PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

Ice Cores and Climatic Change

G. De Q. Robin

Phil. Trans. R. Soc. Lond. B 1977 **280**, 143-168
doi: 10.1098/rstb.1977.0103

Phil. Trans. R. Soc. Lond. B. **280**, 143–168 (1977) [143]
Printed in Great Britain

Ice cores and climatic change

By G. de Q. Robin
Scott Polar Research Institute, Cambridge

[Plate 1]

The paper deals primarily with the use of stable isotopic ratios to determine the former climate of ice sheets. Studies of temperature profiles throughout ice sheets have shown that for at least several thousand years, changes of isotopic δ ratios have been proportional to changes of surface temperatures; this relationship is discussed in terms of the physical processes involved. It is considered reasonable to use a similar relation for earlier periods in Antarctica, but in Greenland the relation may have varied with time. When determining past climates from the isotopic record, allowances have to be made for changes in the flow and thickness of ice sheets during major glacial periods. These factors are considered in relation to major ice cores from Vostok and Byrd stations in Antarctica and from Camp Century in Greenland. Vostok is the simplest case glaciologically, Camp Century the most complex. On purely glaciological grounds it appears that the ice age gave way to present-day climates some 10000 ± 1000 a B.P., the coldest period being 20000 ± 3000 a B.P., when the climate in Antarctica was 6–8 °C colder than at present. Glaciological data suggest a duration of 50000 to 100000 years for the last ice age. Before this period, climates in Greenland and Antarctica appear to have been around 2–3 °C warmer than at present.

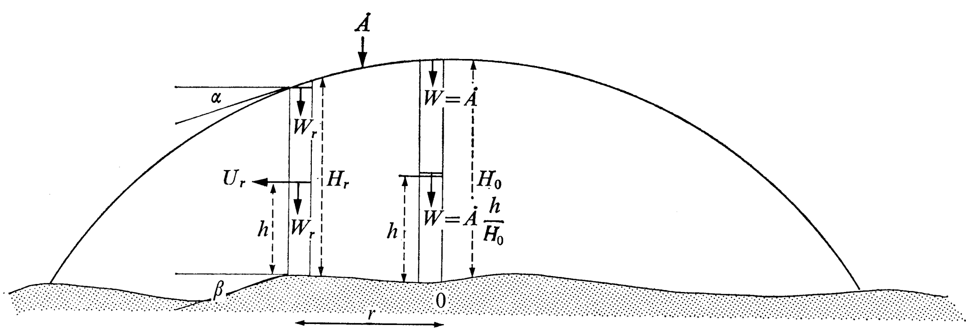
Introduction

The large polar ice sheets consist of atmospheric precipitation that has fallen over a period of some hundreds of thousands of years. In the present paper we discuss information contained in ice cores that have been recovered from Camp Century in Greenland, and from Byrd Station and Vostok in Antarctica. The data so far available covers a period up to about 100000 years ago.

The historical record present in an ice core takes a number of different forms. For example, the ice contains pockets of air, trapped in bubbles and preserved in the ice sheet, which can be recovered for analysis; the composition of the trapped air records the composition of the Earth's atmosphere at the time the bubble was caught in newly forming ice. Impurities, both soluble and insoluble, are present in small quantities of a few parts per million but are clearly distinguishable owing to the extreme purity of the body of the ice. But the most important approach, in the present context, is to study variations of the isotopes of oxygen and hydrogen with depth in the ice itself.

Perhaps the first and least ambiguous contribution from the study of ice cores is that one can measure variations of different parameters from one ice core as a function of depth. We see for instance (Gow & Williamson 1971) that the frequency of dust bands in the ice at Byrd Station increased dramatically at the same time as isotopic δ values showed a general climatic warming at the end of the last ice age. Thompson, Hamilton & Bull (1975) showed similarly that the same warming was accompanied by a decrease in the number of dust particles less than 0.65 μm in size in the ice core. Ragone, Finelli, Leung & Wolf (1972) have studied the variation of soluble impurities with depth in the same Byrd core. There is a vast amount of meticulous work to be done in this field.

For the purpose of the present study, however, we are concerned with collecting facts from various terrestrial environments in order to put together a picture of climatic changes during the Devensian period. In using records from ice cores for this purpose, two major questions need to be answered. First, how closely do isotopic δ values preserve a record of past temperature changes at the surface of the ice sheet? Secondly, a complex question: how do we allow for effects due to carriage of the ice from its point of deposition on the ice sheet to the site of the ice core, bearing in mind that the deepest ice may have been deposited hundreds of kilometres from the coring site and tens of thousands of years ago, and that during this time the size and shape of the ice sheet may have undergone considerable changes. Unless these factors can be estimated with reasonable accuracy, it is difficult to distinguish between the effects of climatic change and the effects of motion within the ice sheet when we interpret changes in isotopic δ values.



at distance r
$$\begin{cases} U_r = \dfrac{1}{H_r}\displaystyle\int_0^r \dot{A}\,\mathrm{d}r \\[2ex] W_r = (\dot{A} + (\alpha - \beta)U_r)\,h/H_r \end{cases}$$

FIGURE 1. Parameters for a two-dimensional cross-section of an ice sheet in a steady state.

MODELLING OF ICE SHEETS

In answering the above points we need to draw on our knowledge of the flow of ice sheets and its effect on temperature distribution. We find that the theory developed for a steady-state ice sheet (that is, an ice sheet of constant size and climatic conditions) gives a good first order prediction of observed temperatures in deep boreholes. A brief comparison of theory and observation will help to show the degree of confidence we may put in the models of ice-sheet flow which are an essential aid to our interpretation of isotopic profiles.

Some parameters for a two dimensional cross section of an ice sheet in a steady state are shown in figure 1. We treat a vertical column in the ice, and assume it remains vertical and undergoes uniform vertical strain throughout the column when it is deformed. At the centre of the ice sheet ($r = 0$), the steady-state assumption implies that the vertical velocity (W_0) at any height h above the bed is given by

$$W_0 = \dot{A}h/H_0 \tag{1}$$

where $\dot{A}$ is the rate of accumulation of ice on the surface and H_0 the ice thickness.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

At a distance r from the centre, the horizontal velocity of our column (U_r), under steady-state conditions will be

$$U_r = \frac{1}{H_r}\int_0^r \dot{A}\,\mathrm{d}r \tag{2}$$

and the vertical velocity (W_r) at height h above the bed will be

$$W_r = (\dot{A} + (\alpha - \beta)\,U_r)h/H_r \tag{3}$$

where α and β are the slopes of the upper and lower surfaces of the ice sheet.

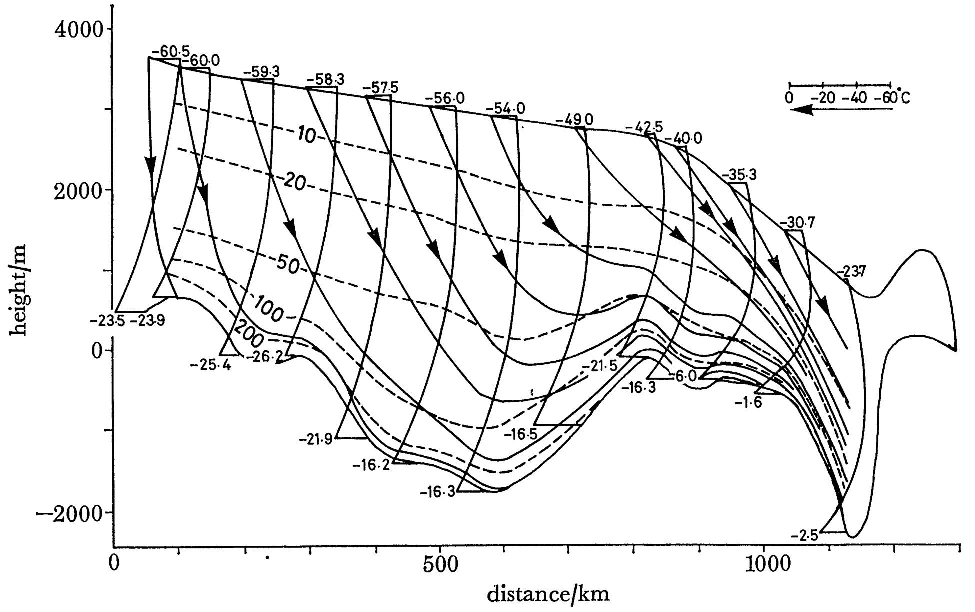


FIGURE 2. Particle paths (⟶), isochrons (10^3 a) (- - - -) and temperature–depth profiles (——) along flowline from Vostok to Wilkes Station calculated on a steady-state model by Budd *et al.* (1971 *a*).

We can match this simple model to heat-flow equations. The rate of change of temperature with time $\partial\theta/\partial t$ at any fixed point in a polar ice mass is determined by a vector equation involving the temperature of ice moving past the point with velocity V through an advection term $\rho c V\,\nabla\theta$, where ρ is the ice density and c its specific heat capacity; by the heat flux divergence at that point $\nabla\,(k\nabla\theta)$, where k is the thermal conductivity of the ice and by the rate of heat generation per unit volume Q within the ice mass. This can be written

$$\rho c\partial\theta/\partial t = \rho c V\nabla\theta + Q - \nabla(k\nabla\theta). \tag{4}$$

A detailed solution to (4), even under steady-state conditions is very complex, and it is necessary to introduce simplifications. The equation can be reduced to one dimension by confining our attention to a single vertical column travelling with the ice, and feeding in appropriate strain rates from (3), as was done in Robin (1955), in order to estimate the temperature distribution in the Greenland ice sheet. The advent of faster computing techniques has been used by Budd, Jenssen & Radok (1971 *a*) to compute temperature distributions along many flowlines in the Antarctic ice sheet. These can be seen in figure 2, which also gives information of the depth of former surface layers after a given time interval.

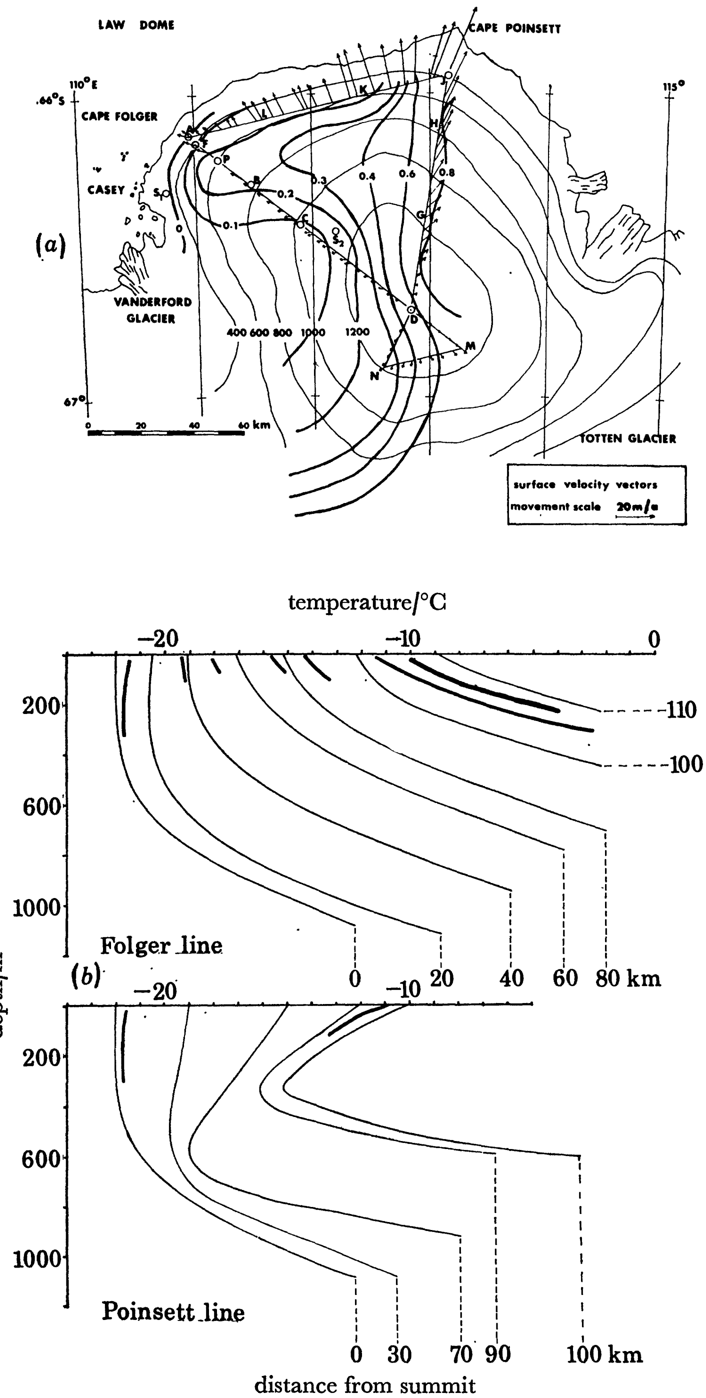


FIGURE 3. (*a*) Surface movement, surface contours and net accumulation on Law Dome, Antarctica. ——, accumulation rate, Mg m^{-2} a^{-1}; ——, elevation, m. (*b*) Calculated and observed (heavy lines) temperature profiles in the Law Dome, Antarctica. The upper figure shows data for slowly moving ice on the line from the summit to Cape Folger. The lower curves show the effect of rapid motion and heavy accumulation on the summit to Cape Poinsett line. From Budd *et al.* (1976). Reproduced from the *Journal of Glaciology* by permission of the International Glaciological Society.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

These models take into account the geothermal heat flux, frictional heat, surface accumulation rates and surface temperatures. Where accumulation rates are heavy and ice motion rapid, geothermal heat is advected away by ice motion before it reaches the surface, with the surprising result that colder ice from further inland produces temperatures that decrease with increasing depth. The effect is seen most clearly on Law Dome, Antarctica, a dome of ice some 200 km across, whose flow is largely independent of the main ice sheet (Budd, Young & Austin 1976). Heavier accumulation rates on the eastern side of the dome will, on the steady-state model, produce negative temperature gradients, but smaller accumulation on the western half results in positive gradients. The agreement between theory and observation is clearly seen in figures 3*a* and *b*. The authors (Budd *et al.* 1976) conclude that 'The temperature profiles measured in the Law Dome conform closely to the profiles calculated from steady-state under the present regime except for a slight warming in the firn zone near the surface of a few tenths of a degree. This warming could be caused by some combination of climatic warming during the last century or so, or slight lowering of the ice sheet.'

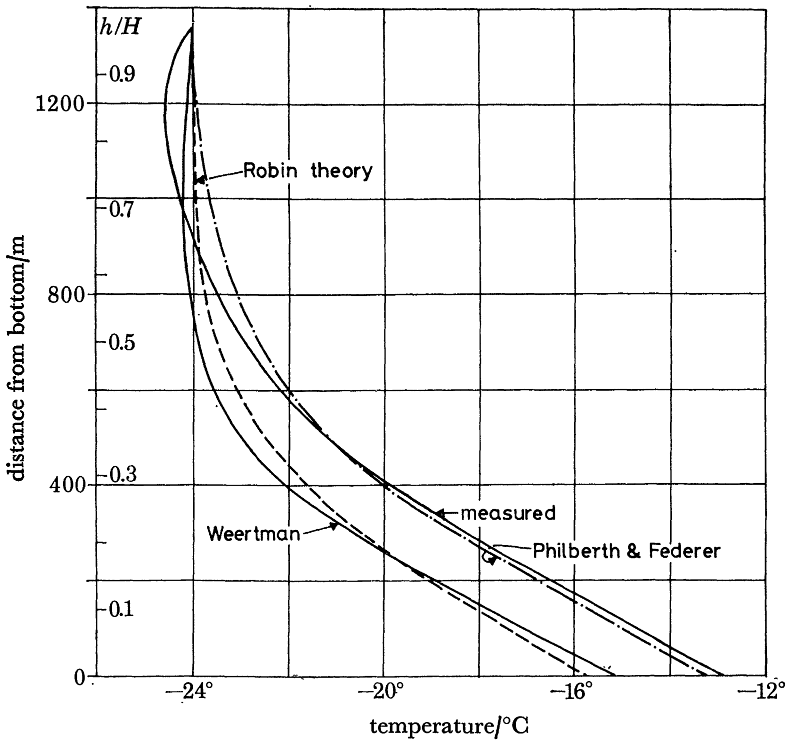


FIGURE 4. Observed and calculated temperature–depth profiles at Camp Century, Greenland (from Philberth & Federer 1971).

When we compare steady-state predictions at Camp Century with observations (figure 4), we find that agreement is limited when using a model involving uniform vertical strain rates (curves for Robin, Weertman) and is considerably improved when using a vertical strain rate that is uniform down to some 400 m or so above the base of the ice, then decreases to zero at bedrock (Philberth & Federer 1971). This type of strain-rate variation is expected when bedrock is frozen. Subsequent studies, shown in figure 5, of the variation with depth of annual layer thickness determined by the isotopic measurements of Johnsen, Dansgaard, Clausen &

Langway (1972) confirm this model of vertical strain rate, provided that accumulation rates have remained constant over the past 8300 years covered by the data.

The temperature models described so far involve input of certain boundary conditions in order to calculate temperature distribution. Budd, Jenssen & Radok (1971*b*) reversed the argument and used the observed temperature distribution, together with the steady-state and uniform vertical strain-rate assumptions to find which boundary conditions gave the best fit to the observed temperatures. The parameters they determined in this way were the accumulation rate, and the rate of change of surface temperature with time. In each of three cases, the best fit was obtained with accumulation rates in good agreement with observed accumulation rates at the present time. This implies that the present-day accumulation rates are typical of those for the last few thousand years. These derived rates of change of temperature with time are plotted in figure 6 against isotopic δ values for Camp Century and Byrd Station. We see good agreement between the trends of temperature and isotopic values over the past 2000–4000 years at these stations, even though the trends are in opposite directions at the two stations. This gives confidence that over this time period, variations of δ values are closely in step with variations in mean temperature at these two stations.

Robin (1976) and Johnsen (1977) have shown that mean temperatures and mean isotopic values have varied in step over the last few decades in Greenland. The most effective illustration of this point comes from Budd (to be published), as a result of the Cambridge Workshop on isotopic and temperature profiles in ice sheets. Using measurements of accumulation rates and temperatures along the flowline upstream of Byrd Station (Whillans 1973), they calculated the steady-state temperature profile for the Byrd Station borehole for comparison with the observed profile. Then leaving the other parameters at their steady-state values, they introduced temperatures that varied in proportion to changes in isotopic δ values. The temperature profile calculated on this basis was a considerable improvement over the steady-state model. Introduction of possible past changes of ice thickness did not improve the fit further.

The general conclusion at this stage is that variations in mean isotopic δ values are proportional to mean temperature changes at individual stations on the Greenland and Antarctic ice sheet. The period over which mean values must be taken should be not less than one year, but preferably decades or centuries.

For a study of the changes of the climate of ice sheets during the Devensian period, we must consider whether or not the relationship between isotopic δ values and temperature derived for the past few thousand years may apply on a longer time scale.

Although an extension of the type of studies already described for Byrd Station to the deeper ice of eastern Antarctica may eventually give fairly direct evidence on temperature changes in the latter part of the Devensian ice age, measurements are not available at present to derive mean surface temperatures from say 10000 to 15000 or 20000 years ago. We must therefore consider the processes involved in isotopic fractionation of water vapour, in order to see if the relationship between isotopic δ values and mean temperatures that has applied during recent millennia on polar ice sheets may also have applied during ice ages.

Figure 7 shows schematically the processes of isotopic differentiation of water vapour. Over the tropics, the δ values of precipitation are close to mean oceanic values. However, as one moves to higher latitudes, evaporation from the oceans falls due to lower temperatures, and there is a net transport of water vapour towards the poles. Since heavier isotopes are precipitated in larger amounts than lighter isotopes, there is a progressive depletion of the heavier

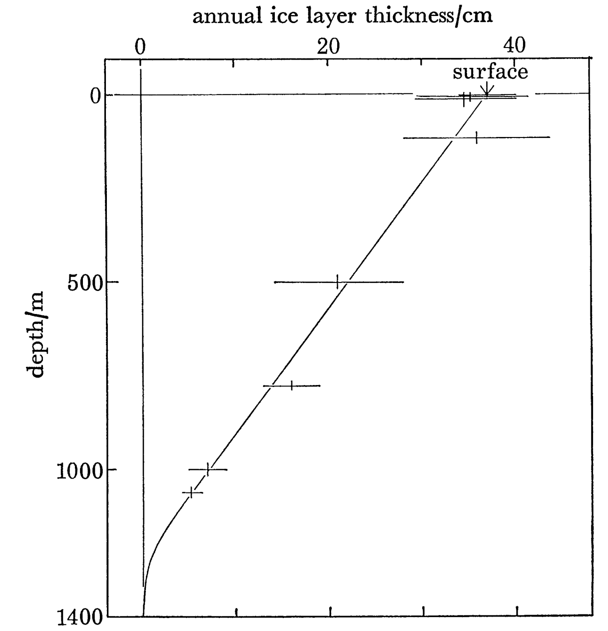


FIGURE 5. Thickness of annual layers determined from isotopic layering studies of the ice core from Camp Century plotted against depth. Data in original form from Johnsen *et al.* (1972) and Dansgaard *et al.* (1973).

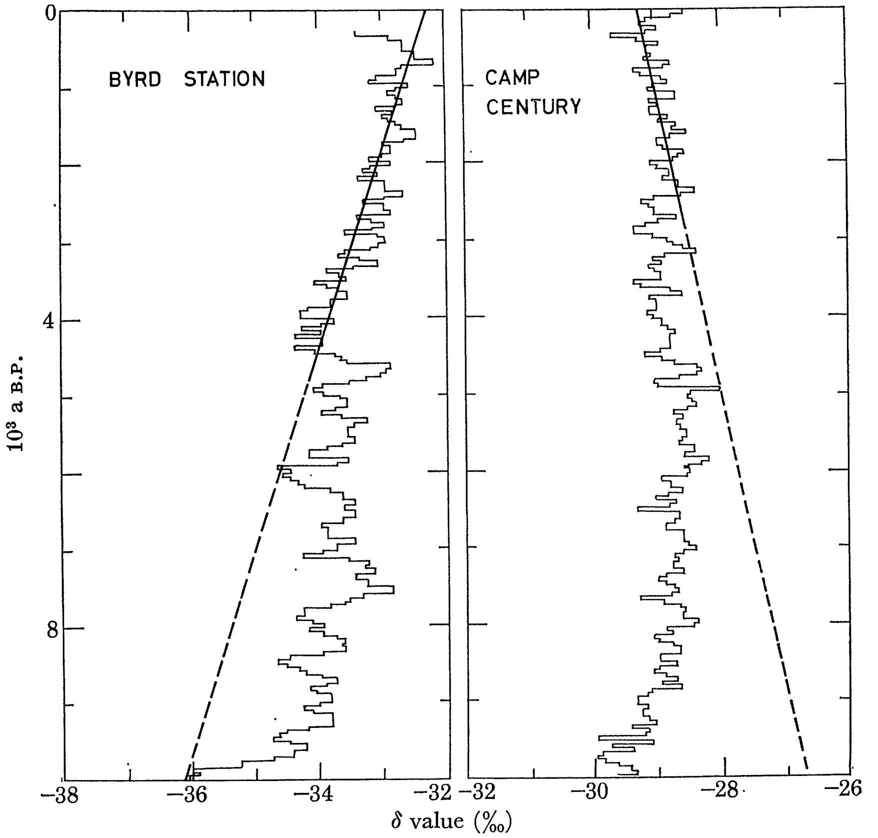


FIGURE 6. Comparison of temperature change with time (straight lines) derived from analysis of temperature–depth profiles at Byrd Station and Camp Century with $\delta^{18}O$ isotopic profiles on a time scale shown by Johnsen *et al.* (1972). From Robin (1976). Reproduced from the *Journal of Glaciology* by permission of the International Glaciological Society.

isotopes, and δ values decrease with increasing latitude, leading to mean δ values around 20 ‰ on the coast of Antarctica. Since evaporation from the surface of the ice sheet is very small owing to low temperatures, the depletion of the heavier isotope proceeds rapidly as an air mass moves over the gradually rising surface of the continent. In these conditions we can apply the equations derived for the Rayleigh condensation conditions, in which we deal with a slow process with the immediate removal of the condensate (falling snow) from the vapour after formation.

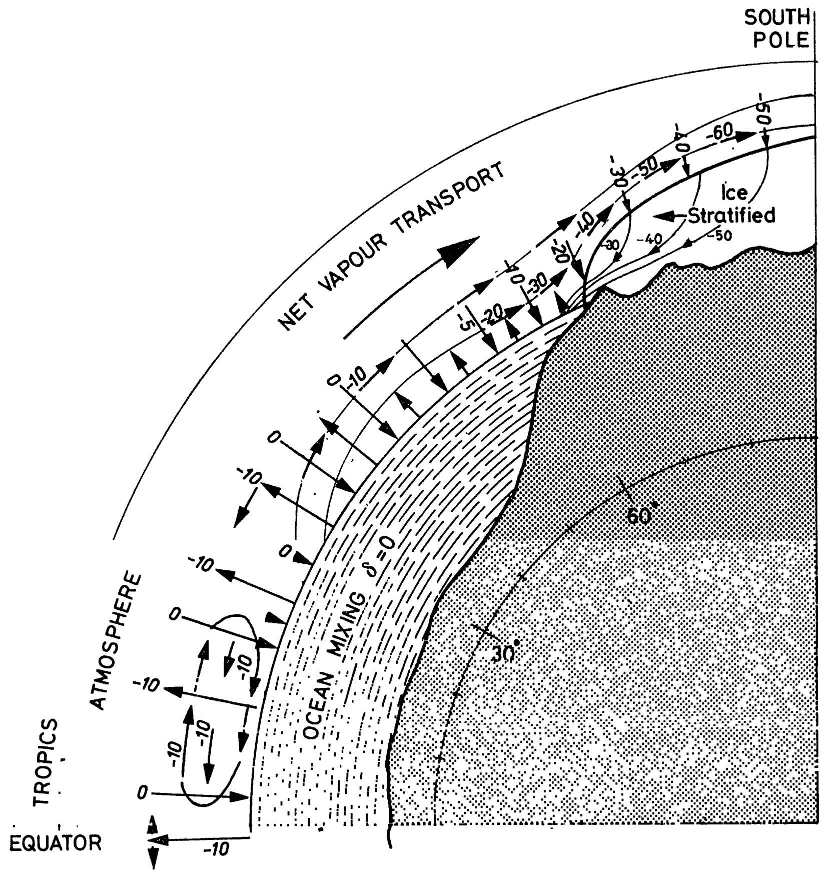


FIGURE 7. Schematic presentation of isotopic cycle of water vapour from the ocean to the Antarctic ice sheet.

Picciotto, De Maere & Friedman (1960) have studied the relationship between $\delta^{18}O$ values and temperatures of the clouds in which the snow was formed over Roi Baudouin Station on the coast of Antarctica, while Aldaz & Deutsch (1967) have studied the same process at the South Pole. Both studies extended over one year, and were used to derive a relation between cloud temperatures and mean $\delta^{18}O$ values.

The results from the South Pole are shown in figure 8, and by line A in figure 9, while line B in figure 9 shows the results from Roi Baudouin. Also shown on figure 9 are two dashed curves showing the change of isotopic composition to be expected, if the air masses over Roi Baudouin were subject to adiabatic cooling while the Rayleigh condensation process applies, as the air mass was forced upwards over the continent. The Rayleigh process gives δ_c the isotopic δ value of the condensate in terms of the remaining fraction of the vapour phase by

$$\delta_c = (\alpha/\alpha_0)\, F_v^{\alpha_m - 1} - 1, \tag{5}$$

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

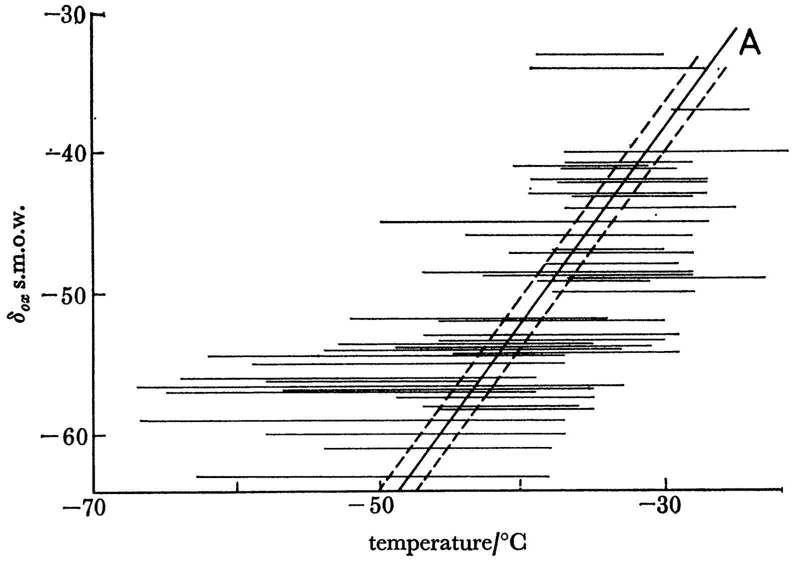


FIGURE 8. Isotopic $\delta^{18}O$ ratios of falling snow at the South Pole plotted against temperatures of clouds from which precipitation fell. Upper and lower temperature of the cloud are shown by ends of the horizontal lines. Based on Fig. 2 from Aldaz & Deutsch (1967).

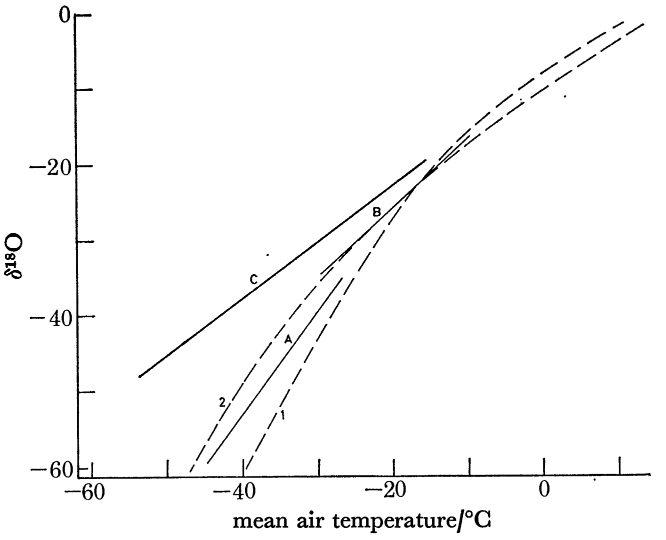


FIGURE 9. Isotopic $\delta^{18}O$–temperature relations. Curve A, clouds at South Pole, as in figure 8. Curve B, clouds at Roi Baudouin Station on the Antarctic coast, from Picciotto *et al.* (1960). Dashed curves 1 and 2, calculated for Rayleigh condensation with adiabatic cooling, based on curve B. Curve 1, α values based on Zhavoronkov *et al.* (1955). Curve 2, α values based on Merlivat & Nief (1967). Curve C, values for isotopic and temperature values for surface layers of firn of eastern Antarctica based on work by Lorius (1975).

where α, α_0 and α_m are the isotopic fractionation factors at condensation temperature t, the initial temperature t_0 and the mean temperature $\frac{1}{2}(t+t_0)$. Since α values have not been satisfactorily determined for ^{18}O at low temperatures, two curves are shown. Curve 1 uses α values obtained by extrapolation of results of Zhavoronkov, Uvarov & Sevryugova (1955) for ^{18}O fractionation at higher temperatures, while curve 2 uses α values measured for deuterium by Merlivat & Nief (1967) but converted to ^{18}O values. The agreement, particularly of observed gradients with curve 2, shows that the explanation of the observations appears satisfactory.

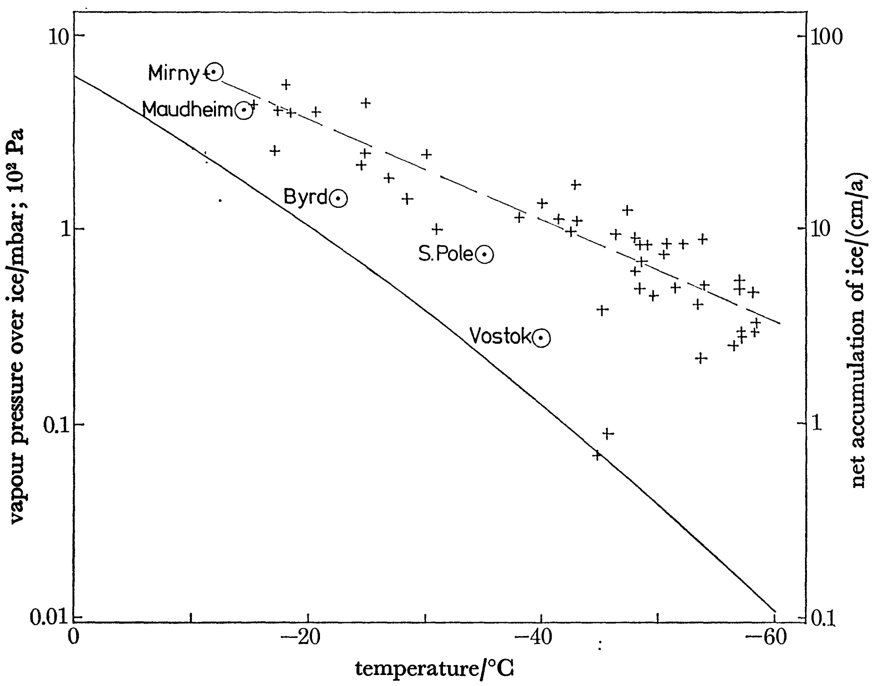


FIGURE 10. Water vapour pressure over ice (line) and annual net accumulation of ice on the Antarctic ice sheet, both plotted against temperature. Crosses show data plotted against mean annual temperature of surface layers of ice sheet. Circles show same data plotted against mean annual temperature of atmosphere above the surface inversion layer.

Also shown on figure 9 is a curve for the relation obtained by Lorius (to be published) for δD values of deuterium values of the surface layers of the ice sheet of eastern Antarctica plotted against mean ice temperatures. His δD values have been converted to δ^{18}O values by the empirical relation $\delta D = 8\delta^{18}O + 10$. The relation between δ values and surface temperatures is more nearly linear than that of cloud temperatures, which are explained by the Rayleigh process. It is, however, this process which determines the isotopic composition of the falling snow, whereas the mean surface temperature is due to the complex heat balance at the surface resulting from the thermal radiation balance and supply of heat from the atmosphere. Since clearer skies increase in frequency as one moves inland from the coast of Antarctica, the resultant surface inversion increases in strength as one moves further inland. Over coastal stations, the mean annual strength of the inversion in the lowest few hundred metres of the atmosphere runs from −0.3 °C at Mirny Station, +3.0 °C at Maudheim, +6 °C at Byrd Station, +13 °C at the South Pole and +15 °C at Vostok Station. Mean temperatures above the inversion are within 4 °C of the effective precipitation temperatures observed in cloud studies. The difference between curves 1 and 2 for cloud temperatures and curve 3 is thus readily explicable in terms of the surface inversion. The linear relation between δ values and mean surface temperatures is thus an empirical approximation to observations, rather than a natural law.

A similar feature can be seen for another parameter which will be useful later. In figure 10 we show by the continuous curve, the relation between the saturation water vapour pressure over ice and temperature. Also shown on the same curve by crosses are the mean annual accumulation at points on the Antarctic ice sheet as a function of mean surface temperature, while the circles show that mean annual accumulation as a function of the mean air temperature above the surface inversion. The trend of the latter shows clearly that the accumulation is primarily governed by the amount of water vapour that can be carried in the air mass above the inversion. The increasing separation between the crosses and circles at low temperatures is due to the increasing strength of the inversion over the cold inland regions – the same effect as we saw in figure 9.

If the whole atmospheric system falls in temperature during an ice age, the numerical relation between δ values and temperature should be similar in value to those observed at the present time. Small changes may arise due to a lower temperature of evaporation of water vapour from the ocean: alternatively, the temperature of evaporation may remain the same, if the main source moves closer to the equator. In the latter case, the amount of isobaric cooling will be greater before air masses reach the Antarctic ice sheet, and adiabatic cooling starts, due to motion over the ice sheets. In either case, once sublimation to snow commences, fractionation dominates the picture and will result in a similar numerical relationship. Dansgaard (1961, 1964) discusses the physics of the processes in detail.

Two other factors that operate in an ice age should be mentioned. The first is a change in the mean isotopic composition of ocean waters, which will increase the $\delta^{18}O$ ratio of the source of water vapour by 1 to 1.5‰. This should be added when temperature changes are estimated from $\delta^{18}O$ ratios during ice ages.

The second factor is changed atmospheric circulation. Such changes are likely to be much greater over Greenland, due to formation of large ice sheets on continents of the northern hemisphere, than over Antarctica, where the geometry of the ice sheet and surrounding areas of pack ice will not change drastically at the end of the ice age. The effect of such changes in Greenland are difficult to estimate.

Variation in size of ice sheets with time

Now that we have seen how much we may deduce about the past temperatures of ice sheets from isotopic δ values, we need to determine how much we can deduce about past climates from our past temperature–isotopic record. Qualitatively, it is broadly probable that the colder the period, the more extensive will be the ice sheets. For quantitative estimates, however, we must determine how much of the changed surface temperatures of ice sheets were due to past changes of ice thickness, and hence surface elevation of ice sheets, and how much was due to broad scale climatic changes at a fixed elevation. An extreme case is that of the highest point of the ice sheet over Canada during major glaciations, which may have been as much as 4000 m in elevation compared with a present-day surface close to sea level. However, for this discussion we need only consider the ice sheets of Greenland and Antarctica.

Surface profiles along flowlines on ice sheets have a certain regularity (Robin 1964) in spite of changes in rates of accumulation and ice temperatures between different locations. In figure 11 we use profiles from two different lines over the ice sheet of western Greenland in showing how the ice thickness along the flowline through Camp Century may have changed during the last ice age. The major effect is due to changes of the outer limit of the ice sheet. In the case of Antarctica, the outer limit is defined by the ocean, either by the line where ice shelves start to float, or by ice cliffs which do not penetrate far beyond the low water line. During ice ages, a fall in sea level of 100–150 m will enable the ice to advance to the new shoreline,

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

but around eastern Antarctica the continental shelf is not wide and such advances will normally not be much more than 100 km. The net effect on surface elevations at the South Pole and Vostok stations will be small, probably not exceeding 100 m. In western Antarctica, the continental shelf is much wider, especially in the Ross Sea and beneath the Ross Ice Shelf. Although the extent of additional grounding is still being studied, it is clear that a rise of a few hundred metres in ice thickness at Byrd Station is likely, together with some change in the flow direction of ice.

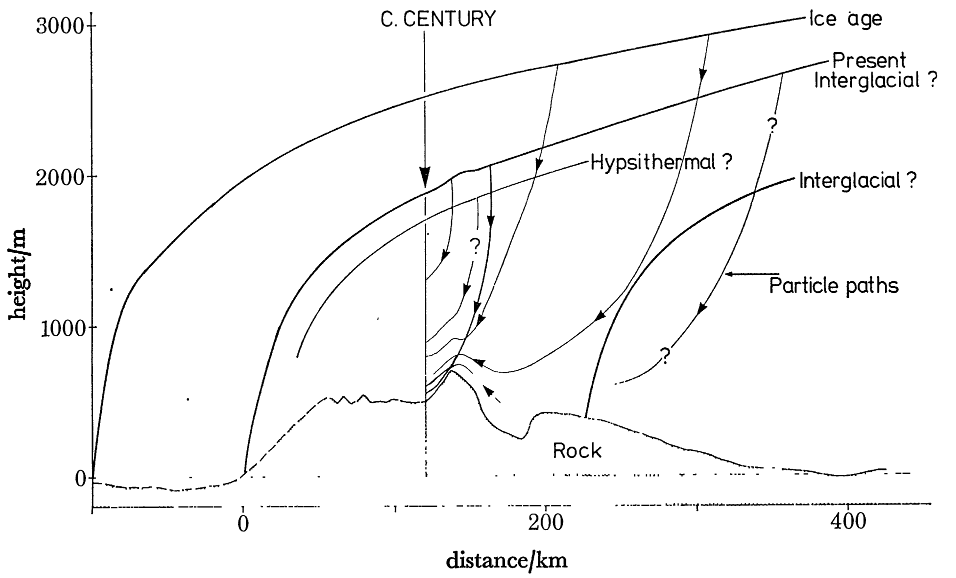


FIGURE 11. Surface and bedrock profiles along the flowline passing through Camp Century, Greenland. See text for discussion of various surface profiles. Lines with arrows show particle paths for various points in the ice core from time ice was deposited on surface to its arrival at Camp Century.

In Greenland, much of the ice now melts before reaching the sea, and effects may be larger than that due to lowering of sea level only. The greatest effect on surface elevation due to the changing size of an ice sheet takes place at the outer boundary, and the effect decreases the further one proceeds up a flowline, as seen in figure 11.

One recently proposed method of studying the elevation at which any piece of ice recovered from an ice core was originally deposited is to determine the amount of air trapped per unit volume of the sample. The volume occupied by air, when the ice ceases to be porous owing to closure of air pockets under increasing loads will be a function of atmospheric pressure and hence elevation, of the temperature at the time of trapping, and of the fractional volume occupied by air at this stage. We can write the fractional volume V of air at s.t.p. at time of closure of pores in terms of the densities of pure ice ρ_1 and of firn at the time of closure ρ_c by

$$V = \left(\frac{1}{\rho_c} - \frac{1}{\rho_1}\right)\frac{T_0 P_c}{P_0 T_c} \tag{6}$$

where T_0, P_0 refer to standard temperature and pressure and T_c, P_c refer to temperature and pressure at the time of closure. Now, if the actual volume of voids at the time of closure is a constant and is independent of temperature and other variables, and we can apply appropriate corrections for temperature from our isotopic δ values, we can determine the elevation at which the ice was deposited. A paper by Raynaud & Lorius (1973) shows relevant data for the Camp Century ice core: the authors conclude that the ice from the Wisconsin period was deposited at elevations 1200–1500 m higher than the present surface elevation at that site. Gow & Williamson (1975) in studying the ice core from Byrd Station report little systematic variation in total gas content 'except possibly for slightly reduced values between 1500 and 900 m'. These depths correspond with the culmination and termination of the Wisconsin cold phase in Antarctica. Most recently, a report of studies in Terre Adelie by Lorius (1975) gives values of V for an ice core obtained from a site at 270 m surface elevation, showing that some 'cold' isotopic ice at deeper levels had come from considerably greater elevations. Comparison of the V for shallower layers from this site with the same figures for Camp Century suggests to the present writer that while the concept is useful, the derived numerical values may be too great by a factor of up to two. A similar conclusion comes from comparison of surface profiles in figure 11 with the 1200–1500 m elevation change derived from measurements of V. These different estimates could be due to the assumption of constant density at the time of closure, as opposed to a density at closure that varies slightly with temperature and/or other factors. However, Raynaud (unpublished) and Lorius (private communication), who have carried out the studies on ice cores from Camp Century and Terre Adelie, consider that the evidence favours a constant density of closure. The problem is not fully resolved, but when understood adequately, measurement of V will provide a powerful tool for the study of ice cores.

Dynamic factors that should, if possible, be taken into account when modelling former ice sheets indicate that for a constant basal stress, ice will be slightly thicker for a greater rate of accumulation. On the other hand, lower temperatures will tend to increase ice thicknesses, if these affect the basal layers of ice. Our knowledge is insufficient to develop accurate numerical models for our present needs, and although this should be possible in the future, an empirical approach based on the form of present-day ice sheets is probably the safer method at the present time.

Age–depth relation in ice cores

We have already discussed vertical strain rates and annual layer thicknesses in relation to temperature studies. Provided a continuous ice core is available, and the annual layers can be satisfactorily determined by isotopic methods, annual layers can be counted just as accurately as tree rings or varved clays. This has been done for a period of 740 years at Dye 3 in central Greenland by Dansgaard, Johnsen, Clausen & Gundestrup (1973), and should be a satisfactory technique for up to 10000 years and longer over many parts of Greenland. Barkov, Gordiyenko, Korotkevich & Kotlyakov (1975) have used a similar technique to a depth of 507 m on the Vostok ice core, although their claim to resolve annual layering seems inconsistent with the views of Johnsen *et al.* (1972) on the disappearance of thin isotopic layers due to diffusion. Oeschger, Stauffer, Bucher & Loosli (1977) are developing methods of down-hole extraction of CO_2 from boreholes in polar ice, in order to use ^{14}C dating to provide a time scale, but no results are available at present for the Devensian period, since there are considerable difficulties in working in ice at greater depths. The oldest age determined at Byrd Station is close to 3000 years for ice at 385 m depth.

When dealing with layers over 3000 years old, we are primarily dependent on glacial knowledge of accumulation and deformation rates to date the ice core. If the simple vertical column model is used, the present layer thickness (l) is related to its thickness at the time of deposition (l_0), by Nye's (1963) formula

$$l/l_0 = h/H_0, \tag{5}$$

where h is the present height of the layer above bedrock, and H_0 the ice thickness at the point of deposition. l_0 will normally be the mean annual accumulation at the site of deposition. Integration by numerical or other means will then give the age to any depth. Dansgaard & Johnsen (1969) and Philberth & Federer (1971) have modified equation (5) to take account of the decreasing strain rates at lowest levels shown in figure 5. Dansgaard, Johnsen, Clausen & Langway Jr. (1971) also used preliminary time scales derived by these methods to find periodicities in δ values, which they then used to date still deeper layers in the Camp Century ice cores. However, further evidence is needed before accepting these time scales, and three factors should be considered.

The first is the likelihood of decreased rates of accumulation during glacial periods, due to the inability of the atmosphere to transport water vapour. We see from figure 10 that a general fall of about 7 °C in world temperature is likely to halve the rate of accumulation on the Antarctic ice sheet. While such factors are usually appreciated, they have not been applied in calculating age–depth relations. Nor have ice depths at the points of deposition been used in many cases, since the information is often not available.

The second factor is to determine whether our models of ice flow are satisfactory in various circumstances. Radio-echo methods of sounding polar ice sheets have shown the presence of internally reflecting layers in the ice mass, and in Robin, Evans & Bailey (1969) one such layer, traced over 40 km, was shown to correspond to the surface layer about 1000 years ago. More recently, Whillans (1976) has compared his studies of surface strain and movement along the flowline upstream of Byrd Station with radio-echo soundings along that line. His calculated depths of surfaces of 2500, 5500 and 30 000 years of age correspond quite closely to internal layers producing reflexions. Since it is difficult to produce any other hypothesis than that of some depositional effect to explain these layers (see Paren & Robin 1975), we appear justified in regarding these as relic surfaces, or 'isochrons' as shown in our figure 2 from Budd *et al.* (1971*a*).

Improved equipment designed in the Technical University of Denmark, and the use of differentiation of the received signal (Gudmandsen 1975), show great detail in the latest soundings obtained by the joint SPRI-TUD-NSF programme in Antarctica (figures 12 and 13). Figure 12 which is along a flowline about 130 km from Vostok Station, shows that the height of an internal layer above bedrock can vary considerably in the deeper layers, although in the upper part of the ice sheet, layering tends to be parallel to the surface. In the vicinity of Dome C (75° S 123° E) where bedrock is generally level, layering is almost parallel to upper and lower surfaces over at least 80 % of the ice depth (figure 13). The radio-echo technique thus offers clear advantages in selecting sites for any drilling in future. It also provides evidence

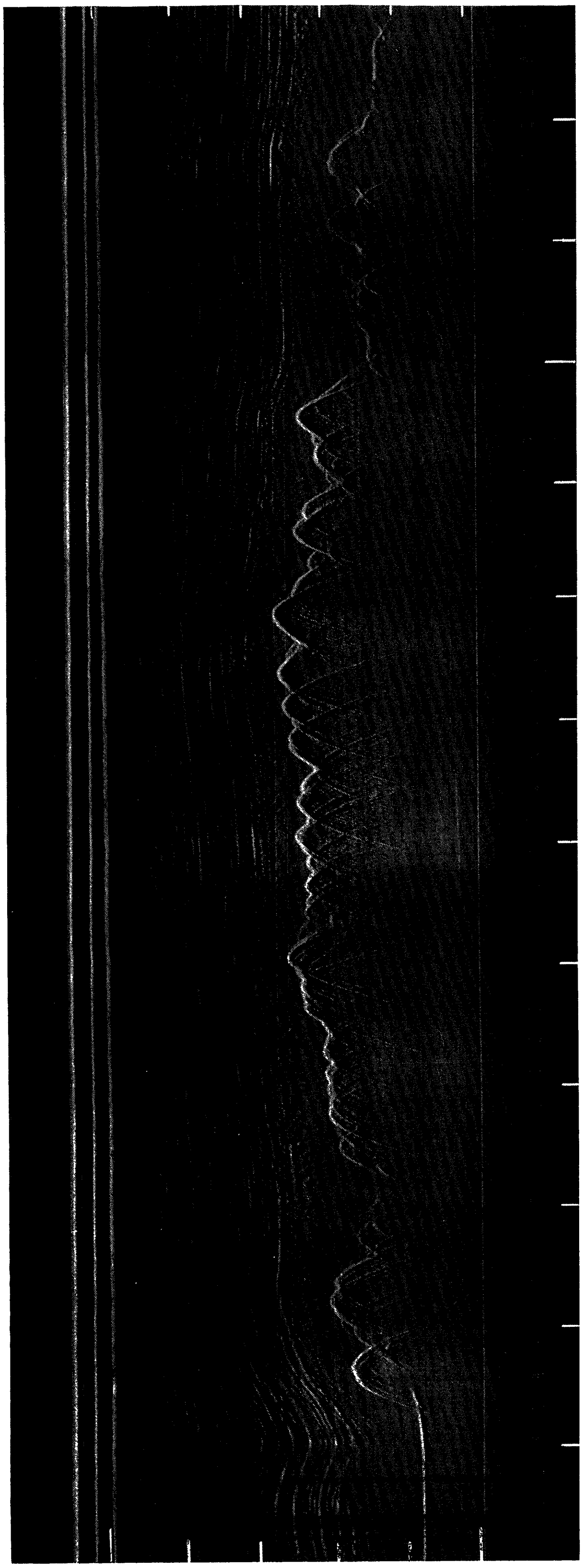
---

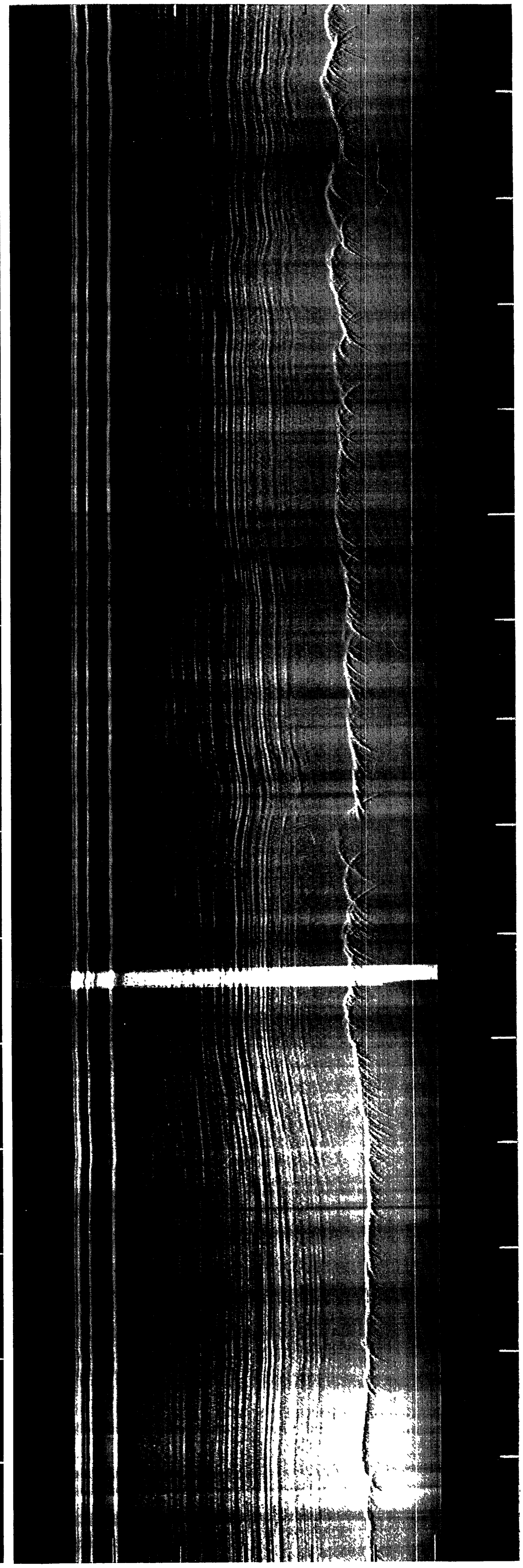
Description of plate 1

Figure 12. Radio-echo sounding record along flowline with similar conditions to Vostok but 130 km north, showing how internal layering heights may vary considerably from that expected from equation (5). Smooth bottom echo on left of diagram is the start of a large sub-ice lake.

Figure 13. Suitable layering conditions for interpretation of deep drilling near Dome C in eastern Antarctica.

Figure 12





Figures 12 and 13. For description see opposite.

(*Facing p.* 156)

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY B BIOLOGICAL SCIENCES

that dating of deep ice layers by conventional glaciological models may be inaccurate over uneven terrain.

The third factor that has received little attention is the question of continuity of an ice core. There is no question of the continuity of precipitation on the upper surface of the ice sheet on the time scales we consider, but where significant changes of ice flow have occurred over the last 100000 years in Greenland or Antarctica, the possibility of discontinuities arises. Ice from near the base of the ice cores from Camp Century and Byrd Stations shows isotopic values similar to those of ice deposited in these areas during the last 10000 years, in spite of the intervening cold periods. If the ice is of relatively local origin, it raises a question of continuity that is most readily discussed in terms of conditions around Camp Century.

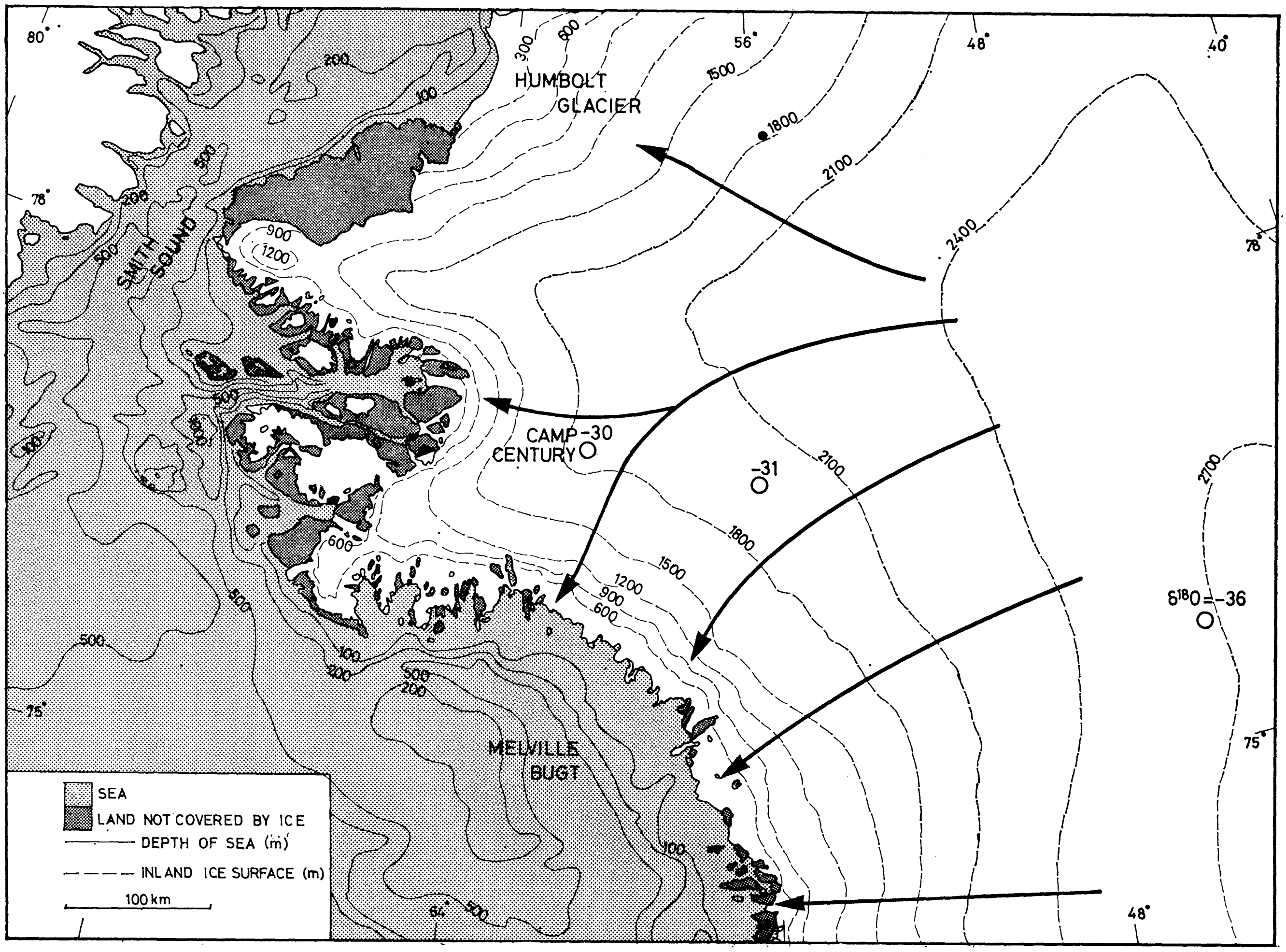


FIGURE 14. Surface contours of the ice sheet of northwestern Greenland from map by Weideck, showing probable flowlines of ice.

In figure 14 we show present-day contours of the ice sheet of northwestern Greenland taken from the map of Weideck (1971). Arrows have been added to show the main directions of ice flow, which follow the generalized contours. Upstream from Camp Century, there appears to be a tendency for the ice flow to be diverted either to the north or east of Camp Century, which leads to the low velocity of flow at Camp Century, which from various pieces of unpublished data appears to be from 5 to 12 m a^{-1}. During a glacial period it seems certain that the ice sheet will expand – at least to the new sea level more than 100 m below the present level – and if ice shelves are formed it would have grounded at depths from 300 to 600 m below the present sea level. In this case, the surface height at Camp Century would be considerably greater, and ice from inland is likely to have moved across the site of Camp Century with less diversion of flow. Figure 11 shows some profiles of possible ice surfaces at different periods that take account of isotopic, temperature, total gas content and flow factors already discussed. Also shown are particle paths to the present ice core from their point of deposition. Although the scales in figure 11 are open to question, and it could be argued that the horizontal scale could be doubled, qualitatively the ideas shown are sound. If no discontinuities are present in the ice column, then the ice sheet during the last interglacial period must have been much smaller (figure 11, lower interglacial? surface), but it would seem fortuitous that the ice now at the base at Camp Century has a similar isotopic value to that now being deposited on the surface. The same argument applies in the case of Byrd Station – which makes the alternative more plausible, namely that the basal ice was deposited during the interglacial at a similar location under similar conditions to the present. This implies that the ice sheets were of similar size and form to the present ice sheet during the interglacial.

Table 1. Forward movement (km) at top of basal layer of ice†

basal layer thickness	1 m	10 m	50 m	100 m
after 10000 a	0.3	3.2	15.8	32
after 50000 a	7.9	15.8	79	158
after 100000 a	3.2	32	158	316

† Basal shear stress 10^5 Pa (1 bar), random crystal orientation, temperature −9 °C, strain rates from Budd *et al.* (1971*a*, Table 3).

It also follows, as can be seen from an alternative particle path to the deepest layers at Camp Century, but coming from local origin, that there must on this model be a discontinuity in the time-depth scale on the ice core, since it involves ice from further inland over-riding ice of local origin. We see on the map in figure 14 that ice deposited around the 2700 m contour has mean $\delta^{18}O$ value 6‰ lower than that of present deposition near Camp Century. If the interglacial ice sheet was similar to that of the present day, then as the ice sheet increased in thickness and carried inland ice over ice of local origin at Camp Century, the lowest levels of inland ice found in the core are likely to have been deposited well inland, and may have a $\delta^{18}O$ value of the order of 6‰ lower than the local ice. This would cause a sharp discontinuity of this amount in the ice column. A number of discontinuities of this magnitude which were found in the lowest levels of ice at Camp Century will be discussed later.

If the lowest layers of ice were over-ridden by ice from further inland, the question arises as to whether such basal ice could remain in place over a period of 10^4–10^5 years. This would not be so on our simple model described by equation (2), but we have seen from figure 5 that strain rates decrease rapidly towards bedrock when the basal temperature is well below pressure melting point.

In table 1 we show the forward movement of ice at different heights above the glacier bed under a shear stress of 1 bar, using constants for polycrystalline ice and the present basal temperature. The present basal stress, based on data from Mock (1968), is about 20 kPa (0.2 bar), but it may well have approached 100 kPa (1 bar) when inland ice was moving over the site. Figures in table 1 should be divided by two if the stress falls from 100 to 60 kPa (1.0–0.6 bar), or by five for a fall from 100 to 30 kPa (1.0–0.3 bar). Although our model is imperfect, and we have neglected the possibility of recrystallization of ice in an easy glide direction, it does appear that the observed 'isotopically warmer' layer around 50 m thick at the bottom of the ice core could be a remnant from the ice before the onset of the last ice age which did not move more than a few tens of kilometres in the intervening time.

When the basal ice is at melting point, we expect some basal sliding, which experience elsewhere indicates may be of the order of half the velocity at the surface of a glacier. However, if sliding of only 2 m a^{-1} were present, basal ice would have moved 100 km in 50000 years. At Byrd Station, the basal sliding may be three times this amount, and one must expect the lowest levels of ice to have originated 100 km or more from the present site of the borehole. The chances of finding basal ice that has moved very little while being over-ridden by ice from further inland are very much less in the Byrd ice core than in the case of Camp Century. This may explain the absence of sharp discontinuities in the Byrd core.

We are now in a position to review isotopic data from the three major ice cores, starting with the most straightforward. We have plotted basic data in figures 15, 16, and 18 on a linear depth scale, and also show the time scale derived by those using the data.

(*a*) *Vostok*

From a climatological point of view this is the simplest to interpret. The surface elevation of the ice sheet is unlikely to have changed by more than 100 m during the last ice age. Surface contours (Scott Polar Research Institute, Map 1974) indicate that the deepest ice beneath Vostok will originate near ice 'Dome B', about 300 km upstream from Vostok with a surface elevation around 300 m higher. The ice core results now available come from the top 26 % of the ice sheet, so corrections for deformation are not large. Present-day motion of the ice sheet at Vostok, from astronomical measurements (Liebert & Leonhardt 1974), is about 3 m a^{-1}, which is close to the steady-state figure we obtain by use of equation (3). The isotopic δ value change with depth due to steady-state flow is estimated at $-1.5‰$ at 1000 m, as indicated by the heavy line in figure 15. Uncertainties due to changing elevation should not exceed $\pm 1‰$.

The major difficulties in interpreting isotopic data in figure 15, kindly supplied by V. M. Kotlyakov of the Institute of Geography, Academy of Sciences, Moscow, are to assess the accuracy of the time scale, and the significance of the rapid fluctuations of $\delta^{18}O$ values with depth. The large fluctuations in $\delta^{18}O$ values could result either from sampling techniques, or perhaps from some irregularities in deposition of ice. Barkov *et al.* (1975), in discussing the study of the top 507 m, report that sample thicknesses varied from 3 to 10 cm, and the spacing between samples ranged from 0.1 to 5 m. They then grouped samples into 20 m intervals for palaeoclimatic interpretation. We have grouped the data in figure 15 into 50 m intervals for this purpose, and have added the heavy line showing the mean $\delta^{18}O$ value over each 50 m interval. A shaded zone around this line shows the standard deviation of the mean as an indication of the error of the mean.

The time scale used in Barkov *et al.* (1975) for analysis of the top 507 m of the core was calculated by taking an accumulation rate of 2.4 cm ice a^{-1}, and applying corrections for vertical strain given in equation (5). This uses the best value of local accumulation rate available. Temperature analyses by Budd, Jenssen & Young (1973) gave a best fit accumulation rate of 2.8 cm ice a^{-1}, which indicates the figure used has applied reasonably well for several thousand

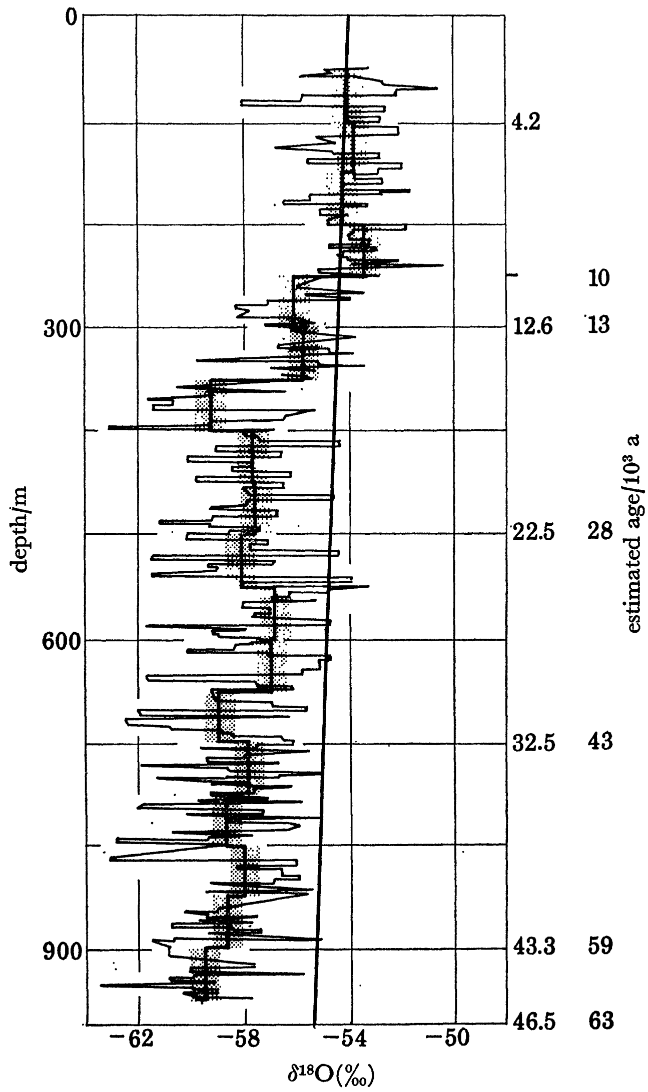


FIGURE 15. Isotopic $\delta^{18}O$ profile from Vostok, Antarctica, plotted on a linear depth scale, supplied by V. M. Kotlyakov and to be published by Barkov *et al.* (in press). Mean values over 50 m depth intervals have been added together with shading to indicate the standard deviation of the mean. The heavy sloping line shows estimated corrections for changing surface altitudes, if the climate had otherwise remained constant. Estimated age scales on the right are those calculated by Barkov and others (inner figures), and estimated in this paper (outer figures).

years. Studies of isotopic layering reported by Barkov *et al.* (1975) at five depths from 48 to 391 m depth (see earlier discussion) showed annual layer thicknesses that decreased from 3.2 to 1.7 cm of ice with a mean value of 2.5 cm ice. They point out that at 391 m depth, a layer of 3.2 cm thickness would decrease to 2.8 cm thickness, which is 40 % greater than that observed. 'This could be explained by lower rates of accumulation, but the illegibility of seasonal variations of $\delta^{18}O$ at horizons 260 and 391 m do not allow us to accept these data with confidence.' They also point out that an average of one year in five or six tends to be missing in any single location around Vostok: hence their figure of 2.4 cm a^{-1} used for determining age. Inspection of figure 15 shows that a figure close to this amount has been used to date ice to 950 m depth (Barkov, Gordiyenko, Korotkevich & Kotlyakov, in press). From our discussion of transport of water vapour in the atmosphere (figure 10), it would appear very likely that a lower rate of accumulation was typical of colder periods, in which case the ages shown on the right hand side of figure 15 should be increased by about 66 % for the ice column below 250 m depth. Inspection of radio-echo records along the flowline upstream from Vostok shows an internal layer at depth around 1700 m from 90 to 20 km upstream of Vostok, increasing in depth to 1900 m over an interval from 20 to 8 km upstream of Vostok. This would involve a layer originally at 850 m depth increasing to 950 m depth between 6000 and 2400 a B.P. When this deformation is also taken into account, we find that the age scale below 100 m should be decreased by 9 % due to this factor. Below 250 m depth this involves a net increase of 50 % in the scale.

When all preceding corrections are taken into account, we conclude that isotopic δ values reached their present-day levels after the ice age 10000 a B.P. with an error of ± 2000 years estimated. The coldest conditions occurred from 17.5 to 20.5×10^3 a B.P. (15.5–17.5 on Barkov *et al.* scale), with an isotopic lowering of 4.5‰ to which we should add a correction for changing oceanic composition, giving a probable temperature lowering of 6–8 °C. Before that, temperatures were a little warmer, with regional isotopic values 2‰ (*ca.* 4 °C) lower than present values from around $32–40 \times 10^3$ a B.P. (or $25–30 \times 10^3$ a B on the time scale of Barkov *et al.*). Colder periods (−4‰ and *ca.* −6 °C) also occurred around 39–43 and $60–63 \times 10^3$ a B.P. (30–32.5 and 43–45 a B.P. Barkov *et al.*).

The Soviet Antarctic Expedition are developing techniques to obtain deeper ice cores from Vostok. The results will be awaited with great interest by all climatologists and glaciologists.

(*b*) *Camp Century*

This ice core has been discussed in more detail than any other, especially by Dansgaard *et al.* (1971) and Dansgaard *et al.* (1973). Earlier discussion in this paper has shown that it is the most difficult to interpret prior to 10000 a B.P., in terms of changing thickness and flow (figure 11), and consequently in dating. Interpretation of the top 25 % of the core deposited since 8300 a B.P. is straightforward, since strain rates appear to have been uniform over this period (figure 5).

The first dating of the ice core used by Dansgaard & Johnsen (1969) was based on a modification of equation (5) to allow for the decrease in strain rate at the lowest levels, but using the present-day accumulation rate and local ice thickness. They found an apparent periodicity in $\delta^{18}O$ values around 2000 years back to 45000 a B.P. (*ca.* 1330 m depth), but before this date the period appeared to decrease to 4000 years. They have assumed that the 2000 year period was present at an earlier date, and used this as a control of earlier dates to determine their time scale.

The above treatment depends on glaciological parameters to 45000 a B.P., but it is not consistent with the model of changes of the ice sheet shown in figure 11. We use this model to estimate $\delta^{18}O$ values, which are plotted against depth in figure 16, as a semi-quantitative estimate of the effect of changing thickness and flow on $\delta^{18}O$ values and on the age scale. The latter assumes an accumulation rate during the last ice age to be two-thirds of the present, and makes a further change by a ratio of 1.7:1.0 to allow for greater ice thicknesses at the point of deposition of ice further inland. The estimates of effects of flow on $\delta^{18}O$ values below 1320 m, shown as approximately −34‰, take account of ice that was deposited during the interglacial

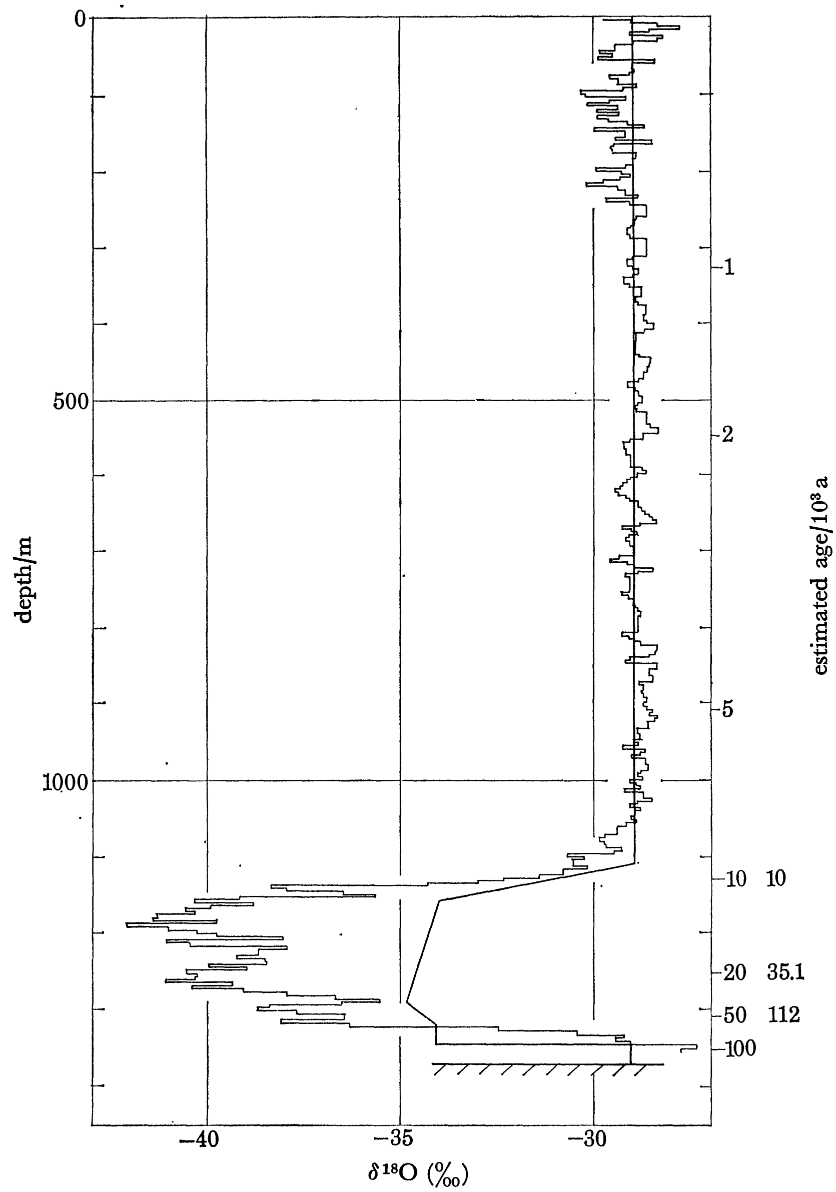


FIGURE 16. Isotopic δ^{18}O profile from Camp Century, Greenland, plotted on a linear depth scale, supplied by Johnsen. The heavy line shows estimated corrections for changing surface altitudes as in figure 15. Age scales on the right are those of Dansgaard *et al.* (1971) (inner figures) and those estimated in this paper based on figure 11 (outer figures).

period that is assumed to have over-ridden ice deposited locally during the interglacial. This in turn assumes that the ice sheet was then of similar size and of similar climate to the present ice sheet. The effect is shown more dramatically in figure 17 which has a linear time scale as in Dansgaard *et al.* (1971). If much of the lower ice column had originated from the inland ice, the period during which ice of −34‰ was deposited will be uncertain. One would, however, expect this to be some thousands of years. Instead, in the profile of Dansgaard *et al.* we find a few sharp isolated peaks of about −34‰, which they interpret as short periods of intense cold before the onset of the ice age, with longer periods of deposition of ice of −27 to −29‰. An alternative explanation is that the lowest 50 m of the ice column does not represent a continuous

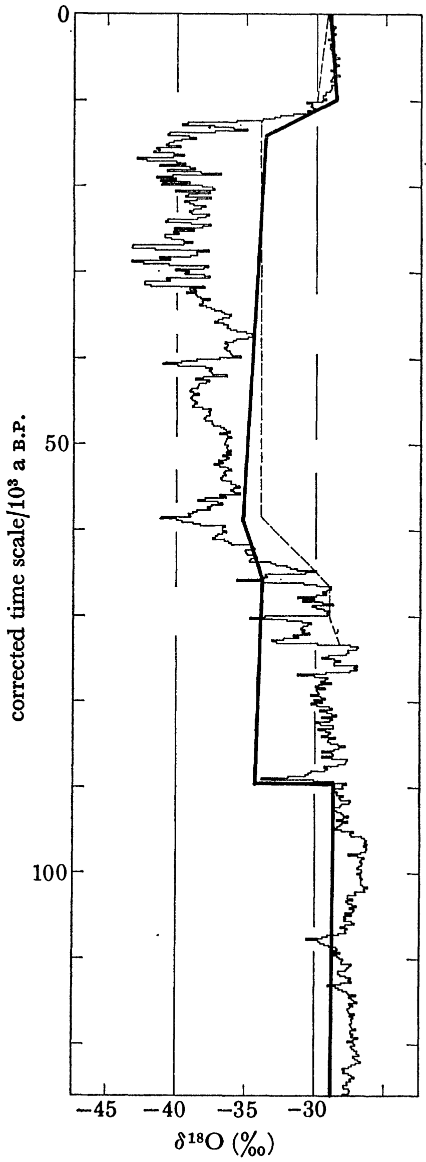


FIGURE 17. Isotopic $\delta^{18}O$ profile for Camp Century, Greenland, plotted on a linear time scale, with dashed lines showing corrections for the influence of changing surface altitudes from Dansgaard *et al.* (1973). The heavy line shows the same corrections as estimated by the author from the model shown in figure 11.

stratigraphy, but shows a mixture of ice from further inland and locally deposited ice. One possible cause of such mixing would be factors associated with the flow of ice in three dimensions which are not taken into consideration on our two dimensional model. Such effects are likely if sharp peaks rise above the glacier bed by the order of a hundred metres or more, in which case ice at lowest levels may tend to flow around a peak, and this may disrupt the stratigraphy when meeting ice flowing over the peak.

The net effect of our discussion so far is to cast doubt on the dating of ice cores by glaciological methods, in cases where the ice was deposited more than 10000 a B.P. If we neglect the glaciological approach, and accept dates for major climatic changes from other sources, then we can calibrate major isotopic changes in the ice core by these dates. A modification of these approaches has been used by Dansgaard *et al.* (1971) to relate events on the isotopic record to known events in the glacial geological record.

Since sampling techniques have been well established by Dansgaard and others, the rapid variations of $\delta^{18}O$ from 15 to 32×10^3 a B.P. on Dansgaard's time scale appear real. It is difficult to determine whether these represent rapid climatic changes, an irregular flow of the ice sheet such as may be associated with surging, or changes in atmospheric circulation and in the sources of water vapour that may not be directly associated with temperature changes.

We can summarize our conclusions on the Camp Century ice core by saying that although information since 10000 a B.P. is dated to within 10 %, before that time dating errors could be as high as a factor of 2.5. The coldest periods on our time scale were centred around 20, 25, 45 and 100×10^3 a B.P. (17, 28, 40, 58×10^3 a B.P. on the scale of Dansgaard *et al.* 1971)). The general climatic effects remaining after eliminating changes of flow and thickness of the ice sheet were maxima of 8, 7, 6.5 and 3.5‰ in $\delta^{18}O$ values, but owing to changes in circulation patterns, an interpretation in terms of temperature is not proposed.

Prior to the onset of the ice age, isotopic δ ratios in the lowest levels of ice were 1–2‰ higher than the present near-surface values (Dansgaard *et al.* 1973). If we accept the argument that this ice was deposited locally, it indicates a climate at that time that was about 2 °C warmer than at present.

(*c*) *Byrd Station*

Of the three major ice cores, more is known about glaciological conditions along the present-day flowline upstream of Byrd Station than is known for the other sites (Whillans 1973, 1976). Primary dating of the ice core in figure 18 is based on this flowline data, and has been carried out independently by Whillans (1976) and Budd and others for the Cambridge workshop (to be published). Results of the two analyses are very similar. Earlier studies by Epstein, Sharp & Gow (1970) used the local accumulation rate and ice thickness to determine ages which were similar to those of Whillans and Budd at 1000 m and about 10000 years older at 2000 m depth. Later work by Johnsen, Dansgaard and others produced three further time scales based on different assumed flowline conditions, which differ considerably from the conditions shown in Whillans (1973). More recently, Thompson *et al.* (1975) have dated the ice column on the basis of the dust content of the ice core, using an apparent annual cycle in dust content. They suggest that the basal ice is approximately 27000 years old, and that the time–depth factor below 1100 m is much too compressed.

The overall impression from these studies is that the ice core at Byrd Station is difficult to interpret and to date. Weight is given to these arguments by current controversy over the stability of the ice sheet of western Antarctica (Hughes 1973; Thomas 1976; Weertman 1976, and others) and by claims of Denton, Armstrong & Stuiver (1971) that the whole of the ice of the present Ross Ice Shelf area was grounded during the Wisconsin period.

There is little doubt that the area of grounded ice in the Ross Sea must have been greater during the last glacial period as a result of a sea level lowering of 100–130 m, and this must in turn have had some effect on the ice thickness and elevation at the site of Byrd Station. In turn, the direction of ice flow may have changed, but we have no evidence of this. Indeed the agreement of the form of the 30000 year isochrons calculated by Whillans (1976), and the form of the internal reflexions at that depth, suggests that the flow near Byrd Station was similar to the present during the last ice age. Unlike Camp Century, Gow & Williamson (1975) report little systematic variation in the total gas content of ice in the Byrd core, 'except possibly for slightly reduced values between 1500 and 900 m depth'. However, the fact that an ice column of about 1200 m in length is available for analysis of conditions prior to the end of the last ice age, compared to the lowest 250 m of the Camp Century ice core, increases the importance of the Byrd ice core. There is no irregularity in the isotopic record (figure 18) to indicate that ice from great inland elevations is present in the ice core. We therefore show a curve for the

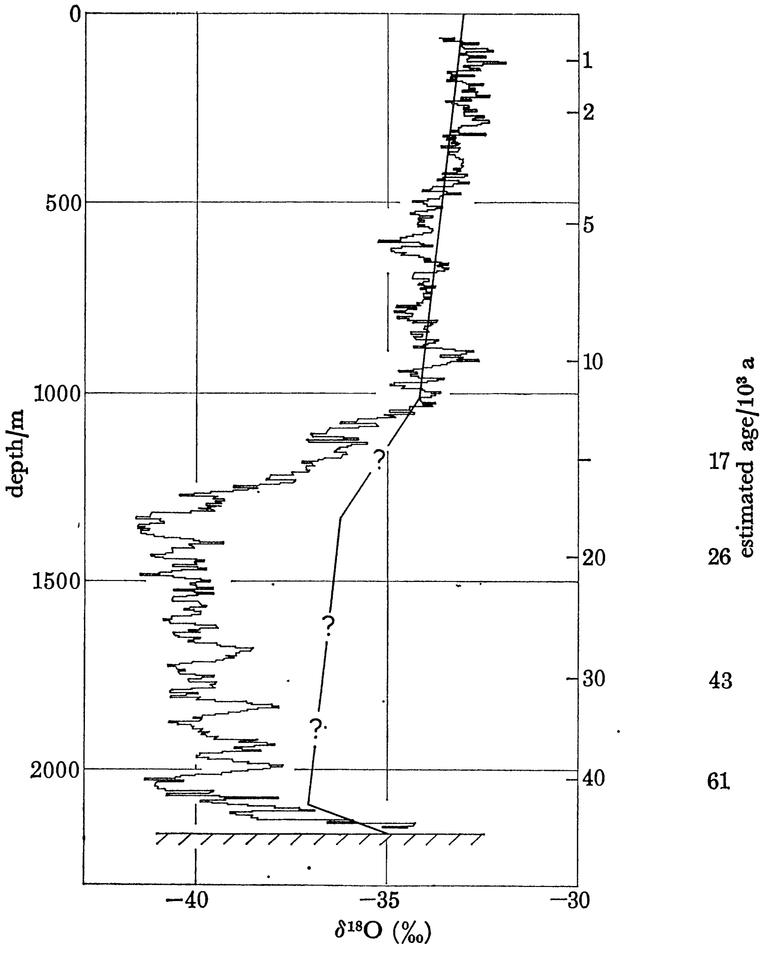


FIGURE 18. Isotopic $\delta^{18}O$ profile for Byrd Station, Antarctica, plotted on a linear depth scale, supplied by Johnsen. The heavy line shows the authors' estimated corrections for the influence of changing surface altitude, if the climate had otherwise remained constant. Age scales on the right are based on information from Whillans (1973) (inner figures) and on estimates made in this paper (outer figures).

effect of changes in thickness and flow, that allows for an increase of ice thickness of 300 m only, and no change in flow conditions from the present day fall of 300 m from the ice divide 150 km from Byrd Station. We also allow for a general decrease of precipitation to 2/3 of the present level, before 10000 a B.P. On this basis we find that the culmination of the last ice age took place at about $22–24 \times 10^3$ a B.P. ($17.5–19 \times 10^3$ a B.P. by Whillans etc.), with a climatic lowering of $\delta^{18}O$ values around 5.3‰, corresponding to a temperature lowering of about 7–8 °C. Although this agrees with the figures for Vostok, the uncertainties of the effect of changing surface levels of ice are much greater. Fluctuations of 2‰ in $\delta^{18}O$ are present at earlier dates, but no prominent colder periods appear until about 65000 a B.P. (41000 a B.P. on Whillans's scale).

Basal ice from before the start of the Devensian ice age shows isotopic δ ratios close to the surface layers of the ice sheet (Epstein *et al.* 1970; Johnsen *et al.* 1972). Since this ice must have been transported by ice movement of 100 km or more from a higher part of the ice sheet, it would appear from the present geometry of the ice that either pre-Devensian climate was about 2–3 °C warmer than the present, or the ice sheet surface was generally lower. The latter condition could also be the product of a slightly warmer climate.

General conclusions

There is a great deal of climatic information to be obtained from ice cores. Use of a relation between isotopic ratios and mean temperatures has been shown to be satisfactory, especially over the past few thousand years. Major difficulties of interpretation arise in determination of the age–depth relation in ice cores older than 10000 a B.P., and in making corrections for changes in the size and flow of ice sheets, in order to determine broad scale climatic changes. Nevertheless the continuity of ice cores is such that one may use dates of major climatic sources to date major changes in isotopic stratigraphy, such as the onset and end of an ice age.

Dates given from purely glaciological considerations put the termination of the last ice age around 10000 ± 1000 a B.P., and the culmination of the ice age around 20000 ± 3000 a B.P. At this time, the lowering of temperature over inland Antarctica was 6–8 °C. Ice age conditions existed continuously for a period of at least 50000 a, possibly 100000 a, although warming of about 3 °C from the coldest conditions occurred at various periods.

Before the onset of this glacial period, isotopic δ ratios in the basal ice at Camp Century and Byrd Station suggest that the climate was slightly warmer than at present – possibly by 2–3 °C.

Many of the ideas on interpretation of temperature and isotopic profiles in polar ice sheets expressed in this paper were initiated or stimulated by a Cambridge Workshop on these topics (see Robin 1973). Thanks are due to all colleagues who participated in the Workshop, the proceedings of which are to be published as a monograph by Cambridge University Press. The Workshop was made possible by a Grant-in-Aid from the Royal Society.

References (Robin)

Aldaz, L. & Deutsch, S. 1967 On a relationship between air temperature and oxygen isotope ratio of snow and firn in the South Pole region. *Earth Planet. Sci. Lett.* **3**, 267–274.

Barkov, N. I., Gordiyenko, F. G., Korotkevich, Ye. S. & Kotlaykov, V. M. 1975 Oxygen isotope study of a 500 m ice core from the borehole at Vostok station. *Inform. Bull. Soviet Antarctic Exped.* No. 90, 39–49.

Budd, W. F., Jenssen, D. & Radok, U. 1971*a* *Derived physical characteristics of the Antarctic ice sheet. Mark I.* Melbourne: University of Melbourne, Meteorology Department. Publication 18.

Budd, W. F., Jenssen, D. & Radok, U. 1971*b* Reinterpretation of deep ice temperatures. *Nature, Phys. Sci.* **232**, 84–85.

Budd, W. F., Jenssen, D. & Young, N. W. 1973 Temperature and velocity interaction in the motion of ice sheets. *First Australasian Conference on Heat and Mass Transfer*, Section I, pp. 17–24. Melbourne: Monash University.

Budd, W. F., Young, N. W. & Austin, C. R. 1976 Measured and computed temperature distributions in the Law Dome Ice Cap, Antarctica. *J. Glaciol.* **16**, 99–109.

Dansgaard, W. 1961 The isotopic composition of natural waters with special reference to the Greenland Ice Cap. *Meddr Grønland*, **165** (2), 1–120.

Dansgaard, W. 1964 Stable isotopes in precipitation. *Tellus* **16**, 436–468.
Dansgaard, W. & Johnsen, S. J. 1969 A flow model and a time scale for the ice core from Camp Century, Greenland. *J. Glaciol.* **8**, 215–223.
Dansgaard, W., Johnsen, S. J., Clausen, H. B. & Langway Jr, C. C. 1971 Climatic record revealed by the Camp Century ice core. In *Late Cenozoic glacial ages* (ed. K. K. Turekian), pp. 37–56. New Haven and London: Yale University Press.
Dansgaard, W., Johnsen, S. J., Clausen, H. B. & Grundestrup, N. 1973 Stable isotope glaciology. *Meddr. Grønland* **197** (2), 1–53.
Denton, G. H., Armstrong, R. L. & Stuiver, M. 1971 The late Cenozoic glacial history of Antarctica. In *Late Cenozoic glacial ages* (ed. K. K. Turekian), pp. 267–306. New Haven and London: Yale University Press.
Epstein, S., Sharp, R. P. & Gow, A. J. 1970 Antarctic ice sheet: stable isotope analyses of Byrd station cores and interhemispheric climatic implications. *Science, N.Y.* **168**, 1570–1572.
Gow, A. J. & Williamson, T. 1971 Volcanic ash in the Antarctic ice sheet and its possible climatic implications. *Earth Planet. Sci. Lett.* **13**, 210–218.
Gow, A. J. & Williamson, T. 1975 Gas inclusions in the Antarctic ice sheet and their glaciological significance. *J. Geophys. Res.* **80**, 5101–5108.
Gudmandsen, P. 1975 Layer echoes in polar ice sheets. *J. Glaciol.* **15**, 95–101.
Hughes, T. 1973 Is the West Antarctic Ice Sheet disintegrating? *J. geophys. Res.* **78**, 7884–7910.
Johnsen, S. J. 1977 Stable isotope profiles compared with temperature profiles in firn and with historical temperature records. Paper presented at International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, 1975. (To be published by International Association of Scientific Hydrology.) (In the press.)
Johnsen, S. J., Dansgaard, W., Clausen, H. B. & Langway Jr, C. C. 1972 Oxygen isotope profiles through the Antarctic and Greenland ice sheets. *Nature, Lond.* **235**, 429–434.
Liebert, J. & Leonhardt, G. 1974 Astronomical observations aimed at determining ice sheet motion in the Vostok station area. *Inform. Bull. Soviet Antarctic Exped.* No. 88, pp. 68–70.
Lorius, C. 1975 *Rapport d'Activité, 1975*. Grenoble: Centre National de la Recherche Scientifique, Laboratoire de Glaciologie.
Merlivat, L. & Nief, G. 1967 Fractionnement isotopique lors des changements d'état solide-vapeur de l'eau à des températures inférieures à 0 °C. *Tellus* **19**, 122–127.
Mock, S. J. 1968 Snow accumulation studies on the Thule Peninsula, Greenland. *J. Glaciol.* **7**, 59–76.
Nye, J. W. 1963 Correction factor for accumulation measured by the thickness of the annual layers in an ice sheet. *J. Glaciol.* **4**, 785–788.
Oeschger, H., Stauffer, B., Bucher, P. & Loosli, H. H. 1977 Down bore hole extraction of gases and dissolved and particulate matter in ice. Paper presented at the International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, 1975. (To be published by International Association of Scientific Hydrology.) (In the press.)
Paren, J. G. & Robin, G. de Q. 1975 Internal reflections in ice sheets. *J. Glaciol.* **14**, 251–259.
Philberth, K. & Federer, B. 1971 On the temperature profile and the age profile in the central part of cold ice sheets. *J. Glaciol.* **10**, 3–14.
Picciotto, E., De Maere, X. & Friedman, I. 1960 Isotopic composition and temperature of formation of Antarctic snows. *Nature, Lond.* **187**, 857–859.
Ragone, S. E., Finelli, R. V., Leung, S. & Wolf, C. 1972 Cationic analysis of the Camp Century, Greenland, ice core. *U.S. Army C.R.R.E.L. Special Report* 178. 15 pp.
Raynaud, D. 1976 Unpublished. Thèse de Doctorat d'Etat. University of Grenoble. (Centre National de la Recherche Scientifique, Laboratoire de Glaciologie.)
Raynaud, D. & Lorius, C. 1973 Climatic implications of total gas content in ice at Camp Century. *Nature, Lond.* **243**, 283–284.
Robin, G. de Q. 1955 Ice movement and temperature distribution in glaciers and ice sheets. *J. Glaciol.* **2**, 523–532.
Robin, G. de Q. 1964 Glaciology. *Endeavour* **23**, 102–107.
Robin, G. de Q. 1973 Cambridge Workshop on temperature and isotopic profiles in polar ice sheets, 1973. *Polar Rec.* **16**, 902–907.
Robin, G. de Q. 1976 Temperature-depth and oxygen-isotope profiles in polar ice sheets. *J. Glaciol.* **16**, 9–22.
Robin, G. de Q., Evans, S. & Bailey, J. T. 1969 Interpretation of radio echo sounding in polar ice sheets. *Phil. Trans. R. Soc. Lond. A* **265**, 437–505.
Scott Polar Research Institute, Cambridge. 1974 Map: Antarctica, Radio echo sounding map series A. 1:5000000. Sheet 3. Ice sheet surface and sub-ice relief. 90° E–180°.
Thomas, R. H. 1976 Thickening of the Ross Ice Shelf and equilibrium state of the West Antarctic ice sheet. *Nature, Lond.* **259**, 180–183.
Thompson, L. G., Hamilton, W. L. & Bull, C. 1975 Climatological implications of microparticle concentrations in the ice core from 'Byrd' station, Western Antarctica. *J. Glaciol.* **14**, 433–444.
Weertman, J. 1976 Glaciology's grand unsolved problem. *Nature, Lond.* **260**, 284–286.

Weideck, A. 1971 Quaternary map of Greenland. 1:2500000. Geological Survey of Greenland, Copenhagen. (Undated: received 1971.)
Whillans, I. M. 1973 State of equilibrium of the West Antarctic inland ice sheet. *Science, N.Y.* **182**, 476–479.
Whillans, I. M. 1976 Radio-echo layers and the recent stability of the West Antarctic ice sheet. *Nature, Lond.* **264**, 152–155.
Zhavoronkov, N. M., Uvarov, O. V. & Sevryugova, N. N. 1955 Some physico-chemical constants of heavy oxygen water. In *Primeneniye mechenykh atomov v analiticheskoy khimii* (ed. A. P. Vinogradov), pp. 223–233. Moscow: Akademiya Nauk SSSR (Inst. geokhimii i analit. khimii).

BIOLOGICAL SCIENCES
THE ROYAL SOCIETY
PHILOSOPHICAL TRANSACTIONS OF

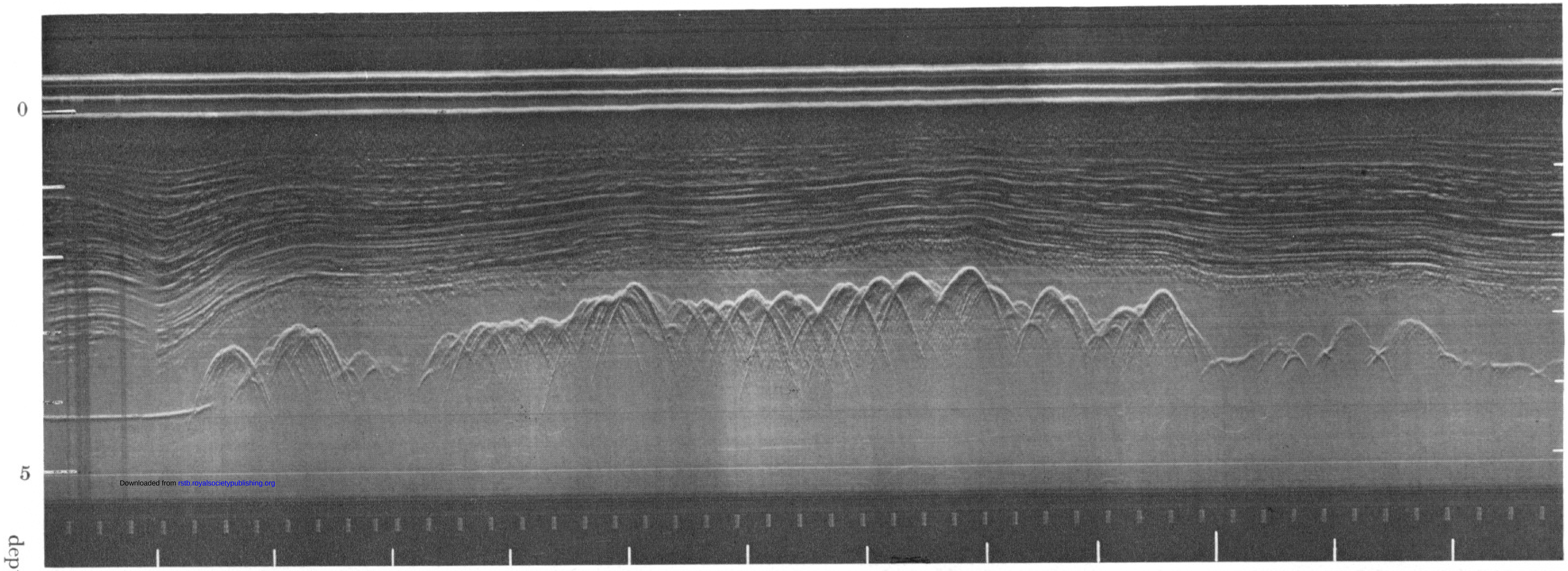


Figure 12

Figures 12 and 13. For description see opposite.