Antarctic sea ice growth and oceanic heat flux

IAN ALLISON
Antarctic Division, Department of Science and Technology, Kingston, Tasmania 7150, Australia

ABSTRACT The classical relationship between the thickness of a floating ice cover and air temperature, the Stefan law, greatly overestimates the growth rate of Antarctic sea ice, because, among other simplifications, it neglects the role of heat transfer from the ocean to the ice at the lower boundary. Sea ice measurements from near Mawson, Antarctica, are used to show that, by modifying the Stefan law to account for this heat exchange, both the growth and decay of the ice cover can be followed. Air temperature and ice thickness measurements are used to estimate the ocean heat flux for several years at Mawson; these estimates agree well with heat flux values calculated from more detailed data. The ocean heat flux shows systematic variation throughout the season and is in part controlled by the ice growth process itself. The flux is greatest initially when brine rejected by the rapidly growing ice sets up thermohaline convection in the underlying water. As the rate of ice growth and brine rejection decreases, so does the convection and the heat flux to the lower ice boundary drops off. A second peak in the heat flux, occurring in September - October, is possibly linked with large scale meridional heat advection in the southern ocean. The oceanic heat flux estimated for several other Antarctic stations shows the same pattern as at Mawson.

The classical relationship between surface temperature and the growth of a floating ice cover, Stefan's law, assumes that heat linearly conducted through the ice is exactly balanced by the latent heat of fusion of newly formed ice. This yields the familiar expression for the ice thickness, $h$, after time $t = T$

$$h^2 = \frac{2k}{\rho L} \int_0^T \theta \, dt$$

where $k$ is the thermal conductivity of the ice, $\rho$ is the ice density, $L$ the latent heat of fusion of the sea ice, and $\theta$ the temperature difference between the lower and upper ice surfaces.

This relationship has often been used as the basis for the derivation of equations predicting sea-ice growth from air temperatures, although it has several serious limitations when applied to the growth of a real sea ice cover. Because Stefan's law applies only to ice of zero specific heat, it makes no allowance for changes of heat content within the ice nor for the time taken for surface temperature changes to vary the heat flux at the lower boundary. Other factors modifying the growth of a real
sea ice cover may include the existence of other heat sources or sinks (for example, turbulent heat transfer between the ice and underlying water or shortwave radiation absorption within the ice), variation of the thermal properties of the ice with depth and time, and the presence of a snow cover insulating the ice surface from the air temperature. The many modified analytical solutions to the Stefan problem which have attempted to account for these factors have generally been too complicated for practical use.

Schwerdtfeger (1964) corrects for the change in heat content in the ice by replacing the latent heat, $L$, in the Stefan law by an effective latent heat, $L_{\text{eff}}$

$$ L_{\text{eff}} = L(1 + (Q_1/Q_P)) $$

where $Q_1/Q_P$ is the ratio of change of heat content in the ice to the heat involved in freezing for a unit increase of ice thickness, and can be expressed as a function of the ice salinity and temperature. The effect of lag between the surface temperature and the ice growth rate can be minimized by considering only long time intervals.

A more serious problem in predicting Antarctic sea ice growth arises because of the substantial turbulent heat exchange at the ice-water interface. Weller (1968), from heat budget considerations at the lower boundary of an ice cover near Mawson, determines a transfer of from 5 to 19 W m$^{-2}$ from the water to the ice; for much of the year this is greater than the latent heat term. The presence of a strong heat flow from water to the ice at Mawson is confirmed by the substantial difference between the measured ice thickness in 1963, and that predicted from surface temperature by a computer analogue of the conduction process (Schwerdtfeger, 1970).

On the large scale, the northward transport of water with Antarctic characteristics, resulting from sea-air-ice interaction, is balanced by a poleward flux of relatively warm saline water. Gordon & Taylor (1975) use an average annual heat loss of Antarctic surface water of about 20 W m$^{-2}$ to derive a heat budget for the Antarctic surface water which requires volume fluxes similar to those required for the salt balance. Parkinson & Washington (1979) use a constant oceanic heat flux of 25 W m$^{-2}$ in their large scale numerical model of Antarctic sea ice. They recognize that a constant flux involves a considerable simplification, especially during periods of rapid ice growth when the discharge of salt from the ice may result in unstable stratification in the mixed layer, and convection with the underlying water.

The effect of the ocean heat flux on sea ice growth can be seen from ice thickness measurements made in an open embayment, West Bay, near Mawson (67°36'S, 62°52'E) during 1969. Water depth at the site was over 200 m and the ice surface was kept relatively snow free by strong katabatic winds. Fig. 1 compares the measured ice growth with that estimated from Stefan's law (equation (1)) using 10-day mean air temperatures. The temperature difference between upper and lower ice surfaces, $\theta$, has been taken as $(\theta_a - 1.8)$, where $\theta_a$, the air temperature is a reasonable
approximation of the ice surface temperature (since the snow cover rarely exceeded 10 mm) and -1.8°C is the freezing point of sea water.

The value of thermal conductivity used, 1.93 W m\(^{-1}\) degree\(^{-1}\), is that obtained from Schwerdtfeger (1963) for the mean temperature and salinity of the Mawson ice cover. The Stefan thickness has been calculated using both a latent heat of 0.28 MJ kg\(^{-1}\) and an effective latent heat of 0.33 MJ kg\(^{-1}\) to correct for heat storage within the ice (Schwerdtfeger, 1964). In both cases the estimated ice growth is much greater than that observed. To obtain a reasonable match to the observed growth the thermal conductivity of the ice would need to be as low as 1.34 W m\(^{-1}\) degree\(^{-1}\) and even so, this could not explain the observed decrease in ice thickness while the temperature difference, \(\theta\), was still positive.

Also shown in Fig. 1 is the ice thickness estimated from the relationship

\[
h = \left(\frac{2k}{\rho L} \Sigma(\theta t)\right)^{-\frac{1}{2}} - \frac{1}{\rho L} \Sigma \theta t
\]

(2)

where \(\Sigma\) is a constant. This is the solution of a Stefan problem, modified by an additional heat flux at the lower boundary; the additional flux being nonconstant but decreasing with time. With an effective latent heat of 0.33 MJ kg\(^{-1}\), a value of \(\Sigma\) of 9 W m\(^{-2}\) gives the best fit to the observed data. The inclusion of an extra heat flux term at the lower boundary not only gives a reasonable fit over the ice growth period, but also explains the decrease of ice thickness despite heat conduction away from the boundary. There remain, however, systematic departures from the observed values, indicative of variations in oceanic heat flux different from the simple model of equation (2).

The growth of sea ice at West Bay, Mawson, for four different years is shown in Fig. 2. Ice thicknesses have been corrected for the measured surface ablation, so that only changes at the lower boundary are shown: the snow cover was slight in all years. The mean monthly air temperatures for the same years are shown in Fig. 3. The annual ice cover at Mawson usually first forms in mid to late March, although breakouts and reformation can occur up till the end of May (as for example in 1969 and 1965). The more significant variations in maximum thickness of the cover are determined by the date of final formation, but differences in growth rate like those between 1976 and 1977, are related to differences in air temperature.

The magnitude of the heat flux from the ocean to the ice, \(Q\), has been estimated for these growth curves from the heat balance

\[
Q + \rho L \frac{dh}{dt} = k \left(\frac{d\theta_i}{dh}\right)
\]

(3)

where \(\frac{d\theta_i}{dh}\) is the temperature gradient in the ice and has been approximated as previously, by \((\theta_a - 1.8)/h\) (where \(h\) is the mean ice thickness for the period considered). Again the effective latent heat has been used to correct for heat storage in the ice. Monthly means only have been calculated to minimize the lag between surface temperature changes and the resultant effect on the conducted flux at the lower boundary.

The monthly values of \(Q\) so calculated, are shown in Fig. 4(a)
Fig. 1  Observed ice thickness at Mawson 1969, compared with thicknesses predicted from Stefan's law (dotted line), Stefan's law using an effective latent heat to compensate for heat storage in the ice (dashed line), and Stefan's law modified to include a decreasing oceanic heat flux (solid line).

Fig. 2  Sea ice thickness, West Bay, Mawson.

Fig. 3  Mean monthly air temperatures, Mawson.
against time since formation of the ice cover. The curves for all 4 years, including that derived from the initial cover in 1965 which broke out in May, show a similar pattern. The heat flux from the ocean decreases rapidly in the first 70 or so days from a maximum immediately after formation. A second lesser maximum of $Q$ seems to occur around September or October, just before the maximum ice thickness is reached.

Fig. 4 Heat flux from the ocean to the sea ice at Mawson. (a) Monthly flux values estimated from mean air temperatures and ice growth. The values for September are indicated by S and the time of maximum ice thickness with an arrow. (b) Flux values calculated using measured ice temperature profiles (indicated points) compared to those estimated as for (a).

Fig. 5 Sea ice temperature profiles at Mawson, for the latter part of 1969.
The validity of estimating the conducted flux from air temperatures and using an effective latent heat to correct for storage terms can be seen from Fig 4(b). Here the fluxes estimated from air temperature and ice growth alone, are compared to fluxes calculated as a remainder in an energy balance of the lower surface. In this case the conduction term is calculated from the measured temperature gradient at the lower boundary (e.g. Fig. 5) using a thermal conductivity which is varied with the measured salinity and temperature of the lower ice.

The results for 1965 are from Weller (1968) while those for 1969 and 1977 are from a series of studies of the thermodynamics of sea ice described by Allison & Akerman (1980). In all cases the values of \( Q \) calculated in this way are in excellent agreement with the earlier estimates, and show the same variation with time. For most of the year the measured ice surface temperature was within \( \pm ^\circ \) of the mean air temperature, and the temperature profiles within the ice were linear. From mid September on, increasing radiation absorption in the upper ice layers leads to curvature of the temperature profiles (Fig. 5) but even for this period the heat flux estimates using air temperatures agree with those from the energy balance (Fig. 4(b)).

Errors in ice thickness measurement of 100-150 mm or temperature errors of 10-15\(^\circ\) would be needed to account for the variations of the ocean heat flux with time and, considering the regular pattern of the variations and the agreement with the energy balance values, the fluctuations must be considered as real.

The initial high values of the oceanic heat flux, and its rapid decrease, can be explained by thermohaline convection caused by salt rejected from the growing ice. The decrease of salt content of the sea ice in 1969 is shown in Fig. 6(a). Brine is rejected from the sea water of salinity \( \sim 33\%/o \) during the freezing process, and the salinity of the ice is further decreased

---

**Fig. 6**  (a) Salinity structure of the seasonal ice cover at Mawson, 1969. (b) Change in salinity of the top 1.5 m of ocean (water and ice).
by brine expulsion and gravity drainage during the growth season (Lake & Lewis, 1970). The decrease in salt content of the top 1.5 m of the ocean is shown in Fig. 6(b). Density instabilities caused by this rejected salt set up penetrating convective plumes in the underlying water, bringing relatively warmer deep water to the ice-water interface (Lewis & Walker, 1970). The period of high oceanic heat flux to the ice coincides with maximum ice growth and hence maximum convection. As the ice growth rate decreases, the rate of salt rejection also drops, and convection and heat transfer to the ice also decrease. For the first 100 days after ice formation, the values of ocean heat flux (Fig. 4(a)) are strongly correlated with ice growth rate, and hence salt rejection.

The second peak in the heat flux of 15-20 W m$^{-2}$ occurring in September-October is more difficult to explain, but some information on processes in the water is available from salinity and water temperature measurements made at the West Bay site in 1977. These measurements were made at 10 m depth intervals to 100 m, although at irregular time intervals.

Between 22 February and 1 April the top 100 m of water is cooled from a mean temperature of about -1.0$^\circ$C to a temperature uniformly near the freezing point. Most of this cooling occurs before the ice formation on 18 March. No measurements are available between 1 April and 20 September, but over this period the water increases in average temperature, despite considerable heat flux to the ice. With an average heat flux of about 3 W m$^{-2}$ to the ice, and the temperature rise of the water equivalent to a further 3 W m$^{-2}$, there needs to be an advective transport to the region averaging 6 W m$^{-2}$. The water salinity over the same period also increases at a rate greater than can be explained by salt rejected from the sea ice, and balance requires a net salt transport to the region.

Between 20 September and 20 October the inferred advected heat has increased to about 17 W m$^{-2}$, and is matched by a corresponding increase in the salt transport. Although there is a mean increase in heat in the top 100 m between these dates, the temperature profiles actually show a temperature decrease in the first 30 m due to heat loss to the ice. The temperature increase in the column from 30 to 70 m deep is a fairly uniform 0.06$^\circ$C and, assuming heat advection to the top 30 m at the same rate as that giving this increase, an average vertical heat loss to the ice of about 8 W m$^{-2}$ is required to explain the observed temperature decrease. Given the crudity of the assumptions used to derive this value, it is in reasonable agreement with the previous estimate for the same period of about 13 W m$^{-2}$ (Fig. 4). September temperature profiles measured at Mawson in 1956 (Bunt, 1960) show a similar net heat loss in the top 20-40 m and a heat gain from 40 to 100 m. Several profiles to 400 m measured further north in 1978 show general advection of heat in the upper 200 m or so, but again with a loss to the ice in the topmost 20 - 50 m.

Between 20 October and 7 December the heat transport has decreased to 10 W m$^{-2}$ while the salt balance becomes complicated by the influx of fresh water from melting continental ice.
Overall there is a continual net horizontal transport of heat and salt to the coastal region, with a maximum transport of both in September-October. This peak in advection is reflected in a maximum in the vertical transport to the ice. While it is beyond the scope of this paper to investigate changes in the large scale north-south ocean circulation that would result in seasonal variation of salt and heat transport, the maximum in September-October can probably be related to the peak in air-sea interaction occurring around the same time, due to a maximum ice extent and rapidly increasing solar radiation (Ackley & Keliher, 1976).

Given that the initial large heat flux to the ice is a function of the ice growth, and that the second maximum is due to large scale effects, then a similar seasonal pattern of heat flow from the ocean to the ice would be expected for other Antarctic sites. Few data are available for sites without an insulating snowcover, but estimates for some suitable sites are shown in Fig. 7. The data for Molodezhnaya (67°40'S, 45°51'E) are from Evseev (1969) and those for a snow free site at Halley Bay (75°30'S, 26°36'W) from Limbert (1970); heat fluxes at the latter site were estimated using the high thermal conductivity of 2.6 W m\(^{-2}\) degree\(^{-1}\) quoted by Limbert. At both sites the ice cover formed later than at Mawson (27 May at Molodezhnaya; late May at Halley Bay) but whereas the maximum thickness of the ice at Molodezhnaya, 1.4 m, was comparable to that at Mawson, the Halley Bay cover attained a thickness of 2.2 m. For both sites the estimated oceanic heat flux shows variations similar to those in Fig. 4, with the second maximum still occurring in September despite the later ice formation. The initial heat flux at both sites is higher than at Mawson as might be expected from the rapid initial ice growth.

Also shown in Fig. 7 are estimates of the ocean heat flux at a different site near Mawson. These are derived from the measured thicknesses of snow free ice in Mawson Harbour, where the water depth is only about 30 m. The results shown are for 1967 and
1970, but similar results were also obtained for the ice growth at this site in 1954, 1955, 1957, and 1958 (all from Mellor, 1960), and in 1969. In all cases there is an initial decrease in heat flux and a second maximum around September-October.

The heat flux from ocean to ice appears to be fairly uniform over large areas. Ice thicknesses measured up to 40 km north of Mawson in September 1977 and to 30 km north in 1978 show a maximum spread of 0.3 m and these variations are almost always directly related to the thickness of snow cover on the ice. In a few areas the bottom topography may cause local effects, for example the sudden decrease of ice thickness of 0.3 m observed both in 1977 and 1978 at 26 km north of Mawson, and not related to a snow cover change, may be caused by locally enhanced upwelling. The sea ice measurements for Mawson, 1963, show a maximum thickness of only 0.72 m (at least 0.6 m less than usual) and indicate a very large heat flux from the water (Schwerdtfeger, 1970). This is also probably due to locally enhanced vertical transport, as the measurements in that season were taken above a prominent shoal.

Ice thickness data from other Antarctic stations cannot be used to give reliable estimates of the ocean heat flux because the presence of a snow cover negates the assumption that the mean ice surface and air temperatures are equivalent. However, although the magnitude of the estimated flux is unreliable, the variation of the heat flux during the season at several snow covered sites shows the same general pattern as in Fig. 4 and Fig. 7. Despite the presence of a continuous snow cover, heat fluxes estimated from observed ice growth curves for Mirny (66°33'S, 93°01'E) 1965 (Petrov & Chikovskii, 1969), a different site at Halley Bay, 1963 (Limbert, 1970) and Davis (68°35'S, 77°58'E) 1979 all showed a rapid decrease from a maximum after ice formation and a second peak in September-October.

CONCLUSIONS

The heat flux from ocean to ice is an important factor in the growth of Antarctic sea ice. For snow free ice the magnitude of this flux can be estimated from mean air temperatures and ice growth.

The heat flux is not constant, but shows a marked seasonal variation due in part to the ice growth itself. Initially, the flux is very high due to thermohaline convection caused by salt rejected from the ice. The magnitude of the flux is proportional to ice growth rate and may be as high as 50 W m⁻² for rapidly growing ice, but drops quickly as the ice growth rate decreases with increasing thickness. A second peak of 10 - 20 W m⁻² occurs in September-October. This is probably due to seasonal changes in large scale meridional heat advection in the southern ocean.

Limited data from coastal sites other than Mawson suggest that this seasonal variation is a general pattern. The mean heat flux over the ice growth season is usually about 10-15 W m⁻² but for large scale models of sea ice a simple parameterization of the initial decrease of heat flux, as a function of ice growth rate, may be preferable to the use of mean fluxes.
ACKNOWLEDGEMENTS  I am indebted to the following members of Australian National Antarctic Research Expeditions, for providing previously unpublished ice thickness data: John Illingworth (Mawson, 1967), Ian McCarthy (Mawson, 1970), Graeme Akerman (Mawson, 1976), John Tann (Mawson, 1977), and Mal Griffin (Davis, 1979). The water temperature and salinity data, and ice thicknesses north of Mawson were supplied by John Tann (1977) and Rob Wills (1978).

REFERENCES


