Journal of Glaciology, Vol. 35, No. 119, 1989

MODELLING ANTARCTIC FAST-ICE GROWTH

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ABSTRACT. An existing thermodynamic ice-growth model (Semtner, 1976) has been tested for its ability to predict the growth of fast ice in McMurdo Sound, Antarctica. Significant discrepancies between observed and predicted ice thicknesses were found to occur, primarily due to the presence of sub-ice platelets and the formation of a snow-ice layer. Although these ice-growth processes are not well enough understood to permit rigorous physical modelling, it is shown that fairly simple modifications to the model greatly improve the accuracy of the thickness predictions, and serve to highlight the importance of these processes in the Antarctic fast-ice environment. Surface flooding and snow-ice formation are assumed to occur immediately upon the establishment of a positive hydrostatic water level, and a surface temperature in excess of a critical value, above which interconnecting channels in the ice matrix permit the flow of water to the surface. The presence of the sub-ice platelet layer is assumed to increase columnar ice growth at a rate proportional to the volume fraction of ice in the platelet layer, a simple technique but one that permits estimates of platelet-enhanced growth without detailed knowledge of oceanographic conditions. The resulting model predictions are in close agreement with measurements of fast-ice growth and decay in McMurdo Sound; however, data suitable for testing the model over a complete range of conditions and over multi-year cycles are not available at the present time.

INTRODUCTION

Numerous thermodynamic models have been developed to predict the growth of Arctic sea ice, ranging in complexity from simple formulae based on freezing degree days (Stefan, 1889; Nakawo and Sinha, 1981) to rigorous treatments of the physics governing energy transfers between the ocean and the atmosphere (Maykut and Untersteiner, 1971; Gabison, 1987). Measurements of Arctic ice growth are also abundant and, in general, good association between observed and predicted ice thicknesses can be obtained. It may seem rational, therefore, to suppose that if the physical processes responsible for transferring energy through the ocean–ice–atmosphere system have been adequately accounted for, then these models will be equally effective at predicting the growth of Antarctic sea ice. However, most models contain parameters, such as albedo and the turbulent-energy transfer coefficients, which are site specific. More importantly, there is always the possibility that processes neglected as being either insignificant or absent in one region make important contributions to ice growth in another. In short, it is dangerous to presume *a priori* that a thermodynamic model which has been proven to predict accurately ice growth in the Arctic will accurately predict ice growth in the Antarctic.

In this paper an existing thermodynamic ice-growth model will be tested for its ability to predict the growth of fast ice in McMurdo Sound, Antarctica, and modified to represent better the special environmental conditions found there. Although developed specifically for McMurdo Sound, and therefore encumbered with some of the problems of regionally based models mentioned above, the resulting formulation will, it is hoped, be better suited to the general problem of Antarctic fast-ice growth than those currently available.

MODEL DESCRIPTION

Since there exists very little detailed atmospheric and oceanographic information of a form suitable as input data in an Antarctic thermodynamic model, the first criterion that our formula must meet is simplicity. Since energy exchanges between the ice and the ocean strongly affect the growth of Antarctic ice (Allison, 1979), the model must also include a simple oceanic heat-flux term. An existing model which meets both of these criteria, and which has the added advantage of being well suited to the modifications required for improved representation of the physical processes influencing Antarctic ice growth, is the "zero layer" model of Semtner (1976). Like several more complex models (Maykut and Untersteiner, 1971; Gabison, 1987), Semtner's formula relies on a calculated equilibrium surface temperature, from which all the necessary surface fluxes can be determined. This is accomplished by assuming that the flux of energy to the surface through the snow/ice slab (F_0) is equal to the sum of all the fluxes transferring energy from the surface to the atmosphere (F_a), that is

$$F_0 + F_a = 0 \tag{1}$$

where an upward flux is considered positive. For bare ice, F_0 is calculated using

$$F_0 = \frac{k_i(T_b - T_0)}{h_i} \tag{2}$$

and for snow-covered ice

$$F_0 = \frac{k_s(T_b - T_0)}{h_s + (h_i k_s/k_i)} \tag{3}$$

where k is thermal conductivity, h is thickness, the subscripts s and i refer to the snow and ice, respectively, T_b is the temperature at the ice/water interface, and T_0 is the surface temperature. The flux to the atmosphere is

$$F_a = F_s + F_\ell + F_{Ld} + (1 - \alpha)F_r + \varepsilon\sigma T_0^4 \tag{4}$$

where the subscripts s, ℓ, Ld, and r refer to fluxes of sensible, latent, incoming long-wave radiation, and incoming short-wave radiation, respectively, α is the surface albedo, ε is the surface emissivity, and σ is the Stefan–Boltzmann constant. In this simplified model, it is assumed that the fraction of incoming solar radiation which penetrates the snow or ice surface (I_0) is zero. This has the effect of increasing the predicted surface temperature, and with the imposed linear temperature profile through the ice, results in a slight underprediction of ice growth. Semtner (1976) corrected this by increasing k_i and k_s by a factor of 1.065, a procedure which has been adopted here.

The first four terms on the right side of Equation (4) must be supplied as input data, allowing T_0 to be determined by a Newton–Raphson or similar iterative process. By substituting T_0 into Equations (2) (or (3)) and (4), the change in thickness at the base of the ice sheet (Δh_b) can then be calculated using

$$\Delta h_b = \Delta t(F_0 - F_b)/q_b \tag{5}$$

where Δt is the time increment, F_b is the oceanic heat-flux term (also explicitly defined), and q_b is the volumetric heat of fusion at the bottom of the ice cover.

If the surface temperature is found to be above the freezing point, then surface melting is determined by computing F_a and F_0 for $T_s = 0$, and setting

$$\Delta h_s = \Delta t(F_a - F_0)/q_s \tag{6}$$

for snow-covered ice, and

$$\Delta h_i = \Delta t(F_a - F_0)/q_i \tag{7}$$

for bare ice, where q_s and q_i are the volumetric heats of fusion for snow and ice, respectively.

However, in this form the model is difficult to apply to Antarctic fast-ice growth. Observations of ice growth in McMurdo Sound during the austral winter of 1986 (Crocker, unpublished) showed two additional processes to be making significant contributions to ice growth: the formation of snow-ice, and the formation of a sub-ice platelet layer.

Snow-ice

In near-shore areas, where snow blown off the Antarctic continent accumulates in large quantities, the resulting overburden pressure can lead to flooding of the ice surface and the formation of snow-ice. It has been observed in some coastal zones that snow-ice is the main constituent of the fast-ice sheet (Kozlovskiy, 1981). Snow-ice has also been found in abundance in the winter pack-ice cover, where high oceanic heat fluxes restrict the annual ice growth to 0.6–0.8 m, and flooding is possible on a large scale even though the snow accumulations are less (Wadhams and others, 1987). If estimates of the snow and ice densities are available, then the equilibrium hydrostatic water level (H_w) can easily be calculated using

$$H_w = \frac{\rho_i}{\rho_w}h_s - \frac{\rho_i}{\rho_s}h_i\left[1 - \frac{\rho_i}{\rho_w}\right] \tag{8}$$

where ρ is density, h is thickness, and the subscripts i, s, and w refer to ice, snow, and water, respectively (Röthlisberger, 1963). Determining exactly when, or if, flooding will occur is considerably more difficult. A reasonable assumption might be that surface flooding will take place as soon as H_w becomes positive. If, as observations indicate, flooding occurs in discrete events but eventually reaches equilibrium levels, then although the timing would be incorrect, the resulting error in the growth calculations would be small. However, the McMurdo Sound observations throw some doubt on the validity of this assumption. Throughout the winter, drill holes revealed consistently positive hydrostatic water levels, but there was a conspicuous absence of large-scale flooding. Because sea ice normally does not crack under thermally induced stress, in order for positive H_w values to be relieved flow must take place through interconnecting channels in the ice. Pounder (1965) suggested that the transition between interconnected and discontinuous brine-drainage channels occurs at a temperature of roughly −15°C. This figure will be highly dependent upon the salinity and crystal structure of the ice, and measurements of ice-surface temperatures in McMurdo Sound during several flooding events showed that a value of −8°C is more appropriate in this specific case. We therefore have two criteria which must be satisfied before flooding can occur: positive hydrostatic water levels, and surface temperatures greater than or equal to −8°C. It is recognized that imposing a constant minimum temperature for flooding is a crude approximation, but very little is known about the relationships between the temperature, salinity, and crystal structure of sea ice and its permeability and, until such information becomes available, the threshold temperature criterion must suffice.

It is assumed that the transport of liquid to the surface has no effect on the thermodynamics of the ice sheet, apart from the obvious effects of changes in h_s and the thickness of the snow-ice layer (h_{si}). Spot measurements made on the ice immediately in front of Scott Base in March 1986 showed the flooding liquid to be high-salinity brine (≈125 ppt), at roughly the same temperature as the snow into which it was flowing (indicating that flooding takes place through the brine-drainage network rather than through stress fractures, and that the liquid in the brine pockets is pushed to the surface ahead of the ocean water below). The snow/liquid mixture produced is therefore highly saline (about 70 ppt), but the temperature profile through the ice remains virtually unchanged, and the amount of advected heat energy is quite small. We have also assumed that the amount of latent heat released as the flooded layer solidifies is negligible. This is of course not strictly true, but McMurdo Sound observations suggest that the amount of latent heat released from the flooded layer is small. Since the temperature and salinity of the flooding liquid do not change when mixed with the snow, the mixture will be in a state of chemical equilibrium for that temperature. The observed salinity of 70 ppt, and temperature of −8°C, indicates that the brine volume in the mixture was about 465 ppt (Frankenstein and Garner, 1967), and therefore the volume of ice was roughly 500 ppt, slightly greater than the 350 ppt occupied by the ice grains in the original snow layer. Since the temperature did not change, most of this increase must be due to compaction of the snow on flooding. In fact, in the absence of any significant decrease in temperature, most of the solidification of the snow-ice layer must be the result of compaction of the snow in the presence of the liquid. Although reductions in the temperature of the flooded layer will result in some freezing of the flooded liquid, the deep snow cover which remains over the "snow-ice" layer acts to moderate atmospheric temperature fluctuations, and in this case the observed brine-volume decrease resulted primarily from snow compaction and drainage of the interstitial fluid after the initial flooding event (the salinity of the snow-ice layer was 70 ppt on 1 March, 44 ppt on 30 March, and 15 ppt several months later when the sample shown in Figure 1 was taken). It is not known to what extent freezing of the flooding liquid contributes to snow-ice formation on a regional scale (this subject should be investigated further); however, we feel that the errors introduced into this simple type of ice-growth model by neglecting the latent heat released from the snow-ice layer will be small.

Equation (8) is based on the assumption that the flooded layer becomes completely saturated. This is clearly not true, as close examination of the ice revealed numerous small air pockets trapped in the ice matrix (see Fig. 1). Qualitative observations of the snow-ice in McMurdo Sound show that its air-bubble content is similar to that of snow-ice formed on lakes. Adams (1981) has measured the density of lake snow-ice and we adopt his value for ρ_{si} of 820 kg m^{-3} (corresponding to an air-bubble content of 10%), and adjust the calculated hydrostatic water levels by increasing H_w uniformly by 10%. Also, because both brine and air are poorer conductors than pure ice, the low density and high salinity of the snow-ice layer will combine to reduce its thermal conductivity. To estimate the magnitude of this reduction, we use Schwerdtfeger's (1963) model for the thermal conductivity of bubbly, saline ice as representative of the thermal conductivity of snow-ice (k_{si}). Then k_{si} can be calculated using

$$k_{si} = k_{bi} - (k_{bi} - k_b)\frac{S\rho_{si}}{\mu\rho_w T} \tag{9}$$

where S is the ice salinity in parts per thousand, T is the ice temperature in degrees Celsius, μ is a constant (−0.0182 °C^{-1}), k_b is the thermal conductivity of brine

$$k_b = 0.4184(1.25 + 0.03T + 0.00014T^2), \tag{10}$$

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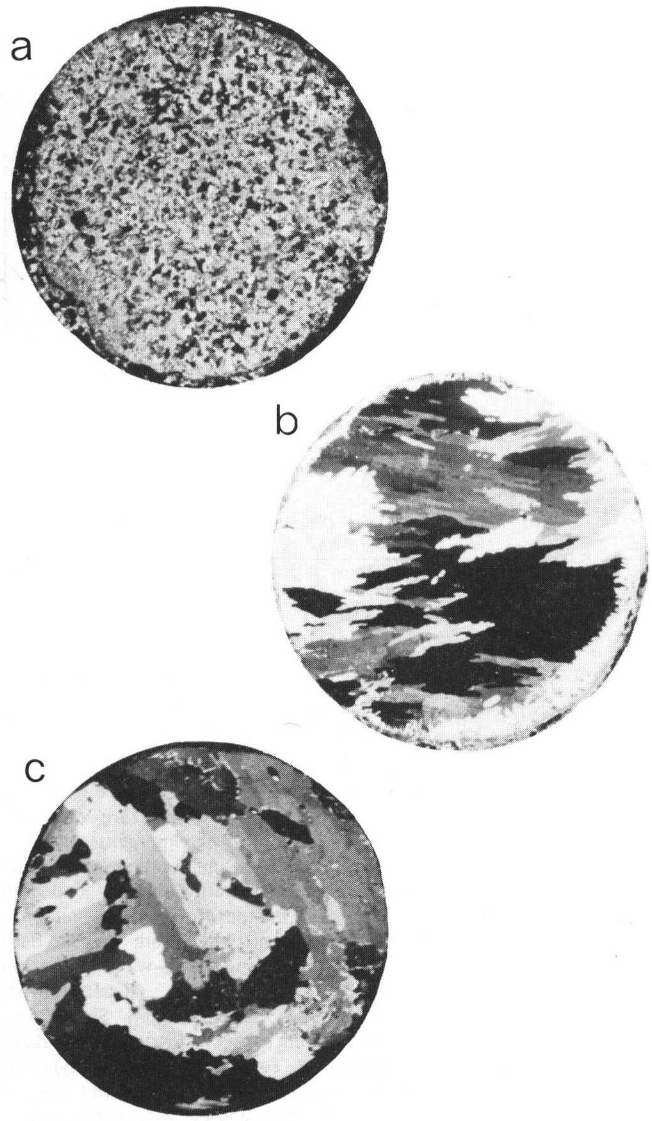

Fig. 1. Vertical crossed-polaroid photographs of (a) snow-ice, (b) columnar ice, and (c) ice formed by columnar growth into a sub-ice platelet layer. All three thin sections are from cores taken in McMurdo Sound during the winter of 1986; (a) and (c) from roughly 1.5 km offshore from Pram Point, and (b) about 100 m from shore in North Bay. Each section is 76 mm in diameter.

k_{bi} is the thermal conductivity of bubbly ice

$$k_{bi} = \frac{2k_0 + k_a - 2V_a(k_0 - k_a)}{2k_0 + k_a - V_a(k_0 - k_a)}, \qquad (11)$$

k_a is the thermal conductivity of air, V_a is the volume fraction of air, and k_0 is the thermal conductivity of pure ice (Schwerdtfeger, 1963). At $-10\,^{\circ}\mathrm{C}$ and a salinity of 20 ppt, Equation (9) gives $k_{si} = 1.66\ \mathrm{W\,m^{-1}\,K^{-1}}$ which is about 10% less than a value of k_i for "normal" sea ice calculated using Untersteiner's (1961) formula

$$k_i = k_0 + \frac{\beta S}{T} \qquad (12)$$

in which β is a constant ($0.117\ \mathrm{W\,m^2\,kg^{-1}}$), and S and T are again in parts per thousand and degrees Celsius. For simplicity, we therefore set k_{si} to $0.9 \times k_i$ throughout the calculations. An additional alteration must then be made to accommodate the three-layer (ice, snow-ice, and snow) system of series-connected conductors. This is accomplished by replacing Equations (2) and (3) with

$$F_0 = \frac{(T_b - T_0)}{\dfrac{h_i}{k_i} + \dfrac{h_{si}}{k_{si}} + \dfrac{h_s}{k_s}} \qquad (13)$$

(see Leppäranta, 1983).

Sub-ice platelets

A second modification which must be made to the model is to incorporate the contribution to growth of the sub-ice platelets; discoid, single crystals of ice formed when supercooled water comes in contact with the base of the ice sheet. They normally take the shape of plates roughly 5 mm thick and ranging in diameter from less than 10 mm to over 150 mm. Supercooled water and sub-ice platelets have been observed several times near ice-shelf margins in the Antarctic (Kozlovskiy and Cherepanov, 1976; Foldvik and others, 1985), and the general mechanism of formation has been described by Robin (1979). Although tidally modified, the currents in McMurdo Sound are principally from under the McMurdo Ice Shelf (Lewis and Perkin, 1985) and therefore fit Robin's explanation. The water flowing into McMurdo Sound originates in the Ross Sea and flows under the Ross Ice Shelf at roughly its mid-point (Jacobs and others, 1979; Pillsbury and Jacobs, 1985). As the water moves southward, it is forced to greater depths by the overlying ice shelf, and may exceed the pressure-reduced melting point at the ice-shelf bottom even if its initial temperature was equal to the surface freezing point. The heat consumed in melting basal ice is drawn from the water, and reduces its temperature. If this heat loss is sufficient to lower the water temperature to below the *surface* freezing point, then on the return journey northward into McMurdo Sound it will become slightly supercooled as it rises to the surface and the pressure is released. The supercooling is relieved when the water comes in contact with the bottom of the ice sheet and nucleation results in the formation of a sub-ice platelet layer. The platelets will form under both sea and shelf ice, wherever a sufficient pressure reduction has occurred to produce super-cooling, but the greatest quantities will be produced where the pressure is lowest, that is at the base of the sea-ice sheet. It is also possible for sub-ice platelets to form at depth (Dieckmann and others, 1986), in which case they would simply float upward and accumulate on the underside of the ice, but there was no evidence to suggest that nucleation was occurring in the water column in McMurdo Sound during the winter of 1986.

It is interesting to see that the thickest platelet layers measured during two transects across McMurdo Sound on 4 August and 17 October correspond to the areas of maximum surface supercooling recorded by Lewis and Perkin (1985). The timing of the platelets' first appearance and growth is also significant. The thickness of the platelet layer was observed to be increasing from early July, when it first appeared, until about mid-September, after which it remained fairly constant. Wright and Priestley (1922) found that platelets grew on their manilla lines between 5 August and 12 October 1911, and between 28 August and 28 September 1912. The later dates of first appearance of the crystals may be the result of the greater distance of their measurement site (which was roughly half-way between Cape Evans and Inaccessible Island (Deacon, 1975)) from the ice-shelf edge; nevertheless, their observations suggest that the phenomenon occurs at about the same time each year. This can be readily explained by referring to Robin's (1979) theory. The water rising out from under the McMurdo Ice Shelf will be *supercooled* only if it were initially near its freezing point as it flowed under the Ross Ice Shelf. For supercooling (as opposed to merely cooling) to occur in the return flow, the water must lose enough heat through basal melting to have its temperature reduced to below the surface freezing point. The absence of platelet growth until mid-winter, and a reasonable estimate of the residence time of water under the ice shelf of 2–3 months, suggests that the water in the Ross Sea is not sufficiently cooled until early winter (roughly the month of May). There is no evidence to indicate that, as Wright and Priestley (1922) hypothesized, changes in the currents themselves are

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responsible for the timing of the phenomenon, although it cannot be ruled out as a possible contributing factor.

Considering the thermodynamics of platelet growth on a regional scale, if we assume that a platelet layer 1.0 m thick, consisting of 50% ice and 50% sea-water, forms under the ice cover over a 60 d period, and that the layer extends horizontally away from the ice-shelf edge a distance of 20 km (figures based on measurements during the 1986 season), then 9×10^6 kg of ice will be formed per metre width of ice shelf. Assuming that the "active" water column extends to a depth of 100 m (Lewis and Perkin, 1985) and the average rate of flow out from under the ice shelf is 0.10 m s^{-1} (Gilmour and others, 1962; Tressler and Ommundsen, 1962; Heath, 1977), then 5×10^7 m^3 of water per unit width will be acting to form the platelet layer over the 60 d period. Given that $L_i = 2.5 \times 10^5$ J kg^{-1}, and $c_p = 4.2 \times 10^6$ J m^{-3} $°C^{-1}$, then the 2.25×10^{12} J m^{-1} released on freezing corresponds to a mean supercooling in the active layer of 0.011 °C, which is of the same order as, though slightly lower than, the values of surface supercooling of 0.03 °C observed by Lewis and Perkin (1985). Using the same assumptions, the ice-growth rate would be 9.6×10^{-5} kg s^{-1} m^{-2}, giving a mean oceanic heat flux of −24 W m^{-2}. However, because only a part of the ice formed in this manner actually becomes incorporated into the ice sheet (see Fig. 1), the remainder forming an unconsolidated mass of platelets beneath the ice which melts when the water column warms in spring, we cannot simply set F_b to −24 W m^{-2}. The influence of the platelets on the growth of the solid ice sheet is therefore as dependent on the conductive flux of heat through the slab, and the columnar growth which results, as it is on the growth of the platelets themselves. A more realistic solution to the platelet problem might be simply to increase the columnar ice-growth rate during the period that the platelets are present by an amount proportional to the volume of the platelet layer occupied by ice. For example, if the platelet layer is 50% ice, then Δh_B would be multiplied by two each day the platelet layer is present. This is of course a great over-simplification because it ignores the effects of energy and salt transfers due to the formation of the platelets, and the physical changes to the ice bottom due to their presence. It has the advantages of being both simple and requiring no detailed information about the *rate* of platelet growth.

Turbulent fluxes

A third, but relatively minor, change from the Semtner model is the calculation, rather than explicit definition, of the turbulent fluxes F_s and F_ℓ. This is achieved using the aerodynamic formulae

$$F_s = \rho_a c_{pa} C_s u (T_a - T_0) \tag{14}$$

and

$$F_\ell = 0.622 \rho_a C_\ell u (f e_{sa} - e_{s0})/p_0. \tag{15}$$

Here ρ_a is the density of air, c_{pa} is the specific heat of air, u is the mean wind speed, C_s and C_ℓ are the sensible and latent energy-transfer coefficients (1.75×10^{-3}; (Overland, 1985)), f is the relative humidity, p_0 is the surface pressure, and e_{sa} and e_{s0} are the saturation vapour pressures of the air and at the surface, respectively. Maykut's (1978) fourth-order polynomial approximation is used to estimate e_{sa} and e_{s0} from the air and surface temperatures, and f and p_0 are assumed to be constant, with typical Antarctic values of 0.6 and 1013 mbar, respectively.

RESULTS AND DISCUSSION

Predicted and observed ice growth

Figure 2 shows measured ice-core lengths from McMurdo Sound along with columnar and snow-ice thicknesses predicted by the original and modified versions of the Semtner model, after setting the initial conditions to observed values. Although many more ice-thickness measurements were made, difficulties in accurately determining the base of the ice sheet in the presence of a sub-ice platelet layer caused an unacceptable degree of uncertainty in the values of h_i, and only the core data can be considered sufficiently reliable for the present purpose. Input data for T_a, u, F_r, and F_{Ld} have been modelled using Fourier transforms of mean monthly data collected from a variety of sources. Air temperature, wind velocity, and incoming short-wave radiation are from Scott Base meteorological records, while incoming long-wave radiation is a synthesis of Antarctic measurements (Rusin, 1961; Thompson and MacDonald, 1962, Zaitseva and Shlyakhov, 1984) and values estimated from observed air temperature and cloud data, using empirical formulae (Oke, 1978). On the basis of observations, α, S, and ρ_s have been set at constant values of 0.75, 6 ppt, and 350 kg m^{-3}, respectively. Changes in snow depth are represented by a linear increase of 0.35 m during the month of May, and held constant for the remainder of the year, while the thermal conductivity of the snow was calculated using the empirical formula of Abel's (1893). The timing and quantity of sub-ice platelet growth were also based on measured values, and the density of the ice was fixed at 910 kg m^{-3} (Anderson, 1960).

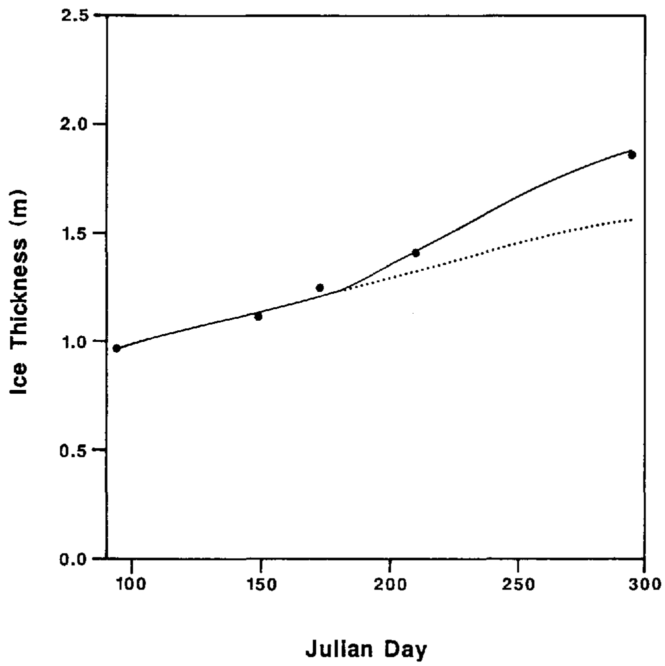

Fig. 2. Ice thicknesses predicted by the original (dotted line) and modified (solid line) version of the Semtner model, along with five measured ice-core lengths (large dots).

The agreement between the observed ice thicknesses and those predicted by the modified version of the model (solid line) is quite good, while the original model significantly under-predicts ice growth. Similar under-predictions resulted when the models of Maykut (1978) and Nakawo and Sinha (1981) were run with the same input data (Crocker, unpublished). As there was no flooding or snow-ice formation during the observation period (and none was predicted by the model), the improved predictions of the modified model are solely a result of including the influence of the sub-ice platelet layer on ice growth. Unfortunately, there are no data currently available which can be used to test the flooding and snow-ice growth parts of the model. Some isolated flooding events were observed at the end of the measurement period in mid-October, when the ice-surface temperature approached the prescribed threshold, but our treatment of the snow-ice growth process remains for the time being unproven. Bearing this in mind, it is nevertheless interesting to look at the patterns of ice growth and decay predicted by the model over the course of several seasonal cycles.

Multi-year projections

In order to extend the thickness predictions through the summer ablation period, we must first define a term accounting for basal melting. From measured summer oceanic heat fluxes (Mitchell and Bye, 1985), summer melting is simulated by setting F_0 to a constant value of +70 W m^{-2} through the months of November, December, January, and February. This heat energy is used first to melt the sub-ice platelet layer, and begins to reduce the thickness of the ice sheet only when all the platelets have been removed. An indication that this rather crude technique is at least generally representative of actual conditions can be found in the observation that both model predictions and measurements (Paige and Lee, 1967) of the timing and magnitude of ice-sheet ablation coincide quite closely.

Figure 3 shows forecast growth and decay cycles over a 5 year period. Snow-ice does not form in appreciable quantities until the second summer, when a deep snow pack, the reduced winter growth accompanying it, and

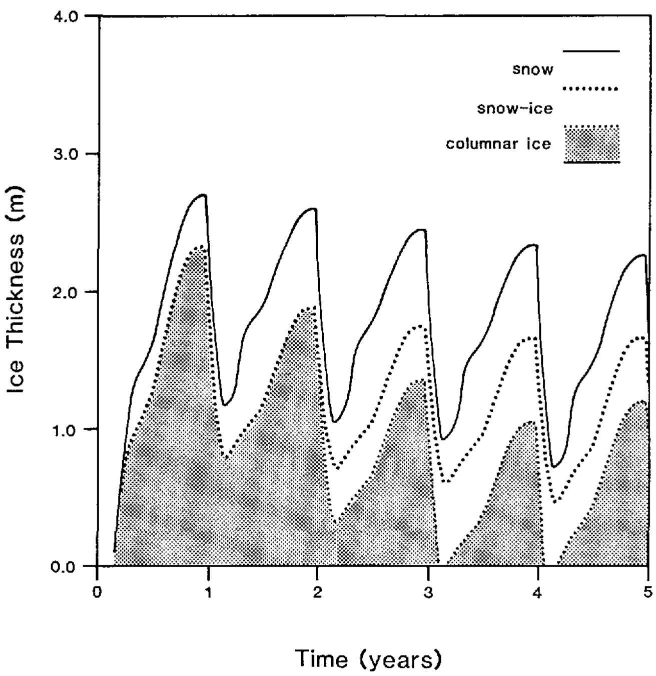

Fig. 3. Predicted growth and decay cycles for snow, snow-ice, and columnar ice in McMurdo Sound over a 5 year period. It is assumed that freeze-up occurs on 1 March in year 1.

basal melt during the summer, combine to produce the necessary conditions of H_w and T_0. There is a consistent reduction in the total ice thickness with time after the first winter, as the increased snow depth, and to a lesser extent the increased snow-ice depth, reduce the flow of heat through the slab. After 5 years, equilibrium conditions are reached, with snow-ice constituting just over 25% of the total thickness at the end of winter peak, and 100% of the total during the summer minimum. The predicted lack of surface melting, a product of the relatively strong winds and dry air prevalent during the summer (Andreas and Ackley, 1982), is consistent with observations, and has been found to be characteristic of Antarctic sea ice. The general pattern and magnitude of the thickness oscillations are also in agreement with observations from the multi-year fast ice in the southern end of McMurdo Sound, although it is difficult to assess the snow-ice predictions without corresponding snow-depth data. As a result, any assessment of the influence of different snow-accumulation regimes on the growth characteristics of the ice can only be considered tentative, but several features of the analysis are noteworthy.

Since snow deposition is normally greatest near the coast, and decreases markedly with distance from shore, the range of snow conditions likely to be found in McMurdo Sound can be modelled by varying the snow accumulation from 0.35 m a^{-1} to extremes of 1.00 and 0.05 m a^{-1}. With a deep snow-pack, true equilibrium conditions are not realized; rather a repeating 3 year cycle is established in which h_i progressively increases and h_{si} is progressively reduced, but on average snow-ice thicknesses are much greater than with 0.35 m of snow per year. Annual accumulations of 0.05 m, on the other hand, result in large initial columnar ice thicknesses (≈3.0 m) and a complete absence of snow-ice until the eighth winter, with an equilibrium not being established for over 20 years. When equilibrium conditions are reached, however, the total ice thickness at the end of winter maximum is only 1.85 m, 0.40 m of which is snow-ice.

Also, in the absence of surface melting, the equilibrium thickness is controlled by the summer oceanic heat flux. For example, reducing F_b by 75% produces a predicted equilibrium thickness (with 0.35 m a^{-1} of snow) of over 10 m, providing a possible explanation for reports of extremely thick, old, sea ice near the edge of the McMurdo Ice Shelf (Hoffman and Stehle, 1963). Unfortunately, crystallographic observations from this or other multi-year ice sheets are not available, and the predicted proportions of columnar and snow-ice cannot be verified.

SUMMARY

The growth of fast ice in McMurdo Sound is strongly influenced by the presence of a sub-ice platelet layer, and by surface flooding under heavy snow loads. The importance of these phenomena, and their effects on ice growth and thermodynamics has been discussed, and it is shown that the thermodynamic ice-growth model of Semtner (1976) can be used to predict ice-thickness changes when modified slightly to account for the presence of the platelets and the formation of snow-ice. The modifications applied here are quite crude, however the phenomena themselves are poorly understood, and at the present time more sophisticated attempts to physically model them are not possible. Agreement between observed and predicted ice thicknesses is good, both before and after a sub-ice platelet layer has formed, but there are still insufficient data to test fully the snow-ice growth predictions.

The model can also be used to predict ice growth and decay over several seasonal cycles, and yields some interesting patterns which should be investigated in future field studies. The model modifications have been aimed specifically at describing growth processes in McMurdo Sound. However, it is hoped that an improved representation of Antarctic fast-ice growth in general has been achieved and, perhaps more importantly, that we have brought to light some of the special processes associated with the growth of Antarctic fast-ice sheets, processes which must be accounted for if more physically rigorous Antarctic ice-growth models are to be developed.

ACKNOWLEDGEMENTS

This project was supported by the U.K. Natural Environment Research Council under grant GR/5713A. G.B.C. was supported in Cambridge by the Natural Sciences and Engineering Research Council of Canada, and in the Antarctic by the Antarctic Division of the New Zealand Department of Scientific and Industrial Research. The assistance of the 1986 winter-over crew at Scott Base is gratefully acknowledged.

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MS. received 18 January 1988 and in revised form 1 September 1988