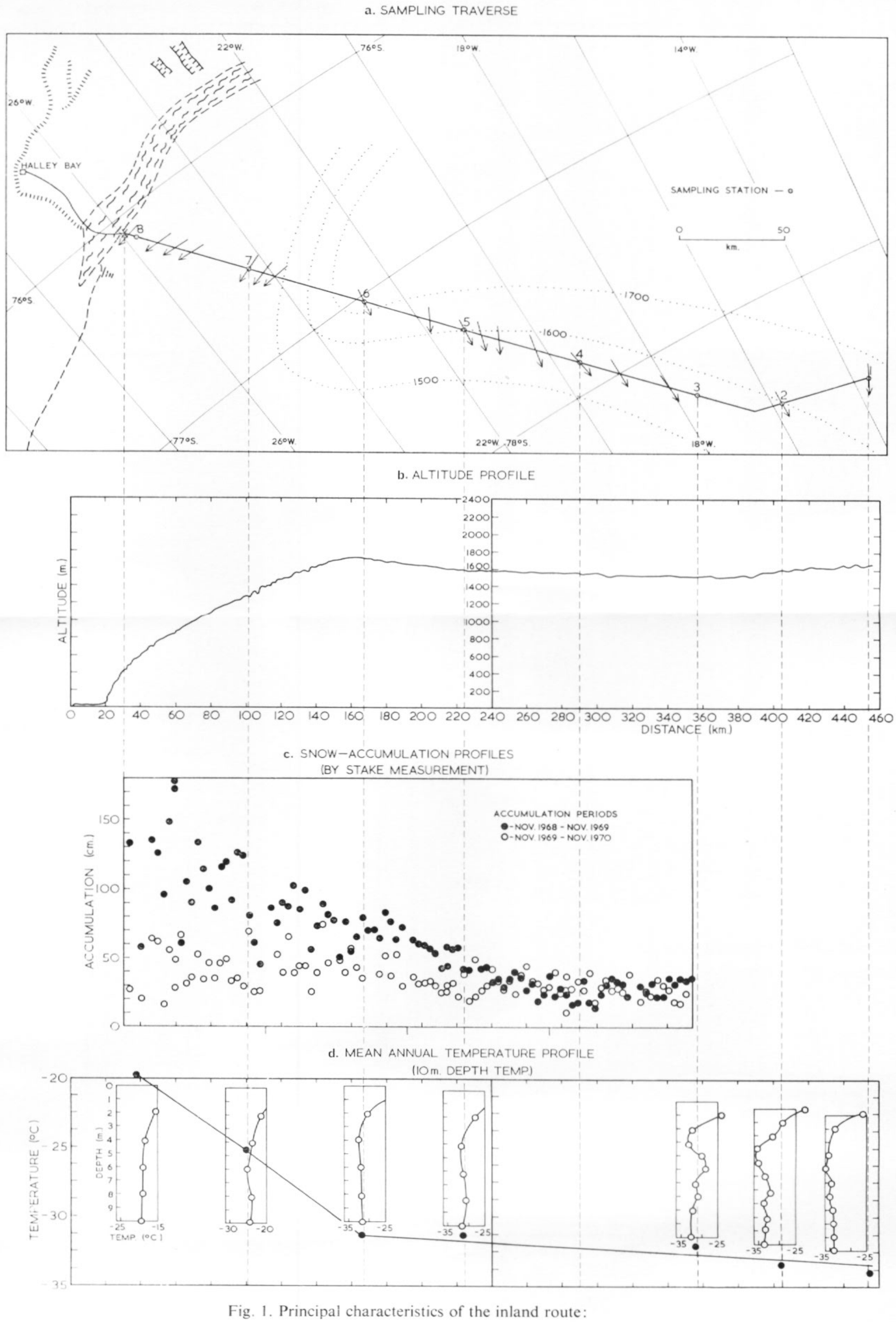

Fig. 1. Principal characteristics of the inland route:

a. Location of sampling stations and orientation of sastrugi (arrows).
b. Surface elevation.
c. Snow accumulation.
d. Snow temperature.

[face page 1]

SNOW ACCUMULATION, CONDUCTANCE AND TEMPERATURE INLAND FROM HALLEY BAY

By D. A. PEEL

ABSTRACT. Snow deposited in Antarctica incorporates a sample of the atmosphere which undergoes little change over long periods of time. Variations with place and time in the concentration of impurities in the snow may be related to the origin and mechanism of transport of these materials. Samples for geochemical analysis were collected from the upper 2 m. of the snow cover along a line stretching 450 km. inland from Halley Bay. Procedures appropriate to the subsequent analyses were used to minimize contamination during collection and transport of the samples. Accumulation profiles deduced from measurements of route-marking stakes were extended to cover the upper 2 m. of the snow cover by conventional stratigraphical methods. Closely spaced aneroid barometer measurements were used to generate a surface-elevation profile which can be correlated with snow temperatures at 10 m. depth. Further samples were collected to investigate the suitability of conductance measurements for identifying the season of deposition of snow strata. Techniques were developed for measuring the contribution of dissolved mineral salts to the conductance in the low range $1\text{–}2 \times 10^{-6}$ ohm^{-1} cm.$^{-1}$ which was found more than 80 km. inland from the coast. This component, which is dominantly of marine origin, is seasonally variable and directly comparable with the stratigraphical profiles up to 200 km. inland.

THE seasonal variation of ionic species in Antarctic snow is little understood because much recent published work (Hanappe and others, 1968; Murozumi and others, 1969; Boutron and others, 1972) has been aimed at determining areal variations in concentration rather than temporal variations; thus mean and not seasonally attributed snow was used. More comprehensive studies have been made in Greenland where Langway (1967) measured conductance and ionic (Na^+, K^+, Cl^-, $SO_4^{=}$) concentrations along a 1·3 m. section of a deep ice core from Camp Century. He found higher salt concentrations in autumn and winter snow than in summer snow, with maxima in the autumn. This trend was confirmed in a selection of samples taken from 1 year's accumulation by Murozumi and others (1969), who obtained a correlation between salt content and ^{18}O content. The only systematic attempt to measure the conductance of Antarctic snow was made by Gow (1968), who discussed the effect on conductance of both place and time (depth) in any given section. However, the time studies that he reported were made on deep ice cores and were aimed at discovering secular rather than seasonal changes. He did not find any systematic variation with time.

THE TRAVERSE ROUTE

The main inland sampling journey took place over the period 15 October to 9 December 1970. The route (Fig. 1a) followed a line laid out in 1968 towards the Shackleton Mountains as far as lat. 78°32′S., long. 16°12′W. From here it was originally planned to link up with the end point of the United States Antarctic Research Program South Pole–Queen Maud Land traverse III at lat. 78°42′S., long. 06°52′W. (Picciotto and others, 1971). However, heavy crevassing blocked the route and the traverse was terminated at lat. 78°36′S., long. 13°06′W.

The route was selected:

i. To follow a line directly inland from the coast since it was required to collect a series of samples with decreasing marine influence.
ii. To avoid exposed rock which might contribute mineral dust components to the snow and thus mask long-range material. The nearest rock exposures at any point on the traverse were Tottanfjella (250 km. to the north-west) and the Theron Mountains (300 km. to the south-south-west).
iii. To cover an area in which previous journeys were well documented so that samples could be collected from virgin territory.

The fact that stakes had been laid earlier lent confidence to the stratigraphical interpretations.

The sample station nearest the coast (station 8) was established on the inland ice sheet where it rises in a series of gentle undulations from the rifts which separate it from the Brunt Ice Shelf. Undulations persisted for the first 30 km. inland from the hinge zone and then the ice rose gradually to a second more marked series of undulations from 80 km. inland. These persisted for 50 km. reflecting considerable relief in the underlying topography. On reaching the top of the slope the surface topography was monotonously flat for over 200 km. with sastrugi rarely more than 20 cm. high. The inland end of the route was marked by a series of gentle undulations terminating in a crevassed area which probably marks the headwaters of Slessor Glacier.

Altitude Profile

Method and measurement

Three Wallace and Tiernan type FA-181 surveying altimeters were calibrated against a precision aneroid (Baromec Mark 1) at the start and finish of the journey. On the outward journey, measurements were made at 200 m. intervals for the first 90 km. and then at 600 m. intervals for the rest of the journey. Distances were routinely recorded using a standard track revolution counter fitted to the "Muskeg" vehicles and corrected by linear adjustment to fit astronomically fixed control points.

The method was similar to that described by Kennett (1965) for use where logistic limitations preclude the operation of a true "leap-frog" system. A combination of time-correlated aneroid observations in our two vehicles which normally travelled close together and occasional static observations allowed partial compensation for general atmospheric pressure variations. A closing error was established from profiles obtained on the outward and return journeys. Systematic errors arising from general atmospheric variations were more clearly shown by estimating the sea-level pressure to ± 3 mbar at selected points on the traverse using interpolated sea-level synoptic charts made available by the Meteorological Office. The altitudes at these points were estimated by assuming a mean air-column temperature calculated from the surface temperature and a dry vertical adiabatic lapse rate, resulting in two further profiles.

Results

Fig. 1b represents an average of the four primary profiles. These are in agreement to ± 15 m. up to 240 km. from the traverse reference point (Fig. 1a). Beyond this point there is a systematic divergence to give an overall spread of 70 m. at 450 km. distance. This appears to be a temperature effect. The discrepancy (<20 m.) between the synoptically corrected curves and the corresponding partially corrected curves can be accounted for by the observed difference between, on the one hand, the arithmetic mean of the Halley Bay and corresponding inland temperatures, and on the other, the mean air-column temperature calculated assuming a dry adiabatic lapse rate. The greater divergence (<70 m.) between the partially corrected profiles is thought to arise from an overestimation of the surface temperature on the return journey which was subject to more intense solar radiation.

The following error limits on the profile (Fig. 1b) are believed to be representative:

i. Between 0 and 240 km. Absolute altitude error $= \pm 20$ m.
ii. Between 240 and 450 km. Absolute altitude error $= \pm 30$ m.

Mean Annual Temperature

Where the mean annual temperature is warmer than $-35°$ C, the temperature at a depth of 10 m. in the Antarctic ice sheet lies close ($\pm 1°$ C) to the mean annual temperature at screen level (Loewe, 1970). In order to characterize mean annual air-temperature changes along the route, 10 m. holes were drilled at each of the sampling stations and temperatures were recorded at various depths using a heavily lagged mercury-iridium thermometer ($\pm 0 \cdot 1°$ C). Intermediate temperatures were recorded at half-hourly intervals until two readings agreed within $0 \cdot 5°$ C. This usually occurred within an hour of placing the thermometer in the bore hole. The 10 m. temperature was recorded after the thermometer had been in position overnight.

The resulting profiles (Fig. 1d) showed the seasonal temperature wave and indicated that it had been attenuated before reaching 10 m. depth. It appears that altitude is the dominant influence on mean annual temperature along the route, this being the normal situation on the Antarctic ice sheet (Bull, 1964). A satisfactory correlation is made by assuming a simple latitudinal effect of 2° C per degree of latitude. The mean annual air temperature then falls with increasing altitude at a rate of $0 \cdot 90°$ C per 100 m. which is close to the dry adiabatic lapse

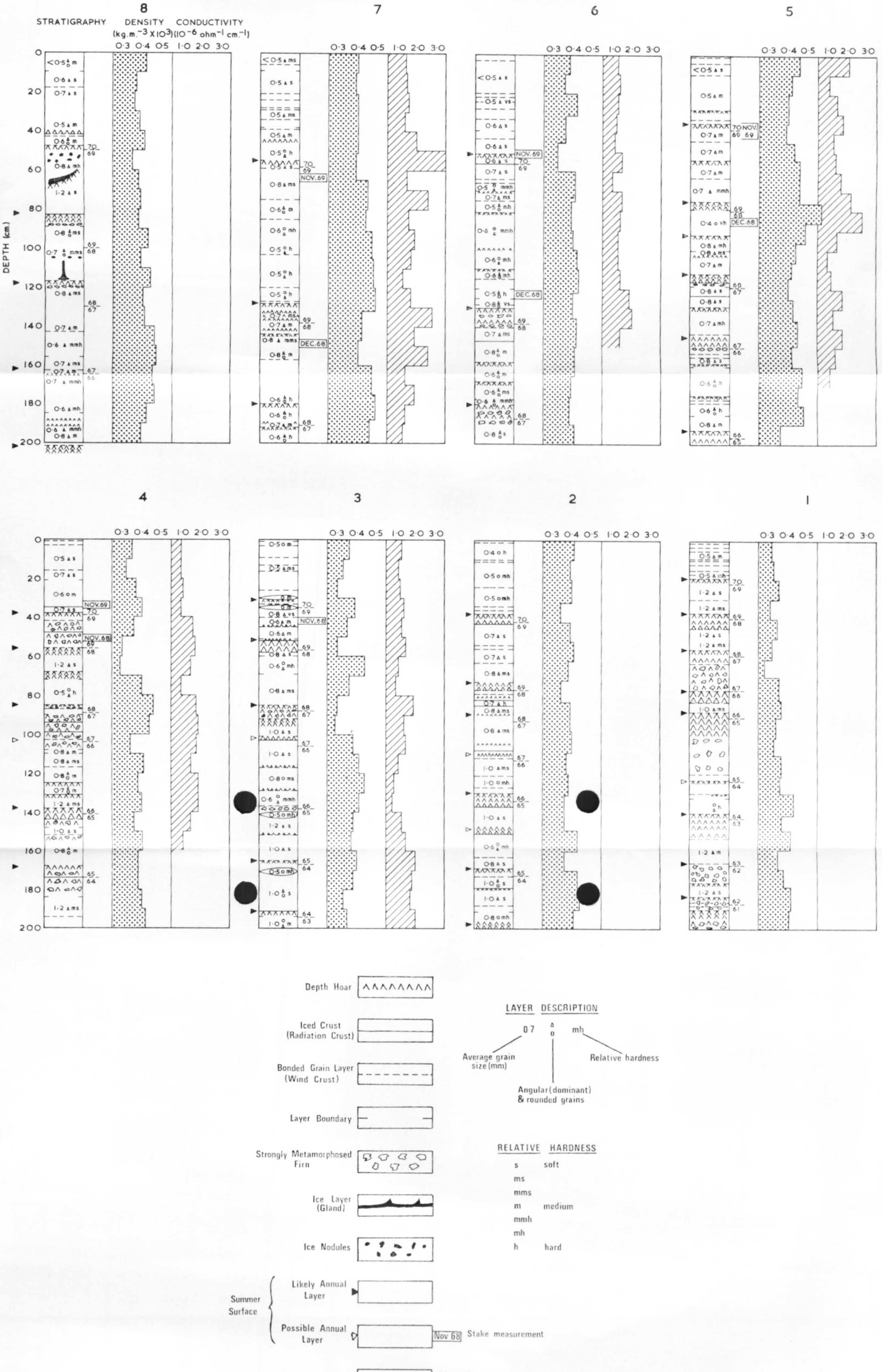

Fig. 2. Snow stratigraphy at the eight sampling stations. Conductance was measured at five stations.

[face page 3

rate. This compares with a value reported by Shimizu (1964) of 0·82° C per 100 m. for another part of west Antarctica.

Accumulation from Stake Measurements

The results of two annual cycles of stake measurements made along the first 350 km. of the route are shown in Fig. 1c. These data have not been corrected for snow settling. An examination of the snow-density increase through these two cycles at the sampling stations (Fig. 2) indicates that the correction is approximately 6 per cent per annum at station 8 decreasing to approximately 2·5 per cent per annum at station 4.

The November 1968–November 1969 data show a marked decrease in accumulation with increasing altitude and distance from the coast up to 240 km. inland, after which the amount remains nearly constant with increasing latitude. The November 1969–November 1970 data exhibit similar characteristics, although the initial decrease in accumulation is less marked, indicating less penetration by cyclonic systems during this period. Inland of 240 km. the two sets of data are similar.

The 1968–69 data show a striking decrease in the scatter of individual accumulation values from a mean value drawn along the route (62 per cent scatter in the area of station 8 decreasing to 30 per cent at station 4), although the variation in proportionate scatter in accumulation values is greater for the 1969–70 data. Correlation of the two sets of data shows that many sites with a tendency to produce high or low values compared with the mean curve show the same tendency in both profiles. This indicates that a significant proportion of the observed variability is an areal variability due to the topography of individual sites.

A detailed comparison of the accumulation profile with the altitude profile shows that for the first 60 km. of the route up the inland slope there is a close correlation between topographical features and the resultant accumulation levels. The lowest accumulation values are found on open slopes followed by values on the nearly horizontal areas of gently convex slopes. The highest values are found on concave slopes and in hollows. A similar influence of surface slopes on accumulation was noted by Swithinbank (1957). Farther inland the correlation is poor probably owing to weaker katabatic winds and to the tendency of the prevailing east-north-east winds to blow along the contours.

Snow accumulation was generally measured by means of wooden stakes, although every 10 km. aluminium stakes were used for comparison. The results obtained from the two types of stake at the same stations agreed within a standard deviation of 5 cm. snow ($\sim 2 \cdot 4$ g. cm.$^{-2}$) for the two profiles, which is a reasonable value for local areal variability (Giovinetto, 1964). This deviation was random inland of 140 km., whereas for the first 140 km. section the wooden stakes gave consistently greater accumulation values, averaging 9 cm. of snow for the 1968–69 measurements and 14 cm. of snow for the 1969–70 measurements. The general effect is probably due to katabatic winds which caused extra drift at the wooden stakes (route-marking stakes approximately 7 cm. wide) because of their greater wind resistance. The relatively large data scatter and the greater discrepancy between measurements of aluminium and wooden stakes in the 1969–70 period indicate more powerful katabatic wind activity during this period. It would be inadvisable to draw a mean accumulation profile from these two sets of observations as the temporal variability is so large, particularly over the first 240 km.

Accumulation from Snow Stratigraphy

Eight sampling stations were established along the route, at each of which a 2 m. deep pit was excavated (Fig. 3). Stratigraphical observations were made along one of the 4 m. long walls of the pit at right-angles to the long axis of dominant sastrugi at the surface. The results of these and associated density measurements, made on rectilinear snow blocks according to the method described by Schytt (1958), are presented in Fig. 2. Evidence of summer melting was observed at station 8 only. At the other stations it was necessary to use criteria developed elsewhere for regions of dry snow facies (Vickers, 1959; Koerner, 1964; Kotlyakov, 1966; Benson, 1970; Bull, 1971). The methods were shown to be suitable in the upper sections of stations 8 to 3 where a comparison could be made with neighbouring stake measurements.

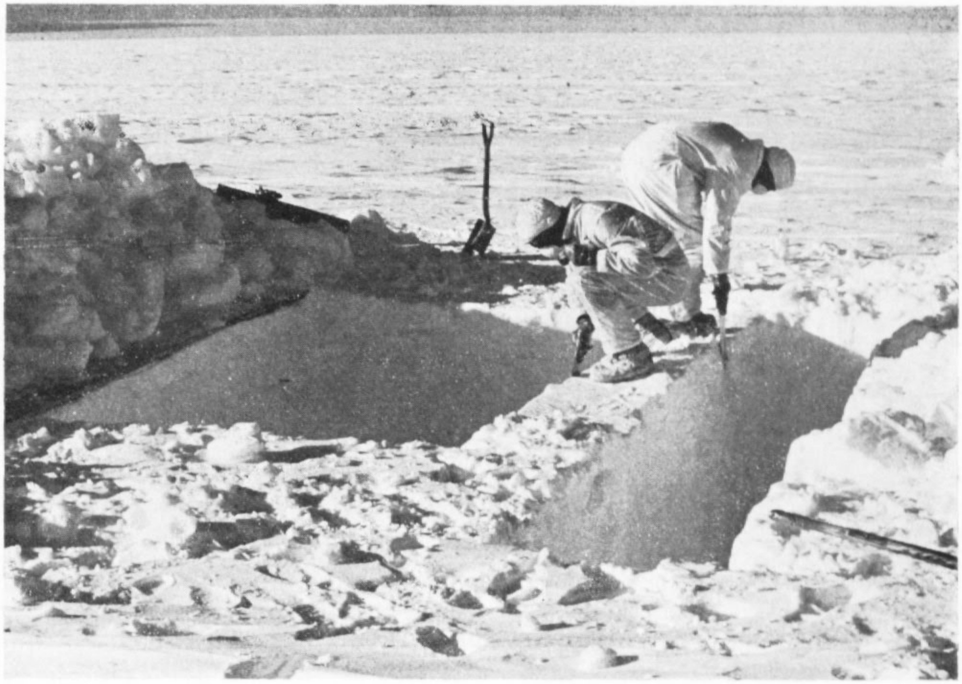

Fig. 3. Trimming the walls from which samples were taken.

The most useful indicators proved to be:

i. *Sublimation layers* (*depth hoar*). These were found to be the most consistently useful single markers at all stations. They form as a result of upward migration of water vapour through the upper layers of snow when there is a rapid lowering of surface temperature which causes large temperature gradients. This occurs most frequently at the end of summer but can sometimes occur at other seasons. A depth-hoar layer generally develops at the top of a high-porosity layer which facilitates vapour transport. However, less well developed layers are produced in the absence of summer snow and may be diagnostically equally important. A reverse temperature gradient can also cause depth hoar to form at the lower levels of a high-porosity layer (Shimizu, 1964, p. 43–44). The exact level and extent of a particular layer depends on the position of snow crusts and on the porosity and temperature gradient in the upper 50 cm. of snow. It does not, therefore, bear a fixed relationship to the summer/winter interface. The annual layers marked in Fig. 2 represent the estimated positions of the summer surfaces at which condensation of depth-hoar crystals was able to take place. On average it is to be expected that these will be separated by 12 months, although for any one cycle the period may lie between 10 and 14 months. Sublimation layers were never considered as possible annual layers unless they persisted across the whole 4 m. section. At station 8 they occurred at a level about 15 cm. above the layers exposed during the hottest part of the summer.

ii. *Radiation crusts*. Radiation crusts were observed intermittently as far inland as station 2. They are believed to form as a result of internal melting of the crystals (by a greenhouse effect) due to solar radiation penetrating vertically orientated snow crystals along their *c*-axes. In general the crusts were 1–2 m. thick and exhibited a milky appearance. Often they had a columnar structure distinguishing them from wind crusts, which are bonded grain layers of similar thickness that may form at any time of the year. Layers of intermediate appearance were classified as bonded grain layers so that undue weight would not be given to the appearance of a doubtful radiation crust. Wherever radiation crusts occurred, they gave an unequivocal indication that the level had been exposed during a summer period.

iii. *Grain-size, hardness and density*. Several studies have shown that cyclic variations in grain-size and hardness are useful criteria for distinguishing between summer and winter snow deposits. Winter snow is in general of smaller grain-size (~0·2 mm.) than summer snow (up to 1 mm.), as a result of which it tends to have a higher density and a lower porosity than summer snow (Shimizu, 1964, p. 46). Because wind velocity is generally higher in winter than in summer, winter snow tends to be more compacted and thus harder. Metamorphosis has been observed to occur mainly in the upper 1–1·5 m. of the snow cover. At greater depths the temperature gradient and porosity are too small to allow appreciable growth in grain-size. New summer snow, although relatively large-grained, is readily broken into smaller particles (~0·5 mm.) which do not grow significantly until the beginning of the winter when large temperature gradients in relatively warm snow cause dominant vapour-transport processes. Winter snow, which is fine-grained and of low porosity, does not grow significantly even when subjected to large temperature gradients as it may be in the spring season. This leads to an increasing differentiation between summer and winter snow with increasing depth in the snow pack.

In our work grain-size and hardness were found to be useful as confirmatory evidence but in general we did not find simple cycles which could be used to delineate annual boundaries. Density measurements, made for water-equivalent determinations, generally correlated with grain-size; soft, large-grained layers normally had a lower density than hard fine-grained layers.

General description of the stratigraphy

The upper section of the snow cover above the 1969–70 summer surface was in general very fine-grained and it had maintained its primary wind stratification. At stations 1 and 2, wind packing of the winter layers indicated by rounded grains (at station 2) and by closely spaced wind crusts (at station 1) yielded a hard winter snow column. The other stations, although exhibiting similar fine grain and evidence of frequent wind crusting, were much softer. It is likely that the snow structure had loosened because of the increased power of the solar radiation and increased air temperature at these stations. At stations 8, 7, 6 and 5 there was evidence of only slight increase (0·5 mm. to 0·6 mm.) in grain-size of the winter layers between the surface and 2 m. depth, the hard-packed layers showing the greatest stability. The summer layers were generally composed of grains of approximately 0·8 mm. diameter throughout the 2 m. section except in the upper parts of the exposed snow where intense metamorphosis had yielded grains of up to a few millimetres in diameter. Correlation between the accumulation measured at neighbouring stakes (up to 1 km. away) and the stratigraphical interpretation at these four stations lay within expected limits of local areal variability. There was no evidence that any complete season's accumulation had been removed during the last 5 years.

Inland of station 5, stratigraphical interpretation became more difficult and more subjective because of the lower accumulation and resultant increase in metamorphosis. For example, stake measurements at station 4 support the view that the gross development of sublimation crystals between 40 and 60 cm. depth was the result of a very small accumulation during 1969. The radiation crust at 48 cm. may represent the upper surface of the 1969 winter or even the 1968 autumn accumulation which was exposed to the 1969–70 summer. The region between 48 cm. and 42 cm. is interpreted as accumulation in late 1969 or early 1970 which was strongly metamorphosed during the autumn of 1970. A similar feature between 90 and 107 cm. at station 4 was interpreted in the same manner.

There was some evidence, at stations 4, 3 and 2, for the maintenance of hardness and grain-size relationships between summer and winter snow, although differences were smaller than those observed at stations nearer the coast. However, the fine-grained hard layers still appeared relatively stable under the conditions prevailing at these stations. In general, both the winter and summer grain-sizes were larger than those for the other stations because a given layer spent a longer period within 50 cm. of the surface and was thus subject to diurnal variations in temperature.

Station 1 was characterized by intense metamorphosis which destroyed many of the characteristics of the original snow layers. The winter-layer grains appeared to have increased in size to approximately 1·2 mm., probably because they remained within the range of diurnal temperature changes for over 2 years. Layers exposed during summer and autumn suffered intense change. The interpretation of this site must be regarded with caution.

Mean accumulation by stratigraphy

The deduced mean annual snow accumulation and derived mean annual water equivalents at each station are given in Table I. An error bar was calculated for each result by assessing the three contributions to variability at each site as shown in Table I.

i. The interpretative variability is the estimated error incurred in assigning annual boundaries to the stratigraphical profile. It assesses the probability of over- or under-estimating the number of complete annual units at a particular site but it is not concerned with the actual position of an assigned annual boundary in a given annual cycle.

ii. The temporal variability is deduced from the stratigraphical profile. It is not a true temporal variability because it includes an interpretative error from the actual placing of an annual boundary.

iii. The areal variability may be split into two terms:

   a. The local areal variability arises from the micro-relief of the surface. A value of ± 6 cm. of snow was given by the difference between stratigraphical observations and neighbouring stake measurements.

   b. A general areal variability arises from the influence of larger topographic features.

The scatter of points on the stake accumulation profile provides a measure of the combined areal variability (Table I). The component standard deviations are brought together in the total variability term which is believed to represent the standard deviation of individual values from a true mean annual accumulation. The standard error of the quoted mean accumulation is slightly less than this because the temporal variability contribution is random and is therefore lowered by approximately the square root of the number of years of observation. The random component of the areal variability is similarly reduced. The total variability is similar for all stations and averages at 47 ± 9 per cent of which 25 ± 5 per cent is the temporal variability, 30 per cent is the areal variability and 22 per cent is interpretative. However, only the temporal variability appears to be randomly distributed along the profile. The combined areal variability tends to decrease and to become less systematic with increasing distance inland, whereas interpretative errors tend to increase in this direction making the overall variability relatively constant. Even at site 1 where stratigraphical interpretation was very difficult the interpretative errors were probably no greater than the combined areal variability. Even along the extensive very flat areas the areal variability was never less than 25 per cent. It would seem that, even with new methods of accurately dating snow sections which may virtually remove the interpretative variability factor, the accuracy of mean accumulation values determined at single stations may improve by less than 50 per cent depending on the number of years that have been studied and on local topography.

For comparison with the snow-stake accumulation profile, mean annual water equivalents have been converted into snow-depth equivalents using the densities observed in pits. The general trend shows a gradual decrease in snow accumulation from the coast into the interior with most of the drop occurring in the first 300 km. This is consistent with the view that much of the precipitation is associated with the penetration of cyclonic systems from the Weddell Sea. The levelling off in accumulation rate occurs where the route lies at a virtually constant altitude. There are indications that for the last section of the route, where the altitude starts to increase again beyond 380 km., accumulation tends to fall off with increasing distance.

TABLE I. SAMPLING STATION CHARACTERISTICS

Station number	*Lat. S.*	*Long. W.*	*Distance from coast* (km.)	*Altitude* (m.)	*Mean annual temperature* (°C)	*Mean accumulation* (g. cm.$^{-2}$ yr.$^{-1}$)	*Interpretative variability* (*S*) (per cent)	*Temporal variability* (*R*) (per cent)	*Areal variability* (*S*/*R*) (per cent)	*Combined variability* (*S*/*R*) (per cent)
1	78°36″	13°14′	395	1,660	−34·2	7·6	33	24	33 *R*	52
2	78°33′	15°51′	361	1,600	−33·6	9·8	29	34	25 *R*	51
3	78°20′	16°50′	319	1,520	−32·6	10·9	29	19	25 *R*	43
4	77°53′	19°10′	256	1,550		10·9	25	32	27 *R*	49
5	77°26′	21°01′	194	1,580	−31·4	16·4	20	26	23 *R*	40
6	77°03′	22°30′	142	1,720	−31·4	24·0	17	23	16 *S* and *R*	33
7	76°34′	24°04′	86	1,280	−24·9	28·2	17	19	31 *S* and *R*	40
8	76°06′	25°31′	28	560	−19·8	16·8	10	21	60 *S* and *R*	64

S represents a systematic error, *R* a random error.

Comparison with U.S. South Pole–Queen Maud Land Traverse (SPQMLT III)

The general trends in temperature and altitude observed over the last section (station 2 to station 1) of our journey accord with those recorded for the last section of SPQMLT III. Assuming that the general surface topography may be extrapolated continuously between the two routes, then an extrapolation of our mean accumulation data to the position of the last station of SPQMLT III agrees with the value (4·5 g. cm.$^{-2}$ yr.$^{-1}$) reported for this point (Picciotto and others, 1971).

However, both a comparison of the measured mean annual air temperatures using the altitude and latitude lapse rates deduced here and extrapolation of the surface gradients at the end points of the two traverses suggest that the SPQMLT III altitudes may have been over-estimated by 300–400 m. at the end point.

Snow-sample Collection

At each of the sampling stations a 2 m. pit was excavated about 1 km. up-wind from the camp. Equipment and personnel were carried in a small vehicle to a point 0·75 km. from camp. The remaining 0·4 km. was traversed on foot to minimize possible contamination of the snow samples by vehicle exhaust fumes and other factors associated with camp life. Normal clothing was worn during the initial excavations. Two pits were dug in order to leave a free-standing wall between them about 75 cm. wide. They were orientated so that both the dividing wall and the stratigraphical wall lay at right-angles to the prevailing sastrugi. A high-contrast wall was selected for stratigraphical observations.

Nylon clean-area garments, face masks and disposable polythene gloves were then put on and the pit was cut back to its final shape using stainless steel snow saws (Fig. 3), cleaned before use by making long saw cuts in virgin snow near the pit. The main samples intended for subsequent organic, particulate and marine ion analysis were extracted from the free-standing wall which was at this stage approximately 50 cm. wide (Fig. 4). They were trimmed to a final size (30 cm. cube) on the windward, virgin side of the pit, wrapped in layers of pre-cleaned aluminium foil and sealed in metal containers. Each sample was referred to dominant stratigraphical features in the free-standing wall which had been previously cross-correlated with similar features in the opposite wall. A detailed stratigraphical analysis was carried out later. This enabled the samples to be extracted rapidly once the wall had been trimmed to its final state.

Following the stratigraphical analysis, a section of wall was trimmed and small (100 g.) duplicated samples were extracted to represent 10 cm. intervals down the face. The snow was collected in pre-cleaned polythene bottles with polythene liners. The bottles had been washed in nitric acid and aged in several changes of boiling conductivity water. After final multiple rinsings in double distilled de-ionized water, the bottles were sealed until sample collection. All the samples were returned to England in the frozen state.

Conductimetric Analysis

Introduction

Visual stratigraphy cannot normally identify snow deposited in a particular season and relies to a large extent on the annual sequence of metamorphic processes. However, annual cycles in gross marine ion content of the snow can be expected to arise from normal annual fluctuations in sea-ice coverage, which change the effective distance of a particular station from the open sea. Closely spaced conductance measurements should provide a back-up for field stratigraphy and may allow interpretation of the season of deposition of the main samples. For organic analysis investigations it was desirable to have a rapid means of assessing relative marine contributions to any given sample. Earlier work (Hanappe and others, 1968; Murozumi and others, 1969; Boutron and others, 1972) had indicated that, within 500 km. of the coast, marine ions are dominant so that a measure of gross ion content should be sufficient.

Sample treatment and measurement of conductance

The samples, still frozen in their original polythene containers, were placed in a small insulated box cooled to −15° C by means of a metal box containing solid CO_2. The insulated

Fig. 4. Handling blocks prior to trimming and packing.

box was then transferred to a glove box maintained at a positive pressure of pure (white spot) nitrogen, in a clean area. The metal cold pad was maintained at a slightly negative pressure to prevent leakage of CO_2 into the system. Sample containers were then opened in turn and where possible a core of snow was removed from the centre of the sample thereby reducing the influence of possible contamination by the polythene. The cores were transferred to pre-cleaned Pyrex tubes where they were allowed to melt. Each tube had been cleaned with acids and distilled water followed by steaming processes. It had then been transferred to the glove box and rinsed with fresh conductivity water (specific conductance $<0\cdot3\times10^{-6}$ ohm^{-1} cm.$^{-1}$). Coring was done with a similarly pre-cleaned Pyrex glass corer that was cooled by placing it in a tube suspended in liquid nitrogen. Recrystallization of some of the samples precluded the use of the corer since the samples had become too hard to penetrate. No discrepancy outside the experimental error ($\pm0\cdot2\times10^{-6}$ ohm^{-1} cm.$^{-1}$) was observed in the conductance determined from duplicated samples prepared by the two methods.

The conductance cell was designed for use in conjunction with a glove box. It was of the pipette type and had lightly platinized electrodes, a cell constant of $0\cdot125$ (determined with standard KCl solutions) and a capacity of approximately 7 ml. The inlet tube of the cell protruded into the glove box through a rubber flange, the bulb and upper sections remaining in the laboratory atmosphere. A small thermocouple strapped to the bulb measured the temperature to $\pm0\cdot1°$ C; this was maintained constant by a small fan which circulated air freely around the bulb area. A two-way tap at the top of the cell enabled it to be flushed with pure nitrogen prior to drawing a sample into the bulb using the other connection to vacuo.

Melted snow samples contain varying amounts of CO_2 ($\not> 0{\cdot}8 \times 10^{-6}$ ohm^{-1} cm.$^{-1}$) depending on recrystallization conditions and variable losses during melting. Such variations are likely to mask seasonal variations because the average specific conductance may be less than 2×10^{-6} ohm^{-1} cm.$^{-1}$ (Gow, 1968). After melting each sample, pure nitrogen pre-saturated with water was passed through a fritted glass probe into the melt water, where it was allowed to bubble for about 20 min. This period was found to be sufficient to remove carbon dioxide introduced into de-ionized water. A Wayne Kerr bridge was used to measure conductance, readings being recorded when they had remained steady for 10 min. after attaining thermal equilibrium. Conductances were measured in the region of 23° to 26° C and subsequently corrected to 25° C. Individual specific conductance measurements were accurate to $\pm 0{\cdot}05 \times 10^{-6}$ ohm^{-1} cm.$^{-1}$. The standard deviation of individual values between duplicated samples was found to be $\pm 0{\cdot}02 \times 10^{-6}$ ohm^{-1} cm.$^{-1}$. The difference does not necessarily imply contamination and is evidently partially due to real variations in each 10 cm. snow section.

Discussion of conductance results

Histograms of the measured snow conductances are plotted in Fig. 2 alongside the stratigraphical profiles. No samples were collected from station 8 because the results would have been upset by extensive percolation of summer melt water. Stations 1 and 2 were not sampled because of the low accumulation values at these points and because of their distance from the coast which was expected to make seasonal fluctuations small. Mean values for these stations will be obtained later from the main samples.

Cyclic changes in conductance are apparent in samples taken from the other stations although variations become less marked as the distance from the coast increases. The three stations nearest the coast (up to 200 km. inland) show a marked correlation between conductance and stratigraphy. Maxima are associated with late summer and autumn layers. Station 6 is less well defined, probably because it had a low accumulation in the summer periods compared with the winters. This is shown both by the stratigraphy (large areas with small grain-size and relatively hard layering) and by extensive low conductance layers. Similar trends are apparent at stations 4 and 3, although they are less marked and it is difficult to make a direct correlation with the observed stratigraphy. This is probably due to a combination of factors including:

i. Annual accumulation increments are only 2–3 times the size of the increments taken for analysis.
ii. Summer/winter ionic concentration differences are much reduced.
iii. Low accumulation makes it easy for wind "gardening" processes to mix up layers from different seasons.

The results indicate that the method is most suitable for characterizing undisturbed winter layers at these stations.

The conductances reported in Fig. 2 have been averaged at each station to give accumulation-averaged conductances ($\bar{\kappa}_A$) which are represented in Fig. 5. These have been calculated according to the equation:

$$\bar{\kappa}_A = \sum_1^N (C_n \rho_n) \Big/ \sum_1^N \rho_n,$$

where C_n is the conductance of the nth increment, ρ_n is the average density of the nth increment, and N is the total number of increments sampled in the section. These values represent true mean annual conductances within limits imposed by the number of annual cycles sampled. They agree within experimental error with the corresponding depth-averaged conductances ($\sum_1^N C_n/N$), showing that high-density layers are not associated with either high- or low-conductance layers. This leads to the conclusion that snow density is seasonally invariant in the region of the traverse.

The average values in Fig. 5 may be compared with average conductances reported by Gow (1968) from other parts of Antarctica if allowance is made for the effect of dissolved CO_2 on his results. In our work it was found that, on average, dissolved CO_2 accounted for $0{\cdot}5 \times 10^{-6}$ ohm^{-1} cm.$^{-1}$ of the conductance measured before flushing the samples with

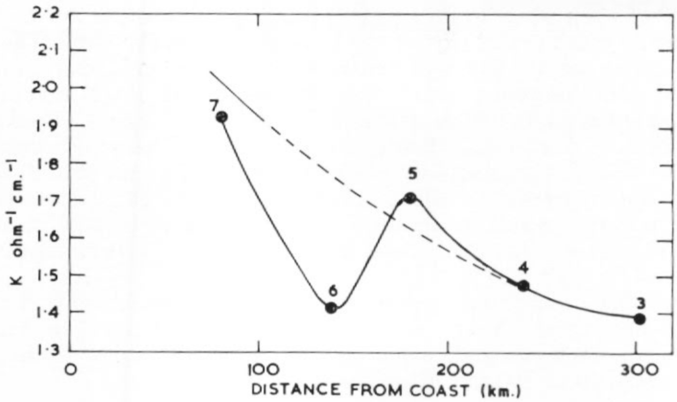

Fig. 5. Mean specific conductance at sampling stations.

nitrogen. Gow reported conductances of snow at "Byrd" station (approximately 700 km. from the nearest summer coastline) which average $1\cdot8\times10^{-6}$ ohm^{-1} cm.$^{-1}$. Subtracting $0\cdot5\times10^{-6}$ ohm^{-1} cm.$^{-1}$ for the estimated CO_2 contribution to his results yields a conductance due to dissolved salts which compares with values reported at our station 3 approximately 300 km. from the summer coastline. His values for Eights Station (approximately 300 km. from the coast) have a corrected mean conductance of $2\cdot3\times10^{-6}$ ohm^{-1} cm.$^{-1}$, higher than values obtained only 80 km. from the coast in our work. This tends to confirm Gow's suggestion that his samples might have been slightly contaminated.

The observed seasonal variation may be explained if it is assumed that a dominant component of the aerosols which contribute to the ionic content of the snow comes from sea spray. This accords with observations from other parts of Antarctica (Murozumi and others, 1969; Matveev, 1970; Boutron and others, 1972) where the observed concentration of Cl^-, assumed to be entirely of oceanic origin, bears a close relationship with other dominant ions. Our work shows the same generally concave relationship between mean gross ionic content and distance from the coast over the first 300 km. Much of the decrease in salt concentration occurs within 120 km. from the coast. Indeed similar studies from Greenland (Langway, 1967; Murozumi and others, 1969) would lead us to expect a factor of 10 decrease over this distance. Recent work more than 300 km. inland in Antarctica (Boutron and others, 1972) shows that the distribution of ions is controlled by factors other than the distance from the coast.

Within 300 km. it is to be expected that seasonal fluctuations will increase in magnitude as the coast is approached. The difference between summer and winter concentrations should approximate to the difference between the mean concentration at a given point and the mean concentration at a point farther inland by a distance equal to the width of the sea ice at its maximum extent from the coastline. The effective period during which "summer" ionic concentrations may be expected in snowfall will also decrease with increasing distance inland as a result of the rapid decline of marine influence.

In Greenland, Langway (1967) observed conductance peaks in autumn and higher conductances in winter than in summer snow. He attributed this to the more turbulent nature of the oceanic surface during autumn and winter periods. Open leads in the north and unfrozen seas in the south can continue to supply spray particles to the storm circulation during the winter season.

In the Halley Bay area of Antarctica the situation is rather different. Minimum sea-ice coverage in the Weddell Sea occurs in February when storm activity tends to be minimal. However, any cyclonic activity during this period can deliver a high concentration of ions into the interior. This probably explains the midsummer peaks at station 7. Falling temperatures in March cause ice to form in the southern part of the Weddell Sea but offshore winds leave an open shore lead. A period of maximum storm activity commences in the autumn. There are still sufficient areas of open water to furnish sea-spray particles to the storm winds and the ice is probably sufficiently thin in most parts to be broken up by ocean swell. This is evidently the period when the highest concentration of marine salts is carried inland from the Weddell Sea. A rapid temperature drop as winter approaches and a lessening of storm activity allows rapid growth of a complete sea-ice coverage extending to the coast. The ice thickens steadily and protects the sea surface from further winter storm activity. This may explain the relatively sharp cut-off in the observed conductance at the end of the autumn period. Because the only approach which cyclonic systems can make into the sampling area is across sea ice, it is to be expected that winter snow would maintain a steady low level of conductance until about November, when the sea-ice boundary once more approaches within about 300 km. of the summer coastline.

The results suggest that conductance measurements may be used in Antarctica as a convenient stratigraphical indicator up to 200 km. inland from the summer coastline. Farther inland the interpretation becomes more subjective but the method may still be useful in defining undisturbed winter layers up to 300 km. from the coast.

ACKNOWLEDGEMENTS

I wish to thank all those members of the British Antarctic Survey scientfic station at Halley Bay during the period January 1970–February 1971 who helped to make preparations for the journey. In particular, I thank fellow members of the traverse party, S. Bean, P. Burton, D. Hoy and M. Vallance for their invaluable assistance in obtaining the data reported here and in collecting the snow samples.

I should also like to record my gratitude to Dr. C. W. M. Swithinbank, who initiated this work, for many helpful discussions and for his valuable criticism of this manuscript.

The interpretation of the traverse data and the laboratory work were carried out in the Department of Chemistry, University of Bristol, and I am grateful to Professors M. Whiting and G. Eglinton for making facilities available and to members of the technical staff for their friendly co-operation.

MS. received 14 September 1973

REFERENCES

BENSON, C. S. 1970. Comparative snow studies on the Antarctic and Greenland ice sheets. (*In* GOW, A. J., KEELER, C., LANGWAY, C. C. *and* W. F. WEEKS, *ed. International Symposium on Antarctic Glaciological Exploration (ISAGE) . . . 1968 . . . .* Gentbrugge, International Association of Scientific Hydrology, etc., 242–53.) [IASH Publication No. 86.]

BOUTRON, C., ECHERIN, M. *and* C. LORIUS. 1972. Chemistry of polar snows. Estimation of rates of deposition in Antarctica. *Geochim. cosmochim. Acta*, **36**, No. 9, 1029–41.

BULL, C. B. B. 1964. Mean annual surface temperature. (*In* BENTLEY, C. R. *and others*. Physical characteristics of the Antarctic ice sheet. *Antarct. Map Folio Ser.*, Folio 2, Map. 5.)

————. 1971. Snow accumulation in Antarctica. (*In* QUAM, L. O., *ed. Research in the Antarctic*. Washington, D.C., American Association for the Advancement of Science, 367–421.) [American Association for the Advancement of Science, Publication No. 93.]

GIOVINETTO, M. B. 1964. The drainage systems of Antarctica: accumulation. (*In* MELLOR, M., *ed. Antarctic snow and ice studies*. Washington, D.C., American Geophysical Union, 127–56.) [Antarctic Research Series, Vol. 2.]

GOW, A. J. 1968. Electrolytic conductivity of snow and glacier ice from Antarctica and Greenland. *J. geophys. Res.*, **73**, No. 12, 3643–49.

HANAPPE, F., VOSTERS, M., PICCIOTTO, E. *and* S. DEUTSCH. 1968. Chimie des neiges antarctiques et taux de déposition de matière extraterrestre—deuxiême article. *Earth Planet. Sci. Lett.*, **4**, No. 6, 487–96.

KENNETT, P. 1965. Use of aneroid barometers for height determinations on the Larsen Ice Shelf. *British Antarctic Survey Bulletin*, No. 7, 77–80.

KOERNER, R. M. 1964. Firn stratigraphy studies on the Byrd–Whitmore Mountains traverse, 1962–1963. (*In* MELLOR, M., *ed. Antarctic snow and ice studies*. Washington, D.C., American Geophysical Union, 219–31.) [Antarctic Research Series, Vol. 2.]

KOTLYAKOV, V. M. 1966. *The snow cover of the Antarctic and its role in the present-day glaciation of the continent*. (U.S.S.R. Glaciology Research, Section 9 of the IGY program.) Jerusalem, Israel Program for Scientific Translations.

LANGWAY, C. C. 1967. Stratigraphic analysis of a deep ice core from Greenland. *CRREL Res. Rep.*, No. 77, 130 pp.

LOEWE, F. 1970. Screen temperatures and 10 m temperatures. *J. Glaciol.*, **9**, No. 56, 263–68.

MATVEEV, A. A. 1970. Chemical hydrology of regions of east Antarctica. *J. geophys. Res.*, **75**, No. 18, 3686–90.

MUROZUMI, M., CHOW, T. J. *and* C. PATTERSON. 1969. Chemical concentrations of pollutant lead aerosols, terrestrial dusts and sea salts in Greenland and Antarctic snow strata. *Geochim. cosmochim. Acta*, **33**, No. 10, 1247–94.

PICCIOTTO, E., CROZAZ, G. *and* W. DE BREUCK. 1971. Accumulation on the South Pole–Queen Maud Land traverse, 1964–1968. (*In* CRARY, A. P., *ed. Antarctic snow and ice studies II.* Washington, D.C., American Geophysical Union, 257–315.) [Antarctic Research Series, Vol. 16.]

SCHYTT, V. 1958. Glaciology. II. Snow studies at Maudheim. *Norw.-Br.-Swed. Antarct. Exped., Scient. Results*, **4A**, 1–63.

SHIMIZU, H. 1964. Glaciological studies in west Antarctica, 1960–1962. (*In* MELLOR, M., *ed. Antarctic snow and ice studies.* Washington, D.C., American Geophysical Union, 37–64.) [Antarctic Research Series, Vol. 2.]

SWITHINBANK, C. W. M. 1957. Glaciology. I. The regime of the ice shelf at Maudheim as shown by stake measurements. *Norw.-Br.-Swed. Antarct. Exped., Scient. Results*, **3B**, 41–75.

VICKERS, W. W. 1959. Antarctic snow stratigraphy. *Trans. Am. geophys. Un.*, **40**, No. 2, 181–84. [*IGY Bull.* No. 23.]