The Melting of Ice in the Arctic Ocean: The Influence of Double-Diffusive Transport of Heat from Below

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ABSTRACT

This investigation was originally prompted by two oceanographic observations: an increased rate of melting of sea ice in the Arctic Ocean, and the advance of an anomalously warm tongue of Atlantic water intruding across the Arctic below the halocline over the past few decades. A series of laboratory model experiments has previously been carried out to explore the possibility that the extra heating at depth could be responsible for the enhanced melting rate. These experiments have demonstrated that a one-dimensional heat flux from below through a series of double-diffusive layers can in principle lead to faster melting of floating ice. However, it is now essential to test these ideas quantitatively under ocean conditions and to compare the results with other possible mechanisms of melting.

A simple calculation shows that there is enough heat in the intruding Atlantic water to melt all the ice in the Arctic in a few years if all the heat could be brought to the surface in this time. The vertical double-diffusive transport of heat is slower than this, but it is large enough to make a substantial contribution to the increased rate of melting over the last three or four decades. Another proposed mechanism for melting is the solar input to the surface mixed layer from the atmosphere. In particular years when detailed measurements and calculations have been made, this atmospheric input can explain both the seasonal cyclic behavior of ice and the increased melting rate. Given the large heat content in the intruding Atlantic layer, however, it seems worth exploring further other advective two-dimensional mechanisms that could transport this heat upward more rapidly than the purely vertical double-diffusive convection. For example, dense salty water produced by freezing on the shelves around the Arctic Basin could flow down the slope and penetrate through the halocline, thus mixing with the warm water and bringing it to the surface.

1. Introduction

The ideas discussed in this paper were first presented at the 16th National Congress of the Australian Institute of Physics in February 2005 and were published as an abstract in the handbook for that meeting (Turner 2005). During the following year a draft of a more detailed manuscript was prepared, but it was not then submitted for publication because of my reservations about how interesting the results would be to expert observers of the Arctic Ocean. However, the extremely low value of sea ice extent in 2007, followed by the recent publication of American Geophysical Union Geophysical Monograph 180 on arctic sea ice decline (DeWeaver et al. 2008), have reawakened my interest in making these ideas more widely known.

In Geophysical Monograph 180, Cullather and Tremblay (2008) give (on p. 192) a Web site reference to a report on a series of three Arctic Observation Integration Workshops sponsored by the National Science Foundation in March 2008. Working Group 1, Documenting and Understanding Sea Ice Change, pointed to uncertainties and needs related to understanding the state of the atmosphere, the thermodynamics and dynamics of the sea ice and, for the ocean, the required measurements and understanding of processes, including (the first two items are quoted verbatim from the Working Group Report):

1) interannual quantification of the oceanic heat fluxes into the Arctic from the Pacific and the Atlantic, and
2) quantification of the processes by which this heat may influence the sea ice (e.g., by upward heat flux

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either by diffusive or eddy processes, or by mechanical, topography-related processes).

These are precisely the processes that are the subject of the present paper.

A brief summary will now be given of the observations that originally motivated this study and the laboratory experiments suggested by them. More detailed discussions have been presented in the two papers by Turner and Veronis (2000) and Turner and Veronis (2004). The original examples given when this paper was first drafted will be retained, but some more recent references will be added where appropriate.

A very clear outline of the measurements of the melting of sea ice in the Arctic was given by Rothrock et al. (1999). They collated data acquired on submarine cruises during 1958–76 and 1993–97 and plotted the changes in draft of the floating ice between the two periods on a map of the whole Arctic Ocean. All measurements revealed a decrease in thickness, the means in three areas being −0.5 to −1.0 m, −1.0 to −1.5 m, and −1.5 to −2.0 m. These represent a decrease of 40% in the volume of sea ice in the Arctic Ocean over three decades (i.e., a mean rate of decrease of more than 1% yr^{-1}). This estimate has been confirmed by Wadhams and Davis (2000), who included in their assessment submarine data from different areas of the Arctic. The more recent data obtained during the summer of 2007 have revealed an even more rapid rate of melting and shrinking in the area of the sea ice in the Arctic Ocean (see, e.g., Deser and Teng 2008).

Although it is not the main focus of this paper, it is worth highlighting a possibly far-reaching implication of an increase in freshwater outflow from the Arctic into the Atlantic Ocean. As pointed out by Stocker and Wright [1991; and since then by others, notably Bryden et al. (2005)], the increase in fresh, light, surface water could cause a rapid transition in the Atlantic's deep circulation by reducing the strength of, or even cutting off, the sinking arm of the "conveyor belt" in the North Atlantic. This is potentially the most rapidly acting consequence of climate change; it could have a profound effect on the climate of the United Kingdom and northern Europe, though the sea level rise due directly to this would be negligible. The indirect effects could be much larger, however. An increase in the area of open water in the Arctic Ocean could accelerate the rate of melting of the Greenland ice sheet, which would certainly cause a rise in sea level.

The intrusion of a tongue of anomalously warm water from the Atlantic into the Arctic Ocean during the 1990s has been documented in detail by Carmack et al. (1997), as it has extended farther and farther across the Arctic Basin during this period. [Previous observations of the intruding Atlantic water, going back to the 1970s, have been discussed by Aagaard and Greisman (1975), whose estimates of the associated heat flux will be discussed further in the next section.] Carmack et al. (1997) have shown that, due to double-diffusive convection, the extending warm intrusion forms persistent multiple layers, 40–60 m thick, spreading out laterally in a coherent manner through the Atlantic water and upper deep waters of the Arctic, below a shallower halocline (see Fig. 1). The authors noted that these layers can support both diffusive and finger convection, and they believe that they are self-organizing and self-propelled by these double-diffusive transports. These observations played a big part in focusing attention on the previously neglected two-dimensional double-diffusive processes and led to the laboratory study of Turner and Veronis (2000).

2. The heat required to melt the Arctic sea ice, compared with that in the intrusion

How much heat is in fact needed to melt the sea ice in the Arctic Ocean, and could this be supplied by the intruding Atlantic water? The heat required to melt 1-m thickness of ice is 8 kcal cm^{-2} (3.35×10^8 J m^{-2}) for a latent heat of fusion of 80 cal gm^{-1}. Aagaard and Greisman (1975) have calculated, using direct observations of temperature and velocity in the inflowing West Spitzbergen Current, that the average net rate of transport of heat into the Arctic by this Atlantic inflow was about 6.7×10^{10} kW, with a mean temperature of the intruding water of 2.2°C. This current accounted for more than half the total heat flux through all the passages into the Arctic during a yearlong period of direct current measurements in 1971–72. Aagaard and Greisman's result is probably an underestimate of the present subsurface advective heat flux from the Atlantic into the Arctic. It does not take into account the later systematic increase in temperature in the core of the Atlantic layer during the 1990s and its extension across the Arctic, as reported by Carmack et al. (1997) and discussed above.

Now let us suppose that the above heat flux of 6.7×10^{10} kW is distributed uniformly over the whole Arctic Ocean, which has a surface area of approximately 10^7 km^2. The basin-averaged heat loss from this layer, assuming a steady state, is about 6.7 W m^{-2} or 2.1×10^8 J m^{-2} yr^{-1}. Hence in 12–18 months there *is* enough heat in the Atlantic layer alone to melt 1 m of sea ice (or all the ice, with mean thickness of 2.5 m, in 4 yr) *if* this heat were able to reach the surface under the ice pack over these time scales. The local melting rate would be faster if an increased heat flux were acting over a more limited region. But the mechanisms that could bring the

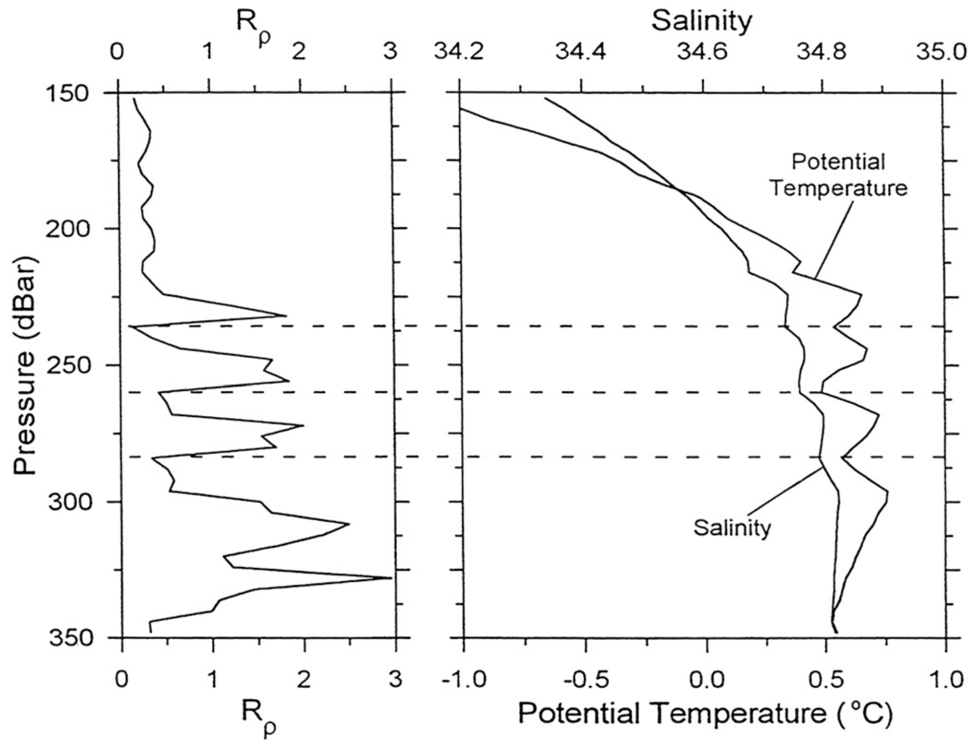

FIG. 1. Vertical profiles of temperature, salinity, and density ratio R_ρ at a station at the edge of the Canada Basin. [Obtained during the Canada/United States 1994 Arctic Ocean Section Experiment, reported by Carmack et al. (1997).]

heat in the subsurface layer to the surface over this very short time scale remain to be properly evaluated.

Turner and Veronis (2004) have carried out a series of laboratory experiments to examine explicitly the effect of the vertical transport of heat from hot layers, intruding laterally below the surface into a salinity gradient, on the melting of ice floating on the surface. The melting rate in these experiments was found to depend systematically on the magnitude of the salinity gradient and on the presence or absence of heating from below. As shown in Fig. 2, the melting rate decreased as the salinity gradient was increased, with the melting rate being higher in the heated runs for low salinity gradients. Above a certain intermediate gradient there was no significant difference between the unheated and heated runs. Although these results demonstrate the potential effect of vertical double-diffusive transport of heat, so far as the ocean is concerned they are only qualitative, and they point to the need for the quantitative study presented in the next section.

3. Quantitative estimate of vertical heat flux due to double-diffusive processes

In various places in the ocean there are clear examples of well-defined layers that are undoubtedly formed by double diffusion. In the Arctic, in particular, Neal et al. (1969) obtained the profiles of temperature underneath ice island T-3, which are reproduced in Fig. 3. The stratification is clearly in the "diffusive" sense, with an unstable temperature distribution and a stabilizing salinity gradient (not shown). With such well-resolved layering it is possible to calculate the vertical heat (and salt) fluxes through the interfaces and the convecting layers between them, using laboratory results and the theory based on them. But rarely are such clearly resolved observations available, and methods have been developed to relate the fluxes to mean gradients of T and S.

Kelley (1984) proposed a dimensional argument, based on the mean gradients, the relative contributions of S and T to the density gradient, and also the molecular diffusivity κ_T, to derive expressions for a typical length scale and the heat flux for the diffusive regime when the layering is poorly resolved. A later review article by Kelley et al. (2003) has amplified and clarified these results, as well as critically examining the applicability of laboratory flux laws from the laboratory to the ocean and the effect of lateral variations.

To derive an expression for the length scale H_0 in a double-diffusive region, we need to know the mean buoyancy frequency N over the depth of interest, the

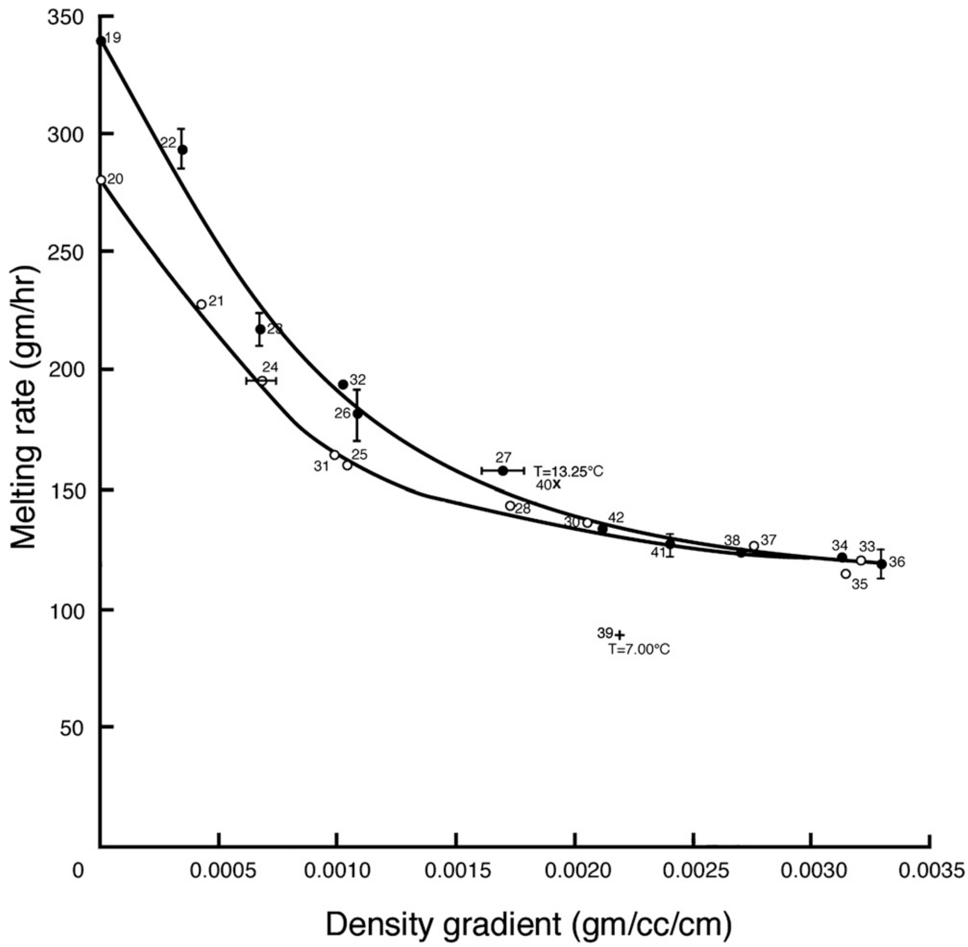

FIG. 2. The melting rate of ice floating on the surface as a function of density gradient for unheated runs (open circles) and heated runs (solid circles), in the laboratory experiments of Turner and Veronis (2004).

relative contributions of S and T to the mean gradient $R_\rho = (\beta\, dS/dz)/(\alpha\, dT/dz)$, and some molecular property to combine with N to give a length. Choosing the molecular diffusivity for heat κ_T, it follows on dimensional grounds that

$$H_0 = \left(\frac{\kappa_T}{N}\right)^{1/2}. \qquad (1)$$

Then the actual layer thickness will be

$$H = GH_0, \qquad (2)$$

say, where G is a function of R_ρ and of the Prandtl number and the ratio of heat and salt diffusivities. For the ocean only the density gradient ratio is varying, and so a plot of the nondimensional layer thickness $H/(\kappa_T/N)^{1/2}$ against R_ρ, using ocean measurements by many observers in a variety of staircases leads to the plot shown in Fig. 4; the letters and numbers identify the data sources used. The curve shown is of the form suggested by the use of laboratory scaling laws based on the 4/3 power law for the interfacial fluxes, as first proposed by Turner (1965) and widely used since. It follows from the 4/3 flux law and the way in which N appears in the layer thickness scale H_0 that both G and the effective diffusivity K_T deduced from Fig. 4 (and the corresponding K_S) are independent of N (i.e., they depend only on R_ρ). From the plots given in Kelley (1984) we find that K_T falls from 500 to 100 times the molecular value κ_T as R_ρ increases from 1 to 2; it is $40\kappa_T$ at $R_\rho = 3$ and $20\kappa_T$ at $R_\rho = 5$.

The temperature profiles presented by Padman and Dillon (1987), using measurements made in the Arctic Internal Wave Experiment in 1985, show that in the steepest part of the thermocline the mean temperature increases by 0.375°C over a depth of 40 m. Molecular diffusion of heat through this region, using $\kappa_T = 1.4 \times 10^{-7}$ m^2 s^{-1} and assuming the profile is smooth with no steps, gives a flux of 5.54×10^{-3} W m^{-2} or 1.78×10^5 J m^{-2} yr^{-1}. A typical value of R_ρ in staircases under the Arctic ice is $R_\rho = 3$, so Kelley's prediction that the heat flux is 40 times the

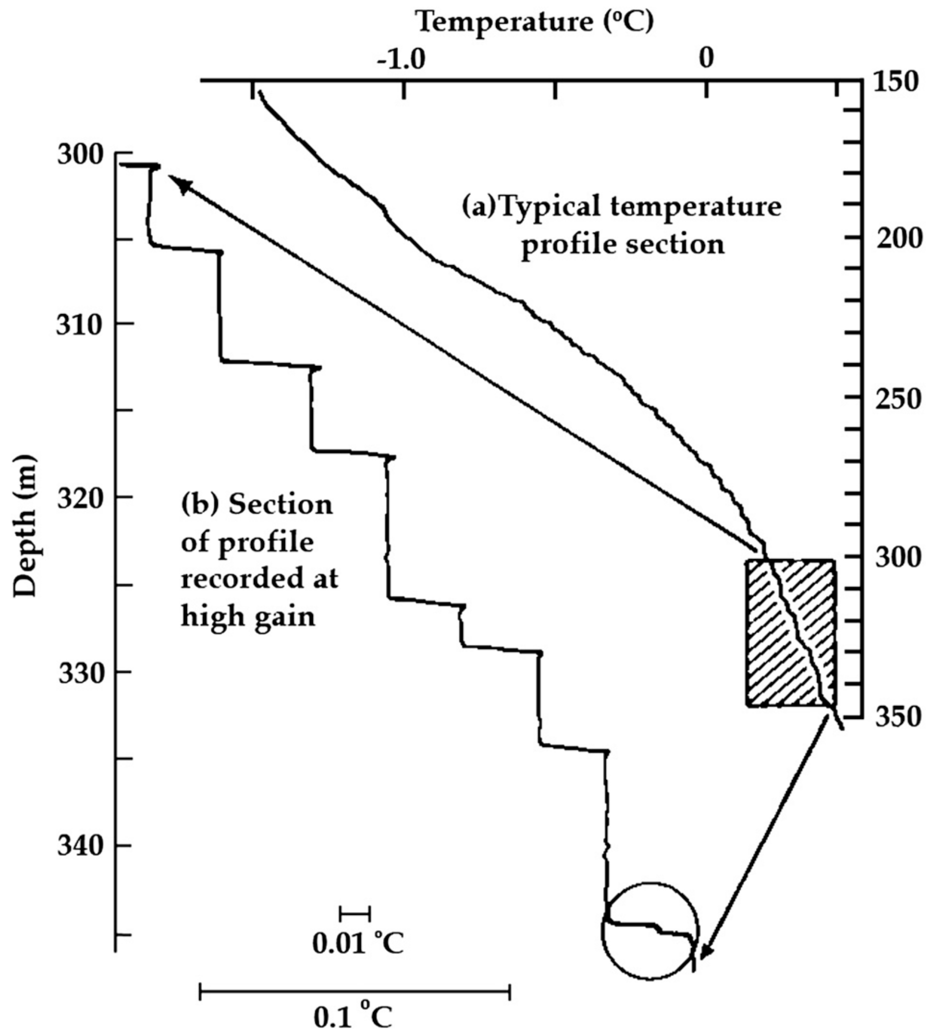

FIG. 3. Temperature profile under ice island T3, showing steps formed by the double-diffusive mechanism. Reproduced from Neal et al. (1969).

molecular value gives a flux of $F_H = 0.22$ W m^{-2} or 7.1×10^6 J m^{-2} yr^{-1}. This is about 1% of the heat flux required to melt all the sea ice in a year [i.e., it is sufficient to account for a significant fraction of the enhanced rate of melting over the past four decades (again with the qualification that this heat must penetrate the halocline)]. Padman and Dillon (1987) also note that the basin-averaged annual rate of heat loss calculated by Aagaard and Greisman (1975) is most likely to be due to mixing processes on the shelves, an idea that is explored further below.

4. Other mechanisms and estimates of the rate of melting of sea ice

a. Heat input from the atmosphere

Another mechanism for the melting of sea ice that has been studied extensively is the heat input from the atmosphere directly to the surface of the ice and also into the surface mixed layer and hence to the underside of the ice. A yearlong interdisciplinary study (October 1997–October 1998) of the Surface Heat Budget of the Arctic Ocean (SHEBA) reported by Perovich and Elder (2002) and Perovich et al. (2003) monitored the atmospheric and ice temperatures and used a network of more than 100 ice thickness gauges in ice of various types and thickness. At every site there was a net thinning of the ice during the SHEBA year. There was an average winter growth of 0.51 m and a summer melt of 1.26 m, consisting of 0.64 m of surface melt and 0.62 m of bottom melt, with some variation according to ice type. This corresponds to an annual average net heat flux into undeformed ice of 7.5 W m^{-2}. Perovich and Elder (2002) also noted that the SHEBA values are 2–3 times larger than prior results, obtained

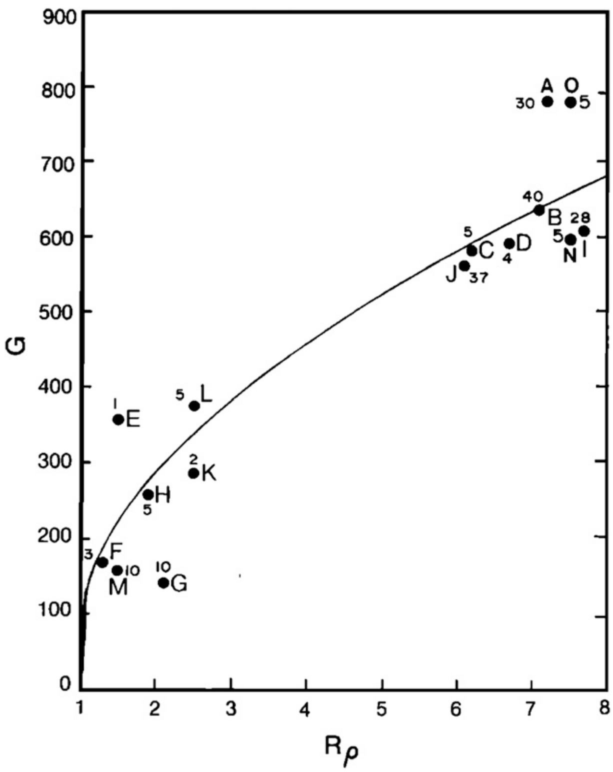

FIG. 4. Oceanic layer thicknesses scaled by $(\kappa_T/N)^{1/2}$ (the function G), plotted against the density gradient ratio R_ρ [reproduced from Fig. 1 in Kelley (1984)].

between 1975 and 1995, which were more typically 3.5 W m^{-2}.

Rothrock et al. (2003) have compared various numerical calculations with observations. The surface forcing included daily air temperature and pressure, geostrophic winds, and downward radiation. Their own model ocean was stratified in salinity, with a surface mixed layer, and a heat balance equation was used to determine the growth and decay of floating ice. Their survey of a wide range of model results showed that while there are many differences in how each modeler chose and incorporated forcing data, and in their internal representation of the physics, there was broad agreement between observations and numerical models representing Arctic ice over the 1990s. In particular years when detailed measurements and calculations have been made (e.g., the period of SHEBA), the input from the atmosphere can indeed explain the increased melting rate observed in recent years; it is certainly much larger than the double-diffusive flux from below.

b. *Mixing by the downflow of dense shelf water*

Finally, however, we return to the question that was left unresolved in section 2. It was shown that there is enough heat in the intruding layer of Atlantic water to melt all the sea ice in the Arctic Ocean in a few years, and the calculations in section 3 indicated that the purely vertical double-diffusive flux of heat could accomplish this over perhaps four decades. But it is certainly worth exploring further alternative advective, two-dimensional processes that could bring this heat more rapidly into the surface mixed layer.

Wells and Wettlaufer (2007) have described a theory and laboratory experiments that illuminate the dynamics of a two-dimensional density current flowing into a two-layer stratified basin. A density current flowing down a slope will dilute due to entrainment of lighter

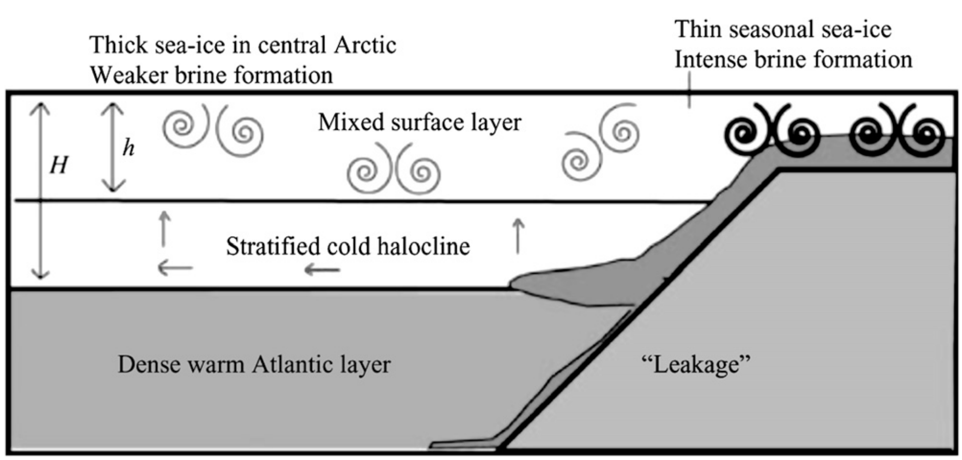

FIG. 5. Sketch of contributions to the dynamics of halocline maintenance [reproduced from Fig. 6 of Wells and Wettlaufer (2007)]. Oceanic freezing and melting in the shelf regions of the Arctic Ocean drives gravity currents that can spread at the halocline. If this bottom current continues and increases its density during the winter, it can break through the interface, mix with the dense Atlantic layer, and bring the warm water to the surface.

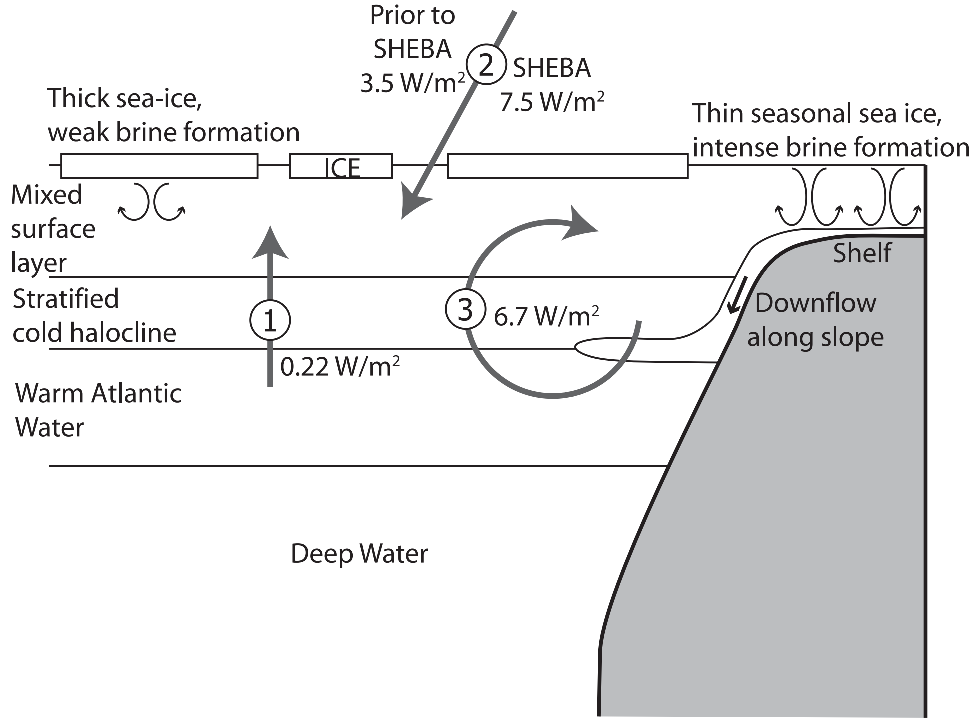

FIG. 6. Cartoon showing the three mechanisms discussed that can contribute to the melting of ice in the Arctic Ocean, and the corresponding estimates of the heat fluxes set out in the text. For comparison, the flux required to melt all the sea ice in a year is approximately 25 W m^{-2}. 1: Upward double-diffusive transport from the warm layer of Atlantic water. 2: Heat flux from the atmosphere into the surface mixed layer, and thence into the ice. 3: Mean surface heat flux over the whole Arctic Ocean required to balance the net input of heat by the inflowing warm Atlantic layer. Can this be driven by overturning due to downflows on the slope, as sketched in Fig. 5?

fluid above. If it is denser than the fluid in the lower layer when it reaches the interface it will continue to the bottom, but if it remains lighter than the lower layer it will spread out along the interface. It will continue to become denser, however, because of the "filling box" effect in the upper layer, and eventually penetrate the interface on a time scale that is a function of the buoyancy flux and the area of the basin.

For the laboratory experiments conducted in a long narrow channel the conditions determining penetration depend on the density step $\Delta_{12} = g(\rho_2 - \rho_1)/\rho_0$ between the two layers and the buoyancy flux per unit width $B = \Delta_{12}V$, where V is the volume flux per unit width. The criterion for breakdown can be related to the values of a Richardson number $\mathrm{Ri} = \Delta_{12}H/B^{2/3}$, where $\mathrm{Ri}^{\mathrm{in}}$ denotes the initial value and Ri^* the critical value at which the density of the outflowing layer becomes equal to that of the lower layer, and H is the depth of the upper layer at various stages of the filling box process. Note that the rate at which Δ_{12} changes is inversely proportional to the length L of the experimental tank, and the time to penetration can be expressed as $t_p = (\mathrm{Ri}^{\mathrm{in}} - \mathrm{Ri}^*)L/B^{1/3}$. The experimental results reported by Wells and Wettlaufer (2007) indicate that Ri* lies in the range 21–27.

In the Arctic context the rejection of brine on the shelves in a cold winter could lead to a dense layer spreading out along the halocline, increasing in density and eventually penetrating below the level of the core of Atlantic water, as shown in Fig. 5, reproduced from Wells and Wettlaufer (2007). This would gradually, or more rapidly, bring the warmer Atlantic waters into contact with the surface sea ice by a combination of upwelling and the erosion of the existing stratification that allows the surface forcing to penetrate more deeply into the water column. However, the application of the laboratory results requires a careful consideration of the differences between the two-dimensional laboratory flow and the three-dimensional shelf regions.

The volume flux of dense shelf water into the halocline over the area A of the Arctic Ocean was estimated to be $V_A = 2.8 \times 10^6$ m^3 s^{-1}. Using $A = 10^7$ km^2 and a value of Δ_{12} at the low end of measured values, namely, $\Delta_{12} = 0.022$ m s^{-2}, the corresponding buoyancy flux per unit area is $B_A = V_A\Delta_{12} = 8 \times 10^{-9}$ m^2 s^{-3}. A crucial step is the recognition that this flux is delivered over a basin scale of $L \approx 1000$ km approximately and thus that $B = B_A A/L \approx B_A L = 0.008$ m^3 s^{-3}. The final conclusion of Wells and Wettlaufer (2007), using the full range of recent oceanographic conditions to determine the initial Δ_{12} and hence $\mathrm{Ri}^{\mathrm{in}}$, is that the penetration time could range between 0.56 and 1.2 yr (i.e., that penetration could happen on a time scale comparable with the seasonal buoyancy forcing).

5. Summary and conclusions

The process that prompted this study of the melting of Arctic sea ice, namely, vertical double-diffusive convection between the intruding warm Atlantic layer and the surface, has been shown to be rapid enough to contribute substantially to the observed increased rate of melting over the past 30–40 yr. Other mechanisms, including the direct input of heat from the atmosphere, have also been assessed, and these lead to much larger rates of melting. The heat fluxes due to each of these processes are summarized and compared in Fig. 6. A particular conclusion of this study is that it seems prudent to give further consideration to the large source of heat in the Atlantic layer, and the potential for it to produce a more rapid rate of transfer of heat to the surface than is possible through a purely vertical transfer. Two-dimensional advective downflows of cold salty water from the surrounding shelves have been shown to be a promising mechanism for mixing this warm water rapidly to the surface. This process alone could produce a catastrophic melting of the sea ice on a much shorter time scale than either of the other mechanisms discussed here.

The dramatic observations of the rapid decrease in the area covered by ice over the past 2 yr suggest that the magnitude and relative contribution of each of these three processes should be more carefully monitored. Alone or in combination, and taking into account heating from both the ocean and the atmosphere, these processes could explain the existing observations and lead to an even greater rate of melting in the future.

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