Paper I

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Abstract

In the marginal ice zones, drifting sea ice encounters large ocean heat fluxes and melting rates. However, as found from modelling studies and observations of ice melting, double diffusive effects at the ice/ocean interface limits the melting rates. In this paper, direct measurements of turbulent heat and salt fluxes from the marginal ice zone during rapid melting are presented. The strength of double diffusion is found to be significant and close to the range suggested from other studies. Calculated melting rates when double diffusive effects are present are compared to melting rates calculated from a traditional bulk parameterization of ocean heat flux for a range of temperatures and friction velocities often encountered within the marginal ice zones. This comparison shows that by ignoring double diffusive effects, melting rates are overestimated by several cm per day, which has a significant impact on a predicted future ice cover.

1 Introduction

In the Arctic, the limiting component of exchange of heat between the ocean and the atmosphere is the sea ice [*Maykut and Untersteiner*, 1971]. The state of the sea ice cover, such as thickness, composition and horizontal distribution, are important factors in predicting future climate. At the ice/ocean interface melting/freezing balances the discrepancy between heat from the ocean and heat conducted through the ice. While the heat conduction in the ice is set by the temperature gradient (and hence the air and ocean temperatures), the ocean heat flux is generally set by the conditions of the under-ice boundary layer.

From the first modelling approaches to Arctic sea ice [*Maykut and Untersteiner*, 1971], the ocean heat flux was set as a constant of 2 Wm^{-2} to model the average ice thickness equilibrium. Later studies have shown the average ocean heat flux in the Arctic ocean to be slightly larger, e.g. 7 – 8 Wm^{-2} for multiyear ice during the SHEBA experiment [*Perovich and Elder*, 2002] and 3.5 Wm^{-2} for the AIDJEX campaign [*Maykut and McPhee*, 1995]. The ocean heat flux has a temporal or seasonal variation, but also a spatial variation over relatively short distances [*Wettlaufer*, 1991], which is linked to the variability in ice drift, under-ice topography and mixed layer hydrography.

Interaction between drifting ice and the underlaying mixed layer becomes increasingly important in the marginal ice zones (MIZ), where ice drift is rapid; ice drift divergence creates open leads and upper ocean temperatures can normally be well above freezing. In these areas ice/ocean/atmosphere interaction is intense, melting ice rapidly as it drifts.

The necessity of relating the ocean boundary layer fluxes to the mean properties of the boundary layer is given by the difficultly of making year round small scale measurements in ice covered areas. Such parameterizations can be based on mixed layer temperature and the relative speed of ice and ocean [e.g. *Josberger*, 1987; *McPhee*, 1983; *McPhee et al.*, 1987].

A key question in the MIZ is how fast ice melts when it drifts into water with temperature well above freezing? During the 1984 MIZEX experiment the wind blew the ice across an ocean front exposing the ice to water with temperatures more than one Kelvin above freezing. However, the observed melting was much less than expected from mixed layer temperatures and ice/ocean interface stress [*McPhee et al.*, 1987]. The proposed explanation for this was found after including effects of double diffusion at the interface, where transfer of heat is more efficient than salt in a thin molecular sublayer close to the ice. Hence salt exchange limits the heat exchange at the interface. Also laboratory studies have shown that molecular effects in the so called molecular sub layer are important even for flow over rough surfaces [*Owen and Thomson*, 1963; *Yaglom and Kader*, 1974]. For predicting ocean atmosphere heat exchange, ice freezing/melting and future ice cover in a best possible way, it seems crucial to include these effects in any parameterization of ice/ocean interaction [*Holland and Jenkins*, 1999; *Notz et al.*, 2003; *Schmidt et al.*, 2004; *Skyllingstad et al.*, 2003].

In this paper we will present data obtained during an ice drift in the MIZ north of Spitsbergen. It includes interface measurements of heat and salt fluxes and provides a first direct estimate of the strength of double diffusion at the ice/ocean interface.

In the following section the theoretical background will be presented in section 2; field experiment, data and processing will be presented in section 3 and results and discussion will be provided in section 4.

2 Background

In a horizontally homogeneous environment, conservation of heat in a small volume close to the ice/ocean interface is given as

$$\rho_i L\dot{h} = k_i \frac{\partial T_i}{\partial z}|_0 + \rho_w c_p \langle w'T' \rangle_0 \qquad (1)$$

where the terms on the right (conductive heat flux in the ice and ocean heat flux, respectively) are balanced by melting or freezing on the left.

ρ_i and ρ_w are the ice and water densities, L is the latent heat of fusion, $\dot{h}$ is the melting/freezing rate at the interface, k_i is the thermal conductivity of sea ice, $\frac{\partial T_i}{\partial z}$ is the temperature gradient in the ice close to the lower interface, c_p is the specific heat of sea water and $\langle w'T' \rangle_0$ is the interface kinematic heat flux. In kinematic form, the equation is

$$\langle w'T' \rangle_0 = w_0 Q_L + \dot{q} \qquad (2)$$

where $Q_L = \frac{L}{c_p}$, $w_0 = \frac{\rho_i}{\rho_w}\dot{h}$ and $\dot{q} = \frac{k_i}{\rho_w c_p}\frac{\partial T_i}{\partial z}$. Similarly, the conservation of salt at the interface yields:

$$\langle w'S'\rangle_0 = w_0(S_w - S_i) \tag{3}$$

where S_w is the mixed layer salinity, $\langle w'S'\rangle_0$ is interface salt flux and S_i is the salinity of the ice.

A bulk parameterization of kinematic heat flux at the interface is given by the mixed layer temperature elevation above freezing, $\Delta T = T_w - T_f(S_w)$ and interface friction velocity u_{*0} [*McPhee*, 1992]

$$\langle w'T'\rangle_0 = St_* u_{*0}\Delta T \tag{4}$$

where St_* is the turbulent Stanton number, a bulk exchange coefficient, and T_w and S_w is the mixed layer temperature and salinity, respectively.

The ice/ocean interface is considered hydraulically rough and dimensional analysis suggests that St_* should have a Reynolds number dependence [*McPhee et al.*, 1999]. Several field experiments with measurements of interface fluxes, stress in the surface layer and mixed layer temperature and salinity suggest that this dependence is not very strong, on the contrary the heat and mass transfer seem to be remarkably constant over a large range of Reynolds number conditions [*McPhee et al.*, 1999]. St_* is found to be in the range 0.0050 – 0.0060 [*McPhee*, 1992].

For a fully turbulent flow, laboratory studies have shown that molecular effects are important even over rough surfaces for high Reynolds numbers [*Owen and Thomson*, 1963; *Yaglom and Kader*, 1974]. For the ocean boundary layer, most of the change in temperature and salinity occurs within the molecular sublayer [*Steele et al.*, 1989] and scalar fluxes across this layer are set by turbulent exchange coefficients, α_h and α_s.

$$\begin{aligned} \langle w'T'\rangle_0 &= \alpha_h u_{*0}\delta T \\ \langle w'S'\rangle_0 &= \alpha_s u_{*0}\delta S \end{aligned} \tag{5}$$

In (5), $\delta T = T_w - T_0$, $\delta S = S_w - S_0$ and T_0 and S_0 is the interface temperature and salinity which are linked by the freezing line relationship, $T_0 = -mS_0$.

The ratio $R = \frac{\alpha_h}{\alpha_s}$ is important for interface fluxes. For $R = 1$, salt and heat will be transferred at the same rate over the molecular sublayer and will keep interface temperature and salinity at the initial T_0 and S_0, hence fluxes are steady. If $R > 1$, heat will be transferred at a higher rate than salt which will lower S_0, hence increase T_0. In this case exchange of heat is limited by the exchange rate of salt in the molecular sublayer. *Notz et al.* [2003] inferred from modelling observed false bottom migration that $R = 70$, which was the upper limit of the range they found from earlier laboratory studies ($35 \le R \le 70$). The presence of double diffusion, incorporated in the ratio *R*, is the main difference between the bulk parameterization of heat flux (4) and the molecular sublayer parameterization. In case of (4), salt is assumed to be diffused at a sufficient rate to keep the heat exchange at a maximum, i.e. α_s limits towards infinity and *R* towards zero.

3 Experiment data

Data presented in this paper was obtained during the WARPS (Winter Arctic Polynya Study) project in 2003, based on the German research ice breaker *FS Polarstern*. During an 18 hours period in 1 – 2 April, the ship drifted passively with the ice in the Whaler's Bay area north of Spitsbergen. The drift started at 80.43N 12.82E and continued 9.7 km towards southwest with an average speed of 15 cm s^{-1}.

A turbulence mast consisting of two Turbulence Instrument Clusters [TICs, *McPhee*, 2008] was deployed in the upper boundary layer through a hydrohole in the ice about 200m away from the ship. In general, the ice cover consisted of heavily ridged multiyear ice with thickness > 2 m, however the turbulence mast was deployed on refrozen lead of approximate size 100 x 50 m, with fairly levelled ice with thickness of 1.1 m [*Fer et al.*, 2004]. The TICs were separated vertically by 4 m, levelled at 1 m and 5 m below the ice/ocean interface, respectively. Each TIC consists of temperature (T), conductivity (stdC) and micro conductivity (μC) sensors, all from SeaBird Electronics in addition to an Acoustic Doppler Velocimeter (ADV) from Sontek/YSI which measures the three dimensional velocity in a small sampling volume close to the tip of the instrument. Fast time-response T and μC sensors and ADV were aligned so that they all sample at the same vertical level, whereas stdC was fixed ~ 20 cm above the others to provide the absolute conductivity.

Data were sampled at 2 Hz, resolving fluctuations in velocity, temperature and salinity well into the inertial subrange. By dividing the time series into 15 min periods we capture the covariance in advected eddies, which typically have time scales on the order of minutes. Within every realization, the velocity is rotated into a streamline coordinate system ($\langle u \rangle = U$, $\langle v \rangle = 0$, $\langle w \rangle = 0$) and linear trends of temperature and salinity are removed in order to find the fluctuating components of velocity, temperature and salinity. Covariance of vertical velocity, temperature and salinity is then calculated within every 15 min interval.

The μC sensor does capture the high frequency variations in the conductivity field, however is subject to a considerable drift over a relatively short time scale. To adjust the absolute value of the μC sensor a method is used where the low pass filtered raw frequency signal of the μC sensor is regressed against the low pass filtered stdC conductivity. This provides new calibration coefficients for the μC sensor in order to calculate the correct absolute salinity. The basis for this method is that both sensors should capture the relatively slow changes in conductivity (periods > 5 min) and therefore can be used for calibration.

The stdC cell consists of a 19 cm long tube in which water flows through when conductivity is measured. In order to not disturb the natural flow, there was no pump configuration on the stdC sensor. This makes the flow through the cell slower, which can have a quite severe low pass filtering effect on the flow through the cell and hence on the data sampled [*Norge Larson, pers. comm., 2007*]. Problems occur when calculating salinity using the measured conductivity and temperature from the T sensor, which has no flow constrictions. This introduces artificial variance from the temperature into the salinity signal, and in order to correct for this, temperature is low pass filtered using a cut off period representative for the low flow velocities in the stdC sensor. For the data presented here, a cut off period of 1 min is used.

4 Results and discussion

During the short ice drift, a 1.1 m long ice core was extracted from the ice and temperatures were measured at 10 cm intervals. Temperature in the ice increased linearly from the surface where it was close to air temperature towards ocean temperatures near the bottom with a mean temperature gradient of -21.7 K m^{-1}, which equals an upward conductive heat flux in the ice of 41.2 Wm^{-2}.

In Fig. 1 the average temperature and salinity profiles from the three CTD stations obtained in the same area as the drift, are shown. Profiles show a mixed layer with depth of 29 m and mean temperature and salinity of -0.74˚C and 34.53 psu, respectively, overlaying the warmer and saltier water which originates from inflow of Atlantic Water below. All three CTD stations were made within 10 km of the drift and the shading in Fig. 1 indicates the range of salinity and temperature at every depth for the three CTD profiles, hence the large horizontal variations in water mass properties within a relatively small area. It also shows the large amount of heat separated from the ice only by the shallow mixed layer.

After the initial data processing, heat flux, salt flux and friction velocity (square root of stress) are calculated as:

$$\begin{aligned} F_H &= \rho_w c_p \langle w'T' \rangle \\ F_S &= \langle w'S' \rangle \\ u_* &= \left(\langle u'w' \rangle^2 + \langle v'w' \rangle^2 \right)^{1/4} \end{aligned} \tag{6}$$

within every 15 min interval and plotted in Fig. 2. Average kinematic heat flux was $6.53 \cdot 10^{-5}$ K ms^{-1}, equivalent to 268 Wm^{-2}, average salt flux was $1.92 \cdot 10^{-5}$ psu ms^{-1}, average friction velocity was $0.94 \cdot 10^{-2}$ ms^{-1} and average mixed layer temperature elevation above freezing was 0.93 K.

As apparent from Fig. 2, temperature in the mixed layer is changing on a relatively short time scale due to horizontal advection of heat. To avoid influence of advection in the analysis of exchange coefficients and double diffusion, only those 15 - min intervals where the mean change in temperature is less than one standard deviation of the temperature within the same interval, is considered. This criterion is satisfied for 24 of 40 15 - min intervals. Choosing one standard deviation is a compromise to avoid advective effect as well as still use a significant fraction of the 15 - min intervals. In the following, unless otherwise stated, only turbulent fluxes and mean values for those 24 intervals are used.

Applying turbulent heat flux, friction velocity and mean temperature and salinity directly in the bulk parameterization (4), results in a turbulent Stanton number $St_* = 0.0084$. This is slightly higher than estimates from other field campaigns [*McPhee et al.*, 1999].

Average temperature, salinity, friction velocity and turbulent fluxes of heat and salt are applied to estimate the ratio of turbulent exchange coefficients of heat and salt as given in (5). From heat balance at the ice/ocean interface, w_0 is calculated from (2) and combined with (3) and the freezing line relationship provides S_0 and T_0. These calculations result in $w_0 = 8.82 \cdot 10^{-7}$ ms^{-1} ($\approx$ 7.6 cm day^{-1}), S_0 = 28.76 psu and T_0 = -1.55 ºC. Calculated interface values give turbulent exchange coefficients, $\alpha_h = 1.13 \cdot 10^{-2}$ and $\alpha_s = 4.0 \cdot 10^{-4}$, resulting in $R = \frac{\alpha_h}{\alpha_s} = 33$. This is close to the lower limit of the range suggested by *Notz et al.* [2003]. The impact of choosing a criterion to avoid horizontal advection effects, is illustrated by choosing other thresholds or use no criterion at all. By setting the threshold to 0.5 (2) times the standard deviation of temperature leaves 15 (35) 15-min intervals that satisfy the criterion, resulting in an exchange coefficient of 23 (37). If all 40 intervals are included, $R = 28$.

Under the encountered conditions, there were two important factors that potentially influenced the measurements, (i) heterogeneity in ice topography and thickness and (ii) horizontal advection of heat and salt. As a consequence of (i), stress at 1 m is 20 % less than at 5 m, somewhat contrary to the expected decrease of stress from the interface towards the mixed layer depth. Measurements at 1 m have a relatively small footprint area which reflects ice roughness within a limited range, whereas the measurements at 5 m are more representative as they reflect conditions also beyond the relatively smooth deployment site. The temperature gradient in the 1.1 m thick ice in which the turbulence mast was deployed is also believed to be larger than what one might expect in the surrounding thicker ice, hence the calculations above might have been different if the measurements have been made in the multiyear ice.

By (ii), temporal changes in temperature and salinity can make it difficult to relate interface fluxes to the mean properties of the mixed layer. This is illustrated by the bulk parameterization (4) where the estimated St_*, from this study, is more than 40 % larger than the common values from literature. However, for the purpose of determining $R = \frac{\alpha_h}{\alpha_s}$, the intervals with the largest temporal changes in temperature are excluded and turbulent fluxes of heat and salt are compared locally which reflects processes at play at the interface. It is therefore believed that effects of horizontal heterogeneity will not alter this ratio. Measured turbulent fluxes, stress and conductive heat flux are all representative of the small area in which they were sampled, which formed a small "natural laboratory" within the multiyear ice pack. Hence the present approach of determining the R can be justified using the measurements even though they are not representative for the entire ice floe or the MIZ in general.

As discussed, turbulent fluxes and melting rates of ice can be estimated from mixed layer temperatures and interface friction velocities, estimated from e.g. observed ice drift velocities. To quantify the consequences of choosing one parameterization of heat flux over another, melting rates are calculated using the bulk parameterization (4), $w_{0,bulk}$, and by including double diffusion effects with $R = 33$ (5), $w_{0,DD}$. Fig. 3 shows the difference in melting rates $w_{0,bulk} - w_{0,DD}$ in cm day^{-1} for a range of mixed layer temperatures and friction velocities commonly encountered within the MIZ. Mixed layer salinity is set to 34.44, equal to the average of measurements presented here and conductive heat flux is assumed to be zero, which is typical for melting conditions in the Arctic. α_h is chosen so that heat fluxes from both (4) and (5) are equal in case of no double diffusion, $R = 1$. Fig. 3 shows that using the bulk parameterization in areas where mixed layer temperature is high or interface stress is large, clearly overestimate the melting rate by up to several cm day^{-1}. This is in line with observations where melting rates were much smaller than estimated by more primitive parameterizations [*McPhee et al.*, 1987]. Over the course of an entire melting season in the marginal ice zone this effect can add up to a significant error in ice extent, ice thickness and ocean/atmosphere heat exchange.

Data presented here is to the author's knowledge the first direct measurements of both heat and salt fluxes in a regime where ice is melting rapidly. Recent studies [*McPhee et al.*, 2008] focussed on the freezing process and concluded that double diffusive processes are not important during freezing. This study shows that the ratio of turbulent exchange coefficients during melting is around 33 which favours double diffusive processes. Although this study only presents a short snapshot, the data provides important insight into ice/ocean interaction and points out the importance for future predictions of sea ice extent and thickness.

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Figures

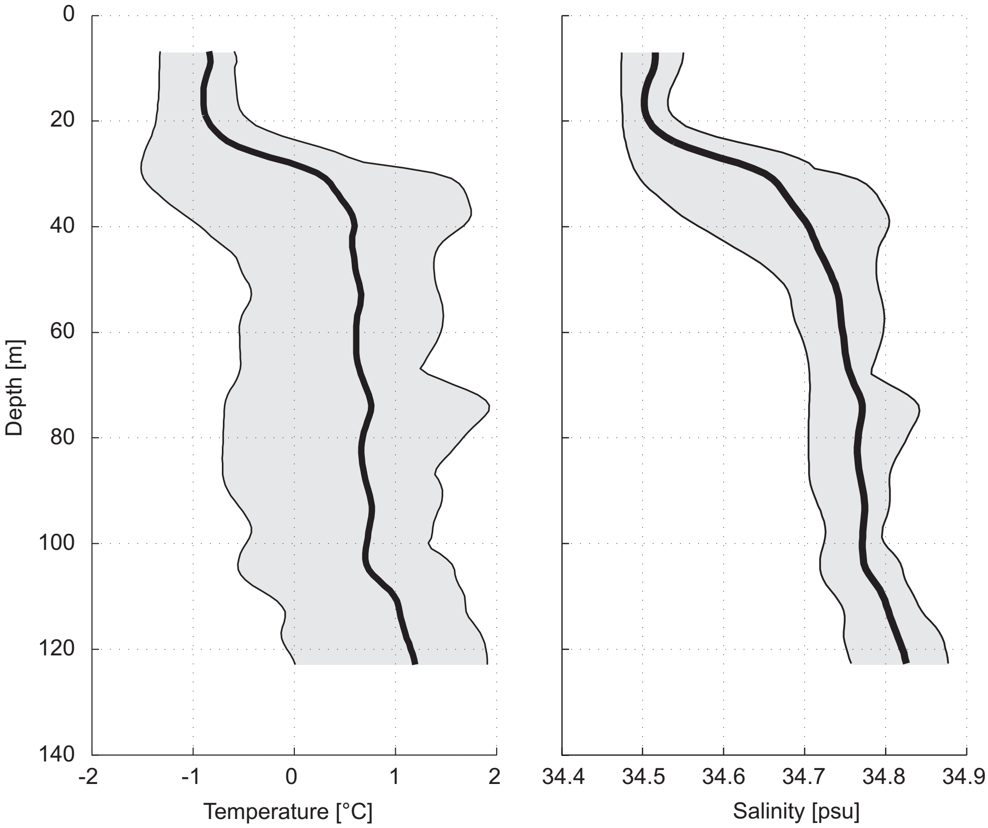

Fig. 1 Average temperature and salinity profiles from the CTD casts made in the study area. Thick lines are averages of the three CTDs, while shading indicates the range of temperature/salinity encountered at every depth which indicates the variability in between the three CTD casts.

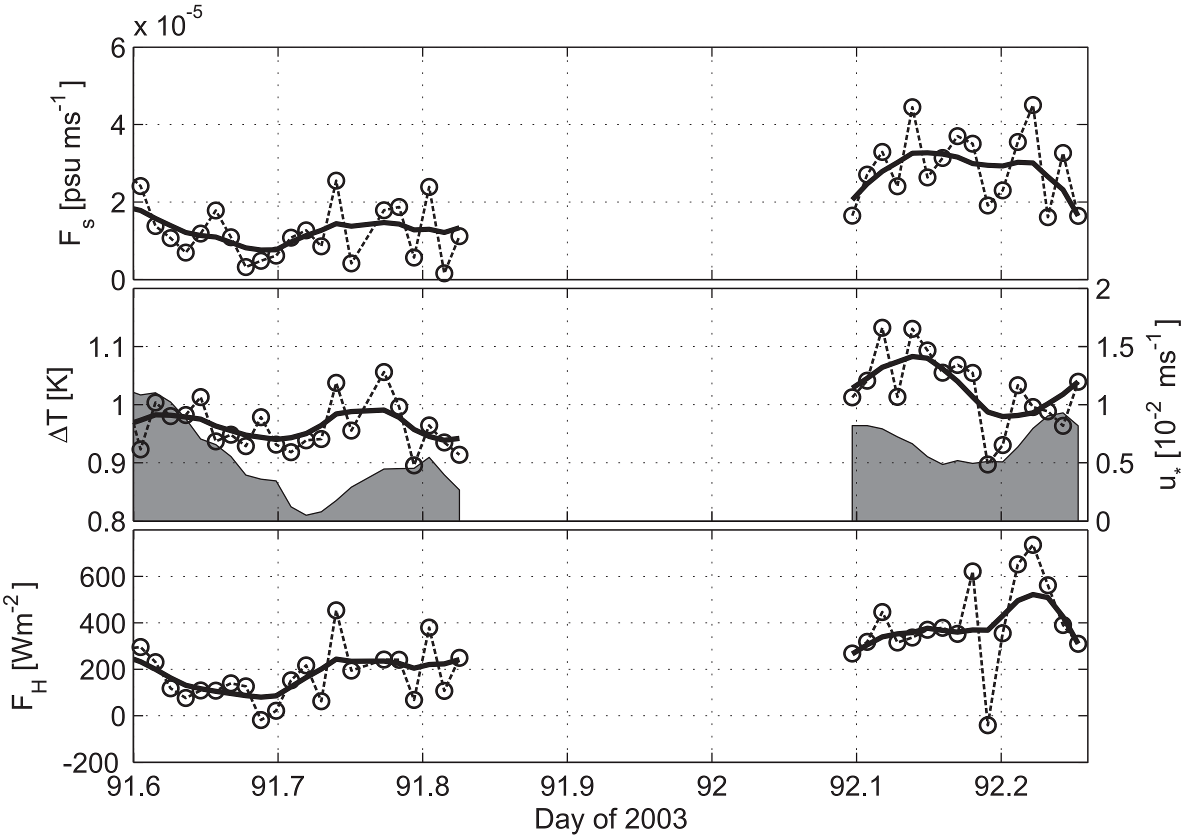

Fig. 2 Figure shows 15 min averages of salt flux (F_S), mixed layer temperature elevation above freezing (ΔT, gray shading), friction velocity (u_*) and heat flux (F_H) at 1 m below the ice/ocean interface. Thick lines are 1 hour running mean.

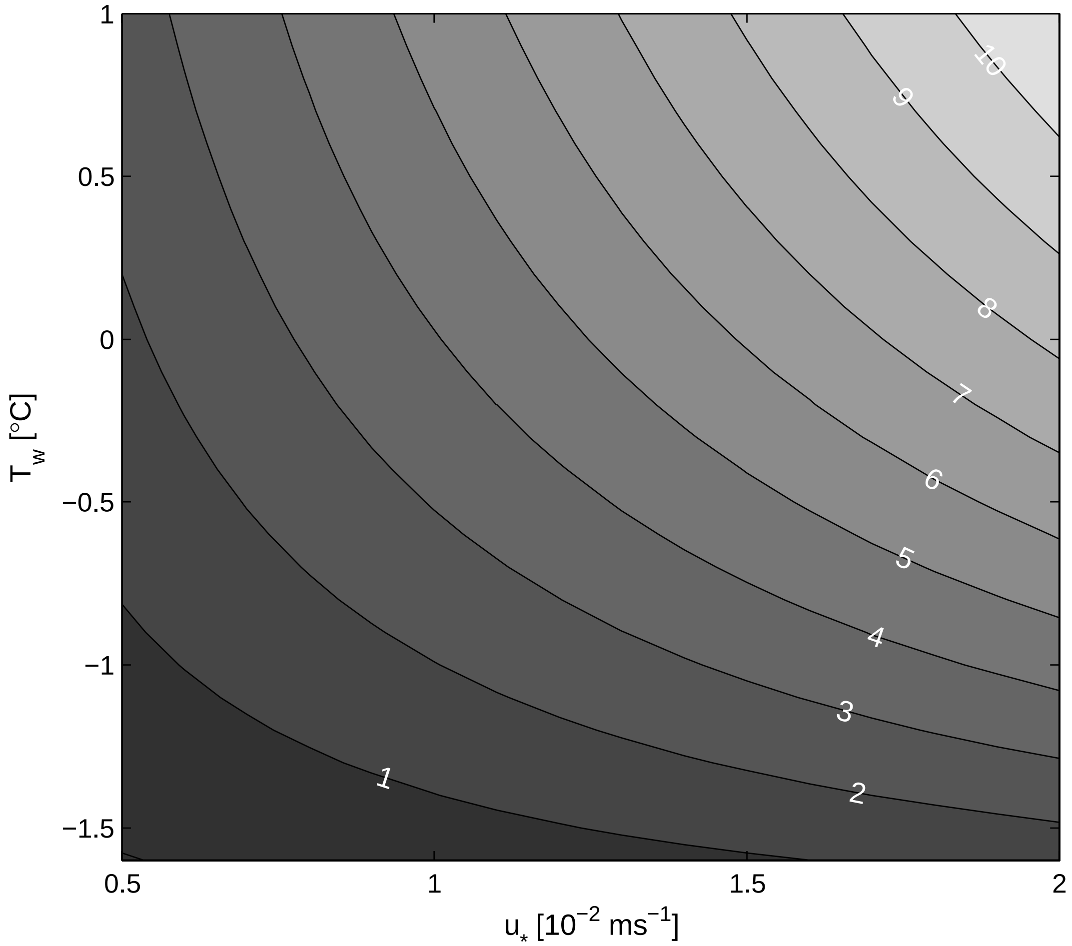

Fig. 3 Contours show the difference in melting rates, $w_{0,bulk} - w_{0,DD}$ [cm day^{-1}] estimated from the bulk formula (equation 4) and by including double diffusive effects on the ice/ocean interface with R = 33. The turbulent exchange coefficient is calculated so that heat fluxes are equal for the given set of friction velocities and temperatures in the case of no double diffusion, R = 1. Mixed layer salinity is set to 34.44 and conductive heat flux in the ice is assumed to be zero.