Supercooling at the Surface of an Arctic Lead[1]

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4 January 1973 and 7 May 1973

ABSTRACT

Supercooling of the surface water on open leads in the Arctic Ocean has been suggested as a possible source of water for the observed subsurface freezing. Observations with an infrared radiometer at the shore-lead outside Pt. Barrow, Alaska, in April 1972, verify this supposition.

1. Introduction

There has been much discussion of supercooling in the near surface water in polar regions as an explanation for observed underwater freezing (e.g., Untersteiner and Sommerfeld, 1964; Littlepage, 1965; Coachman, 1966; Countryman, 1970; Lewis and Lake, 1971; see the latter reference for Russian literature on the subject), and some corroborating experiments and observations have been cited (Untersteiner and Sommerfeld, 1964; Littlepage, 1965; Dayton *et al.*, 1969).

Lewis and Lake (1971) question the reality of *in-situ* supercooling of the bulk water, but these authors and Coachman (1966) agree that it is very likely that supercooling occurs at the surface of open leads during periods of appreciable heat loss. During measurements of radiative surface temperature at the offshore lead at Pt. Barrow, Alaska, in April 1972, evidence of a supercooled surface layer was found.

2. Instrumentation and procedures

An infrared radiometer (PRT5, Barnes Engineering Co.), which responds to radiation in the atmospheric window 8–14 μm, was positioned over open water at heights of 20–60 cm depending on the thickness of the surrounding ice (see Fig. 1). The radiometer has a 2° viewing angle making the observation spot 1.5–4 cm in diameter. The sensing head of the instrument was fitted with an auxiliary protecting cover and heaters in order for the internal reference cavity to be able to maintain a constant temperature. The data were collected on magnetic tape in a nearby heated tent.

The purpose of the expedition was to obtain measure-

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[1] Contribution No. 285, Department of Atmospheric Sciences, University of Washington.

ments of the temperature difference across the region of molecular heat conduction in the water. This layer is typically 1 mm in thickness and sustains a temperature drop of the order of 0.5C (Ewing and McAlister, 1960; Häussler, 1956). If the regions of conductive and convective heat transfer in the water are joined over a shallow transition region, one can approximate the temperature profile by a top boundary layer with a constant temperature gradient (under steady-state conditions), and a lower deep mixed layer of constant temperature T_b. To obtain the temperature difference across the boundary layer, $\Delta T = T_0 - T_b$, where T_0 is the surface temperature, the radiometer was positioned over the lead to measure T_0 continuously. Periodically the boundary layer was destroyed or made extremely thin by stirring with a submerged long-handled brush. This reveals T_b to the radiometer. Thus, ΔT is obtained on the magnetic tape within a short time span with all the environmental influences remaining constant.

That the stirred radiation temperature approximates T_b can be seen from Saunders' (1967) hypothesis. From dimensional reasoning he deduces that the thickness δ of the conduction region is given by

$$\delta \propto \nu(\rho_w/\tau_w)^{\frac{1}{2}}, \qquad (1)$$

where ν is the kinematic viscosity of water; τ_w the Reynolds stress due to wind, convection or mechanical stirring in the water; and ρ_w water density. Stirring serves the function of increasing τ_w in the water and decreasing δ. With a very thin conduction region, a much smaller ΔT is able to sustain the same heat flux, according to the heat flux equation written for a linear temperature gradient:

$$-\lambda\frac{\Delta T}{\delta} = Q, \qquad (2)$$

where λ is thermal conductivity and Q the vertical heat flux. The assumption is made that during vigorous stirring ΔT becomes negligibly small, at least of the order of hundredths of a degree Celcius.

The accuracy of the Precision Radiation Thermometer (PRT5) is given as 0.2C by the manufacturer. From a recording of the data, the difference obtained over a short time interval can be ascertained within 0.1C. Some differences noted between stirrings are due to the difficulty in doing this evenly by hand for any length of time.

Calibration of the radiometer was performed on location by obtaining bucket samples of the saltwater, and mixing this water with ice plus alcohol to obtain lower temperatures, and with water for slightly higher temperatures. The calibration baths were kept in a thermos bottle, which was vigorously stirred before a radiation measurement was obtained. This procedure established that the radiometer was working normally, and its output was linear in spite of the severely cold environment.

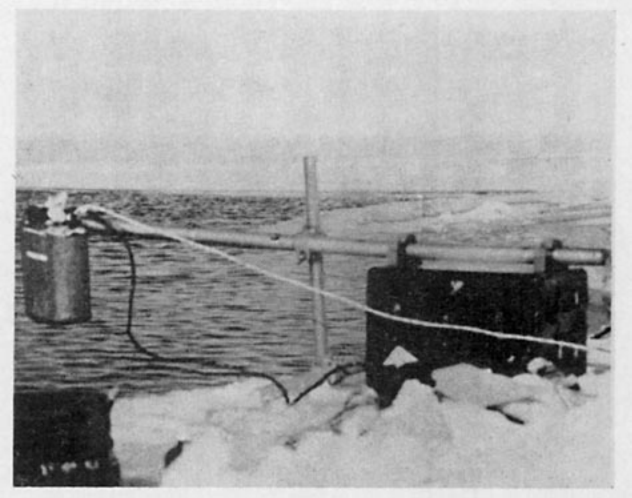

FIG. 1. The PRT5 infrared radiometer with auxiliary heating unit and electronics box in operating position at a lead west of Pt. Barrow, Alaska, 16 April 1972.

Using the simple method of stirring in the lead to obtain the bulk temperature, rather than by measuring with a mercury thermometer in a bucket sample, was necessary since both the measurements of T_0 and T_b are then subject to the same amount of sky reflection from the water surface, and to the same instrumental errors. The correction due to sky reflection was larger in magnitude than ΔT, and would be difficult to calculate with certainty. Comparison of the *in-situ* radiometer measurement of bulk water temperature with the conventional method gives the correction due to reflection as 0.5C. This agrees with the discrepancy between the equivalent blackbody temperature T_e, and T_o obtained by inverting the radiation equation:

$$N(T_e) = \epsilon N(T_o) + (1-\epsilon)N(T_s), \qquad (3)$$

where $N(T)$ is radiance integrated over the frequency response of the PRT5 radiometer at temperature T, ϵ the emissivity of the water surface, and T_s the average temperature of the sky dome. When using the integrated form of the radiation equation, approximations have been made. Here $N(T_s)$ is assumed independent of wavelength and ϵ independent of temperature and wavelength. Specifically ϵ is assumed equal to 0.98 and T_s is taken as -50C for clear sky with some low haze. Data obtained by Witte (1968) indicate this to be a reasonable average sky temperature for the existing meteorological conditions in the arctic environment.

Salinity samples were obtained on the site with a modified Van Doren bottle, taking care to keep the sampling bottle heated in the tent before the collection was made, and flushing it thoroughly in the lead before the ports were closed.

3. Results

A record of the radiometer trace during three short periods of stirring and returning to the undisturbed T_0

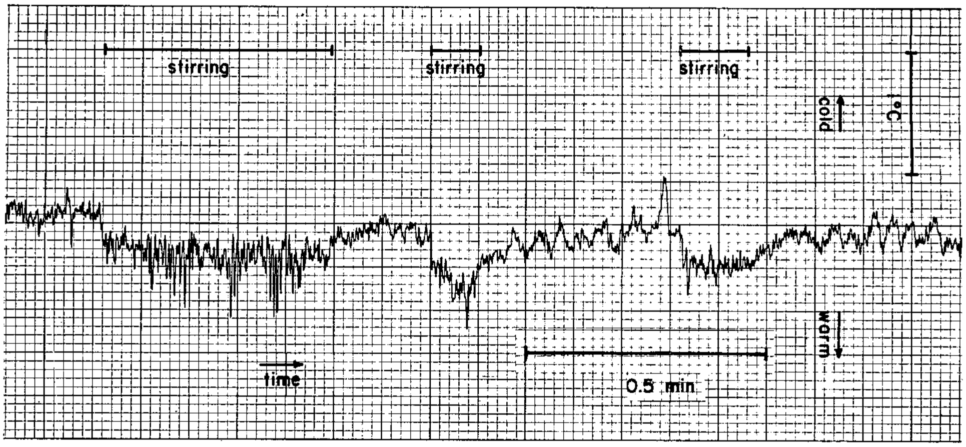

FIG. 2. Trace of the Precision Radiation Thermometer (PRT5) output on 16 April 1972 at the offshore lead at Barrow, Alaska. The data were recorded with a 3-Hz filter, and played back with a 5-Hz filter. There was "young" ice on most of the lead with only a small pond of open water next to the older ice. Even with 10 m sec^{-1} winds this short fetch did not allow the formation of large waves. Longer period fluctuations in the record are possibly due to convection, which was evidenced by an active streaming pattern.

is seen in Fig. 2. It is evident that the subsurface temperature was several tenths of a degree Celsius higher than that of the surface. The observation was repeatable. Data were recorded with a 3-Hz filter and played back with a 5-Hz filter.

Values of ΔT, salinity, and meteorological information averaged for each period of observation are presented in Table 1. The meteorological data were obtained from the National Weather Service station in Barrow, about 10 km distant.

The temperature of the bulk water measured with a mercury thermometer was -1.8 ± 0.1C on all three days. According to Miyake (1939) the freezing temperature of seawater of 33.3‰ salinity is given as -1.87C. Knudsen's (1903) values are about 0.08C higher, and Countryman (1970) tends to believe that the latter are closer to the truth based on results from his own field observations.

That the seawater was very close to its freezing temperature was verified by letting the water samples freeze and measuring their temperature with the thermometer used in the field just as the last bits of ice were melting. (When salt water freezes, the ice expels a concentrated brine which lowers the freezing temperature of the remaining solution.) The testing of the melting point was done in a thermos bottle, so that the last stage of melting took place with a minimal temperature gradient between the ice and the water. The "freezing," or better still, melting point measured in this way was -1.8 ±0.1C indicating that the bulk saltwater at Point Barrow was within 0.1C of its melting temperature in April 1972. For the surface layer to be 0.2–0.3C colder than the bulk water and not be supercooled, it would also need to be 4–6‰ saltier than the bulk water.

4. Discussion

What are the possibilities that the surface film was 4–6‰ saltier than the bulk? What other explanations exist for the radiation temperature's being lower than the freezing temperature of the bulk water, besides supercooling?

A saltier boundary layer is expected to exist near the water surface during evaporation since only the water molecules escape to the atmosphere. The depth of the layer in which salt is transported by molecular diffusion, δ_s, is expected to be less than that for molecular heat conduction (Prandtl, 1910) since the coefficients of molecular transport of heat and salt in water differ by two orders of magnitude. The ratio δ/δ_s, according to Prandtl's theory, is proportional to $(\kappa/D)^{\frac{1}{3}}$, where κ and D are the molecular diffusivities of heat and salt, respectively. This makes $\delta_s\approx\delta/5$, and of the order of 0.02 cm. (Typical values of δ_s and $\Delta S=S_0-S_b$, where S_0 is surface salinity and S_b is the bulk value, need yet to be established by experiment.)

The expected value of ΔS under steady-state conditions can be calculated from the diffusion equation

$$\rho_w D\frac{\Delta S}{\delta_s}=ES_0, \tag{4}$$

where E is the evaporation rate, and ES_0 an expression for the continuous source of salt at the interface due to the evaporation. If the evaporation rate is calculated with the bulk aerodynamic formula to obtain an order-of-magnitude estimate for the conditions of 16 April 1972, it is found to be 2.3×10^{-6} gm cm^{-2} sec^{-1}. For 18 April it is 4.6×10^{-6} gm cm^{-2} sec^{-1}. These evaporation rates bracket a value deduced from humidity observations at an open lead in March 1962 by Badgley (1966). With $D=1.6\times10^{-5}$ cm^2 sec^{-1}, $S_0=33$‰, $\delta_s=0.02$ cm,

TABLE 1. Data obtained at Pt. Barrow on 14, 16 and 18 April 1972. The radiative surface temperature shows fluctuations of ±0.2C.

			Average meteorological conditions[a]								
					Wind						
Date	Time	Location	Air temperature (°C)	Dew point temperature (°C)	Speed (m sec^{-1})	Direction	Sky conditions	Salinity[b] (‰)	T_b (°C)	Average[c] ΔT (°C)	Remarks
14 April	1200–1400	SW of Barrow, 1 km off coast	−19.0	−24.4	5	SW	clear	33.338[d]	−1.8	−0.25	On this day a hole was dug in "young" ice. Ice was continuously reforming at the surface.
16 April	1100–1400	NW of point	−19.0	−24.4	4–8	NE	clear	32.855[d] 32.848[e]	−1.8	−0.35	Observations were made in a natural pond of water at the edge of the offshore lead which had been partly blocked off by a large floe.
18 April	1300–1500	NW of point	−17.2	−22.2	8–10	NE	~4/10 Ci; blowing snow	no sample	−1.8	−0.1 to −0.3	The lead was a several kilometer wide expanse of open water. Waves were present, wavelength ~5 m, height ~0.4 m.

[a] From National Weather Service Office, Pt. Barrow.

[b] Analysis performed at Department of Oceanography, University of Washington.

[c] Error limits ±0.1C.

[d] These samples were obtained 40 cm below the surface.

[e] This sample was skimmed off the surface.

and the evaporation rate of 16 April, Eq. (4) gives a ΔS of 0.9‰. Since convection was occurring and the diffusion of salt is a very slow process, the molecular diffusion layer may not have been in steady state. However, the convective value of ΔS would be smaller for the same conditions.

Since ΔS is linearly dependent on both evaporation rate and thickness of the layer of molecular diffusion, one can conclude that on all three days of observation the surface salt concentration could not be sufficient to explain the observed surface temperature without supercooling. An attempt was made to skim off the surface film on 16 April although this is very difficult to do. If the saltier film were only 0.02 cm thick, it would have accounted for only a fraction of the water obtained. At any rate, the analysis shows only a small difference in salinity, and in the wrong direction (see Table 1).

It might also be argued that if freezing were taking place at the surface, the crystals would act as a foreign film on the water allowing a temperature gradient through the thin ice layer, and the observed ΔT could then be explained without supercooling. However, no indications of crystals could be seen in the water under the radiometer on 16 and 18 April with the naked eye, even at very close range and with good lighting, nor were any crystals seen in the sampling bucket or bottle immediately after withdrawing the samples. On 16 April, new ice was seen to be forming in dendritic structures at the edge of the existing ice while 18 April was a stormy day, and no growth of ice could be observed at all. Observations on 18 April were made on the downwind side of an open lead which was so wide, (i.e., of the order of kilometers) that the other side could not be discerned through the steam fog. Foam and bubbles within the field of view may have contaminated the measurement of ΔT. However, both T_b and T_0 were subject to this, and the only effect seems to have been to make the record more noisy.

On 14 April, the observations were made in a 1.5 m×1 m freshly dug hole, in which ice was continuously reforming at the surface and was skimmed off. The salinity measurement in this case is probably slightly contaminated by crystals in the water.

A depression of the radiation temperature below the thermodynamic freezing temperature of the bulk water has also been found at the surface of both fresh and salt-water tanks of several concentrations in cold-room experiments. In fact, this seems to be the rule before freezing commences in a situation of strong upward heat flux from a free surface. Foster (1969) also found that supercooling existed before freezing in laboratory experiments. He controlled the upward heat flux by a metal block in contact with the surface, and found that the supercooling depended on the smoothness of the metal surface. A free water surface would probably present even fewer nucleating points than his smooth metal block.

5. Conclusions

Supercooling of the conduction region in the water immediately below the interface of an open lead in the Arctic Ocean has been observed. Further experiments are being carried out in a cold room to try to understand how the supercooling depends on the boundary layer structure and the heat flux.

An extension of this problem would be to investigate whether the supercooled film can be convected downward without loosing its supercooling and how this would occur.

Acknowledgments. The cooperation of the staff of the Navy Arctic Research Laboratory at Pt. Barrow is greatly appreciated. I wish to thank Profs. Joost A. Businger and Franklin I. Badgley for reading the manuscript.

This work was supported by the Office of Naval Research under Project NR 083-012, Contract N00014-67-A-0103-0014.

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