The Impact of Melting Ice on Ocean Waters

Adrian Jenkins

British Antarctic Survey, Natural Environment Research Council, Cambridge, United Kingdom

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ABSTRACT

Ice melts when it is in contact with ocean waters that have temperatures above the in situ freezing point. The product is a mixture of meltwater and seawater having properties intermediate between those of the two components. Density is one of the properties that is affected, and this has important implications for how the melt-induced changes are eventually manifested. Although the direct impact of melting is to cool and dilute the ocean, subsequent convection can carry the products of melting to parts of the water column where they are comparatively warm and salty. These principles are illustrated with a set of observations from the continental shelf of the Amundsen Sea. Measurements made near a floating glacier are used to calculate the concentration of meltwater in the water column. Concentrations approaching 2% are associated with comparatively high temperatures, low dissolved oxygen concentrations, and negligible stable isotope anomalies. The impact of drifting icebergs on the Southern Ocean is discussed. Over most of the area to the south of the Polar Front, melting effects a transfer of heat from the Circumpolar Deep Water to the overlying winter water. The resultant net heat flux over the entire area is small, but locally it may exceed 100 W m⁻².

1. Introduction

Runoff from the Antarctic Ice Sheet annually contributes about 2.6×10^{15} kg of freshwater to the Southern Ocean (Jacobs et al. 1992). All but about 2% of this discharge enters the ocean as ice, which subsequently melts over a wide area. This paper addresses the impact of the melting process on seawater properties. Throughout the analysis, the ice–seawater system is assumed to be closed, precluding any consideration of atmospheric influence on either ice or seawater. The discussion is therefore applicable only to processes occurring below the surface mixed layer. Conservative properties of a meltwater–seawater mixture, χ , are then determined simply by the mass of ice, Q_i , and of water, Q_w , that contribute to the mixture, and by the properties of the two components, χ_i and χ_w :

$$Q\chi = Q_w \chi_w + Q_i \chi_i, \tag{1}$$

where $Q = Q_i + Q_w$ is the total mass of the mixture.

Among the properties of ice that impart distinctive characteristics to the water into which it melts are a negligible salinity, a pronounced depletion in the isotopically heavy forms of water, a mixture of incorporated gases that is closely related to their concentrations

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in air, and a low specific enthalpy. The latter is a less convenient thermodynamic variable to work with than potential temperature, but temperature is not conserved according to (1). When ice is added to it, the ocean is cooled by three distinct processes: heat conduction from water to ice that serves to warm the ice from its farfield temperature, θ_i , to the freezing point, θ_f , absorption of latent heat during the phase change, and subsequent mixing of the meltwater, which has a temperature equal to the freezing point, with the ambient water. This leads to an effective potential temperature defined by

$$\theta_i^* = \theta_f - \frac{L}{c_p} - \frac{c_i}{c_p} (\theta_f - \theta_i)$$
(2)

that appears in place of θ_i in the potential temperature form of Eq. (1) (Gade 1979). In (2), c_i and c_p represent the specific heat capacities of ice and seawater, respectively, at constant pressure, and *L* is the latent heat of fusion.

Useful tracers of melting ice are those displaying a large ice–seawater contrast and, because this difference is normally much larger than any spatial variations within the ice sheet, a mass of floating ice can generally be regarded as being uniform. If the seawater can also be assumed to be uniform and both ice and ambient water properties are known, (1) can be used to derive the meltwater content associated with any particular observation χ :

$$\frac{Q_i}{Q} = \frac{\chi_w - \chi}{\chi_w - \chi_i}.$$
(3)

Corresponding author address: Dr. Adrian Jenkins, British Antarctic Survey, Natural Environment Research Council, High Cross, Madingley Rd., Cambridge CB3 OET, United Kingdom. E-mail: a.jenkins@bas.ac.uk



FIG. 1. Bivariate plot illustrating observed properties (χ^i, χ^2) produced as ice with properties (χ^i_i, χ^2) , melts into a single water type, having properties (χ^i_w, χ^2_w) . The locus of all possible observations forms the straight line drawn between the ambient and ice properties. Numbered points indicate the meltwater fractions associated with selected positions on the mixing line.

It follows that mixtures of meltwater and seawater will have conservative properties that plot as a straight line on a bivariate graph (Fig. 1). In this idealized situation, the impact of melting ice is easily discerned and the concentration of meltwater associated with any particular observation can be determined from where the measurement falls on the mixing line. Should one of the two properties be unknown in either the ice or the ambient water, extrapolation of the mixing line to concentrations of zero or one can be used to evaluate the unknown properties.

These principles have been used in many studies of ocean waters under the influence of melting ice. The slope of the meltwater mixing line in potential temperature–salinity (θ –S) space has been derived by a number of authors (Gade 1979; Greisman 1979; Nøst and Foldvik 1994). Nøst and Østerhus (1998) used this knowledge to infer changes in the salinity of waters causing melting beneath an ice shelf. They assumed that the temperature of the source waters would always correspond to the surface freezing point and extrapolated the mixing lines passing through measured θ -S characteristics of outflows observed at different times to their intersections with the freezing point curve. The $\delta^{18}O$ salinity trend has also been used to study the evolution of water masses near ice shelves and to calculate the mean isotope ratio of melting ice by extrapolation of the mixing line to zero salinity (Weiss et al. 1979; Potter et al. 1984; Jacobs at al. 1985). Enrichment of dissolved oxygen and helium in water masses in the Ross and Weddell Seas has been attributed to melting at the base of the large ice shelves that cover the southern extremities of the continental shelves (Jacobs et al. 1970; Schlosser 1986), and a combination of stable isotope and helium data have been used to calculate the fraction of meltwater contained in ice shelf water (ISW) and Weddell Sea Bottom Water (Schlosser et al. 1990; Weppernig et al. 1996).

The influence of melting ice on a nonuniform water column can be complex because the cooling and dilution brought about by melting entail changes in the seawater density. To quantify the observable impact of melting, it is necessary to compare the properties of a meltwaterseawater mixture with those found at its equilibrium level in the ambient stratification, rather than with those of the source water. This was recognized by Neshyba (1977), who suggested the possibility of drifting icebergs leaving a trail of nutrient-rich surface water in their wake. His argument was not that the ice was rich in nutrients, but that upwelling caused by melting would raise nutrient-rich waters from below the pycnocline. Jenkins (1993) used simple plume theory to model the upwelling induced by melting icebergs and demonstrated the possibility of local warming of the water column, a phenomenon observed by Potter et al. (1988) and Hellmer et al. (1998). These latter observations highlight the problems of interpreting measurements made in a nonuniform environment.

This paper describes a method for calculating meltwater concentrations that permits the assumption of nonuniform ambient properties. The analysis follows that used by McDougall (1990) to evaluate the impact of "hot smokers" on an arbitrary, linearly stratified water column. Observations from the continental shelf of the Amundsen Sea provide an illustrative example of how the method may be applied. The same principles are then used to evaluate the impact of melting ice on the near-surface waters of the Southern Ocean by a consideration of how scalar properties evolve on isopycnal surfaces.

2. Melting into a nonuniform environment

In the presence of arbitrary variations in ambient water properties, χ_w cannot be assumed constant and calculation of the meltwater fraction using (3) is no longer possible. The change in density caused by melting drives convective motion near the ice, and this in turn causes mixing and further melting. As a result ambient waters with a whole range of properties are combined with the meltwater in unknown quantities to produce the final observation, χ . This section shows how progress can be made under an assumption much less restrictive than that of constant χ_w . The requirement will be for two properties whose trends throughout the ambient waters are linearly related.

On the continental shelves of the Ross and Weddell Seas the ambient waters are close to the surface freezing point almost everywhere, so potential temperature and salinity are approximately linearly related. On the shelves of the Amundsen and Bellingshausen Seas and in the upper levels of the deep ocean basins around Antarctica, where many icebergs melt, the ambient water column consists of Antarctic Surface Water (AASW) overlying a temperature maximum core of Circumpolar Deep Water (CDW). The lower stratum of the AASW, the temperature minimum layer referred to as winter water (WW), and the CDW both have properties that vary slowly horizontally. Regardless of the spatial structure of the pycnocline separating these two water masses, concentrations of most conservative tracers within the pycnocline are approximately linearly related, because they are set by the same mixing processes.

Consider an ambient water column made up of mixtures of two well-defined end member water types. On a bivariate plot, measurements of any two conservative properties made in the ambient waters will lie on a straight line with a gradient given by

$$\left(\frac{\partial\chi^2}{\partial\chi^1}\right)_w = \frac{(\chi^2_w - \chi^2_{w0})}{(\chi^1_w - \chi^1_{w0})},\tag{4}$$

where the subscript 0 indicates reference values measured at an arbitrary point in the water column. Mc-Dougall (1990) pointed out that this simple fact can be exploited in the definition of a composite property:

$$\psi^{2,1} = (\chi^2 - \chi^2_{w0}) - (\chi^1 - \chi^1_{w0}) \left(\frac{\partial \chi^2}{\partial \chi^1}\right)_w.$$
 (5)

The composite property is itself conservative because both constituent properties are individually conserved, but its real utility lies in the fact that it is zero everywhere in the ambient water column. The conservation equation [Eq. (1)] for any such property therefore simplifies to

$$Q\psi^{2,1} = Q_i \psi_i^{2,1}.$$
 (6)

With the application of (6) it does not matter what quantities of ambient waters from various parts of the water column contribute to the final measured properties because these are determined only by the ice properties.

The geometrical interpretation of this is shown on the bivariate plot in Fig. 2. Lines of constant $\psi^{2,1}$ lie parallel to the ambient trend. Only a nonconservative process or an admixture of water with properties that do not lie on the ambient line can produce nonzero values of $\psi^{2,1}$. Melting can produce properties anywhere within a triangle bounded by the ambient trend and by meltwater mixing lines passing through the two points representing the pure end members of the ambient water column. Properties toward the pure ice apex of the triangle are unattainable in practice, because of a physical limit placed on the meltwater content by the amount of heat available for melting in the ambient water. Once sufficient ice has melted to cool the water to the in situ freezing point, no additional meltwater can be added to



FIG. 2. Bivariate plot illustrating observed properties (χ^1, χ^2) produced as ice with properties (χ^1_i, χ^2_i) melts into a mixture of ambient waters, having combined properties $(\chi^1_{ef}, \chi^2_{ef})$. If the ambient waters have properties that lie on a single line, the locus of all possible observations forms the triangle shown, with the ambient properties forming one vertex (bold line) and the ice properties the opposite apex. Points (some numbered) indicate the meltwater fractions associated with selected positions on three mixing lines. Contours of the composite property $\psi^{2,1}$ [defined in Eq. (6)] are indicated by dash-dotted lines.

the mixture. A subsequent change in depth would alter the freezing point and allow further phase changes, but equilibrium would again be reached at the new freezing point temperature. Equation (3), with the effective potential temperature, θ_i^* , substituted for the ice property and θ_f for the mixture property, implies the following limit on the meltwater fraction:

$$\frac{Q_i}{Q} \le \left[1 + \frac{L + c_i(\theta_f - \theta_i)}{c_p(\theta_w - \theta_f)}\right]^{-1}.$$
(7)

In polar oceans the second term in the parentheses is much larger than 1, and the meltwater fraction is limited to about 1% (°C)⁻¹ of temperature elevation above freezing ($\theta_w - \theta_f$) in the ambient water.

For any point within the triangle, the meltwater fraction may be calculated from the geometry of Fig. 2. Addition of meltwater forces water properties to evolve along a straight line toward the point (χ_i^1, χ_i^2) . An effective source water $(\chi_{ef}^1, \chi_{ef}^2)$ can therefore be divined by extrapolating the meltwater mixing line passing through the observed properties (χ^1, χ^2) to its intersection with the ambient trend. The position of the observed point along this line indicates the meltwater fraction, just as in Fig. 1. The same is true for any point on any meltwater mixing line, and the points of equal meltwater fraction line up along contours of constant $\psi^{2,1}$ [Eq. (6)].





FIG. 3. (a) Map showing the coastline and grounding line of Antarctica (British Antarctic Survey 1993). (b) Enlargement of the boxed area in (a) showing the locations of stations (with station numbers) at which the data under discussion were collected. The coast is shown as a solid line, while the shading indicates grounded ice. Ice front positions are from 1966 (British Antarctic Survey 1993).



FIG. 4. Salinity profiles measured to within a few meters of the seabed at the stations shown in Fig. 3b. In each panel the left-hand profile is plotted true to the indicated scale, while the remaining profiles have been shifted sequentially by 0.5 psu to the right.

It is important to note that measurements taken through a water column that is affected to varying degrees by meltwater may exhibit a bivariate trend quite unlike the simple meltwater mixing line shown in Fig. 1. The position of each measured point is determined not only by the fraction of meltwater, but also by the proportions of ambient water from different levels in the water column that go to make up the effective source water. A further point is that if the ambient trend line happens to be the same as the meltwater mixing line, the meltwater fraction is indeterminate. In practice, nonuniformity of the end member water types over a study area, or nonlinearity of the trend between them, could impart a finite width to the ambient line. The width would translate directly into an uncertainty in the observed value of $\psi^{2,1}$, and hence an uncertainty in the calculated meltwater fraction.

3. An example: Tracing meltwater in Pine Island Bay

a. Observations

In March 1994 a number of hydrographic stations were occupied in and near Pine Island Bay in the southeastern Amundsen Sea. The details and implications of this dataset have been presented and discussed in some detail by Jacobs et al. (1996), Jenkins et al. (1997), and Hellmer et al. (1998). Here the data will be revisited in order to give a practical illustration of the principles outlined in the preceding section. Three stations were situated within 0.5 km of Pine Island Glacier, where a considerable meltwater impact might be expected, while others sampled the far-field ambient conditions in the bay and on the outer continental shelf (Fig. 3). All showed shallow haloclines at depths of about 50-100 m (Fig. 4), which will be regarded here as the limit of any ongoing atmospheric influence on the water column. The analysis below will be concerned with the waters below the halocline, where conservative mixing can be assumed to be the dominant control on water properties.

b. Interpretation of $\psi^{\theta,S}$ values

In θ -S space the stations on the outer continental shelf show a simple structure (Fig. 5a), with WW, characterized by temperatures close to the surface freezing point, overlying CDW. In these autumn observations there is no sign of any warm, surface layer, which must already have been extinguished by the onset of freezing. Nearer Pine Island Bay a similar structure is observed, but with the addition of warm features punctuating the layer of WW (Fig. 5b). These warm intrusions are too deep in the water column to be remnants of summer surface warming (Fig. 6), and their likely origin will become apparent below. The three stations occupied near the calving front of Pine Island Glacier look dramatically different (Fig. 5c). Winter water is almost entirely ab-





FIG. 5. Potential temperature vs salinity plots for stations (a) 89–91, on the outer continental shelf, (b) 95–96, on the inner continental shelf, and (c) 92–94, at the calving front of Pine Island Glacier. Station locations are shown in Fig. 3b. Dots indicate averages of data collected over sequential 1-db pressure intervals. Numbered solid lines are isopycnals, labeled with values of σ_0 , and the solid line near the bottom of the diagram (labeled T_j) indicates the surface freezing point. In (c) dashed lines represent approximations to the ambient trend in the main thermocline (bold) and a meltwater mixing line passing through the deepest waters, while dash–dotted lines are contours of meltwater fraction.

sent, and warm intrusions appear in the main thermocline at depths of up to 500 m on stations 93 and 94 (Fig. 6). At station 92 the ambient thermocline itself is absent and the shallow surface layer tops a water column exhibiting an altogether different θ -S trend.

The distribution of data in Fig. 5c can be understood by comparison with Fig. 2. The observations fall within the upper apex of the triangle that may theoretically be generated by melting. One edge of the triangle is formed by the ambient trend in the lower part of the water column, an approximately straight line connecting CDW with the most saline WW. The large number of points, primarily from station 92, that fall close to a second straight line suggest that these characterize the upper edge of the triangle, formed by mixing between meltwater and the most saline ambient water. An extrapolation of this trend to zero salinity gives an intercept of -92.5° C, the effective potential temperature of ice at -15° C. Hellmer et al. (1998) fitted a slightly steeper line to these data and derived an ice temperature of -20° C, the large discrepancy indicating the insensitivity of the gradient to the unknown ice temperature.

With the ambient trend and the ice properties both identifiable, the meltwater fraction can be calculated for every point. This is simply the observed value of the composite property $\psi^{\theta,S}$ normalized by the value in ice, $\psi_i^{\theta,S}$ [Eq. (6)]. A problem arises in that the theoretical triangle is truncated at its intersection with the surface



FIG. 6. Potential temperature profiles measured to within a few meters of the seabed at the stations shown in Fig. 3b. In each panel the left-hand profile is plotted true to the indicated scale, while the remaining profiles have been shifted sequentially by 1.5° C to the right.

freezing point line. The pure ambient water column clearly has a bilinear characteristic (Fig. 5a) so that the composite property $\psi^{\theta,S}$, defined in terms of the θ -S trend through the main thermocline, is nonzero in the WW. Although pure WW is hardly apparent at all beneath the surface halocline on stations 92–94, with the exception of the cold layer near 300 m on station 94, any admixture of ambient water that lies off the deep θ -S trend will contaminate the signature of melting contained in the measured $\psi^{\theta,S}$ values.

c. Interpretation of $\psi^{\delta,\theta}$ values

The properties of WW are set by the cycle of melting and freezing at the surface, a process that changes the salinity of the water while maintaining the temperature at the freezing point. This gives rise to the problematic bilinearity of the ambient θ -S trend. If a tracer existed that either remained constant during freezing or varied in the same manner as salinity, it would be possible to find a composite property that was close to zero throughout the ambient water column. The latter seems unlikely, but stable isotopes behave approximately like the former. Although there is some fractionation on freezing, for the range of WW salinities encountered at the stations discussed here the changes caused by fractionation should fall within the precision of the δ^{18} O measurements. All WW observations should therefore collapse to a single point in $\delta^{18}O-\theta$ space and the whole WW-CDW water column should then conform to a single straight line.

Figure 7 shows data from stations 92–94. The cluster of points near the surface freezing point includes values from the surface mixed layer and the WW at 300-m



FIG. 7. Plot of δ^{18} O vs potential temperature for stations at the calving front of Pine Island Glacier. Station locations are shown in Fig. 3b. Dashed lines represent approximations to the ambient trend between CDW and WW (bold) and a meltwater mixing line passing through the deepest waters, while dash-dotted lines are contours of meltwater fraction. Vertical bars indicate measurement precision.

depth on station 94. The ambient trend has been fitted through these points and through the two pure CDW values near 1°C. The melt line drawn on the diagram trends from the CDW values toward a δ^{18} O of -30%, the approximate mean value of surface accumulation in the inland parts of the catchment basin (Hellmer et al. 1998), at a potential temperature of -92.5° C. There are several points, mostly from station 92, that fall close to this side of the triangle that may theoretically be generated by melting. Using the ambient trend to define the composite property $\psi^{\delta,\theta}$, lines of constant meltwater fraction may be drawn on the $\delta^{18}O-\theta$ diagram shown in Fig. 7. The precision of the δ^{18} O measurements $(\pm 0.03\%)$, compounded with the sparsity of data with which the ambient trend can be defined, leads to an uncertainty of $\sim 3.5\%$ in the inferred meltwater fractions.

d. Interpretation of $\psi^{o,s}$ values

A surprising alternative is suggested by the O_2 -S trends of the outer shelf stations (Fig. 8a), which are approximately linear throughout most of the water column. Conservative mixing is presumably the source of the linearity in the main thermocline (salinity ≥ 34 psu), but it is not clear why this trend should be mimicked so closely in waters with more recent atmospheric contact. A small break in slope can be discerned near 34 psu, but an assumption of complete linearity only leads to small errors in $\psi^{O.S}$ except at the very top and bottom of the water column. Preservation of a similar ambient trend over a large part of the continental shelf (Figs. 8a and 8b) is





FIG. 8. Dissolved oxygen vs salinity plots for stations (a) 89– 91, on the outer continental shelf, (b) 95–96, on the inner continental shelf, and (c) 92–94, at the calving front of Pine Island Glacier. Station locations are shown in Fig. 3b. Dots indicate averages of data collected over sequential 1-db pressure intervals, while stars in (c) indicate measurements on discrete bottle samples. In (c) dashed lines represent approximations to the ambient trend through the CDW and WW (bold) and a meltwater mixing line passing through the deepest waters, while dash–dotted lines are contours of meltwater fraction.

probably helped by the perennial sea ice that normally covers much of the region (Jacobs and Comiso 1997).

The stations closer to Pine Island Bay show additional features (Fig. 8b), with distinct layers of low dissolved oxygen concentration apparent on station 95. Comparison of Figs. 6 and 9 indicates that these features are associated with the warm intrusions noted within the WW. Once again, the data from stations 92–94 (Fig. 8c) look dramatically different. The ambient trend is barely apparent, with most data points falling significantly below it. Variations in dissolved oxygen measured at stations 93 and 94 appear to be negatively correlated with potential temperature variations (Figs. 6 and 9), but a similar correspondence is less apparent on station 92, probably be-

cause of the noisy oxygen trace. A meltwater trend is drawn in Fig. 8c, extending from the most saline ambient water and intersecting the zero salinity axis at a dissolved oxygen concentration of 25.2 ml 1^{-1} , the value for pure meltwater (Hellmer et al. 1998). There are many data points, particularly from station 92, that fall along this side of the triangle of possible O₂–S values.

e. Results of analyses

The meltwater fractions inferred from Figs. 5c, 7, and 8c are plotted as a function of depth on each station in Fig. 10. Below the surface halocline the various calculations are in good agreement. The small differences



FIG. 9. Dissolved oxygen profiles measured to within a few meters of the seabed at the stations shown in Fig. 3b. In each panel the left-hand profile is plotted true to the indicated scale, while the remaining profiles have been shifted sequentially by $1.5 \text{ ml } 1^{-1}$ to the right.

apparent between the results obtained using the bottle dissolved oxygen data and those obtained from the continuous θ -S measurements are greater than can be accounted for by errors in the observations and reflect the uncertainties introduced by the nonlinearity of the ambient trends. Where values obtained from bottle and continuous measurement of $\psi^{O,S}$ differ, this is the result of the lower precision of the continuous measurements. This is particularly apparent on station 92 and in the surface mixed layer on all stations. The relatively large discrepancy between calculations based on $\psi^{\theta,S}$ and $\psi^{O,S}$ immediately below the surface layers is probably caused by contamination of the former signal by an admixture of WW having nonzero $\psi^{\theta,S}$. Results above the surface halocline should be ignored, as here any signature of melting will be obliterated by interaction with the atmosphere.

The generally good agreement apparent in Fig. 10 lends support to the hypothesis that the composition of the water column observed at these stations is a result of conservative admixing of meltwater to an ambient water column that would otherwise display simple bivariate trends akin to those observed at the outer shelf stations. At station 92 the water column between 600 and 100 m comprises four layers with meltwater fractions rising from about 7 to about 18‰. At stations 93 and 94 concentrations of meltwater are generally lower and more variable, with discrete layers interspersed by ambient waters appearing at 94. The warm features below the surface halocline on 93 and 94 (Fig. 6) are apparently associated with meltwater fractions of \sim 10‰, while the cold layer on station 94 contains no meltwater.



FIG. 10. Profiles of calculated meltwater fraction for the three stations next to the calving front of Pine Island Glacier (Fig. 3b). Lines indicate calculations based on continuous measurements of $\psi^{0,S}$ (dashes; see Fig. 5c) and $\psi^{0,S}$ (dots; see Fig. 8c). Crosses indicate results obtained from discrete measurements of $\psi^{\delta,\theta}$ (with error bars; see Fig. 7) and stars are calculations based on discrete $\psi^{0,S}$ observations (see Fig. 8c).

4. Tracer anomalies produced by melting

A comparison of the profiles in Fig. 10 with those in Figs. 4, 6, and 9 indicates the impact of melting ice on the ambient water column in Pine Island Bay. At any particular depth, the presence of meltwater is associated with prominent intrusions of warm water, having a low dissolved oxygen content, while the influence of the glacier is less apparent in the salinity profiles. Prominent features between depths 50 and 300 m on station 95 (Figs. 6 and 9) and a less prominent feature between depths 100 and 250 m on station 96 also have a high temperature, low oxygen signature, which is probably produced by melting. Considering the low effective potential temperature of the melt and that it is three times supersaturated in oxygen, these observations may seem surprising. They are an indication of the role played by upwelled CDW in setting the properties of the meltwater-seawater mixtures. The impact of melting in Pine Island Bay is different to that observed in the Ross and Weddell Seas, where melt-laden outflows from beneath the large ice shelves are the coldest waters on the continental shelves (e.g., Foldvik et al. 1985; Jacobs et al. 1985).

The aim of this section is to derive general expressions for the sign and size of the tracer anomalies produced by melting. These can be quantified as the difference between the observed value of a property and the value of that same property in the undisturbed ambient water column on the same isopycnal (McDougall 1990). Because composite properties are zero everywhere in the ambient water column, (6) can be written

$$Q(\psi^{2,1} - \psi^{2,1}_{\sigma}) = Q_i \psi^{2,1}_i, \tag{8}$$

where the subscript σ indicates the ambient property at the equilibrium density level of a meltwater–seawater mixture having property $\psi^{2,1}$. Considering first the active tracers and using the definition of the composite property $\psi^{\theta,s}$ given in (5), Eq. (8) can be rewritten as

$$Q\left[\Delta\theta - \Delta S\left(\frac{\partial\theta}{\partial S}\right)_{w}\right]$$
$$= Q_{i}\left[\left(\theta_{i} - \theta_{w0}\right) - \left(S_{i} - S_{w0}\right)\left(\frac{\partial\theta}{\partial S}\right)_{w}\right], \quad (9)$$

where the isopycnal anomalies in potential temperature and salinity are given by $\Delta \theta = \theta - \theta_{\sigma}$ and $\Delta S = S - S_{\sigma}$. Introducing the local isopycnal slope, defined by

$$\left(\frac{\partial\theta}{\partial S}\right)_{\sigma} = \frac{\Delta\theta}{\Delta S},\tag{10}$$

and the slope of the meltwater mixing line passing through the reference point, given by

$$\left(\frac{\partial\theta}{\partial S}\right)_{i0} = \frac{\theta_{w0} - \theta_i}{S_{w0} - S_i},\tag{11}$$

the salinity anomaly can be expressed as

$$\Delta S = S_{w0} \frac{Q_i}{Q} \left[\frac{(\partial \theta / \partial S)_w - (\partial \theta / \partial S)_{i0}}{(\partial \theta / \partial S)_\sigma - (\partial \theta / \partial S)_w} \right], \qquad (12)$$

where the fact that the ice salinity is zero has been used. The local isopycnal slope is almost always positive, the only exception being for the case of near-freezing waters fresher than ~24 psu, where a density maximum occurs in the liquid phase. Away from this limited region of θ -S space the sign of the potential temperature and salinity anomalies generated by melting ice are the same [Eq. (10)]. Whether they are positive or negative depends on the precise geometry of the three lines whose gradients appear in (12), but irrespective of the sign, the size is always proportional to the meltwater content, Q_i/Q .

Figure 11 illustrates the principles for the region of parameter space most relevant to the Southern Ocean. The near-surface layers are primarily salt stratified, so the ambient θ -S trend invariably has a gradient that is lower than that of the local isopycnals. The denominator in (12) is then positive. Melting cools and dilutes the ambient waters, and the properties evolve along a meltwater mixing line. Because this line also has a gradient that is less than the isopycnal slope in this region of θ -S space, a mixture of meltwater and ambient water will rise through the water column. If the ambient trend in θ -S is steeper than the melt-induced trend [Ambient 1, numerator of (12) positive], the mixture will be neutrally buoyant at a level in the water column where it is warmer and saltier than the surroundings. The greater the concentration of meltwater, the more pronounced



FIG. 11. Potential temperature vs salinity diagram illustrating properties of two idealized ambient water columns (bold lines), with meltwater mixing lines (dashed lines) extending from the top, the bottom, and the midpoint of each. Open circles on $\sigma_0 = 27.4$ indicate possible properties produced by melting into ambient 1, while open squares on $\sigma_0 = 27.2$ indicate the same for ambient 2.

the cooling and dilution of the effective source water and the higher in the water column, relative to the source level, the intrusion occurs. This leads to a greater apparent warmth and saltiness in the melt-laden water. The steepness of the isopycnals means that the salinity anomalies are much smaller than the potential temperature anomalies, although this effect is most evident near the surface. At greater depths, the increased pressure has the effect of reducing the isopycnal slope and hence increasing the size of the anomalies in salinity at the expense of those in potential temperature. By analogous arguments, if the ambient trend is less steep than the meltwater mixing line [Ambient 2, numerator in (12) negative], cold, fresh anomalies result. If the two lines happen to coincide (numerator is zero), no anomalies are seen.

Waters on the continental shelves of the Amundsen and Bellingshausen Seas conform to Ambient 1, the θ -S trend between CDW and WW being steeper than the meltwater mixing line. Hence the warm anomalies observed in Pine Island Bay and reported by Potter et al. (1988) near the front of George VI Ice Shelf, in the Bellingshausen Sea. By contrast, in the Ross and Weddell Seas the θ -S trend in the shelf waters follows the freezing point relationship, which has a θ -S gradient close to zero (Fig. 11). Melting produces cold, fresh anomalies and, because the ambient waters are already at the surface freezing point, the product of melting must be a water mass that is potentially supercooled, thereby conforming to the conventional definition of ISW.

It is harder to quantify the likely anomalies in passive





FIG. 12. (a) Contour plot of $(\partial \theta / \partial S)_w$ between the temperature maximum (CDW) and temperature minimum (WW) layers [data are from Olbers et al. (1992); units are °C/psu]. Regions where these two layers cannot be defined are indicated by white (north of the Polar Front) and black (near the continental shelves) shading. (b) Vertical heat transport associated with ice melting within or below the main CDW–WW thermocline as a function of $(\partial \theta / \partial S)_w$.

tracers because the gradients along isopycnals [equivalent to (10)] are entirely arbitrary. However, for the near-surface, near-freezing waters of primary interest to this discussion, salinity is by far the most important active tracer. To a reasonable approximation, the isopycnal salinity anomaly, ΔS , can be taken to be zero, and (9) can be rearranged to give an approximate version of (12), appropriate for any property, χ :

$$\Delta \chi = S_{w0} \frac{Q_i}{Q} \left[\left(\frac{\partial \chi}{\partial S} \right)_w - \left(\frac{\partial \chi}{\partial S} \right)_{i0} \right].$$
(13)

The separation of slopes in χ -S space can therefore give a good indication of the sign and size of the property anomalies. The geometry is equivalent to that shown in Fig. 11, but with the isopycnals approximated by straight lines perpendicular to the salinity axis. In Pine Island Bay, the high (negative) slope of the ambient O_2 -S trend compared with that induced by melting (Fig. 8c) is the reason for the observed negative oxygen anomalies. In contrast the geometry of the $\delta^{18}O-\theta$ diagram (Fig. 7) can give no indication of the isopycnal anomalies of either tracer. Hellmer et al. (1998) show that for salinities ≥ 34 psu all data from stations 92–95 collapse onto a single line in $\delta^{18}O-S$ space. The lack of any discernible separation between the ambient $\delta^{18}O-S$ trend and the meltwater mixing line means that, in this particular environment, the isopycnal $\delta^{18}O$ anomalies generated by melting ice are negligible.

From the above analysis it is possible to outline some of the effects of melting icebergs on the upper layers of the Southern Ocean. Icebergs of shallow draft that melt entirely within the WW will always cause cooling because the ambient θ -S trend closely follows the freezing point curve. However, larger bergs are thick enough to reach into or through the main thermocline, and melting at these depths will generate warm intrusions wherever the θ -S trend between CDW and WW is steeper than the meltwater mixing line. For typical CDW characteristics and iceberg temperatures, the meltwater mixing line has a gradient in the range 2.5–2.8°C/psu. Figure 12a shows $(\partial \theta / \partial S)_{w}$ for the main thermocline throughout the Southern Ocean between the Polar Front and the continental margin. The 2.5°C/psu contour is the most interesting as it marks the approximate boundary between regions where melting ice generates warm intrusions and regions where melting causes cooling. For most of the Southern Ocean the former condition holds (Fig. 12a). An estimate of the net effect on the WW heat budget can be obtained from Eqs. (10) and (12), a rearrangement of which yields an expression for the heat flux associated with melt-driven intrusions in the main thermocline

$$Qc_{p}\Delta\theta = Q_{i}c_{p}\left(\frac{\partial\theta}{\partial S}\right)_{\sigma}S_{w0}\left[\frac{(\partial\theta/\partial S)_{w} - (\partial\theta/\partial S)_{i0}}{(\partial\theta/\partial S)_{\sigma} - (\partial\theta/\partial S)_{w}}\right].$$
 (14)

Figure 12b illustrates the behavior of this expression for the range of $(\partial \theta / \partial S)_w$ shown in Fig. 12a, assuming values of 34.7 psu for the salinity of CDW, 4000 J kg⁻¹ K⁻¹ for the specific heat capacity of seawater, 2.7°C/psu for the slope of the meltwater mixing line, and 15°C/psu for the isopycnal slope. The rise of the vertical heat transport toward infinity as the denominator in Eq. (14) approaches zero is unrealistic and serves to highlight the limitations of (14). The isopycnal temperature anomaly, $\Delta \theta$, is inferred from the separation in slope of the ambient and melt-induced θ -S trends, but the ambient trend can be assumed to be constant only within the thermocline. For intrusions that appear within the layer of WW where the θ -S trend is different, (14) represents an overestimate that becomes more extreme as the thermocline stability decreases. Nevertheless, Eq. (14) can be used to place a useful upper bound on the vertical heat flux.

The data contoured in Fig. 12a give a mean $(\partial \theta / \partial S)_w$ between the temperature maximum and temperature minimum layers of 4.7°C/psu, which corresponds to a vertical heat transport of 400 kJ kg⁻¹ of melted ice (Fig. 12b). Jacobs et al. (1992) estimate the total calving flux from Antarctica to be 64×10^6 kg s⁻¹, and if half of this ice melts at depths below the temperature minimum, the mean upward heat flux over the 35×10^6 km² between the Polar Front and the continental margin amounts to 0.4 W m⁻². Although this is not a significant heat flux overall, icebergs are not uniformly distributed throughout the Southern Ocean. They tend to follow well-defined drift tracks determined by prevailing winds and currents and along which their local impact could be significant. An upper limit on the local vertical heat flux can be estimated by assuming that the area influenced by upwelling is equal to that of the icebergs themselves. Taking a mean density of 850 kg m⁻³, a mean thickness of 250 m and a half-life of 3-6 yr for icebergs in the Southern Ocean (Jacobs et. al. 1992) gives a total area in the range 4-8 ($\times 10^4$ km²). Venting all the heat through this much smaller area would imply a local vertical heat flux in the range 150-300 W m⁻².

5. Summary

Melting ice may have a complex effect on any given oceanic water column. Although the direct effect will always be one of cooling, dilution, depletion in heavy isotopes, and enrichment in most dissolved gases, the subsurface dilution causes upwelling (at least in cold waters, where the associated cooling has little impact on the density). Convection promotes mixing and further melting and this continues until the mixture of ambient water and meltwater has attained neutral buoyancy. If the meltwater content of the mixture is to be evaluated, its properties must be compared with those of the effective source water, but these reflect an arbitrary combination of ambient waters. The problem only becomes tractable for observed tracers that exhibit both invariance throughout the ambient water column and a large ice-ocean contrast. Composite tracers, formed in the manner suggested by McDougall (1990), prove useful in this respect. Provided a linear bivariate trend can be identified in the ambient water, a composite tracer that satisfies the condition of invariance can be defined. To the condition of a large ice-ocean contrast in the

value of the composite property must be added the requirement that the ambient and melt-induced bivariate trends differ. In the practical example discussed above, the combinations of δ^{18} O and potential temperature and of dissolved oxygen and salinity have proved useful tracers. In the latter case, the availability of continuous measurements of both components and the wide separation between the ambient and melt-induced trends in O₂/S space (Fig. 8c) have made this composite particularly valuable. Such utility is rather surprising, considering that dissolved oxygen is strictly nonconservative, but beneath an ice cover nonconservative processes that influence oxygen levels may be largely inactive.

Near a floating ice mass the water column will most likely consist of one or more melt-laden flows interleaving in the ambient waters. To quantify the changes brought about by melting, it is necessary to compare the properties of the intrusions with those of the ambient waters that they displace. Without a knowledge of tracer variations throughout the parts of the water column that contribute ambient water to the intruding mixtures, neither the size nor the sign of the isopycnal anomalies can be predicted. The access of CDW to the cavity beneath Pine Island Glacier induces the highest observed melt rates beneath any Antarctic ice shelf (Jenkins et al. 1997), and the water column adjacent to the glacier is up to 3-6 times more concentrated in melt than ISW observed in the Weddell Sea (Schlosser et al. 1990). Nevertheless, the melt-laden outflows are comparatively warm, salty, and low in dissolved oxygen, while their δ^{18} O signature is indistinguishable from that of their surroundings. Melt-induced warming is particularly significant as it represents a flux of heat to the base of the surface mixed layer that may impact the evolution of the sea ice cover. Conditions throughout most of the Southern Ocean are such that the melting of large icebergs will have a similar effect on the WW.

This paper has focused on how melting ice influences the structure of a given water column and has exposed some effects that may appear counterintuitive. The intuitive assumption that melting ice causes cooling, dilution, depletion of heavy isotopes, and enrichment of dissolved gases is valid only if applied in an integral sense to a volume of water that at least reaches the maximum draft of the ice and covers a surrounding area of unknown extent. Care is therefore needed in inferring the presence of ice in the ocean from paleo-temperature and -isotope records. If the proxies only sample a limited part of the water column, the trends that are associated with changes in the concentration of meltwater may not be clear. Care is also required in assessing the impact of ice discharge fluctuations on ocean circulation. The upwelling of deep ocean properties through the pycnocline is a potentially important result of the interaction between ice and ocean, and in this respect the impact of melting ice differs from that of an equivalent freshwater flux at the ocean surface.

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