satisfactory spectrum showing clear summer-winter oscillations (i) a larger horizontal cross-sectional sample area is required and (ii) cutting of samples should be along summer-winter boundaries. This latter requirement can only be achieved after other analyses of the core have been carried out.

Pollen data could be used as an independent check on stratigraphic interpretation. More important, however, is the potential of pollen variation data (in conjunction with oxygen isotope data) for throwing light on air-mass circulation patterns.

Some core processing was carried out in the science laboratory in Frobisher Bay, the remainder was done in the Environment Canada cold rooms in Ottawa.

CORE ANALYSES

All core handling was carried out with double plastic gloves and a face mask. Samples were cut with a cleaned stainless-steel saw and then placed in "whirl-pak" plastic bags for melting.

Core stratigraphic interpretation was carried out simultaneously with sample preparation. The densities of the firn, iced firn, or ice layers were determined,

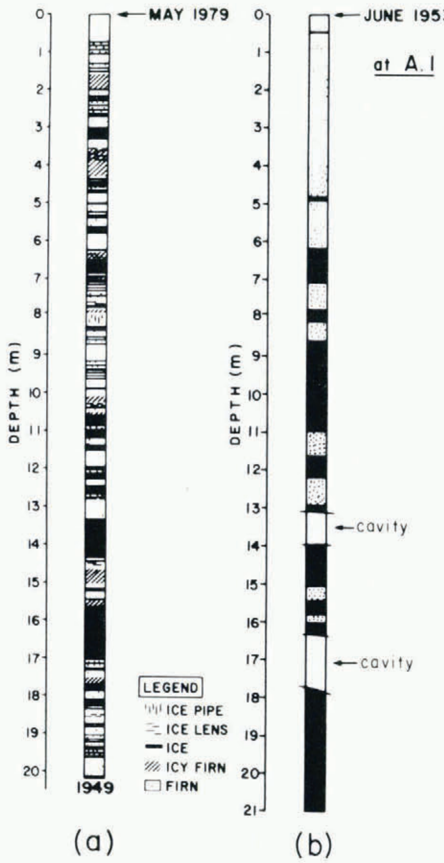

Fig.3. Visual core stratigraphy for (a) the core retrieved in 1979 and (b) the core retrieved in 1953 at site A-1. Upper blank sections are snow, lower blank sections in (b) are cavities. Note that in (b) the apparent lack of detail compared with (a) is due to interpretative differences. The net length of the 1953 core is 18.5 m.

where possible, by weighing, in air, selected samples of the core. The gross stratigraphy of the core, and that of an earlier core (Ward and Baird, 1954) are shown in Figure 3.

Analyses for oxygen isotopes were carried out on 200 samples cut at 10 cm intervals along the complete core. The measurements were done at the Geophysical Isotope Laboratory, Copenhagen, and are shown in Figure 4.

Tritium concentration measurements were made on the water remaining from the oxygen isotope sampling. These measurements were carried out at the Radiation Protection Bureau, Ottawa. The results are shown in Figure 5.

Analysis of the complete core for conductivity and pH was carried out with special attention being paid to the depth interval 9.5 m to 12.5 m which contains a large electrolytic conductivity disturbance (Fig. 6) thought initially, but erroneously, to be caused by the Mt Agung volcanic eruption of March 1963.

The pH measurements presented in Figure 7 represent two independent sets of data obtained from separately cut sequences and using different pH meters. The major trends in both pH and electrolytic conductivity are all reproducible.

Conductivity and particulate analyses were carried out on a quarter section of the core, at the laboratories of the Polar Continental Shelf Project (PCSP), Ottawa. Sample intervals were irregular as a result of separating firn from ice. Most sample lengths were close to 5 cm. Figure 8 shows that this procedure resulted in a much higher variation in concentration values compared with the values shown in Figure 6.

Some limited chemical analyses were carried out. Figure 9 shows the complete Na^+ concentration profile. The analyses were done at the Environment Canada Laboratories (ECL), Ottawa. Despite the fact that no ultra-clean procedures could be adhered to, the values for most of the core are at the same general level as those given by Busenberg and Langway (1979) for some Greenland sites, and are therefore thought to be free of gross contamination.

In order to examine the interval 10-12 m in more detail, remaining core was recut at 10 cm intervals and resubmitted to ECL for analysis for Na^+, K^+, Ca^{2+} and Mg^{2+} (Fig. 10). The levels of Na^+ are consistent with those of Figure 9.

As a final check, principal cation analyses were performed on the samples submitted to the PCSP laboratory. The results (Fig. 11) show much greater variation in adjacent values than do the corresponding data in Figure 10. This is because the sampling interval for the PCSP samples was much less than the previous arbitrary 10 cm sampling interval. Generally, the higher ionic concentrations in Figure 11 apply to the solid ice samples. This is an expected result, since salts would tend to be leached out of snow and concentrated in the refrozen melt water. There are, however, significant differences between the means of the corresponding data sets, and the reasons for this are assumed to be related to sample preparation and measurement. Compared with Figure 11, the data set in Figure 10 has lower means over the interval 10-12 m. These are for Na^+, 0.08; K^+, 0.013; Ca^{2+}, 0.053; and Mg^{2+}, 0.024 (p.p.m.).

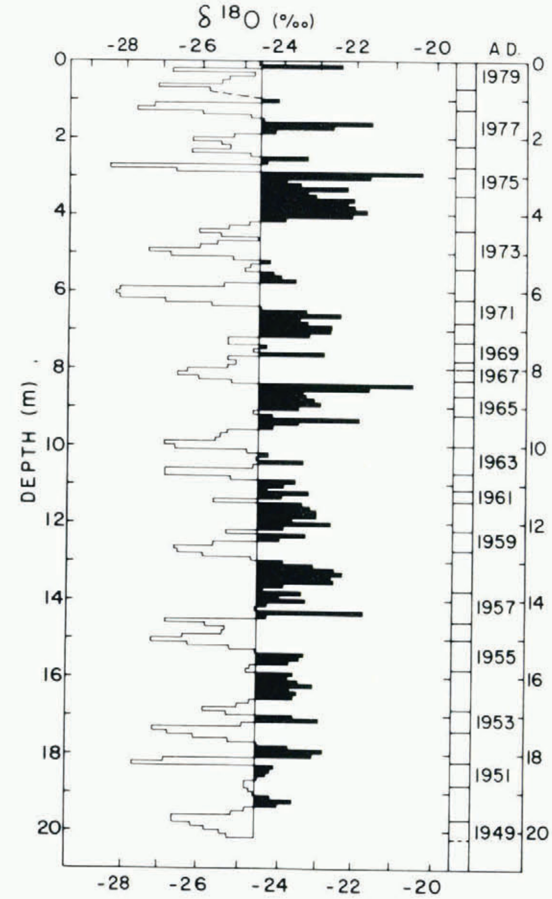

Fig.4. Depth variations of $\delta^{18}O$ with a sample interval of 10 cm. Mean δ value is -24.6 $^o/oo$. Seasonal variations in δ are locked in and have been used, in conjunction with other time control methods (tritium concentration, ion chemistry), to derive the time scale shown. Data were provided by the Geophysical Isotope Laboratory, Copenhagen (W. Dansgaard).

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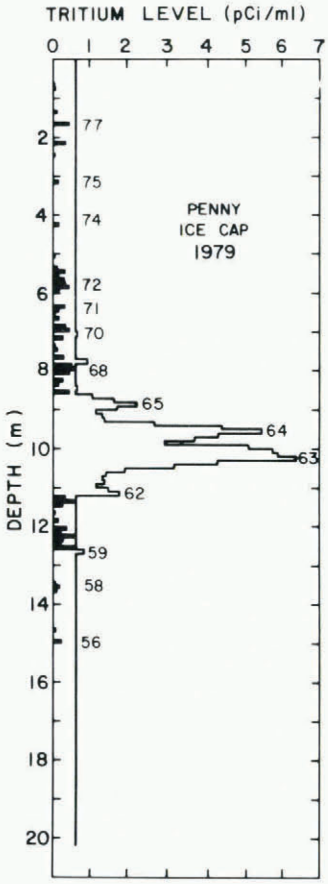


Fig.5. Depth variation of tritium concentration in the ice core, with a sample interval of 10 cm. Levels below one standard deviation (0.6 pCi/ml), marked in black, have been used, since they correspond with peaks in gross β activity variations seen in core samples from the Yukon Territory. Data were provided by the Radiation Protection Bureau, Ottawa (F. Prantl, D. Meyerhof).

DISCUSSION OF CORE ANALYSES

Isotope data

The $\delta^{18}O$ plot (Fig. 4) shows that seasonal δ variations are locked in, and that the core, when interpreted with the help of other data, covers a period of about 30 years. This time scale has been calibrated absolutely at several points, using the tritium concentration diagram (Fig. 5). This shows calibrating peaks in 1956, 1958, 1959, 1962, 1963 (corresponding to pre-moratorium atmospheric nuclear warhead tests), 1964, 1965, 1966, 1968, and 1969. Although it has only been demonstrated at a few points, it is assumed that the isotopic seasons are everywhere in phase with climatic seasons. Prior to 1956, there may be an error of ± 1 year in the time scale.

General stratigraphy

The base of the core is seen to be penetrating snow deposited in about 1949. The interesting result is that a splice may now be made with an 18.5 m core extracted 26 years previously, from a site of similar altitude, 25 km to the south-east (Location A-1, Fig. 2). The core stratigraphy from Ward and Baird (1954) is shown re-drawn in Figure 3(b). Their value for the 1952-53 net accumulation (0.43 m water-equivalent) is exactly the same as the 1949-79 average deduced from the present work.

The well-known cold years of 1972 and 1973, as well as the cool period 1949-53 are correctly reflected in the δ data (Fig. 4) and are confirmed by the stratigraphy (cooler periods are marked by less ice). Figure 3(b) shows that the first 6 m of the core is almost devoid of ice, indicating cool summers. Although Ward and Baird (1954) do not provide a chronology, it is probable that this period extends from 1945-53 (giving a mean net accumulation rate of about 0.40 m a^{-1}). The onset of heavy ice layering below 6.3 m would then logically correspond to the warmest period (1935-45) observed so far this century for most North American localities (Budyko, 1974; Mitchell, 1961).

The cavities which occurred in the 1953 core are interesting, being evidently due to vertical creep of snow bridges spanning a crevasse. During warm summers, the presence of melt water would accelerate the deformation of the bridges. Subsequent winter snow would form cornices over the old iced bridges and a cavity would result. The A-1 core, allowing for the cavities, has a reduced length of 18.5 m and probably extends back to about 1922, which is near the beginning of the general warming period this century (Budyko, 1974). It is possible that in periods much warmer than the last 30 years, the $\delta^{18}O$ record may not be so clear in

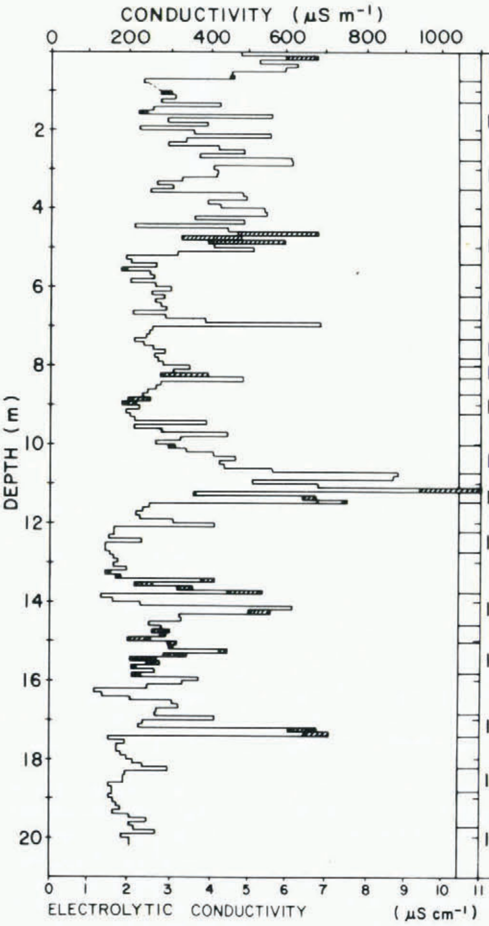


Fig.6. Electrolytic conductivity variations with depth in the core. Values were determined using a Radiometer CDM3 conductivity meter. Shaded lengths represent the major differences in values between two independent determinations made using independently cut samples at 10 cm intervals on different quarter core lengths. Figure 8(a) shows a third data set provided by an independent laboratory. Most major peaks in the well dated mid-core region appear to correspond to mid-year marks and this observation has been used in the lower core to help provide time control.

exhibiting significant seasonal variations. This would not, however, present an insurmountable problem in core interpretation. In a climatology section below, the correlation between the amount of ice in the core and the mean monthly maximum (July/August) temperatures will be discussed.

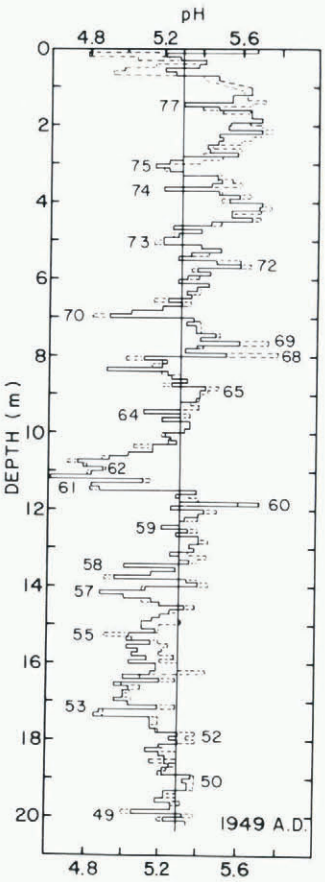


Fig.7. Depth variation of pH determined using a CANLAB H5503-1 meter (dashed lines) and an Orion Research 407A pH meter (solid lines) on two independent sampling runs. Values apply at an atmospheric pressure of 999 mbar and a temperature of 20°C. The major feature of this plot is a major acid peak at 11 m depth. This corresponds with the conductivity peak at the same depth (Fig. 6).

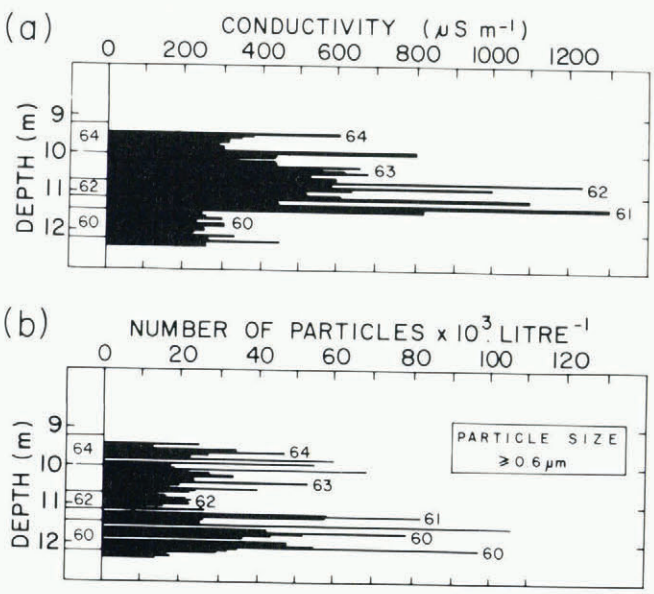


Fig.8. Conductivity (a) and particulate count (b) over the depth interval 9.5 - 12.5 m. Sample interval averages 5 cm. At this cut length double peaks per year are apparent (time control from Fig. 5) (data from Polar Continental Shelf Project Laboratory, Ottawa).

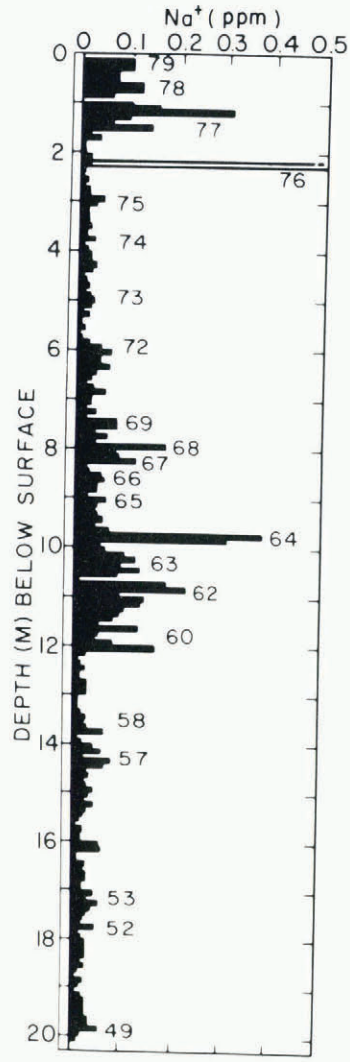


Fig.9. Concentration of Na^+ with depth. Shallow values (corresponding to pit samples) may be influenced by contamination. Values between 9.0 and 12.0 m, are significantly higher than background levels but are thought to be real. Annual peaks (at about mid annual layer) seem to occur for greater than 2/3 of the time span covered by the core. (Data: Environment Canada Laboratory: IL251 AA (flame) spectrophotometer).

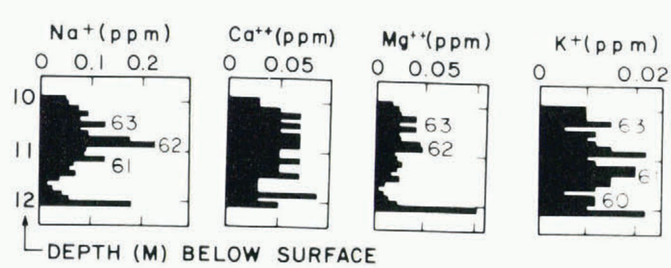
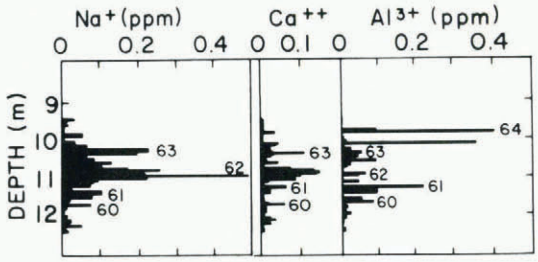


Fig.10. Concentrations of Na^+, K^+, Ca^{2+}, and Mg^{2+} at 10 cm intervals between 10 and 12 m, showing peaks, which in the case of K^+ are predominantly annual (Data: Environment Canada Laboratories).



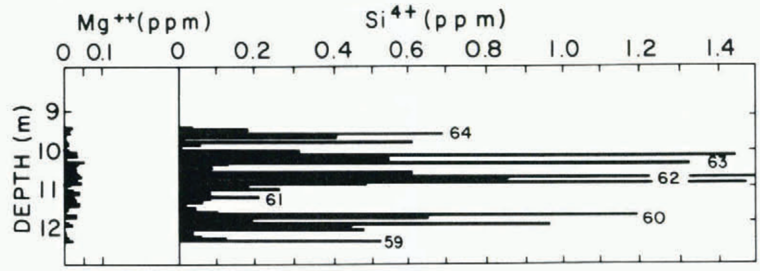


Fig.11. Concentration of Na^+, Ca^{2+}, Mg^{2+}, Al^{3+}, and Si^{4+}, between 9.5 and 12.5 m. Sample intervals are approximately 5 cm. Annual peaks are indicated using the time control established (Fig. 5). Si^{4+} data shows the existence of double peaks per annual layer. This also occurs in the particulate data (Fig. 8(b)). (Data: Polar Continental Shelf Project Laboratories: Perkin Elmer 603 graphite furnace AA spectrophotometer).

Electrolytic conductivity and pH measurements

The tritium results give an unambiguous 1963 date on ice just above the conductivity disturbance (see Fig. 6) at about 11 m depth. Any fallout from the large Mt Agung event would be expected to arrive in 1964. (There is a minor but significant conductivity peak at 9.5 m depth which corresponds to mid-1964).

A possible source of the major disturbance in both the conductivity and the pH profiles, peaking in 1961 and 1962 and continuing through 1963, is northern-hemisphere volcanism, which, according to recent records (Simkin and others, 1981), began a new active phase in the Pacific North-West in the early 1960's compared with generally low activity in the latter part of the previous decade. A similar pattern in conductivity and pH was found in a core from the Yukon Territory (G. Holdsworth, unpublished data) and these core data are correlated primarily with volcanic eruptions in continental Alaska, the Alaska Peninsula, the Aleutian Islands, Kamchatka Peninsula, and the Kurile Islands. In particular, Mt Trident (lat. 58°14'N; long. 155°07'W.), which began a new series of major sulphurous emissions in 1961 (written communication from J. Kienle; Simkin and others, 1981) and continued through the sixties, is thought to be the main source of the conductivity and pH disturbance between 11.5 m and 10 m depth in the core. The electrolytic conductivity values shown in Figure 8(a) confirm the general features of the profile in Figure 6.

Particulate content of the core

Figure 8(b) shows dust peaks in the early 1960's corresponding roughly with the conductivity peaks. Furthermore, there are *double* peaks, which occur about the middle of the year in each case. This type of distribution was found by Koerner (1977[b]) for the Devon Ice Cap core. Hammer (1977) has also found that particulates in ice cores peak at least once per year (and usually mid-year), so that annual layers may be identified by this method. Some of the work of Thompson (1977) also supports these observations.

Core chemistry

The common ion chemistry may be used to further interpret the core. From the base of the core to a depth of 2 m, the background Na^+ concentration (Fig. 9) is close to about 0.02 p.p.m. with a significant departure from this level in the interval from about 12 m to about 6 m. The highest concentrations are to be found between 12 m (1960) and 9.5 m (1964), the interval that broadly coincides with the anomalies in conductivity, pH, and particulates. Moreover, certain Na^+ peaks in 1964 (9.8 m), 1976 (2.2 m), and 1977 (1.2 m) also occur at the same dates as those in a Na^+ concentration profile for a core spanning 20 years obtained from the Agassiz Ice Field, Ellesmere Island (R.M. Koerner, unpublished data). This would seem to indicate, at least for some of the Na^+ fallout, that the two sites were receiving aerosols from the same source.

The consistently higher Na^+ levels in the upper 1.5 m could be due to contamination of the snow-pit samples, and therefore interpretation in this interval is avoided.

Annual layer identification

Seasonal or annual peaks in cation (and anion) concentrations have been recognized in firn cores (Langway and others, 1977; Busenberg and Langway, 1979) and can be used to determine the thickness of annual layers along the core. The established depth-time scale (Fig. 4) has been used to determine if seasonal (annual) variations in ionic species exist. The time annotations on Figures 8, 9, 10, and 11 show to what extent these variations can be recognized. The data obtained from the shorter sample length (approximately 5 cm) is clearly superior to those obtained from the samples with an arbitrary length of 10 cm. This indicates that for this work, at least eight samples per annual layer should be cut (Langway and others, 1977). Conductivity peaks and, to a less convincing degree, particulate concentration peaks occur approximately at the centre of an annual layer from 1960 to 1964 (Fig. 8).

Figures 9 and 10 show only weak annual signatures, with the more convincing signals being found in the K^+ concentrations. Figure 11 shows that some annual signatures exist in the Na^+, Ca^{2+}, Al^{3+}, and Si^{4+} data. The latter, however, exhibit the best signatures with almost unambiguous peaks (or double peaks) occurring from 1959 to 1964. The existence of the annual double peaks in both the particulate and the Si^{4+} concentration profiles is mutually consistent and may be useful in identifying annual layers in deeper core.

PROXY CLIMATE INFORMATION DERIVED FROM THE CORE

A functional relationship is assumed to exist between $\delta^{18}O$ values and the corresponding air temperature at the time of precipitation (Dansgaard and others, 1973). This relationship might be extended to apply to annual means of individual values of δ and temperature. The data and time scale in Figure 4 may

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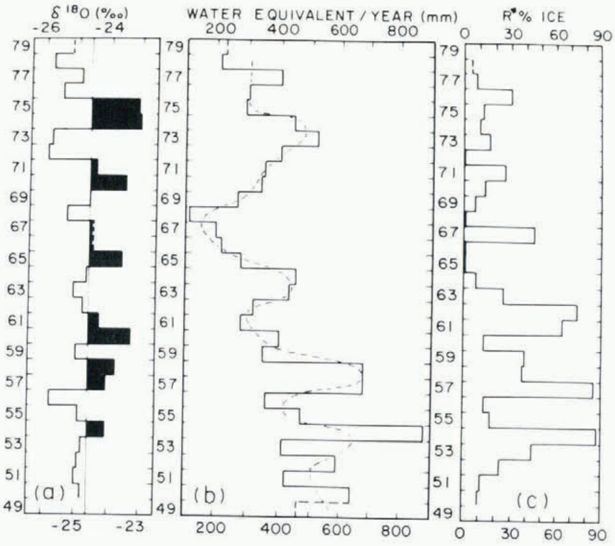


Fig.12. (a) Mean annual δ18O (unweighted) 1949-79; (b) net water-equivalent per year 1949-79; (c) percentage of melt-water ice per annual layer in the core.

be used to compute $\overline{\delta}$, the mean annual values of δ. If the precipitation pattern were roughly the same in each year (but not necessarily evenly distributed within a year) then this $\overline{\delta}$ time series might be expected to show some correlation with a mean (annual) temperature time series for the site, or in the absence of such data, with a corresponding temperature time series for the nearest climatological station.

Because the $\overline{\delta}$ values thus derived (Fig. 12a) are naturally weighted in terms of the schedule of precipitation on the ice cap, it may appear unrealistic to attempt a *direct* correlation with mean annual air temperatures at instrumental stations along the coast (Fig. 1). Therefore, the mean annual temperatures for Broughton Island and Cape Dyer were recomputed by weighting mean monthly temperatures according to the monthly precipitation. This series yielded weaker cross-correlations with the ice cap $\overline{\delta}$ time series than the unweighted temperature time series. It could be concluded that the annual precipitation schedule on the ice cap is significantly different from the precipitation schedule at the nearest coastal stations. For this reason, no weighting has been applied to any of the time series.

Figure 13 a and c shows the mean annual temperatures from 1960-79 for Cape Dyer and from 1959-79 for Broughton Island, respectively. Total annual precipitation for each station (Fig. 13 b,d) is also shown to indicate the great local variability in magnitude and time. Figure 13e shows the much longer time series for mean annual temperature for Frobisher Bay, 300 km south-west of Pangnirtung (Fig. 1). Cross-correlation coefficients between the $\overline{\delta}$ time series and the time series of mean annual temperature (Fig. 13 a,c,e) are given in Table I. Cross-correlation coefficients between instrumental stations are seen to be high whereas the coefficients corresponding to the $\overline{\delta}$ series are low, even when using five-year running mean values. However, the fact that the coldest year (1972) in the last two decades and the cool period 1949-53 can be seen in the δ profile, suggests that extremes and longer-term trends might be seen in a longer time series.

Dansgaard and others (1973) give data on $\overline{\delta}$ versus altitude for west mid-Greenland firn that indicate that the Penny Ice Cap snow (which has a three decadal mean of -24.6°/oo) is slightly less depleted in $H_2{}^{18}O$ than at the corresponding altitude (*c.* 2000 m) in Greenland. The site in Greenland most comparable to the summit of Penny Ice Cap is Dye 2 which is slightly over 2000 m in altitude and has a mean annual *firn* temperature of -16.7°C (Herron and Langway, 1980).

A mean annual *air* temperature for the Penny

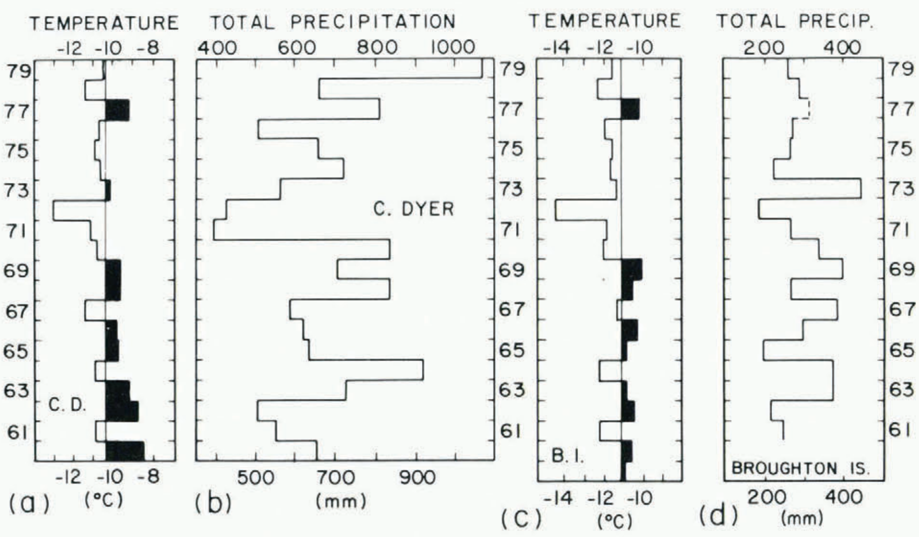
TABLE I

CROSS-CORRELATION COEFFICIENTS BETWEEN TIME SERIES

	Frobisher Bay (MAT*)	*Broughton Is.* (MAT)	*Cape Dyer* (MAT)	(L)
Ice cap ($\delta^{18}O$)	0.333 ± 0.366	0.284 ± 0.436	0.278 ± 0.448	(0)
		0.373 ± 0.486	0.368 ± 0.500	(5)
Frobisher Bay (MAT)		0.874 ± 0.436	0.829 ± 0.448	(0)
Broughton Is. (MAT)			0.899 ± 0.558	(0)

	Frobisher Bay MMDT ø	*Broughton Is.* MMDT	*Cape Dyer* (MMDT)	(L)
Ice cap (R^*_{ice})	0.273 ± 0.372	0.329 ± 0.448	0.411 ± 0.458	(0)
	0.460 ± 0.384	0.562 ± 0.472	0.631 ± 0.486	(3)
	0.747 ± 0.400	0.563 ± 0.500	0.553 ± 0.516	(5)
Frobisher Bay MMDT		0.505 ± 0.436	0.482 ± 0.448	(0)
Broughton Is. MMDT			0.746 ± 0.448	(0)

R^*_{ice} = percentage ice in core per annual layer
* MAT = Mean annual temperature; ø MMDT = Mean maximum daily (July) temperature
Cross-correlation coefficients: $R(L)$ ± 2 standard deviations.
L = length (years) of moving averages. All cases are for zero lag.
Underlined values are significantly different from zero at two standard deviations.



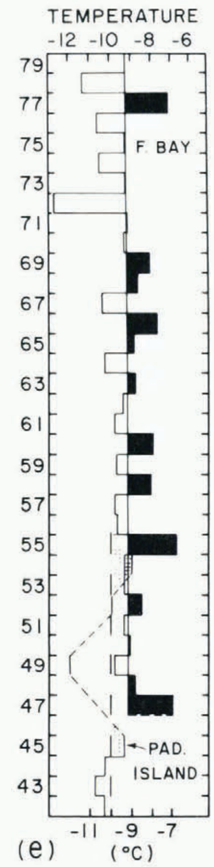


Fig.13. (a) Mean annual temperature (°C) and (b) total precipitation (mm water per year) at Cape Dyer; (c) mean annual temperature and (d) total annual precipitation at Broughton Island. (e) Mean annual temperatures for Frobisher Bay (1947-79) and Padloping Island (1942-55).

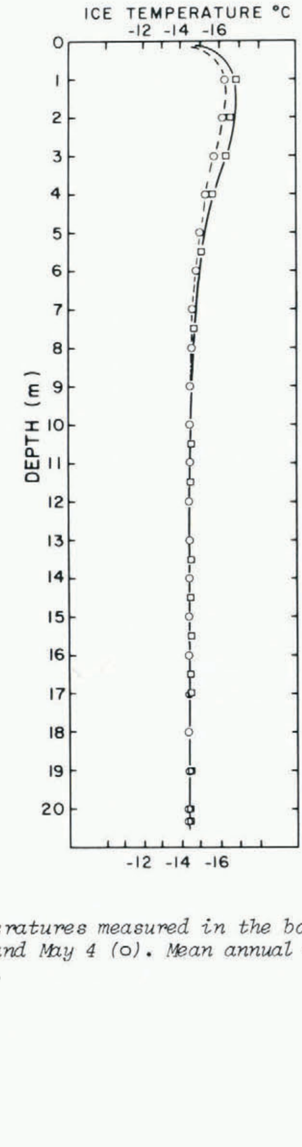


Fig.14. Temperatures measured in the bore hole on May 2 (□) and May 4 (○). Mean annual ice temperature is -14.4°C.

Ice Cap site might be estimated from the mean annual air temperature at Cape Dyer (-10.4°C) (Fig. 13a) and an estimate of the mean annual adiabatic lapse rate. Orvig (1954) gives a value of about 0.0056 deg m^{-1} for the May-August period between 400 m and 2000 m a.s.l. Winter radiosonde data for Frobisher Bay and Arctic Bay (Orvig *in* Rand Corporation, 1963) indicate the existence of strong inversions, so the mean annual lapse rate could be considerably less than and possibly half this value. Taking the upper limit, a mean annual air temperature at 1980 m could be as cold as -19.6°C or, in the lower limit, colder than -15°C.

The temperature-depth profile in the bore hole (Fig. 14) shows the mean annual ice temperature to be close to -14.4°C. Because melt water refreezes *in situ*, latent heat generated is partly retained by the snow-pack thus raising its temperature. A calculation given in Appendix A indicates that this heating of the snow-pack might be by several degrees. Weber and Andrieux (1970) report a mean annual (firn) temperature of about -13°C at 1838 m altitude. This value seems compatible with the present data.

If it is assumed that the slope of the graph of $\overline{\delta}$ versus mean annual temperature ($\overline{T}$) is close to that observed for cold dry snow (approximately 1.0‰ deg^{-1}) (Dansgaard and others, 1973), then the equation for the Penny Ice Cap would be

$$\overline{\delta} \approx \overline{T} - b$$

where b has a value between 5 and 8‰. It is still assumed that δ is the unweighted value.

Figure 12b shows the water-equivalent of the net annual increments in the core. A two-year smoothing has been applied to the data (dashed line) since a possible ± 1 year error exists in parts of the time scale. There would then be no reason to apply lags to the series before attempting to cross-correlate with other time series. It turns out that correlations between any of the precipitation time series (Figs 12b, 13 b and d) are hardly significant, although it is possible, visually, to connect some extreme events. This result is not unexpected since the series in Figure 12b represents *net* accumulation and that in Figure 13 *total* accumulation, and because of the failure to improve temperature cross-correlations using precipitation-weighted temperature data.

The net accumulation values 1969-74 have no correlation with specific net mass balance data given for nearby "Boas" Glacier (Weaver, 1975) for the same period. This indicates that mass-balance data taken from different glaciers and from different elevations in this area cannot be compared over these time spans.

Figure 12c shows the percentage (R^*) of ice in each annual increment of the core. A relationship might be expected to exist between R^* and the number of days during the summer when daily temperatures were at or above a value sufficient to cause surface melting. The parameter closest to conveying this information is the Mean Maximum Daily Temperature (MMDT) (computed for each month and published in station climatological summaries). In this case the MMDT value is usually a maximum in July. Sometimes the August value is greater, in which case this value is taken. Figure 15 shows available time series of MMDT for Frobisher Bay, Cape Dyer, and Broughton Island, which have the best correlations with R^* when data are smoothed by 3-5 year moving averages (Table I).

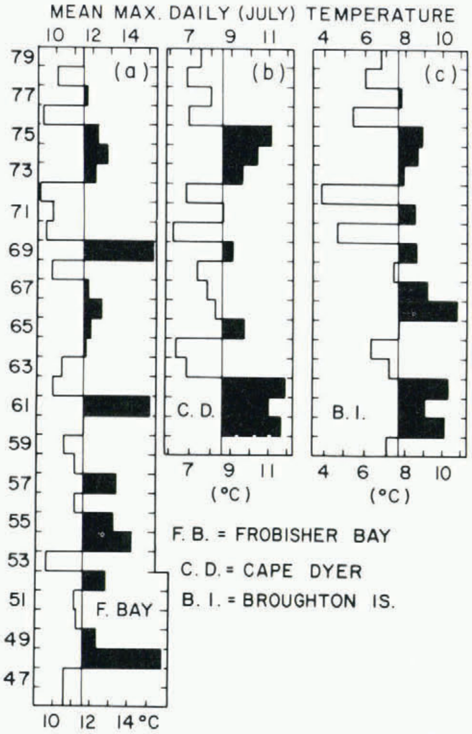


Fig.15. Mean maximum daily temperatures in July (or August, if greater) for Frobisher Bay, Cape Dyer, and Broughton Island. Values above the respective means are shown in black.

By analysing Baffin Island station records for the period 1960-69, Bradley and Miller (1972) found that generally, the "ablation" period (June-August) was becoming cooler, whereas the "accumulation" period (September-May) was becoming warmer by about the same amount (2 deg). The data in Figure 12c seem to show a response to this, in that, if averaging is made over two-year intervals, there is an indication of decreasing ice in the firn over this period. The correlations between R^* and station MMDT values are significantly higher than correlations between δ and MAT values (Table I). These latter correlations might be improved by a weighting procedure applied to the δ's based on knowing the precipitation regime on the ice cap.

TEMPERATURE PROFILE IN THE BORE HOLE AND THE ICE CAP

The temperature-depth profile (Fig. 14) shows temperature values measured by thermistors in early May, 1979. The 10-12 m firn temperature (-14.4°C) is expected to be several degrees warmer than the mean annual air temperature at the site (Appendix A).

Assuming that the present 12 m temperature is approximately representative of the long-term value, that the average accumulation rate was approximately constant over the same period, that the divide has essentially been stationary, and that the ice thickness has been roughly constant over 10^3-10^4 years, it is possible to estimate the basal temperature of the ice cap using a steady-state solution for an ice divide (Appendix B). The result is that the basal ice temperature is well below freezing (-9 to -10°C).

ICE THICKNESS MEASUREMENTS

Ice-cap thickness profiles were determined independently by surface and airborne operations in May 1979.

Surface survey

The surface measurements were made using the radar unit described by Goodman (1975). The equipment (operating at a frequency of 620 MHz) was moved by sled within a 1 km^2 grid covering the divide (Fig. 16a). The time-amplitude spectra (A-scope trace) was observed on a Tektronics 422 oscilloscope. Excellent echoes, with steep fronts in most cases, were interpreted to be from the base of the ice cap (Fig. 16b). In some cases, intermediate echo pulses, partly obscured by the transmit pulse, could be detected at depths slightly greater than half the ice thickness (Fig. 16c). These pulses varied in shape and amplitude as the antenna orientation was changed. The bottom pulse was generally sharpest when the axis of the antenna was perpendicular to the ice divide. Basal return pulse travel times were between 5.5 and 6.2 μs (± 0.1 μs). Allowing for internal electronic delays

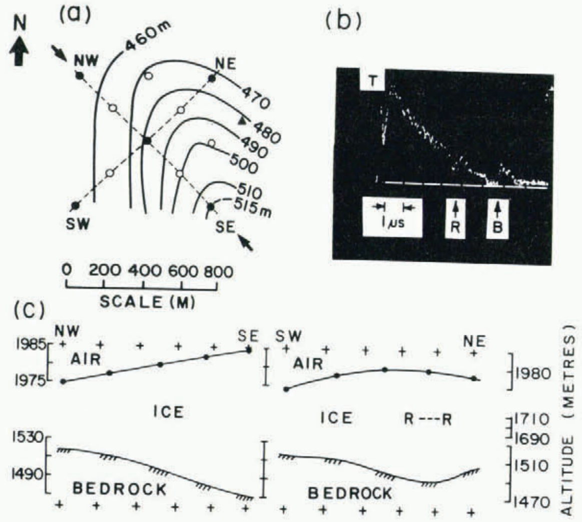


Fig.16. Radar depth results for the divide region: (a) Isopach map. The divide lies along the NW-SE line. (b) Typical A-scope trace showing transmitted pulse (T), an internal reflection (R) and base reflection (B). (c) Surface and bedrock profiles along the NW-SE and NE-SW lines. Reflector horizon R-R corresponds to a depth of about 270 m. Note different vertical scales used.

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and for the height of the antenna above the snow, and assuming an average electromagnetic wave speed of 168.5 m μs^{-1} (Weber and Andrieux, 1970; Jones, 1972), total ice thicknesses are found to lie betwen 457 and 516 ± 10 m. Figure 16a shows the ice thickness map with two profiles given in Figure 16c. Ice thickness is increasing towards the nearest dome to the south-east, a result that is confirmed by the results of the airborne survey.

Airborne survey

On 9 May a radar overflight was carried out. A C-130 aircraft, operated by the U.S. Navy VXE6 squadron in conjunction with the U.S. National Science Foundation and the Technical University of Denmark made a series of passes over the entire ice cap during a 7.5 h period. An inertial navigation system was used and the camp served as a navigation check point. Two transmitting frequences (60 and 300 MHz) were used simultaneously at a pulse width of 250 ns (giving a range resolution of 20 m). Data have been supplied by Overgaard (unpublished) although only a small percentage of the total data are reproduced here. Figure 17 shows two cross-sections produced from the photographic record of the continuous Z-scope traces for the divide area near the core site. Ice depths are seen to be in good agreement with the values determined using the surface equipment. Of particular significance are the existence of internal reflections seen at both frequencies. The reflecting horizons are not continuous nor of equal strength. Near the divide, a reflection was detected at a depth of about 270 m (Figs 16 b, c and 17). This corresponds to a date of 940 ± 50 A.D. and may be a major volcanic or climatic time horizon. Gudmandsen (1975) has described similar reflecting layers for the Greenland Ice Sheet.

A TIME-DEPTH SCALE FOR THE ICE CAP

A number of assumptions must be made in order to estimate the age of the ice at a given depth. First, since the site is similar to Dye 2 (Herron and Langway, 1980), the firn-ice transition is assumed to be at 50 ± 5 m. Secondly, the vertical component of velocity at the surface is assumed to be constant and equivalent to the long-term accumulation rate. Thirdly, a steady-state ice thickness is assumed.

As an approximation, the observed form of the variation $v(z)$ of velocity with depth for Devon Ice Cap (Paterson, 1981, p. 70) has been used with the specific constraint that at depth z = 20 m, t = 30 a. Equivalent empirical expressions for $v(z)$ for Penny Ice Cap are:

$$v(z) = 0.187 \ln(1+z) + 1.2 \quad 0 \leqslant z \leqslant 50\,\text{m}, \qquad (1)$$

$$v(z_i) = V_{is}(1 - z_i/H_i)^m \quad 50 \leqslant z \leqslant 480\,\text{m}. \qquad (2)$$

Where V_{is} is the ice equivalent vertical velocity at the ice equivalent surface of the ice cap, $z_i = z - z_0$ which is the vertical distance measured from the present surface to the ice equivalent surface, H_i is the ice equivalent thickness of the ice cap, ($H_i = H - z_0$), and m has a value of about 1.3. The age at a given depth z is obtained from

$$t = \int_0^z v(z)^{-1}\,dz. \qquad (3)$$

The upper curve (1) meets the lower curve (2) at about 50 m (the firn-ice transition) where by integration of

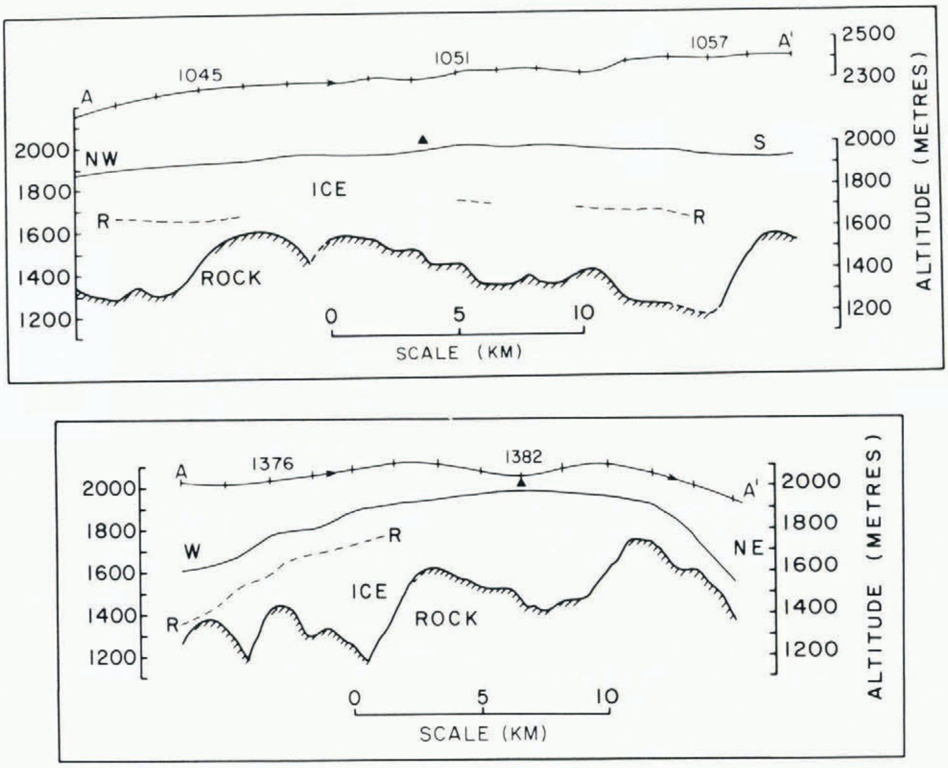

Fig.17. Snow-surface and bedrock profiles along two flight tracks approximately parallel and perpendicular to the axis of the ice cap (see Fig. 2). Reflector horizon R-R corresponds approximately to the reflector horizon shown in Figure 16. Data obtained at 60 MHz frequency. Profiles AA' represent aircraft flight tracks and numbers identify position of aircraft and radar data. The black triangle marks the position of the core site and camp.

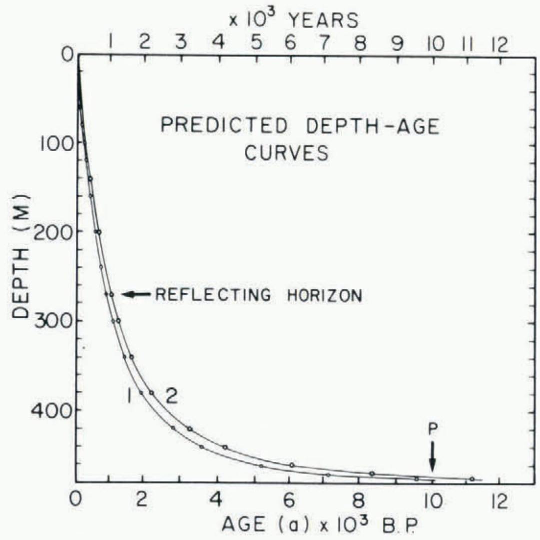

Fig.18. Predicted depth-age relationship for the divide site for a total ice thickness of 480 m. Curve 1 corresponds to a uniform vertical velocity at the surface of 0.47 m a^{-1}; curve 2 to a vertical velocity of 0.4 m a^{-1}. The reflecting horizon at 270 m depth corresponds to an age of 1036 years on curve 2. Point P marks the Holocene-Pleistocene boundary, for curve 1 at 5 m above the bed, and for curve 2 at 8 m above the bed.

(1) according to Equation (3) the age of a layer (z = 50) is about 88 years (± 10 a). Taking z_0 = 16 m, and ignoring further densification, the age of the ice in the depth interval 50 m $< z <$ 480 m (34 m $< z_i <$ 464 m) is:

$$t(z_i) = t_{z_i=34} + \int_{34}^{z_i} V_{is}^{-1}(1 - z_i/H_i)^{-m}\, dz_i. \qquad (4)$$

Figure 18 shows the predicted depth-time curves for m = 1.3 and for two values of v_{is}. The first is equivalent to the present surface net balance and the second is about 14% less. The selected value of m is thought to be very close to the one that best defines $v(z)$ for steady-state during essentially most of Holocene time. A value of $m \approx 1.25$ produces the best fit to the depth time scale for Devon Ice Cap (Paterson and others, 1977). There, Paterson (1976) found that a surface velocity v_s 13.6% less than the presently observed net balance value best fitted the derived time-depth curve.

Corresponding to the major internal reflecting horizon at a depth of 270 ± 10 m and for v_{is} = 0.40 m a^{-1}, the most likely age is 943 A.D. ± 50 a. This is seen to bracket the date of the massive Icelandic eruption of Eldgjá in 934 ± 2 A.D. which for central Greenland has an acidity signature far greater than any other event in the last 7600 years (Hammer and others, 1980). Millar (1982) refers to the recognition of a similar acid-volcanic reflecting horizon at Camp Century, Greenland, originally identified by Robin and others (1969). In addition, this depth-time curve shows that pre-Holocene ice should be encountered within 8 m above the base of the ice cap. This result is in accordance with the findings of Hooke (1976[b]) for the nearby Barnes Ice Cap. A comparison might also be made with the similarly cold-based Devon Ice Cap (Paterson and others, 1977) where pre-Holocene ice was penetrated about 5 m off the bottom. There, the total ice thickness was only 299 m compared with 480 m for Penny Ice Cap. Other, similar curves may be generated for different values of v_{is} and m.

The annual layer thickness λ is another parameter required. If λ_0 is the initial ice equivalent surface annual layer thickness, it may be shown that

$$\lambda(z) = \lambda_0 \exp\left\{\int_0^t \dot{\varepsilon}_z(z)\, dt\right\} \qquad (5)$$

in which

$$\dot{\varepsilon}_z(z) = dv(z)/dz = -\frac{m\, v_{is}}{H_i}(1 - z_i/H_i)^{m-1}$$

is obtained from the velocity curve and a value of depth corresponding to t is obtained from the depth-time curve (no. 2) (Fig. 18) or its analytical equivalent (Equation (4)). The depth at which λ is thus reduced to 0.01 m lies at about 450 m.

CONCLUSIONS

On the Penny Ice Cap divide a suitable site exists for retrieving a core which would cover the complete Holocene time period. The ice below about 450 ± 5 m depth will have annual layers compressed to under 1 cm thickness. Pleistocene ice should be encountered at from 5 to 8 m above the bed. Oxygen-isotope variations due to seasonal temperature changes appear to be locked into the ice, and by diffusion these oscillations will probably be erased in several thousand years (Johnsen, 1977).

The climate data to emerge from the core will require careful treatment. It is evident that the percentage ice (R^*) per annual layer has a significant correlation with summer temperatures as Koerner (1977[a]) found for Devon Ice Cap. This information however is dependent on successfully identifying summer melt layers in the core, a task which becomes increasingly more difficult with depth. The relationship between oxygen isotope ratio and air temperature is not simple, although certain features characteristic of the last few decades of the Arctic climate may be recognized. In order to clarify this relationship a considerable amount of data processing would seem to be necessary.

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APPENDIX A. ESTIMATE OF THE AMOUNT OF HEATING OF THE SNOW PACK BY EXPULSION OF LATENT HEAT

In order to provide an estimate of the temperature *rise* in the snow (firn), and hence to find the *difference* between the mean annual air temperature and the mean annual firn (ice) temperature, due to freezing of melt water, a simplified model is used, in which a thin, infinite plane heat source is considered to exist at the surface of a semi-infinite solid. The equation of one-dimensional heat flow in the (downward) z direction is taken to be:

$$\partial T'/\partial t = \kappa\, \partial T'^2/\partial z^2 \qquad \text{(A1)}$$

where T' is the temperature rise (or perturbation) above the ambient state without latent heat generation, t is time and κ is the thermal diffusivity of the snow.

A solution of Equation (A1) for the above model is:

$$T' = (4\pi\kappa t)^{-\frac{1}{2}}\, Q\, (\rho c)^{-1} \exp(-z^2/4\kappa t) \qquad \text{(A2)}$$

for the case of a plane heat source Q per unit area at $z = 0$. The density of the snow is ρ and its heat capacity c. In reality, heat Q is generated within the first 1 m snow depth. For modelling purposes heating shall be considered to take place within the thin layer at the surface, but initially with all the heat transfer downwards. This imposes an upper limit on the estimated temperature rise in the firn (and hence in the ice lower down). It was determined that 25% of the annual snow layer was converted into ice on average. If snow at density ρ is now converted to ice of density ρ_i then the total heat released in the refreezing of the water is approximately $(\rho_i - \rho) \times 0.25\, L$ J m^{-2} where L is the latent heat of fusion (3.34×10^5 J kg^{-1}). The heat thus released is $Q = 4.5 \times 10^7$ J m^{-2}. Equation (A2) is now evaluated at $z = 0$ for $t = 1$ a, to yield $T \approx 3.9$°C.

This process is an annual event, thus the mean annual firn (ice) temperature will be consistently warmer than the mean annual air temperature by about 4°C. If half the heat is lost vertically upwards then the temperature difference will only be about 2°C. The actual difference is likely to lie between these limits.

This estimate is consistent with the estimates of Hooke (1976[a]) for the Barnes Ice Cap divide.

APPENDIX B. STEADY-STATE TEMPERATURE DISTRIBUTION THROUGH AN ICE DIVIDE

A solution for the temperature distribution below a divide in a model ice sheet was given by Robin (1955). In the model, a constant vertical strain-rate was assumed throughout. Such a vertical strain-rate distribution is not consistent with a cold-based glacier and the computed basal temperatures tend to be colder than observed, where data are available. For thin ice caps, the discrepancy is not serious (0.6 deg for a thickness of 500 m) but for thicker ($>$1000 m depth) ice sheets the discrepancy may reach several degrees.

An expression for the vertical flow rate through the divide was obtained previously (Equation (2)). Substituting this into the original heat-flow equation (e.g. Paterson, 1981, equation (16)) the following equation is obtained after some simplification:

$$d^2T/dz_i^2 - (v_{is}/\kappa)(1 - z_i/H_i)^m\, dT/dz_i = 0 \qquad \text{(B1)}$$

Where T is temperature, v_{is} is the equivalent vertical velocity of the ice surface, κ is the thermal diffusivity of the ice, H_i is the equivalent ice thickness, and z_i is depth measured downwards from the surface.

Using the boundary conditions

$$T = T_s \text{ at } z_i = 0$$

and dT/dz_i = G = the geothermal gradient at $z_i = H_i$ and putting $a = v_{is}/\kappa$, the depth at temperature z_i is then given by:

$$T(z_i) = T_s + G\int_0^{z_i} \exp[(-aH_i/(m+1))\,(1 - z_i/H_i)^{m+1}]\, dz_i. \qquad \text{(B2)}$$

For m=1, this equation reduces to the Robin (1955) solution, and may be evaluated in terms of the error function. Numerical solutions of Equation (B2) recast in the incomplete gamma function form, for a range of values of a, H_i, and m for steps of $0.1 < z_i/H_i < 1.0$ have been obtained. Using values of v_{is} = 0.40 m a^{-1}, κ = 36 m^2 a^{-1}, m = 1.3, G = 0.019 ± 0.004 deg m^{-1}, T_s = 14.4°C ± 1 deg and H_i = 464 ± 10 m, the basal temperature is about - 9.1 ± 2°C. More sophisticated modelling is possible (Paterson and Clarke, 1978) when deeper data are obtained.

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