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Ice Pumps and Their Rates

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An ice pump is a heat engine, driven by the change of freezing point with pressure, which will melt ice at depth in the ocean and deposit it at a shallower location: it is self-starting. Calculations of the maximum magnitude of this effect are made which show good agreement with field data available for sea and lake ice. The discussion is applied to the general case of a moving pack ice sheet with a well-mixed surface layer and to floating ice shelves. The rate of melt from an 11-m-deep pressure ridge keel due to ice pumping is estimated as 26 cm/year, and that from the front of the Ross Ice Shelf at McMurdo Sound, Antarctica is estimated as 5 m/year for the level of water movement noted in the authors' field observations. Far from the ice front, pumping between shelf areas of different thickness will still occur, with tidal motion providing the necessary water exchange, but its magnitude is now limited by the ability to remove the potentially stable layer of melt water out of the system. It is important to realize that the pumping does not depend on the availability of sensible heat in the water column and its effects are additional to any melting caused by the advection of warmer water to the ice-water interface.

1. INTRODUCTION

To the best of our knowledge the first mention of the potential for ice growth due to supercooling of waters in the immediate vicinity of an ice sheet was by *Untersteiner and Sommerfeld* [1964]. They measured supercooling of a few millidegrees in the vicinity of the ice island T3 floating in the Arctic Ocean and noted an accretion of ice onto existing crystals in a tray suspended beneath the sea ice sheet. The mechanism for production of this supercooled water was further explored by *Foldvik and Kvinge* [1974] in their study of oceanographic conditions near the Filchner Ice Shelf, Antarctica. They pointed out that if sea water that had been in contact with the shelf ice at depth was raised, then as the pressure reduced, it became supercooled in relation to the in situ freezing point and could provide a heat sink for ice growth within the water column. The converse argument yields the result that waters descending, after having been in contact with surface ice, will have sensible heat available to cause the melting of ice at depth. This constitutes an "ice pump" as described by *Lewis and Perkin* [1983], which is a heat engine working between different depths of an ice-water interface driven by the change in freezing point as a function of pressure, $\partial T_f/\partial p$, which has a value of 7.53×10^{-3}°C per bar (1 bar = 10^5 Pa and is the pressure at approximately 10-m depth). Lewis and Perkin applied the concept to freezing-melting processes under the pack ice–covered Arctic Ocean. In that case, downward water movement originating at a lead, where ice growth was taking place owing to heat loss into the atmosphere, could cause melting from a pressure ridge keel, and the corresponding upwelling of water from keel depth causes deposition of additional ice within the lead. They determined temperature profiles beneath frozen leads in the pack ice and found a surface layer isohaline and isothermal to at least 30-m depth where temperatures averaged 0.006°C below the surface freezing point. This well-mixed surface layer thus has the average freezing temperature of a 16-m-thick layer of water under the ice sheet.

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The same concept was applied by *Robin* [1979] to melting and deposition on the lower surface of a floating ice shelf where advective water movement will cause melting from deep ice with deposition at sites of shallow draft. Melting would be augmented by any sensible heat available in this advecting water at above freezing temperatures. *Robin* [1979] notes that fresh ice melted from the shelf will be deposited as sea ice at a shallower depth and these two ice-water interfaces produce different reflection coefficients for radio echo sounding. In this way it is possible to interpret echo records to demarcate areas of melting and deposition, respectively, and this has been done for the Filchner-Ronne ice shelves by *Robin et al.* [1983]. A general conclusion by these and many other authors is that bottom melting is very significant in terms of ice shelf dynamics and breakup and that these processes are of major importance in determining the stability of the Antarctic ice sheet, which leads directly into consideration of sea level variations and climatic change.

Under these circumstances it is becoming increasingly important to be able to make estimates of ice pump rates, which is the intention of the present paper.

2. FUNDAMENTALS

The process of ice pumping may be clarified by consideration of Figure 1. A tank of seawater in an adiabatic enclosure has one of its walls made of ice, the ice-water interface being initially along the line AA. If the tank is taken to be a little more than 10 m deep the interface temperature at the bottom of the tank would be 0.008°C below that of the top of the tank. Suppose that infinitesimal melting occurs near the tank bottom. The reduction in density of the sea water immediately next to the interface will cause it to rise, and as it rises it will increasingly constitute a heat sink for ice growth. Should this growth occur near the top of the tank, the full 0.008°C potential will have been utilized to move ice from the bottom to the top of the tank. The salt rejected during ice growth will cause a downward movement of water which will in turn lead to more upwelling. A self-starting ice pump is in operation, and as time progresses, the interface AA will tend to move through BB until eventually equilibrium is reached with the interface horizontal in the CC

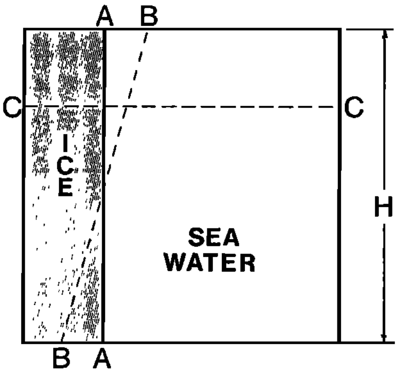

Fig. 1. Schematic of a tank of seawater of depth H in an adiabatic enclosure. One wall of the tank is made of ice with the interface initially along the line AA. See text for discussion of the case for $H \sim 10$ m.

position. The system has then reached its lowest energy state, and no further potential for pumping exists in the given tank. The speed with which such a pump can operate depends on two factors: the rate at which melting can occur given a maximum potential $\Delta T_f = \partial T_f/\partial p \; \Delta p$ and the rate at which water cooled by this melting, thus tending to reduce this differential temperature across the fluid boundary layer, can be transported to the location at lower pressure to cause ice deposit. Thus although the pump can generate its own convective movement, its rate of operation may be greatly increased by changed boundary layer dynamics and the stronger circulation provided by most oceanic regimes.

Two examples of such oceanic circulation are the aforementioned convection beneath a lead and tidal motion beneath an ice shelf. In the former case, $\Delta T_f \sim 0.008$°C for 10-m keels, and even without advection there is good hydrodynamic exchange between the lead and surrounding keels driven by salt rejection during ice formation. The keel melt rate will be limited by the heat flow rate possible from the well-mixed surface layer, which would develop about a 0.004°C temperature difference across a turbulent boundary. In the presence of relative motion between ice and ocean water, the cooled, freshened water moves upward from the keel toward the lead by a combination of convection and turbulent diffusion. Under low turbulence conditions, convective transfer of boundary layer water may dominate. Such interface currents may encounter sharp ice edges that act as turbulence trips and so augment the ice-water heat exchange at those locations. Such places have often been noted to have "underwater ice" deposits consisting of whiskery platelet structures [*Lewis and Milne*, 1977]. This shows that instead of the pump operating over the potential ΔT_f from keel to lead, the process can be broken down into a number of steps. Such platelets eventually grow through the boundary layer and become subject to melting. The ice is then "pumped up" to a shallower location until finally, on average, the rate of deposition must be equal to the rate of keel melt.

In the case of an Antarctic ice shelf 400 m thick adjacent to sea ice ~2 m thick, a potential of about 0.27°C exists to cause transfer of ice between the two. For a steady state pump to work, meltwater must be taken away from the ice-water interface at the same rate as it is being produced so that the pump is not quenched by an increasing layer of meltwater adjacent to the ice. This can be accomplished by ocean currents transporting meltwater from under the shelf or by turbulent entrainment into the underlying water. The maximum pump rate occurs in the absence of this quenching and is then determined by the ability to transfer heat from the water to the ice through a planetary boundary layer, the formation of a density step by meltwater being prevented by turbulent exchange with the water from below. The actual pump rate for ice shelves will thus be controlled by local water movement, whereas for the pack ice case the actual rate is probably nearly always close to its maximum because of the well-mixed surface layer. Antarctic ice shelves extend over hundreds of kilometers and are of variable thickness, so that only close to the ice front would upward drainage to the sea surface be always possible, crevasses excepted. Pumps working at partial potential are to be expected, transforming deep glacial ice to sea ice at a shallower level. Layers of frozen sea water 158 m thick were reported by *Morgan* [1972] for the Amery ice shelf, far in excess of any deposition that could be expected due to heat flow from the water to the ice sheet in response to the latter's cold core. The ice pump provides a possible explanation.

Except close to the ice front it is unlikely that convection based on density difference (as in Figure 1) plays a significant role in water circulation. Horizontal tidal exchanges beneath the ice shelf would usually allow some level of pump operation, provided the ice sheet thickness varied significantly over the distance of the tidal excursion. In particular, waters within a certain distance of the ice shelf front could reach the sea ice at the surface. Steady currents can also exist, as has been demonstrated for the Ross Ice Shelf by *Jacobs et al.* [1979], presumably allowing pumps to operate at a variety of potentials and rates far from the ice front. Tides, by causing vertical mixing, may also have the effect of introducing comparatively warm waters into the basal boundary layer by turbulent entrainment from below and thus enhance melting [*MacAyeal*, 1984]. If the temperature of this warmer water close to the interface, T, is above the range of freezing temperatures available to the ice pump, it is important to realize that for given water circulation conditions increased melting will occur compared with the case where T is within the range, with a decrease in deposition. If the circulation allows the warm water to impinge directly on the locations of both ends of the pump, the need for a reduction in ice deposition is obvious. If the warm water is available only at the greater depth and is cooled to that freezing temperature $T_f(p_2)$ by ice melt, then the deposition at $T_f(p_1)$ will be the same as before. If the warm water, after causing ice melt at depth, still is above $T_f(p_2)$ upon rising, then less deposition will occur. Should the advent of the warmer water be accompanied by a change in the circulation, the above argument may not apply.

In the case of the lead and pressure ridge keel, the water column was found to be well mixed, and a weighted average value of T_f developed across the boundary layers at lead and keel. For the ice shelf this is most unlikely to be true. Local but externally imposed circulation patterns will dictate where and with what waters the melt water from depth and descending water from the surface ice deposit regions will mix. Conditions are subject to change at tidal frequencies, and nearby oceanic water masses may modify rising-descending pump waters near the ice front. For example, *Lewis and Perkin* [1985] report waters of 0.047°C below the surface freezing point arriving at the sea ice in McMurdo Sound immediately adjacent to the Ross Ice Shelf which has a total potential of 5 or 6 times that value. This may be due to ice growth at intermediate intervals, as was discussed above, or due to intermixing with warmer local water masses, or both.

The overall conclusion is that whereas practical pump rates for the Arctic Ocean surface may be calculated quite generally, those for ice shelves will be highly variable and will depend on detailed local knowledge of water movement and properties.

3. MELTING AT THE ICE-WATER INTERFACE

The basic problem is that of calculating heat flow through a turbulent boundary layer in the presence of buoyancy produced by melting. *Gebhart et al.* [1983] considered the case where water movement is controlled by the natural convection induced by melting, but the general case of melting within an externally controlled field of water movement is more suited to the present purpose. The problem has been addressed by *McPhee* [1983], who gives a model for the melting of a wind-driven ice floe floating in "warm" water with Monin-Obukov scaling to account for the buoyancy flux. This scaling length is proportional to the distance below the ice-water interface where the energy generated by buoyancy fluxes is equal to that generated by shear stress. The situation is sketched in Figure 2. The planetary boundary layer consists of an Ekman spiral with a logarithmic profile of relative velocity close to the ice. There is a heat flux from the water, a function of the difference between the interface and far-field water temperatures, which on the assumption that transfer processes for heat and momentum are similar results in a model with a temperature profile similar to that for velocity (curve A, Figure 2). This heat flux is balanced by augmented heat flow into the ice (if any) plus interface melting. The meltwater mixes downward without restrictions so that the heat contained in the whole water mass beneath the ice becomes available for melting. This corresponds to the complete removal of meltwater, which as discussed in the preceeding section, is the condition for maximum heat exchange. In practice, *McPhee* [1983] notes that a fairly sharp temperature gradient often develops at the bottom of a mixed layer which then would determine the maximum depth of downward penetration of the melt. This mixed layer depth would clearly have some relation to the Monin-Obukov length: one may visualise that the buoyancy would be too great to allow further downward shear mixing. Curve B in Figure 2 illustrates this more practical case, which if it is to be a steady state condition requires that the meltwater be advected out of the system. Should this not be possible, curve C represents the final "quenched off" state with little or no further heat flow into the ice being possible. This might correspond, for example, to a domed region under an ice shelf where the buoyancy of the melt-diluted water would not allow advective currents to remove it from the system. Our analysis is for curve A, utilising equations (3), (6), and (7) from *McPhee* [1983] which give nondimensional velocity, temperature, and heat flux, respectively, across the planetary boundary layer. To convert these to practical dimensioned parameters involves selecting a value of Z_0, the surface roughness of the ice. Many estimates exist for the roughness length appropriate to a sheet of sea ice. *McPhee* [1979] found a value of 0.1 m to be appropriate to the planetary boundary layer under pack ice, a value which must include the form drag of keels. Since the heat transfer process considered here takes place close to the ice-water interface, a more appropriate roughness element would be on the scale of the skeletal layer platelets at the growing ice-water interface. Logarithmic layers beneath sea ice have been measured by *Untersteiner and Badgley* [1965] and *Johannessen* [1970]. They found that a value of $Z_0 \simeq 0.01$ m most closely fitted their data, but the roughness elements involved in their flows were difficult or impossible to identify. *Langleben* [1982], measuring stress under flat first-year sea ice, obtained measurements consistent with a hydrodynamically smooth surface ($Z_0 \simeq 2 \times 10^{-5}$ m). Taking the laboratory measurements of *Antonia and Luxton* [1971] on square cross-section roughness bars to simulate platelets along with a platelet height estimate from *Langhorne* [1982] (0.005 m), one arrives at a roughness estimate of $Z_0 = 0.001$ in the fully turbulent to transition regimes. This would be an upper limit, as the platelets are more closely spaced than the experimental obstacles. In Figure 3 the resultant relationship between melt rate and current speed for this "rough" ($Z_0 = 0.001$ m) surface has been contrasted with that for $Z_0 = 0.00005$ m, the values for a "smooth" surface derived from *Schlichting* [1968]. The dependence of the melt rate on the temperature difference ΔT between interface and water is approximately proportional to ΔT, so that at an acceptable level of accuracy it has been possible to plot melt rate per day per degree Celsius against current speed for rough and smooth interfaces.

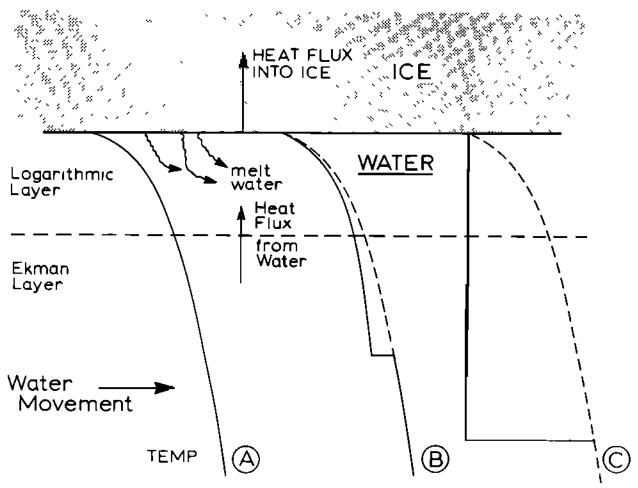

Fig. 2. Temperature and velocity profiles for the ice-water interface. Curve A gives the same shaped curve for velocity and temperature on the thesis that momentum and heat transfer from water to ice occur by the same mechanism (Monin-Obukov similarity). Curve B is the temperature curve when the meltwater buoyancy is sufficient to prevent further downward mixing, thus producing a step. Curve (C) is the "quenched off" case when all the available heat has been removed from the water, with the buoyancy preventing further mixing. In the case of a simple ice pump the difference in temperature between the interface and the far field would be ΔT, as defined in the text.

Bogorodsky and Sukhorukov [1983] have measured melting from the bottom of first-year and multiyear sea ice in the Arctic Ocean over a period of months with and without snow cover. They also measured temperature gradients within the ice, and water temperature and velocity to determine a heat transfer coefficient between water and ice. They produced a relationship (their equation 2) which has been used to generate the upper line in our Figure 3. *Josberger and Meldrum* [1985] have published their preliminary discussion

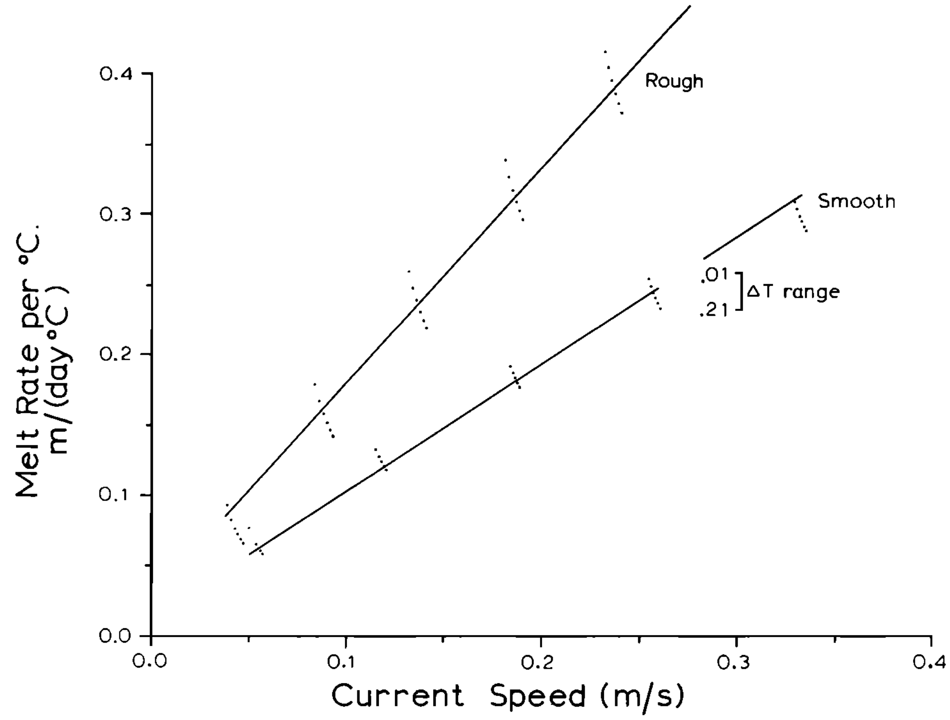

Fig. 3. Melting rates of ice faces in flowing water as sketched in Figure 2. Melt rates are given per day per degree Celsius of ΔT, the temperature difference between the ice-water interface and the far-field water, as a function of current speed. The dots are values calculated from *McPhee* [1983] using a surface roughness Z_0 of 0.001 m (rough) and 0.00005 m (smooth), respectively. The rates are dependent on the surface roughness of the ice boundary but are only slightly dependent on the actual value of ΔT, the range utilized being shown. The line through the rough dots is from the experimental results of *Bodorodsky and Sukhorukov* [1983] for sea ice of the Arctic Ocean and the line through the smooth dots is from *Josberger and Meldrum* [1985] working in the Bering Sea.

on the ablation of sea ice based on measurements made for a few days in the Bering Sea. They have used their data to calculate the value of a dimensionless heat transfer coefficient taking a linear relationship between melt rate, differential temperature, and water velocity. Their result is plotted to give the lower line in Figure 3. For subsequent calculations on sea ice we have preferred to use the "rough" value for Z_0 as the data of *Bogorodsky and Sukhorukov* [1983] are by far the more extensive. A heat transfer coefficient between fresh ice and moving water has been determined by P. F. Hamblin of Canada Centre for Inland Waters and colleagues (personal communication, 1986) for a Yukon lake. Their value is the same as that obtained by *Josberger and Meldrum* [1985] for sea ice. It is thought that the surface roughness of melting glacial ice would be similar to that for lake ice, so that the melting of ice shelves would be represented by the lower line on Figure 3. These three sets of experimental data give us considerable confidence in our analysis.

4. APPLICATION TO THE SEA ICE OF THE ARCTIC OCEAN

Figure 3 has been constructed so as to express the total heat flow into the ice resulting from a given ΔT as its potential for melting. In reality, these melt rates would be achieved only under a uniform sea ice sheet when there was no additional heat flow into the ice associated with conduction to the surface. In winter the normal growth rate under a 1.5-m-thick first-year sea ice sheet is about 1 cm/day. The effect of the ice growth on the boundary layer is taken up by a change in the Monin-Obukov length which results in a small change in the heat flux through the boundary layer.

Consider the case of an ice pump acting between a recently opened system of leads and a uniform (without ridges) ice sheet, with its bottom surface at a depth of 1.5 m. The temperature of the well-mixed layer can be approximated as the average of the freezing temperatures at 0 dbar and 1.5 dbar; $\Delta T_f/2 = 5.6 \times 10^{-4}$ °C. Such a small temperature difference will make no appreciable difference to the vigorous ice formation rate in the lead. Beneath the ice, assuming an average ice-water relative speed of 10 cm/s, a rate of 0.18 m/(day/°C) can be taken from Figure 3, giving a melt rate of 0.01 cm/day. Against a background ice growth rate of 1 cm/day during winter, this figure is not significant. In the case of uniform multiyear sea ice 3 m thick with leads, the melt rate of 0.02 cm/day represents a 6.5% reduction in the background growth rate of 0.3 cm/day. Under broken multiyear ice, pumping constitutes a selective process for the reduction of sea ice growth in the lower half of a sea ice sheet in winter, with a corresponding augmented accretion in the upper half. Multiyear ice is usually associated with extensive pressure ridging. Pressure ridges are consolidated by the freezing of interstitial water between the cold ice blocks pushed down at the time of formation, and as they age, temperature gradients within the ridges approach zero. Thus it is probable that pressure ridge keels can be pumped up equally both in summer and winter. For ice keels 11 m deep with a well-mixed water column 0.004°C below the surface freezing temperature and an average ice-water shear of 10 cm/s, Figure 3 gives a melt rate of about 26 cm/year.

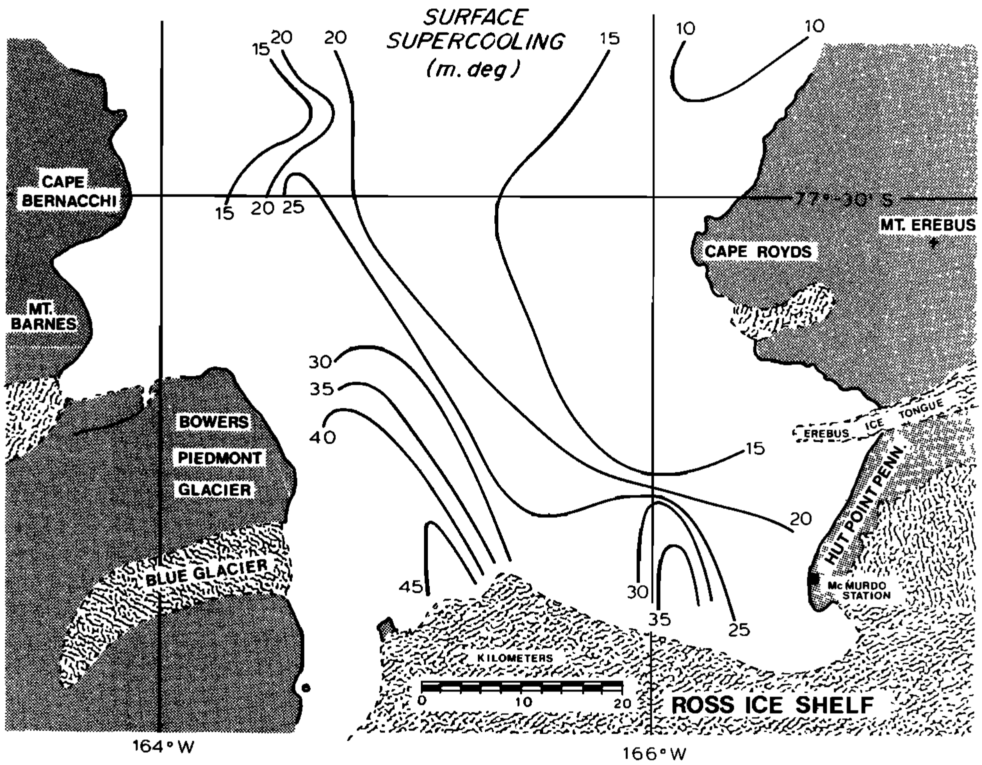

Fig. 4. Contours of surface supercooling 2 m beneath the bottom of the sea ice sheet in McMurdo Sound, Antarctica. Such supercooling must be produced by water coming into contact with ice at depth then upwelling to the surface. The supercooling is relieved at the sea ice–water interface, with salt rejected into the water helping to augment the vertical circulation. From *Lewis and Perkin* [1985].

It is clear that a simple model may be applied to pumping in the pack ice. Starting from a given thickness probability distribution function $G(h)$ and on the assumption that the surface waters are well mixed down to the maximum keel depth, it would be possible to calculate the weighted mean freezing temperature and apply this value to determine the heat exchange at all depths per unit area of ice sheet. This heat would be used to reduce or increase ice growth or cause melting or growth according to the value of the temperature gradient within the ice. It is to be emphasized that the above calculation would not include melting due to sensible heat in the water, which may or may not be allied with a corresponding ice deposition at the thin places of the ice sheet.

5. Application to the Melting of Ice Shelves

As has been discussed, ice pumping from beneath a shelf is very dependent on the local water circulation patterns. Because of the authors' interests, attention will be focussed on the Ross Ice Shelf, an area of immediate importance to the stability of the Antarctic Ice Sheet. In McMurdo Sound, *Lewis and Perkin* [1985], confirming many earlier reports, found that supercooled water coming from beneath the ice shelf caused ice formation beneath the contiguous sea ice. Figure 4 shows the locations just below the sea ice of arrival of this supercooled seawater. Two such foci for upwelling may be noticed; at these locations the supercooling is relieved within 2 m of the sea ice–water interface, and owing to the resulting salt input the water column is marginally unstable to a depth of 190 m, so that resulting convection is capable of reaching a great depth. It is not known how far conditions in McMurdo Sound may be thought to typify those near the Ross Ice Shelf. There is a smooth reduction in the thickness of the shelf ice as one moves towards the sea ice in McMurdo which contrasts with the ice cliff normally seen along the greater part of the Ross Ice Shelf front. Clearly, in McMurdo Sound, melting is taking place near the ice shelf front, causing a sloping interface between shelf ice and sea water which provides for easy exchange of buoyant water from beneath the shelf with the surface waters of the sound. Oceanographic profiles taken near the ice shelf–sea ice interface indicate that at a depth of 360 m (i.e., at the bottom of a 400 m thick shelf) the water is very close to its surface freezing point (that is, about 0.27°C above its in situ freezing point), and this potential is therefore available for ice melt. Current measurements at 200-m depth close to the ice front gave speeds of 5 cm/s. Supposing that this water movement existed beneath the ice shelf and that the fresh water produced by the melting ice could be moved out of the system entirely, using the "smooth" results from Figure 3 gives a pump melt rate of over 5 m/year. The measurements used to construct Figure 4 show that the maximum supercooling at the surface is 0.047°C, which is the freezing point depression corresponding to a depth interval of about 60 m. If the water were coming from 360-m depth, this would indicate a "staged" pump, but as the mean thickness of the ice shelf over 12 km from the front is about 60 m [*Kovacs et al.*, 1982], it probably means that the water circulation pattern involves movement only over that horizontal scale. Alternatively, colder water from the greater depth could be mixing with warmer water en route to the surface, though consideration of temperature and salinity profiles near the ice front and the possible *T/S* mixing lines make this last possibility improbable.

The thickness of the ice sheet is variable and except near the ice front will probably allow under ice "pooling" of meltwater to occur and thus limit pumping. Such appears to be the case at borehole J9, which is within a thin area of the shelf, there 420 m thick. Indeed *Zotikov and Zagorodnov* [1980] report temperature profiles in the sea beneath the ice shelf that bear a resemblance to curve C of our Figure 2. *Foster* [1983] also reports a two-layer system beneath the ice shelf and occasionally measured supercooling of 0.005°C in the water of the upper layer. *Zotikov et al.* [1980] noted 6 m of sea ice formation on the bottom at J9. *Jacobs et al.* [1979] state that currents of up to 18 cm/s were measured beneath the shelf, apparently mainly due to tidal oscillations. Assuming a mean current of 10 cm/s and supercooling of 0.005°C, Figure 3 would allow a maximum pumped deposit of 13 cm of sea ice per year assuming no quench off. Deposition resulting from this process could be far more significant than that due to the temperature gradient at the ice shelf–sea water interface, which with due allowance made for geothermal heat flux from the sea bed is calculated to be 1 cm/year [*Clough and Hansen* 1979]. However, *Zotikov et al.* [1980] estimated only 2-cm/year growth of sea ice at J9, which would indicate minor if any pump activity in practice.

Foster [1983] reports that at J9 there was a layer of fresher water about 30 m thick next to the ice, but if this is to be a region of net deposition, an increase rather than decrease in the salinity of the interfacial seawater layer might be expected. This argument would be the same irrespective of the heat sink for ice growth, and we believe existence of the fresher water must be attributed to melting at greater depths caused by the sensible heat of advected water. *Jacobs et al.* [1979] report finding warmer, more saline water below 500 m at J9, and *MacAyeal* [1984] has calculated melting from this source to be between 0.05 and 0.5 m/year, which is additional to pumping by tidal currents. To emphasize the point, tidal circulation without any available sensible heat will augment pumping; with sensible heat it will cause additional melting that is not, as in the case of the pump, necessarily balanced by addition deposition.

Lewis and Perkin [1985] reported on the extent of underwater ice to be found in McMurdo Sound while the data for Figure 4 were being collected in October–November 1982. Underwater ice is nucleated within the water column and grows from the heat sink provided by supercooled water; it was described in McMurdo Sound by *Paige* [1966] and has a definitive crystallographic structure making it readily identifiable. The relief of supercooling can be either by production of this underwater ice in the water column or directly by growth onto the existing floating ice sheet. Beneath the shelf, direct relief is thought more likely, as the upward moving waters must be close to the ice sheet at all times and pumping would occur over a depth interval smaller than that at the ice front. Both methods of relief occurred in McMurdo Sound in 1982 where underwater ice floated upwards to form a layer up to 1 m thick of loose ice discs 1–2 cm in diameter just beneath the 3-m-thick solid ice sheet. Underwater ice also comprised about 40% of this sheet, having become incorporated into the sheet by the growth of "congelation" ice, which is the crystal type usually found in the Arctic Ocean, a result of heat loss to the atmosphere. In contrast, *Gow et al.* [1982], reporting on their October–November 1980 measurements, found much fewer subsheet floating ice discs and practically no incorporation of underwater ice into the 2-m-thick main ice sheet. The conclusion that there was an increased availability of supercooled water in McMurdo Sound between 1980 and 1982 seems unescapable and must indicate a changed circulation pattern of waters beneath the local part of the Ross Ice Shelf which increased pump rates. Whether this is a normal variation or is indicative of a trend is still to be determined.

Basal melt rates such as are estimated herein raise the question of the mechanical stability of the ice front. Selective melting, increasing as the ice shelf front is approached, might require that some portion of the ice near the front be supported as a cantilever, thus inducing bending stresses that might augment calving.

6. CONCLUSIONS

The ice pump mechanism acts to make a floating ice sheet become of uniform thickness and thus tends to reduce the effect of mechanical deformation on the thickness of a sea ice cover. Experimentation and theory agree that about 26 cm of ice per year would be pumped off an 11-m-deep pressure ridge keel with an equivalent deposition at shallower depths. Depending on the nature of the hydrodynamic connection between keel and lead, convective, advective, shear turbulent, or a combination of these, the pump may be broken down into a number of stages removing ice from the deepest levels and depositing it temporarily at intermediate depths before final deposition.

The pump rate is set by a combination of two factors: (1) the heat flow available through a turbulent boundary layer at the ice-water interface, which is determined by the relative velocity between ice and water, the temperature difference between the mass of moving water and the melting interface, and its surface roughness, and (2) The rate of removal of meltwater from the system as it is produced determines the steady state condition that develops.

The first factor has been calculated following *McPhee* [1983] and sea ice melt rates measured by both Soviet and U.S. investigators agree well with the theory for reasonable values of surface roughness. This agreement also indicates, as appears intuitively obvious from the existence of the mixed layer, that turbulence under a moving pack ice field is sufficient to remove all meltwater and so make the influence of the second factor negligible. This turbulent vertical water exchange prevents "quenching off" of the pump and gives a maximum melt rate, but at the same time, by producing a well mixed surface layer of water adjacent to the ice, it causes the effective temperature differential to be reduced to about $\Delta T_f/2$. The actual value can be obtained by calculating an areally weighted average of the freezing temperature given a specific ice sheet thickness distribution. In the Arctic Ocean, $\Delta T_f/2$ has been measured in the range 0.003–0.006°C [*Lewis and Perkin*, 1983].

For floating ice shelves the second factor appears dominant, but it is probable that near the ice front, horizontal and vertical water exchange is still sufficient to allow the maximum pump rates to be applicable. In the case of melting at the front of the Ross Ice Shelf in McMurdo Sound a figure of at least 5 m/year for basal melting is appropriate for the austral spring of 1982. Observations of ice condition in 1980 [*Gow et al.*, 1982], however, suggest qualitatively that there was far less flow of supercooled water from beneath the ice shelf in that year so that the pump rate was then limited. Far from the ice front, ice pumps can work removing glacial ice from greater depths to deposit sea ice nearer the surface.

Tidal currents provide the necessary water movement and may entrain warmer water from beneath to enhance melting. Melting due to this sensible heat entrainment has been estimated in the range 0.05–0.5 m/year for the Ross Ice Shelf by *MacAyeal* [1984]. It is in addition to and independent of melting due to the ice pump, which is driven solely by the pressure dependence of the freezing temperature.

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