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ICE SHEETS: ISOTOPES AND TEMPERATURES

G. de Q. ROBIN

1

Introduction

The ice sheet of Antarctica covers an area of 13 million km^2 and is more than 4 km deep in places. The Greenland ice covers 1.8 million km^2 and exceeds 3 km in depth. Both these ice sheets contain a wonderfully detailed record of the atmospheric history of the Earth that extends back for hundreds of thousands of years.

Each snowfall that adds a layer to the body of an ice sheet deposits also a climatic record, but the oldest parts of this record are being removed ceaselessly by the outward flow of ice to regions where it melts and mixes with ocean waters. The continuing distortion of ice that takes place as the ice sheets are squeezed outward under their own weight also distorts the historical record. If we are to learn to read the history effectively, it is first necessary to understand the way in which a piece of ice has been transported and transformed since it fell as snow many thousands of years ago and perhaps many hundreds of kilometres from the point where it is recovered. The development of the Antarctic ice sheet has taken place over 20 million years or more, but chances of finding pockets of ice dating from its earlier period are very small. However, near bedrock in Greenland and Antarctica we do find ice that was deposited before the last ice age. Since that time vast ice sheets have grown and covered large areas of North America and Eurasia and then retreated. Although present-day polar ice sheets record changes of the Earth's atmosphere during the last ice age, they must themselves have varied in size and flow at the same time. These changes need to be taken into account when interpreting evidence from the lower levels of ice sheets.

Access to the long historical record in polar ice sheets has been made possible by the application and development of deep drilling techniques over the past three decades. Ice cores of up to 15 cm in diameter are brought to the surface, catalogued, and sections are then taken to specialised laboratories in different countries for detailed analysis. The longest records obtained before 1980 were the two cores from the surface to bedrock of ice sheets recovered by the US Army Cold Regions Research and Engineering Laboratories (CRREL) from Camp Century in north west Greenland (1388 m) in 1966, and from Byrd Station, Antarctica (2164 m) in 1968. Each of these cores covers an estimated period of about 100 000 a. Drilling to a depth of nearly 1000 m has been carried out at Vostok Station on the East Antarctic ice sheet by Soviet scientists and engineers, and analyses of the top 950 m cover an estimated period of about 50 000 a.

In 1976–7 a French group, with logistic support from the US National Science Foundation, obtained ice cores at Dome C (74° 40′ S 124° 10′ E, 3240 m elevation) to a depth of 906 m. It is estimated that these cores cover a time span of 32 000 a.

In 1980 and 1981 Soviet scientists obtained an ice core in good condition to about 1500 m depth at Vostok Station, Antarctica. In 1981, the Greenland Ice Sheet Project, run by US, Danish and Swiss scientists, recovered an ice core to bedrock at 2037 m depth at Dye 3 near the southern dome of the Greenland ice sheet. It will be some time before comprehensive results from these two projects are available. In this volume particular attention is given to results from the earlier drilling projects at Camp Century, Byrd Station and Vostok.

The historical record present in the ice takes several forms, not all of which have been studied adequately to date.

Ice from the central regions of Antarctica or Greenland contains only a few parts per million by weight of impurities such as dust, some of which may come from extra-terrestrial sources. However, thanks to the purity of the body of ice, even minute traces of impurity can be determined, as for example the variation of lead content of the Greenland ice sheet over past centuries (Figure 1.1). One can see the increase of atmospheric lead due to man's activities since the early nineteenth century, and particularly since the introduction of leaded petrols at about the middle of this century. Isolated peaks at earlier dates may be due to natural events, such as volcanic eruptions.

A careful study by Boutron & Lorius (1979) of the abundance of twelve elements (impurities) deposited in central Antarctica at Dome C shows evidence of marked fluctuations over the past 60 years. It appears that in Antarctica, industrial pollution has had no significant effect. Impurities of marine origin (Na, Mg) are the most abundant. Another group of impurities (Al, Fe, Mn, K and Ca) come from a continental crustal source, while the heavier elements (Pb, Cd, Cu, Zn and Ag) which show a large variability with time are most likely to be of volcanic origin.

Ice sheets also preserve a continuing sample of the atmospheric gases. Fine snow crystals falling on the surface soon recrystallise into a coarser structure known as firn, a porous solid. In spite of the absence of melting in a cold polar ice sheet, the firn gradually increases in density as it is buried more deeply until it becomes impermeable at a depth of 50–200 m, and is then termed ice by glaciologists. At this point it contains about ten

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per cent of air by volume, which is then trapped in bubbles and preserved in the ice. This air can be recovered for analysis from ice samples obtained by deep drilling. Composition of air samples from ice of various ages shown in Table 1.1 indicates that only relatively small changes can have taken place in the composition of the Earth's atmosphere over the past 40 000 a.

Isotopes and climate

The most detailed record of climatic change preserved in the ice is provided by the changing proportion of heavy to light atoms (isotopes) of oxygen and hydrogen which form the ice mass. These ratios are measured by means of mass spectrographs in laboratories organised for the purpose. By regular sampling of the ice along a core, we can determine the way in which isotopic ratios vary with depth, and therefore with time, provided we know the relationship between age and depth. Figure 1.2 shows a very detailed record of the variation of isotopic ratios down the ice core obtained at Byrd Station, Antarctica. This book considers the extent to which a record of past temperature changes can be inferred from such isotopic profiles. In other words, do our isotopic ratios provide us with an isotopic 'thermometer' which records the temperatures at the time the ice fell as snow and, if so, how reliable is this record of temperature?

Table 1.1.* *Composition of air trapped in polar ice at Camp Century, north west Greenland.*

Depth, metres	Estimated age, years	N_2, %	O_2, %	Ar, %
108	300	78.2	20.8	0.95
555	2 360	78.3	20.8	0.94
898	4 600	78.1	20.9	0.95
1080	8 000	78.2	20.9	0.95
1201	(14 000)	78.3	20.8	0.94
1300	(20 000)	78.3	20.8	0.95
1345	(40 000)	78.4	20.7	0.95
Atmosphere present day:		78.1	21.0	0.93

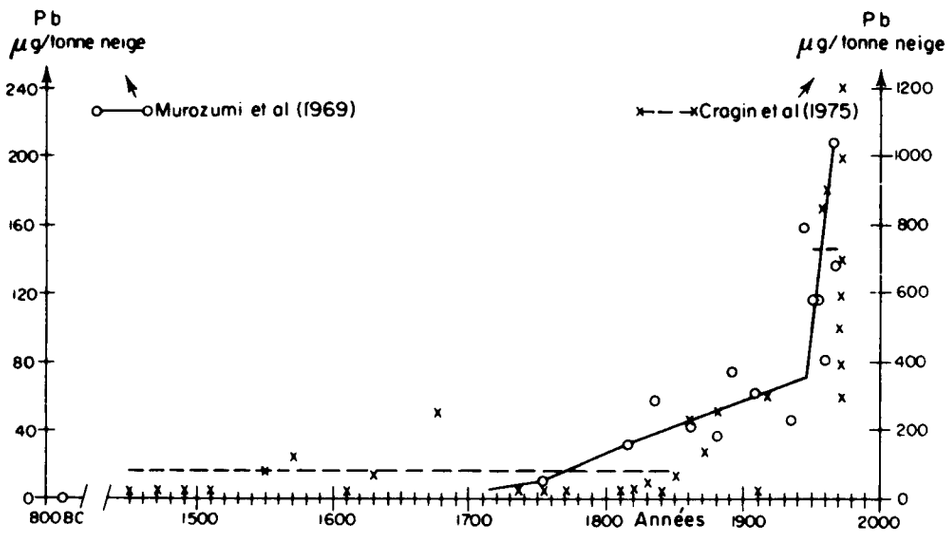
* Based on Table 1 of Raynaud & Delmas (1977). Ages shown are based on Figure 1.9 except for the two deepest samples, which is based on curve 4 of Figure 5.22. Figures for CO_2 content in the original table have been omitted and the paper by Delmas, Ascencio & Legrand (1980) should be consulted. These indicate that CO_2 concentrations were almost half the present level at the end of the last ice age.

Figure 1.1. Variation of lead content of snow deposited in Greenland during historical time. From Lorius & Briat (1976). The left-hand scale and open circles show data from Camp Century (77° N) in north west Greenland from Murozumi, Chow & Patterson (1969). The right-hand scale and crosses refer to results from Dye 3 in central Greenland (65° N) (Cragin, Herron & Langway, 1975) which lies closer to the latitude of the industrial zones.

The relationship between the isotopic ratio and the temperature of snow is due to the tendency of heavier molecules of water to evaporate less rapidly than the normal (lighter) molecules, and for the heavier molecules to condense more readily from the vapour. We can consider these processes as starting from the ocean, the major source of atmospheric water vapour. The oceans, which are of nearly constant isotopic composition, contain 98 per cent of the free water on earth. Of this water, some 99.7 per cent is composed of normal hydrogen (H) and oxygen (^{16}O) atoms in the form of H_2O, as opposed to heavier hydrogen (D, deuterium) or heavier oxygen (^{18}O) atoms. About 0.2 per cent of ocean water consists of $H_2{}^{18}O$ and about 0.03 per cent of $HD^{16}O$.

Instead of presenting measurements of the proportion of heavier molecules in water, water vapour or ice in absolute terms, it is customary and more convenient to use isotopic ratios defined in relation to an arbitrary standard known as 'Standard Mean Ocean Water' (SMOW).* The ratios are presented as δ values for $\delta^{18}O$ or deuterium δD given in relation to SMOW by

* Standard Mean Ocean Water is defined in terms of a water mass held at the National Bureau of Standards (NBS-1) in Washington, DC, USA, and by secondary reference standards at the International Atomic Energy Agency, Vienna.



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$$\delta^{18}O = \frac{R^{18}O_{Sample} - R^{18}O_{SMOW}}{R^{18}O_{SMOW}} \quad (1.1)$$

where R refers to the ratios of $H_2{}^{18}O/H_2{}^{16}O$ in the sample and in SMOW.

In Figure 3.2 we see that, apart from the Arctic Ocean, measured $\delta^{18}O$ values of oceanic surface waters lie between −1.1 and +1.0 parts per thousand ($^0/_{00}$), as should be expected if the primary standard SMOW has been carefully collected. Although the present mean δ value of oceanic waters is close to zero, depletion of the lighter isotopic component due to the formation of continental ice sheets means that the average oceanic $\delta^{18}O$ values rose to about $+1.5^0/_{00}$ when Pleistocene ice sheets reached their maximum extent. On the Greenland ice sheet, mean annual $\delta^{18}O$ values of snow now being deposited range from about $-23^0/_{00}$ to approximately $-38^0/_{00}$, while on Antarctica mean values of current deposition run from $-18^0/_{00}$ to about $-60^0/_{00}$. Some laboratories have concentrated on measuring δ values for deuterium. A sufficient number of comparative measurements have been made to show that for ice and snow of polar regions, the relationship between $\delta^{18}O$ and δD values can be considered for practical purposes as linear and given by

$$\delta D = 8.0\delta^{18}O + 10^0/_{00} \quad (1.2)$$

(Lorius & Merlivat, 1977).

We can now consider the physical processes governing the relationship between δ values and temperatures in polar ice sheets. These are shown schematically in Figure 1.3 with figures giving a rough indication of mean annual $\delta^{18}O$ values for water molecules involved at different stages of the cycle ocean–atmosphere–ice sheet–ocean. The relatively uniform composition of ocean waters is due to mixing during their circulation, together with the fact already mentioned that the oceans contain most of the free water on Earth. During evaporation from the oceans, water is depleted of its heavier component molecules by an amount depending on temperature, but of the order of $10^0/_{00}$ in $\delta^{18}O$ content. Greatest evaporation takes place over the tropics, but these areas are largely separated from the circulation patterns in

Figure 1.2. Former climate indicated from measurements of oxygen isotope ratios on the ice core at Byrd Station. Ice at 1000 m depth was deposited around 11 000 a BP, while the lowest ice is of the order of 100 000 a old. From data supplied by S. J. Johnsen.

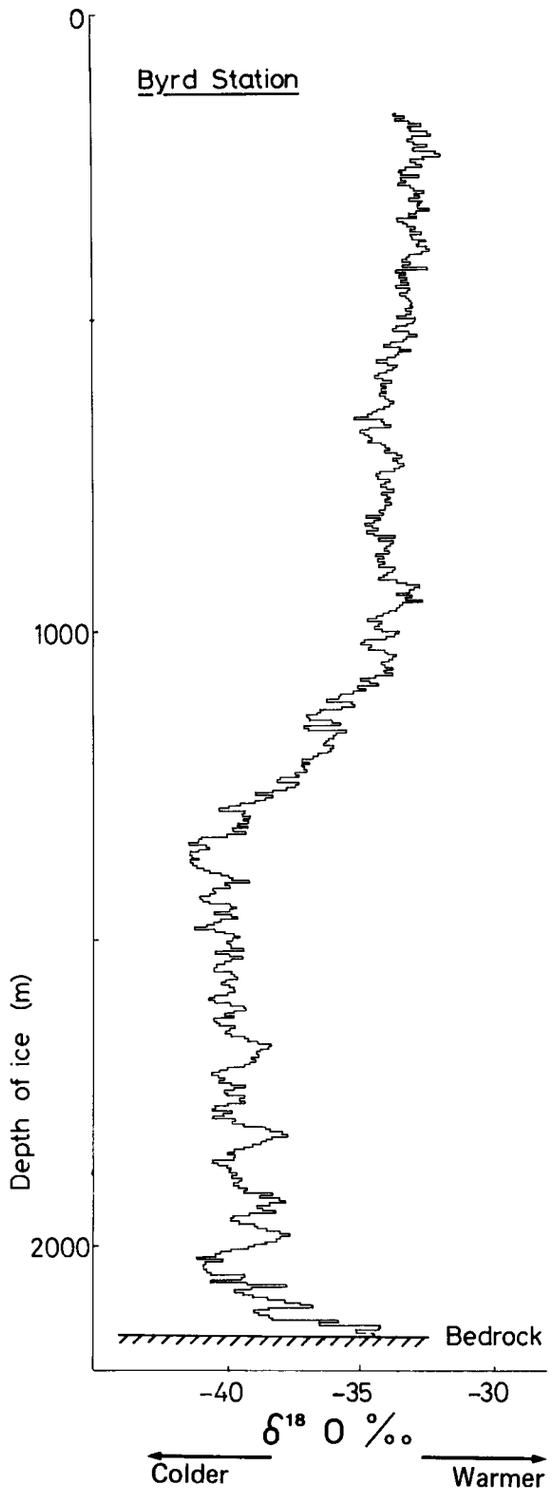
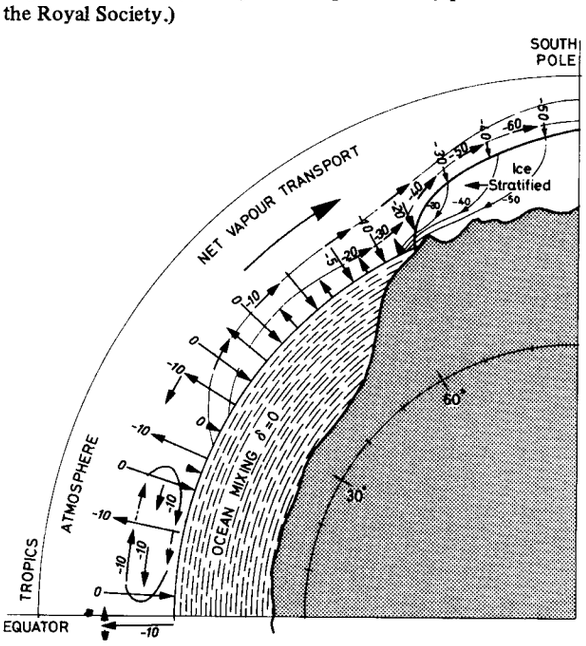


Figure 1.3. Schematic presentation of isotopic cycle of water vapour through ocean–atmosphere–ice sheet–ocean. The figures show approximate $\delta^{18}O$ values in parts per mil at different stages of the cycle. (Robin, 1977. Reproduced by permission of the Royal Society.)



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mid-latitudes by the sub-tropical high pressure areas. However, there is considerable transport of air between mid-latitudes and the polar regions. Furthermore, as one moves to higher latitudes, sea surface temperatures are lower, evaporation from the sea surface decreases, and there is a net transport of atmospheric water vapour towards the poles.

In the tropics, condensation from the vapour enriches the water in the heavier isotopes by an amount similar to the depletion during evaporation, hence tropical precipitation has δ values similar to those of the oceanic water, i.e. $\delta^{18}O \approx 0$. As one moves from mid-latitudes towards the poles, however, the original water vapour is not fully replaced by evaporation from the sea, as in the tropics, so the δ value of the atmospheric water vapour falls increasingly as the air mass moves towards the pole. Over the Antarctic ice sheet evaporation is too small to have much effect on the atmospheric composition. Rough mean $\delta^{18}O$ values on the tangential arrows in Figure 1.3 indicate the magnitude of this depletion as one moves over Antarctica.

A more detailed discussion of the distribution of surface δ values in ice sheets and related mean surface temperatures is given in chapter 3, sections 3.2 (Greenland) and 3.3 (Antarctica). While these sections present evidence of a regular variation of δ values with mean ice temperatures, we see also that the relationship between the two varies to some extent between different geographical regions. Nevertheless, because of their regularity of form, which is discussed in chapter 2, polar ice sheets, more than any other topography, favour the development of consistent relationships between δ values and ice temperatures.

In temperate and sub-polar regions, precipitation is mainly produced as the result of frontal activity between different air masses. Instability in individual air masses, and interactions between the surface and the air mass, such as cooling due to upslope winds, may also produce precipitation. Over mountainous regions, snow falling from one system of clouds is likely to have much the same physical characteristics, whether it falls in a valley or on a mountain. Consequently, a regular relationship between isotopic δ values and elevation or mean temperature is less likely to be found in regions of rough topography.

In peripherar regions of polar ice sheets, especially where surface elevations are less than 1000 m, mean δ values of precipitation are less regular in relation to location and elevation than further inland, where upslope winds are an important cause of precipitation. At a distance of more than 100 km from the periphery, ice-sheet slopes rise continuously towards the centre at a gradient of 1:200 or less in Greenland and still more slowly in central Antarctica. This regularity of form produces greater regularity in the pattern of cooling of air masses. This in turn leads to a regularity in the δ values for precipitation which can be related to temperature and climate.

The δ values of an ice sample recovered from an ice core are determined originally by the processes of precipitation. Although direct condensation of rime on to the surface of ice sheets does take place, the main bulk of precipitation comes from clouds. Studies have been made in Antarctica to measure $\delta^{18}O$ values of falling snow in relation to the cloud temperatures at which the snow was formed. Figure 1.4 shows the data obtained by Picciotto, de Maere & Friedman (1960) at Roi Baudouin Station on the Antarctic coastline (70° 26′ S 23° 19′ E). The results, together with those from a second study made at the South Pole by Aldaz & Deutsch (1967) are shown by lines B and A in Figure 1.5. Also shown on this diagram are dashed curves 1 and 2 which link results

Figure 1.4. Isotopic composition of oxygen in snowfalls with respect to the range of temperature in the corresponding cloud sheet. (From Picciotto *et al.*, 1960. Reproduced by permission of *Nature, London*.)

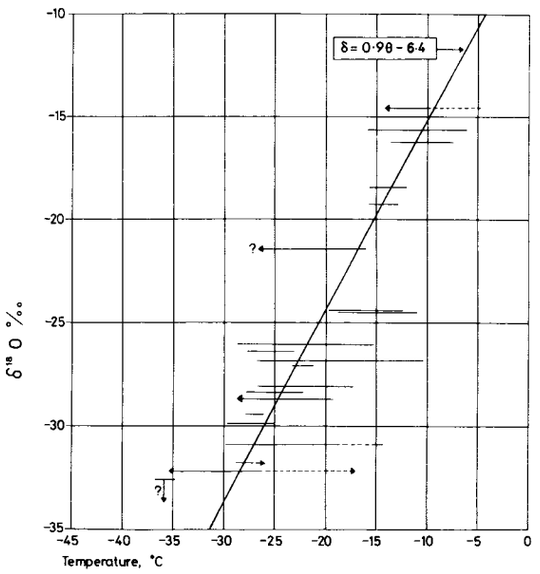
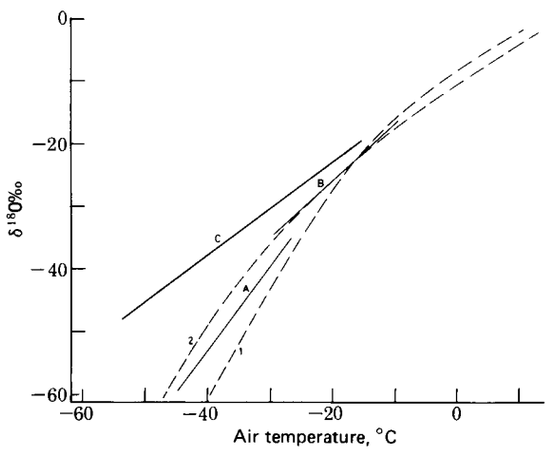


Figure 1.5. Isotopic $\delta^{18}O$–temperature relationships. Line A, clouds at South Pole, from Aldaz & Deutsch (1967); line B, clouds at Roi Baudouin Station on the Antarctic coast, from Figure 1.4. Dashed curves 1 and 2, calculated for Rayleigh condensation with adiabatic cooling, are based on line B; the fractionation factor used to calculate curve 1 is derived from Zhavoronkov *et al.* (1955), that for curve 2 derived from Merlivat & Nief (1967). Line C gives the isotopic–temperature relationship for surface layers of firn of eastern Antarctica (see Figure 3.11). (Based on Aldaz & Deutsch, 1967, Fig. 3. Reproduced by permission of *Earth and Planetary Science Letters*.)



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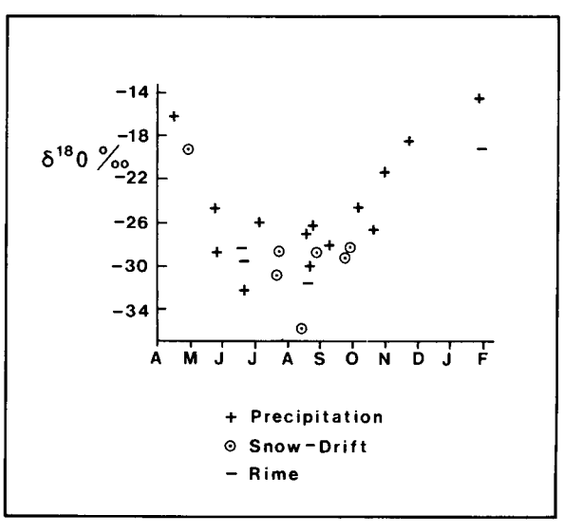
at Roi Baudouin with those at the South Pole by assuming that air from Roi Baudouin undergoes moist adiabatic expansion as it travels up over the Antarctic ice sheet until it reaches the South Pole. Since there is some uncertainty over the correct isotopic fractionation factor to use in the calculations, curves 1 and 2, which use different physical constants, are both shown. These are tied to curve B and are seen to bracket curve A, the δ value–cloud temperature curve for the South Pole; so the physical model appears to be approximately correct. The heavy line C in Figure 1.5 is taken from Figure 3.11. This shows the relationship between δ values of samples of the surface layers of the ice sheet and the mean temperatures of the surface layers. Conversion from δD to $\delta^{18}O$ has been made by equation (1.2). The increasing difference between the δ value of the surface temperature curve and δ value–cloud temperature curves as one moves to lower temperatures is due to increasing differences between the mean surface temperature and the mean temperature of the clouds in which the precipitation is formed. This agrees with meteorological studies of the cold surface inversion over the Antarctic ice sheet, which is caused by intense radiative cooling from the surface throughout the year. As the relatively warm air moves over the continent from the surrounding oceans, strong radiative cooling from the snow surface cools the air above whenever the sky is clear of cloud. Schwerdtfeger (1970) gives mean values of the temperature difference between the surface and the air above the cold inversion (from 450–1000 m above the surface) which have been measured during regular radio-sonde ascents at different stations. At Vostok Station in eastern Antarctica (elevation 3700 m), the air above the inversion is warmer by an average of +15.7 °C than the recorded surface temperature throughout the year; at the South Pole (2800 m), the warming averages +14.1 °C, and at Mirny on the coast, although inversions are often present, air at 500 m elevation is on average 0.3 °C colder than the surface temperatures. At Byrd Station (1500 m), in western Antarctica, air above the inversion averages +6.4 °C warmer than at the surface. Mean annual temperatures at these stations are Vostok −55.6 °C, South Pole −49.3 °C, Byrd −27.9 °C, and Mirny −11.5 °C. The tendency for δ values in western Antarctica to be somewhat higher than in eastern Antarctica (section 3.3) for the same surface temperatures may well be due to heavier cloud cover and hence weaker surface inversions over western Antarctica. Of course, we should not assume that the mean air temperature above the inversion is the same as the mean cloud temperature at which snow is formed. However, the difference is not large at the South Pole where the mean temperature above the inversion is −35 °C compared with the mean condensation temperature of −39 °C (Aldaz & Deutsch, 1967).

The range of temperatures shown in Figure 1.4 is partly due to variations between air masses, but mainly due to seasonal variations as shown in Figure 1.6 where isotopic δ values are plotted as a function of time. We see that these vary from a mean around $-16^0/_{00}$ in summer to about $-29^0/_{00}$ in winter months at Roi Baudouin Station on the coast of Antarctica. In Figure 1.7 we show similar information from Camp Century, but in this case, the data come from a study of closely spaced samples from various depths below the surface of the ice sheet. The seasonal variation is seen as a variation in $\delta^{18}O$ values with depth, the position deduced for summer and winter deposits being marked by S and W respectively.

The range of δ values between summer and winter firn is approximately $15^0/_{00}$ in the top 0.5 m, similar to the range shown in Figure 1.6 for Antarctic coastal precipitation. However, this range decreases with depth and averages about $5^0/_{00}$ from 100 m to depths approaching 1000 m, when a further decrease occurs. These changes are due to mobility of the water molecules in the firn and ice. At shallow depths larger crystals grow rapidly at the expense of smaller, due to transport of molecules mainly by sublimation. This mobility will tend to smooth out seasonal and shorter-period variations of δ values with depth. However, as ice becomes less porous, this type of mobility decreases and is of little significance for densities greater than 550 kg m^{-3} that are reached at 10–20 m depths. The further decrease at depths around 1000 m is due to molecular diffusion within the main body of ice, which becomes more effective as layer thicknesses are reduced and temperatures increase. These factors are discussed in more detail in section 3.4.

The changes of layer thickness are due primarily to the continuous addition of a mass of snow on the top surface of the ice, and the consequent downward and outward squeezing of the ice mass which causes the flow of the ice sheet. We can follow the mechanics of the process if we consider the history of a vertical column of ice, and assume for the purpose of our model (Figure 1.8) that the column stays vertical as it is pushed outwards by the flow of ice. We treat a two-dimensional case only, but the argument is similar for a three-dimensional case. For the present, we limit our study to the case of an ice

Figure 1.6. Seasonal variation of $\delta^{18}O$ values of surface accumulation at Roi Baudouin Station, Antarctica, April 1958–Feb. 1959. (Gonfiantini & Picciotto, 1959. Reproduced by permission of *Nature, London*.)



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sheet in a steady state, that is one in which the size and shape of the ice sheet does not change with time, so that the flow of ice will just be in balance with the rate of accumulation or ablation (loss of ice) at each point on the surface of the ice sheet. For such discussions we neglect seasonal changes of snowfall and ablation and consider only their annual mean values.

At the centre of the ice sheet, if we are adding ice at the rate A_0 on the top surface, the surface must be sinking at a rate $w_0 = A_0$ if the thickness is to remain constant. To compensate for this downward motion we allow our column to spread out horizontally at a uniform rate at all depths. This corresponds to a vertical velocity that falls uniformly from w_0 at the surface to zero at bedrock, and at any height z above bedrock, the velocity w_0 is given by

$$w_0 = A_0 \frac{z}{Z_0} \tag{1.3}$$

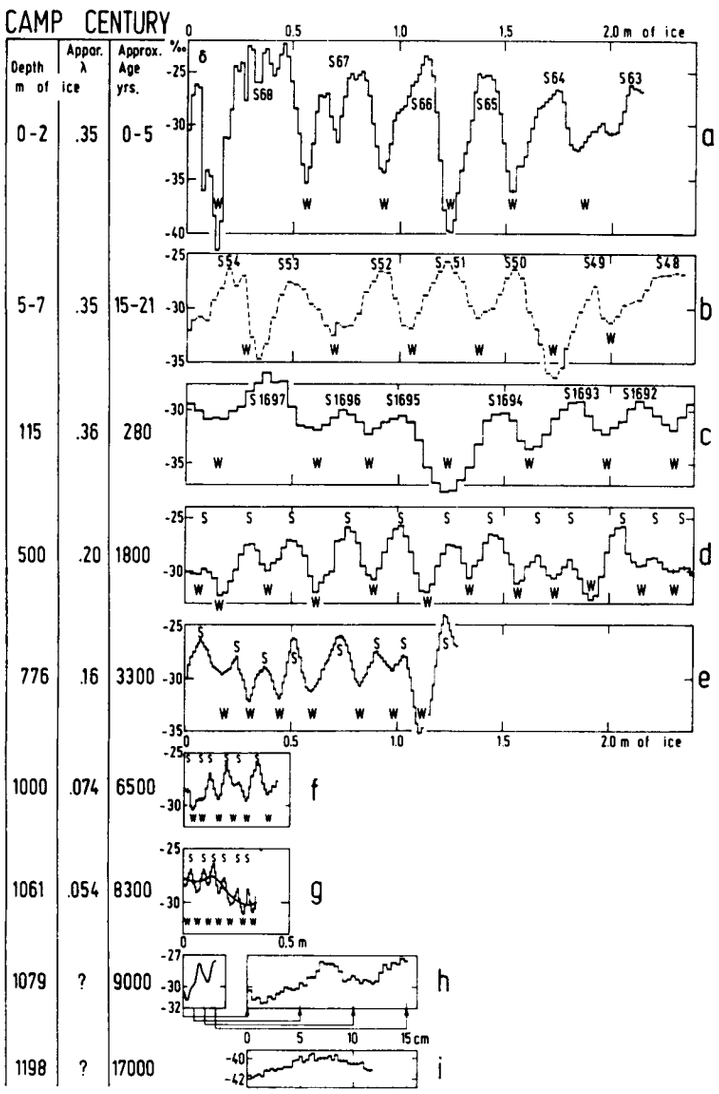
where Z_0 is the ice thickness at the centre of the ice sheet. This in turn implies that the vertical strain rate is uniform at all depths. Then, if the annual accumulation (A_0) has not changed with time (steady-state climate), the model indicates that the annual layer thickness λ should decrease uniformly with depth from the surface to zero at bedrock at the centre of an ice sheet. At a height z above bedrock we will have

$$\lambda = \bar{A}_0 \frac{z}{Z_0} \tag{1.4}$$

We show later in Figure 1.9 that this is approximately the case in the ice core from Camp Century.

We can continue to use our vertical column model away from the centre of an ice sheet. The form of ice sheets is discussed in detail in chapter 2, but for the present we need just note that ice flowing under gravity piles up into a vast, relatively flat dome. The surface slope α, which increases slowly from the centre to the periphery of the ice sheet, in approximately inverse proportion to the ice thickness, controls the flow of ice in much the same way as the slope of a river surface

Figure 1.7. The $\delta\,^{18}O$ oscillations in firn and ice cores from Camp Century at the depths below surface indicated to the left of the figure. S and W indicate interpretations of summer and winter layers, respectively. As the ice sinks towards the bottom, the thickness (λ m) of the annual layers is reduced due to deformation during flow. (From Johnsen *et al.*, 1972. Reproduced by permission of *Nature, London*.)



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determines the flow of water. Bedrock slopes β have important but relatively local effects on ice deformation and flow.

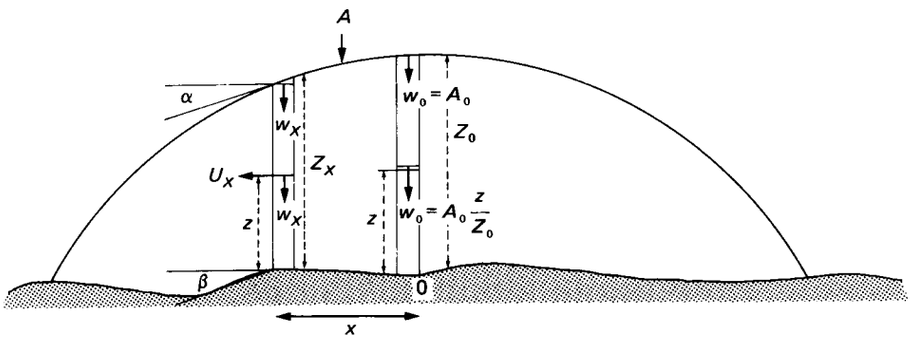
At a distance x from the centre, the two-dimensional steady state requires that the horizontal velocity of our column (U_x) be given by

$$U_x = \frac{\bar{A}x}{Z_x} \tag{1.5}$$

where $\bar{A}$ is the mean mass balance along the line x from the centre, and Z_x is the ice thickness at that point. The height of the column will also be changing due to changes of ice thickness at a rate

$$\frac{dZ_x}{dt} = (\alpha - \beta)U_x \tag{1.6}$$

Figure 1.8. Parameters for a two-dimensional cross-section of an ice sheet in a steady state.



Thus an annual layer thickness A_1 deposited at a distance x_1 from the centre of the ice sheet on ice of total depth Z_1, will have a thickness λ at a distance x from the centre defined by

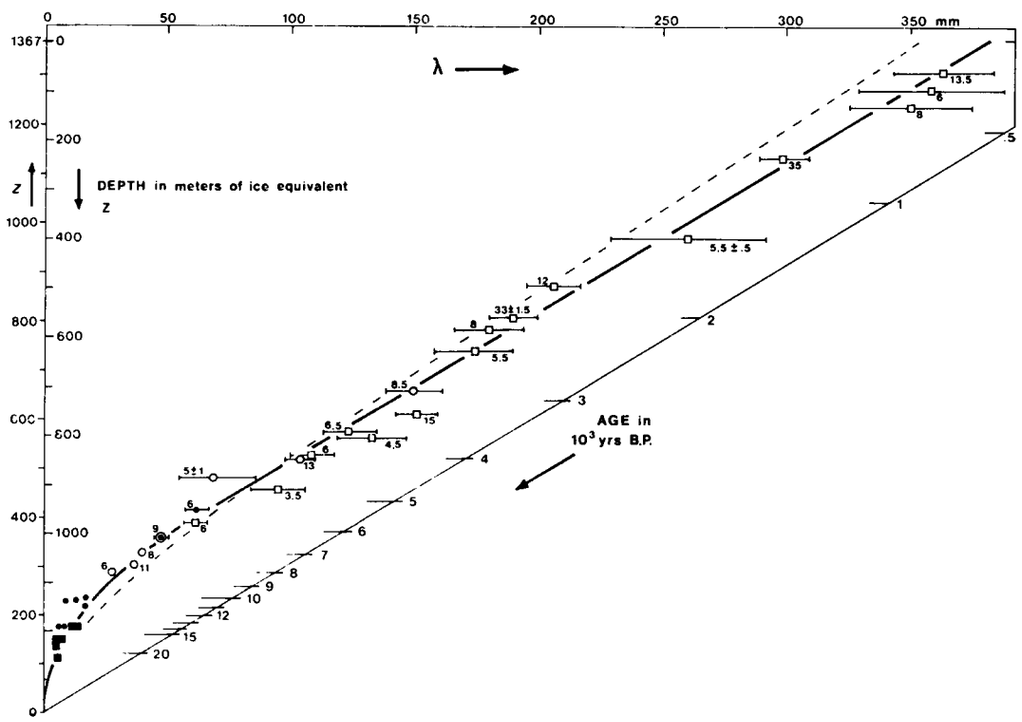
$$\lambda = A_1 \frac{z}{Z_1} \tag{1.7}$$

where z is its height relative to bedrock. The vertical velocity w will be

$$w = [A_x + (\alpha - \beta)U_x] \frac{z}{Z_x} \tag{1.8}$$

Thus the layer thicknesses on this model will still decrease with depth from A_x at the surface to zero at bedrock, but not linearly as at the centre of the ice sheet since A and U_x will vary over the ice sheet.

In Figure 1.9 (from Hammer *et al.*, 1978), layer thicknesses at Camp Century seen in Figure 1.7, together with depth to 400 m above bedrock, corresponding to a uniform vertical strain rate. Below this level the strain rate decreases linearly to zero at bedrock, which implies $\lambda \propto z^2$ over this interval. The time scale shown on the right is derived from the observed layer thickness. (Hammer *et al.*, 1978. Figure reproduced from the *Journal of Glaciology* by permission of the International Glaciological Society.)



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with other observations of layer thicknesses, are plotted against depth in the ice sheet. The heavy line fitted to the observations is identical (apart from a minor thickness change) to that used for the basic model for calculations in section 5.2. The assumption that the vertical strain rate falls to zero over the lowest 400 m takes account of the ice being frozen to bedrock. This would cause ice movement to fall to zero and hence reduce the vertical strain rate to zero also, since there will be no horizontal spreading to permit vertical strain at bedrock. The agreement between the heavy line and the observations is good down to a depth around 1100 m, that is to an age of approximately 9000 a. Below that depth the proportional difference between observed and calculated layer thicknesses increases and the dating appears less reliable.

When we consider the effect of horizontal velocity on vertical movement in equation (1.8) and its consequent effect on the thickness of layers transported from further inland, it is surprising that the calculated thicknesses fit the data in Figure 1.9 so well above 1100 m depth. The simplest explanation is that the accumulation rate recorded in the ice core has varied little over the past 9000 a and that the second term $(\alpha-\beta)U_x$, in equation (1.8) is negligible. Figure 4.6 shows little change in ice thickness for 20 km up the present flowline from Camp Century, that is $(\alpha-\beta)\approx 0$. If the mean velocity of $U_x < 2\ \mathrm{m\ a^{-1}}$, which seems possible along an ice divide, the ice at Camp Century down to 1000 m depth will have been deposited in this section of uniform thickness, and would be of relatively local origin.

In general, we use the term 'steady-state ice sheet' to describe an ideal model in which both the size and climate of an ice sheet do not change with time. These assumptions together with our model make it possible to calculate not only the velocity distribution, but also the temperature distribution throughout a large ice sheet. Using this model, we find rough agreement between calculated and observed temperature distributions in the same way as we find agreement between calculated and observed layer thickness (Figure 1.9). Refinements of these calculations to more exact steady-state models, and development of more realistic models that take account of past variations of size of the ice sheet are dealt with in chapter 5. In that chapter we try to determine just what these past variations have been by using these models and the evidence available from isotopic, temperature, and gas-content studies, as well as from glacial geology.

Temperature

A large proportion of the ice of glaciers in lower latitudes is at, or close to, the melting point, and these glaciers are referred to as 'temperate'. In contrast to temperate glaciers, on which there is considerable surface melting during summer months, polar glaciers by the glaciologists' definition are 'cold'. This defines a polar glacier or ice sheet as one on which the surface temperature does not reach the melting point at any time of the year. Temperature of the upper layers of a polar ice sheet are governed by the thermal conductivity of the firn and ice. Ten metres below the surface the seasonal temperature variations are almost damped out, and temperatures remain close to the mean annual temperature of the surface of the ice sheet. This surface temperature is usually a degree or so colder than the mean annual temperature recorded in a meteorological screen 1 m above the surface.

In Figures 4.1 and 4.3 we show the mean annual (10 m) temperature distribution of the surface layers of the ice sheets of Greenland and Antarctica. We see that temperature is primarily related to surface elevation, but that the latitude and continentality of a location are also factors. The latitude effect is seen more readily in the case of Greenland, while continentality is one of the factors that cause the extremely low temperatures in central Antarctica.

Direct knowledge of bedrock temperatures of Greenland and Antarctica is more limited. At Camp Century the bedrock temperature was −13 °C, while at Byrd Station water rose into the borehole indicating a basal temperature of −1.6 °C, the pressure melting point under 2164 m of ice. Radio-echo sounding of the Antarctic ice sheet has shown the presence of a number of lakes beneath 3–4 km of ice in central areas (Oswald & Robin, 1973), indicating pressure-melting temperatures of about −2 °C to −3 °C. So far, no lakes have been reported beneath the Greenland ice sheet, but the bedrock topography of the ice sheet is not particularly favourable for collection of water. In any case, theoretical predictions which were approximately correct for basal temperatures in the Camp Century and Byrd Station boreholes, indicate that the basal ice over large areas of Antarctica and Greenland will be frozen to the bedrock.

Heat flow, ice flow and temperature distribution in steady-state ice sheets

Ice sheets, like any part of the Earth's crust are affected by the flow of heat from the interior of the Earth. This heat flow does not vary greatly over the Earth's surface. On continents it averages around $28\ \mathrm{mW\ m^{-2}}$, or about $38\ \mathrm{cal\ cm^{-2}\ yr^{-1}}$, enough to melt about $\frac{1}{2}$ cm of ice in a year. It is slightly higher on the ocean bed, and generally higher in the vicinity of mid-ocean ridges. Heat flow is determined by measuring the temperature gradient in rocks, and multiplying this by the thermal conductivity of the rock. As would be expected, therefore, the variation in temperature gradient between different types of rock is much g eater than the variation in heat flow. Typical temperature gradients in soft sediments of the sea bed are $0.05\ ^\circ\mathrm{C\ m^{-1}}$; in limestone, $0.03\ ^\circ\mathrm{C\ m^{-1}}$; in granite, $0.02\ ^\circ\mathrm{C\ m^{-1}}$; and in ice (static), $0.025\ ^\circ\mathrm{C\ m^{-1}}$. The observed temperature gradients in the Earth's crust help us to deduce processes that may be taking place at greater depths in the crust and mantle of the Earth.

In polar ice sheets we have a similar problem, but with certain advantages. We have more knowledge of the deformation (e.g. Figure 1.9) and movement of the ice mass than we have in the case of the Earth's crust and interior. At two locations, Camp Century and Byrd

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Station, we have temperature measurements throughout the entire depth of thick ice sheets and many observations at lesser depths. Isotopic data may provide us with a record of past surface temperatures at which the ice at any depth in an ice core was laid down, even though the ice has moved far to reach the borehole site. The motion of the ice itself has such a large influence on the temperature distribution within the ice mass that we can use temperature gradients to determine approximately whether or not an ice sheet is in a steady state, both with regard to size and to temperature distribution. The mathematical and computational methods used in detailed studies are complex and are developed in chapter 5. In this section we aim to give the general reader some understanding of the processes and factors governing temperature distribution, so that he may appreciate the significance of deductions made about the size and motion of ice sheets, present and past, from temperature and isotopic profiles, even if he does not follow through the detailed computations in chapter 5. The simple vertical column model (Figure 1.8) is sufficient to describe the main features. The scale of the various factors involved, such as ice depths, heat flow and time, must be kept in mind.

While the upper-surface temperatures are of great importance in determining the temperature distribution in a polar ice sheet, the problem is best approached by first considering what happens to the geothermal heat entering the base of an ice sheet which is in steady state. If the ice mass is not in motion, then a steady-state condition is only possible if the amount of geothermal heat flowing upward through the ice mass does not change at any level. Otherwise there would be accumulation or loss of heat at levels where the heat flow changed and hence also changes of ice temperature. A uniform geothermal heat flow of 38 mW m^{-1}, then, requires a constant temperature gradient throughout the ice mass of 0.024 °C m^{-1} provided we regard the density, specific heat and thermal conductivity of ice as constants and neglect their small changes with temperature. Then, if the ice is 1000 m thick, the temperature at bedrock will be 24 °C warmer than at the surface (Figure 1.10). The latter is, however, determined by the surface climate, since the geothermal heat flux is very small in comparison with solar and atmospheric heating effects at the surface. Then, if the mean annual surface temperature is −30 °C, the bedrock temperature will be −6 °C. If the surface temperature is −24 °C, the bedrock will be at the melting point, 0 °C (neglecting effect of pressure). If the surface is warmer than −24 °C, say −12 °C for example, the bedrock temperature will be at the melting point and the mean temperature gradient through the ice can then be only 0.012 °C m^{-1}, which is only half the geothermal gradient. The excess heat reaching the base of the ice will not be conducted upward, and will melt about 2.5 mm per year off the base of the ice. Our ice mass is then no longer of constant thickness and, strictly speaking, the problem changes into one of an ice mass in motion, though in this case the change is small. If the surface temperature is 0 °C, the ice mass will all be at the melting point.

We now consider the effect of motion. Under the steady-state assumption, the temperature at any given height above bedrock will not change with time. However, our steady-state model also shows us that the cold, surface ice is moving downwards towards the warmer bedrock with a velocity given in equation (1.3) or (1.8). The steady-state temperature condition can only be met if the ice moving downwards is heated at the correct rate to maintain a constant temperature at each level. The main source of heat deep in the central regions of an ice sheet is the geothermal heat itself, which is therefore used and absorbed in heating the ice. However, after a given layer of ice has absorbed some of the geothermal heat entering its base, the heat flow which penetrates the next layer above will be decreased by the amount absorbed. The temperature gradient will then decrease in each successive layer as we move upward from the base. The effect is shown quantitatively to fit equation (1.3) for the centre of a steady-state ice sheet 3000 m thick with an accumulation rate of 0.32 m a^{-1} of ice in Figure 1.11. We see here that the upward flow of heat (shown by the length of upward arrows) decreases to zero by 2000 m above bedrock, so that the ice above this level is isothermal with the surface temperature.

Figure 1.10. Approximate temperature–depth profiles in a static ice mass 1000 m thick, neglecting effects of pressure melting, for the following cases:
(1) basal ice at 0 °C and entire heat flux carried upwards through ice;
(2) same heat flux as (1) with mean surface temperature −30 °C;
(3) mean surface ice temperature at −12 °C;
(4) mean surface ice temperature at 0 °C.

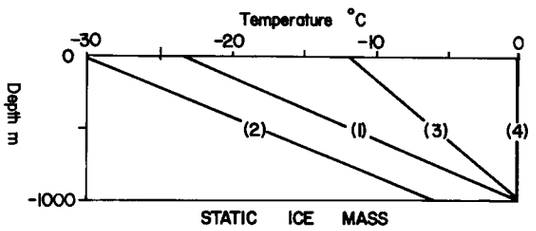
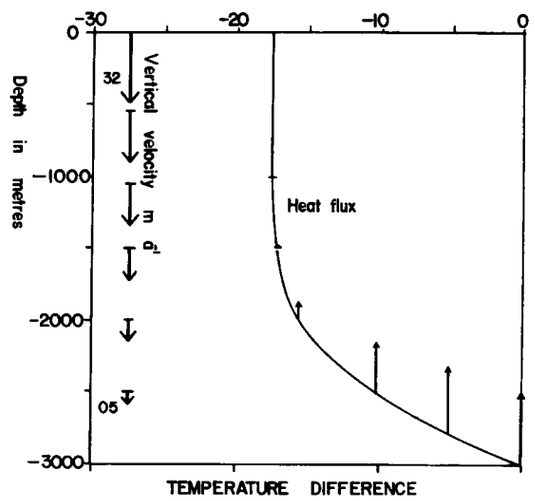


Figure 1.11. Temperature variation with depth in central region of an ice sheet in a steady state with 0.32 m a^{-1} accumulation of ice. The basal temperature gradient (geothermal heat flux) is the same as in Figure 1.10.



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As in the previous example, after calculating the shape of the temperature–height curve from the geothermal gradient and the ice motion, we tie the curve to the surface temperature to predict the actual temperature at any depth. In Figure 1.12 we show the form of the temperature–depth curves for a number of different rates of accumulation, using the same geothermal heat flow (i.e. basal-temperature gradient) in each case. We note that the greater the rate of accumulation, the less the temperature difference between surface and bedrock, and hence the colder will be the basal temperature for a given surface temperature. Although our model does not deal quantitatively with horizontal transport of heat due to the outward movement of our ice column (Figure 1.8), we can readily see what has happened to the geothermal heat reaching the base of the ice. This is best described by the meteorologists' term 'advection', the transport of heat by mass movement of the atmosphere or, in this case, the outward mass movement of ice. The heavier the accumulation, the faster will be the mass movement and the more rapidly will basal heat be carried away by horizontal motion of the ice sheet, so that less heat will flow to the upper layers of the ice sheet.

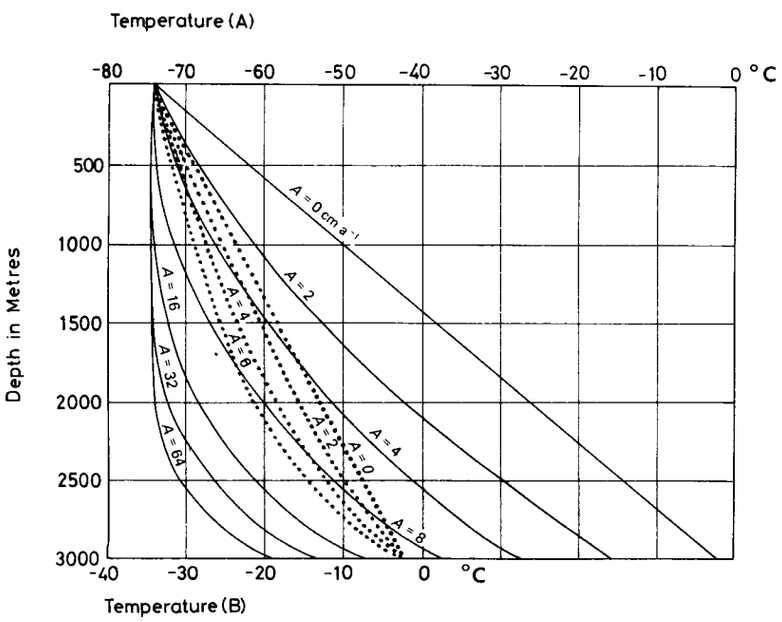
We have not yet dealt with the question of the heat produced by the work done in deforming an ice sheet under the action of gravity. The rate of heat production in any element of volume is proportional to the product of the effective stress by the effective strain rate (see section 5.4). The amount of heat involved due to the uniform vertical strain of our model is relatively small and does not greatly influence temperatures in the upper part of the ice sheets of Greenland and Antarctica. However, shearing or sliding of the ice over bedrock produces large strain rates in the lowest 10 or 20 per cent of the ice. One may regard this as frictional heat concentrated at bedrock, in which case the work done is equal to the product of ice movement by basal shear stress. Typical values of basal shear stress for ice sheets involves a heat production equal to the geothermal flux for ice movement of from 20–50 m a^{-1}. Strictly speaking, this heat production will be spread through the lower levels of ice where deformation is considerable, the actual distribution being more concentrated towards bedrock when this is at the melting point so that sliding takes place. This extra heat supply will increase the temperature gradients in the lower levels and will therefore generally increase the temperature difference between the surface and bedrock.

An effect that opposes the increased temperature differences of the previous paragraph comes from the thinning of ice and hence of our 'vertical column' as the ice moves outwards (Figure 1.8). This thinning involves vertical strain in the ice, in addition to that due to accumulation on our steady-state model, and its effect can be estimated from Figure 1.8 by adding the two effects together as indicated by equation (1.8). For example, an ice motion of 50 m a^{-1} over flat bedrock with a surface slope of 1 in 200 will add an extra downward motion of 0.25 m a^{-1} to the surface of a moving column in a steady-state ice sheet. The resultant effect on deep ice temperatures will also depend on the ice thickness and on accumulation rate, so will vary with location. The balance between the opposing effects of increased vertical strain due to ice thinning versus increased frictional heating due to faster ice motion must be calculated for any given situation.

Changing surface temperature with time

Even on our steady-state model in Figure 1.8, the surface temperature at the top of our moving column will not be constant except at the centre of the ice sheet, since as the column moves away from the centre, the surface elevation becomes lower and the mean

Figure 1.12. Calculated temperature–depth curves (as in Figure 1.11) for a surface temperature of −74 °C (upper scale A) and −34 °C (lower scale B) for different values of accumulation of ice shown in cm a^{-1}. The dotted curves relate to scale B only, since ice temperatures cannot be greater than the pressure melting point (−2.2 °C at 3000 m).



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