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Using Sea Ice to Measure Vertical Heat Flux in the Ocean

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Results of an experiment performed at drifting ice station FRAM I in the Arctic Ocean northwest of Spitzbergen during March–May 1979 indicate that sensible heat flux from the ocean to the ice cover was less than 2 W m^{-2}. The estimate is based on measurements of temperature gradient, growth rate, and salinity of young sea ice. Uncertainty in the magnitude of the heat flux results more from evidence of horizontal inhomogeneity in the growing ice sheet than from measurement errors.

INTRODUCTION

Direct observations of the vertical eddy flux of sensible heat in the ocean are notoriously difficult. Eddy flux meters of various designs have been used, but their delicacy and sensitivity to fouling makes them unsuitable for observations lasting for more than a few hours or days. Use of the profile method is limited primarily by the difficulty in measuring small temperature and mean velocity gradients in the water.

If the ocean is covered by a sheet of sea ice, a convenient way exists to replace these difficult observations with simple ones [*Untersteiner and Badgley*, 1958; *Untersteiner*, 1961; *Lake*, 1967]. Given the thermal conductivity and temperature gradient at a particular reference level in a horizontally homogeneous ice sheet, the vertical heat flux can be calculated, and its integral over time is equal to the algebraic sum of changes in heat content in the ice below the reference level, the latent heat released by freezing at the ice-water interface, plus heat absorbed from the oceanic mixed layer. By making straightforward measurements of ice growth, ice temperature distribution and salinity (which affects thermal properties), the average oceanic heat flux can be calculated directly. The method lends itself particularly to long time intervals (weeks or months) during which the change of ice thickness is bound to be large compared with the errors of each single reading.

The value of knowing the vertical heat flux under the ice lies in the important role it plays in determining regional and seasonal variations of the ice thickness. The thermodynamic model of *Maykut and Untersteiner* [1971] yields a realistic mean equilibrium ice thickness in the central Arctic under the assumption of an oceanic heat flux of 2 W m^{-2}, but it is probable that near the ice edge, and particularly under Antarctic sea ice, much higher values may be found (e.g., *Allison* [1979] reported average oceanic heat flux to be 10–20 W m^{-2} near Mawson, Antarctica, based on measurements of growth in snow-free ice and air temperature records over several years).

The relatively small heat flux in the central Arctic probably results from the presence of cold, saline water positioned between the low-salinity mixed layer and the thermocline marking the upper boundary of comparatively warm, slightly saltier water of Atlantic origin. The sharp density gradient in this cold layer is very effective at insulating the Atlantic water from mixing effects associated with surface stress or convection. No corresponding layer exists in the Antarctic: there the pycnocline and thermocline coincide, thus mixing is much more effective at upward heat transport.

EXPERIMENT

During the period March 11, to May 13, 1979, ice station FRAM I drifted between latitudes 85°N to 83°N and longitudes 11°W and 7°W, (see inset, Figure 1) in a vinicity traversed by only one previous U.S. station (ARLIS II). Prior to its deployment, it was speculated that the oceanic thermal regime at FRAM I might be quite different from the central Arctic because of its proximity to the Atlantic water inflow through Fram Strait, about 400 km to the southeast. In this part of the Arctic, the overall stability of the water column was thought to be less; furthermore, calculations by *Treshnikov and Baranov* [1972], who considered the change in heat content and the circulation of Atlantic water, had indicated upward heat loss by the Atlantic water corresponding to an average flux in excess of 11 W m^{-2} in the region just north of FRAM I. With this in mind, oceanographic work at the station emphasized the thermal and density structure of the upper ocean. This note presents results from one component of the program in which ice growth and thermal gradients in a recently refrozen lead were measured in order to estimate the oceanic heat flux.

Figure 1 is a schematic of the apparatus, with three sites arranged as shown near the center of a 150 m wide refrozen lead about 1 km from camp. Sites 1 and 2 were deployed on April 5, 1979 (day 95); site 3 was emplaced 8 days later. Setup was delayed at first by difficulty in finding suitable ice and later by the frequent break-up of thin ice near the station. By the time of deployment, the lead was covered by drifted snow 3–5 cm thick, which was restored after the equipment was in place.

The depth gauges were patterned after a design by *Untersteiner* [1961] in which a metal crossbar is suspended below the ice on a wire. The wire is allowed to freeze into the ice between readings, leaving the site relatively undisturbed. To check the thickness the wire is freed by applying an electric current, the crossbar is lifted firmly against the ice, and its change of position is noted on a meter stick frozen into the ice at the surface. A portable 110 VAC generator supplied current to the circuit which was closed by providing a

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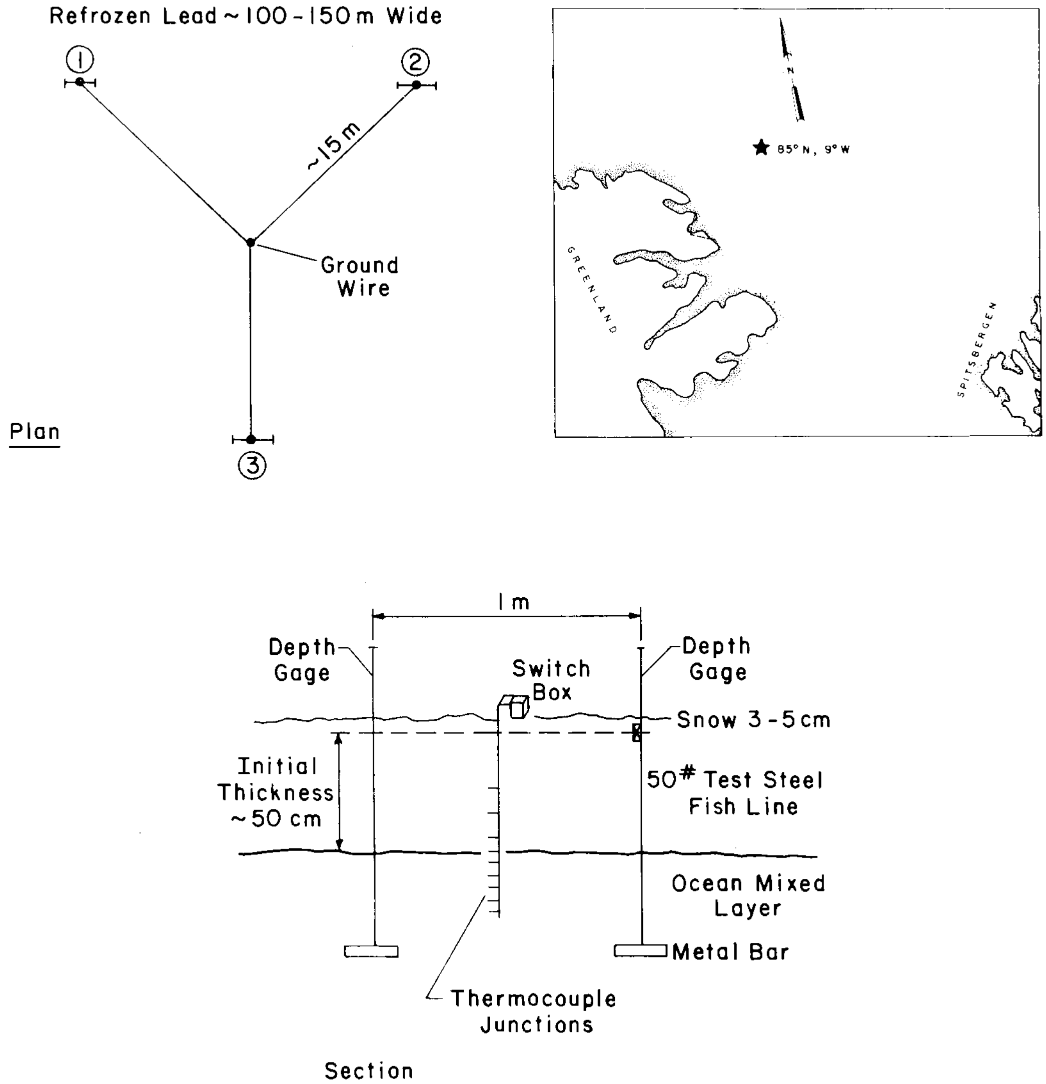

Fig. 1. Schematic of experimental apparatus. Each of three sites included one thermocouple string and two thickness gauges. Inset shows approximate location.

seawater ground about 15 m away. Changes in thickness were checked every third day.

Thickness data are summarized in Table 1. Values for absolute thickness reflect errors in the initial thickness determination of 0.5–1.0 cm associated mainly with determining the level of the upper ice surface. However, changes in ice thickness were measured more accurately, probably to within a few millimeters: thus, while the first and second lines of Table 1 might have considerable experimental error, the station-to-station variation shown in lines 3 and 4 is much larger than the measurement error. The differences in ice growth at the various locations over the same time periods must therefore result from horizontal variations in heat flux in the growing ice sheet. This provides a measure of the uncertainty in our assumption of horizontal homogeneity and how accurately the oceanic heat flux can be estimated from a temperature profile at a particular location. In our case, the uncertainty in ice growth associated with the measured vertical temperature gradients is apparently of the order of 1.5–2 cm, which over a time interval of 30 days is equivalent to errors in the heat flux of 1.5–2 W m^{-2}. Presumably, this source of error in the average flux would decrease if the measurement duration was increased.

The thermocouple junctions were arranged at 5 cm intervals (except for the top three, which were spaced at 10 cm) on a wooden dowel which was frozen into a 5 cm hole drilled through the ice. The leads were fed to a nine-position switch at the surface. Potentials across each switch position were read directly as temperature by a Fluke Digital Thermometer (model 2100 A-03). Temperatures were measured daily. In the cold air, output of the electronic thermometer drifted rapidly; however, the switching arrangement allowed sequential, evenly spaced readings so that instrument drift could be removed. As long as one junction remained in the

TABLE 1. Summary of Ice Growth Data

	Site Duration (Julian days)					
	1, 98–130		2, 98–130		3, 106–130	
	A	B	A	B	A	B
Initial thickness, cm	54.0	52.2	47.0	45.8	63.2	58.3
Final thickness, cm	95.7	failed	90.9	88.1	89.3	87.0
Total growth, cm	41.7	failed	43.9	42.3	26.2	28.7
Growth, Days 106–130, cm	25.2	failed	26.0	24.2	26.2	28.7

Each of the three sites included two thickness gauges labeled A and B. Note that site 3 was emplaced 8 days after sites 1 and 2.

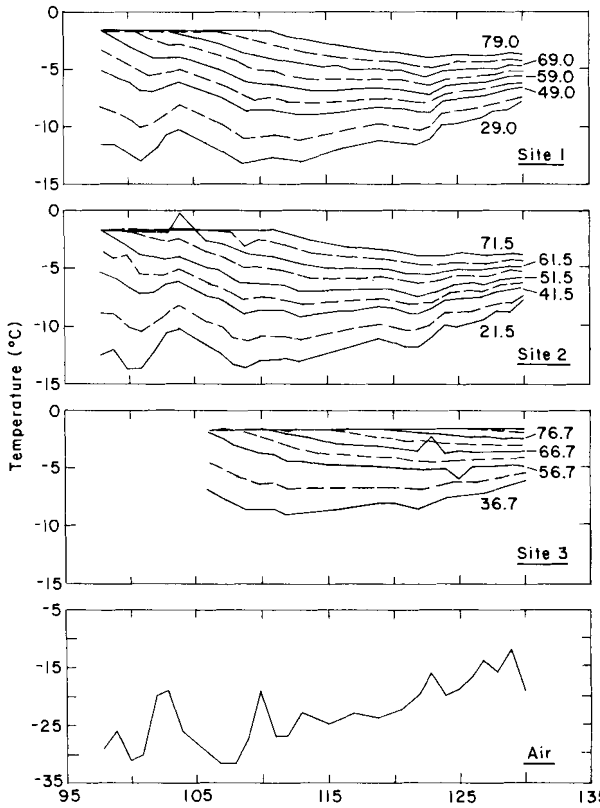

Fig. 2. Temperature records adjusted so that bottom-most thermocouple junction agreed with mixed-layer temperature. Numbers to right are the depths in centimeters from the upper surface of the ice.

water, a constant calibration factor was applied to all the temperatures such that the bottom-most temperature was the same as the mixed-layer temperature (which was known from oceanographic measurements). At sites 1 and 2 the bottom thermocouple was eventually engulfed by the ice, after which the correction factor was determined by extrapolating the mixed-layer temperature up to the level of the bottom sensor by using least squares fitted polynomials to estimate vertical changes in the temperature profile.

Results of the temperature measurements, adjusted so that the bottom sensors agreed with the mixed-layer temperature, are shown in Figure 2. Occasional anomalous spikes in the data give an indication of overall data quality, and deviation among temperature readings in the mixed layer provide an idea of how much variation occurred from junction to junction. Overall, the temperature sensing arrangement performed well, especially at site 1. Short-term changes at various levels are clearly reproduced at all sites and can usually be identified with changes in the surface thermal forcing.

Salinity was measured in ice cores taken at two different times near the apparatus. The bottom 50 cm of each core was divided into five sections, each of which was analyzed. While the various samples showed considerable scatter, there was no discernible vertical or temporal trend. The 10 samples yielded a mean salinity of 6.6 ‰ with a standard deviation of 0.8 ‰. Density was not measured, but was taken to be 0.92 Mg m^{-3} on the basis of analysis of first-year ice of similar salinity by A. Gow (personal communication, 1981).

Interpretation of Measurements

Since sea ice is a mixture of pure ice and brine pockets, its thermal conductivity and specific heat are functions of temperature and salinity [*Malmgren*, 1927]. Using observations that the salts in sea ice are in solution for temperatures above −8.2°C, *Schwerdtfeger* [1963] showed that its specific heat is given by

$$c_s = -\frac{s_i}{\alpha T^2} L_f + \frac{s_i}{\alpha T}(c_w - c_i) + c_i$$

where s_i is the ice salinity (i.e., the mass of dissolved solid material per unit mass of melt solution), T is temperature (°C), L_f is the latent heat of formation of pure ice (333.9 kJ kg^{-1}), c_i is the specific heat of pure ice (2.01 kJ kg^{-1} °C^{-1}), c_w is the specific heat of pure water (4.23 kJ kg^{-1}), and α is a constant (−0.0182°C^{-1}).

The latent heat of sea ice formation, L_s, depends on the relative proportions of pure ice and brine in the newly formed ice. Following *Schwerdtfeger* [1963] it is given by

$$L_s = \left[1 - s_i - \frac{s_i}{s_w}(1 - s_w)\right] L_f$$

where s_w is the mixed-layer salinity (32 ‰).

Thermal conductivity, k_s, depends not only on the amount of brine (i.e., salinity) but also the arrangement of brine pockets. *Schwerdtferger* [1963] has given a rigorous treatment of these effects, but for the present purpose a simplified expression for the thermal conductivity of sea ice [*Untersteiner*, 1961] will suffice:

$$k_s = k_i + \frac{\beta s_i}{T}$$

where k_i is the thermal conductivity of pure ice (2.04 J m^{-1} °C^{-1} s^{-1}) and β is a constant (117.3 J m^{-1} s^{-1}).

Given a record of temperature distribution and ice growth, the average oceanic heat flux $\bar{F}_w$ may be found from

$$\bar{F}_w = \frac{1}{\Delta t}[Q_f + Q_s - Q_L] \tag{1}$$

where Q_f is the time integral over Δt of heat flux through a particular reference level at which k_s and dT/dz are known, Q_s is the total change in heat content of ice below the reference level, and Q_L is the latent heat of ice formation. Since the specific heat depends on temperature, Q_s depends on the thermal history and must also be integrated in time.

As was mentioned above, second-order polynomials were fitted to temperatures measured within the ice so that the least squares difference between the data and the quadratic curves was minimized. These curves are shown in Figure 3 along with measured temperature and ice growth. For reference, the mixed-layer temperature was −1.75°C. While the departure from linear is small, the curvature was consistently observed at all sites, and may be due to thermal lag associated with an overall warming trend in air temperature.

For the oceanic flux calculations, the initial ice interface was taken as a reference level. A time step of 1 day was used in the integration of heat flux and heat content change, which was determined by summing the change in 1 cm layers of ice below the initial interface. The calculations are summarized

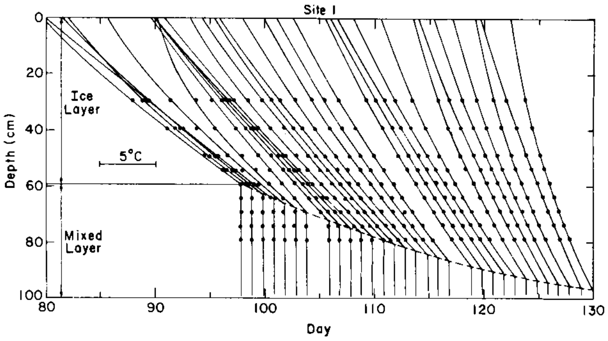

Fig. 3*a*

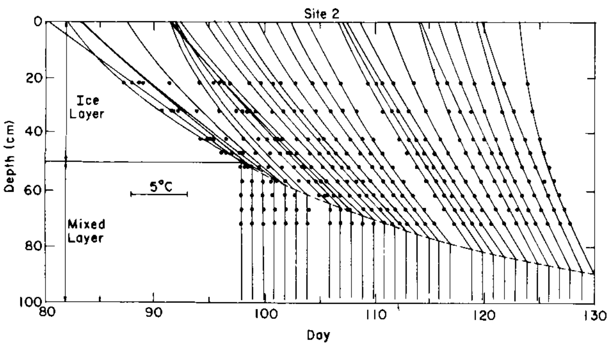

Fig. 3*b*

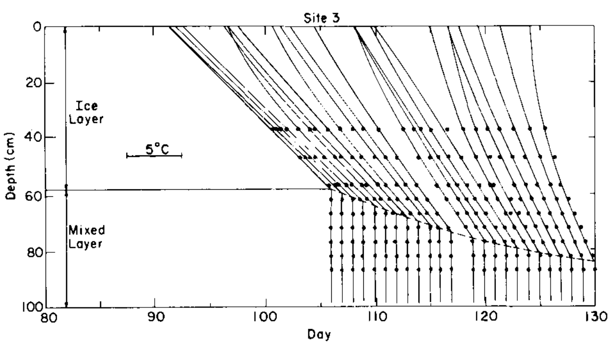

Fig. 3*c*

Fig. 3. Temperature profiles fitted to data points daily. Dashed curve shows ice growth. Reference mixed-layer temperature for each profile is −1.75°C.

in Table 2, for values of salinity ranging one standard deviation each way from the mean.

Discussion

Our results indicate that the magnitude of oceanic heat flux during the drift of FRAM I was small, less than 2 W

TABLE 2. Summary of Oceanic Heat Flux Calculations for Each Site, Over a Range of Ice Salinities Corresponding to One Standard Deviation About the Mean

	s_i, ‰	Q_f, MJ m^{-2}	Q_s, MJ m^{-2}	Q_L, MJ m^{-2}	$\bar{F}_w$, W m^{-2}
Site 1, 32 days	5.8	106.6	−10.9	96.2	−0.2
	6.6	105.4	−12.2	93.3	−0.0
	7.4	104.3	−13.4	90.4	0.2
Site 2, 32 days	5.8	111.7	−11.8	100.2	−0.1
	6.6	110.6	−13.2	97.1	0.1
	7.4	109.4	−14.5	94.1	0.3
Site 3, 24 days	5.8	72.0	− 6.4	65.7	−0.1
	6.6	71.0	− 7.2	63.7	0.0
	7.4	70.0	− 8.0	61.7	0.2

See text for further details.

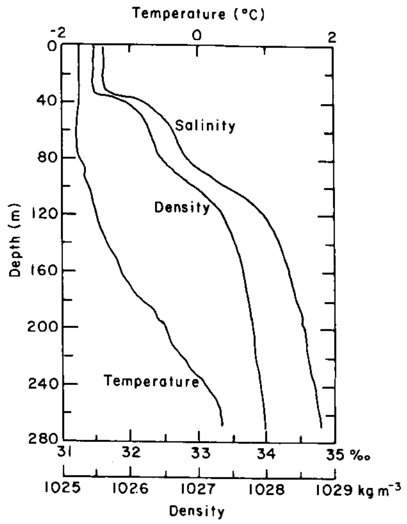

Fig. 4. Temperature, salinity, and density profiles in upper ocean at FRAM I. Note the layer of cold, saline water between the mixed layer and warmer Atlantic water.

m^{-2}. This corroborates our inference that there was little upward heat flux in the water column, as demonstrated by a typical salinity-temperature-density profile (Figure 4). Note that at depths from about 35 to 80 m, salinity and density increase dramatically, but temperature falls slightly (following the freezing line). An upward heat flux through this layer would have to be counter-gradient, which would require localized regions of active interchange between the mixed layer and deeper water. No such features were seen, and our evidence suggests that the advective layer of cold, saline water that exists beneath the mixed layer over most of the Arctic Ocean is indeed an effective insulator.

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