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The sea surface temperature deviation at very low wind speeds; is there a limit?[1]

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ABSTRACT

Laboratory and field measurements of the temperature drop across the boundary layer at the upper surface of evaporating and cooling water in turbulent free convection show that greater values are possible than previously indicated. An upper limit on the surface temperature deviation is imposed only by the maximum heat loss occurring in the earth's most extreme climates.

1. Introduction

That the interfacial temperature of a water body losing heat to the air above is typically colder than lower turbulent layers by several tenths of a degree centrigrade has been established in laboratory studies (e.g., Häussler, 1956; Ramdas & Raman, 1946; Spangenberg & Rowland, 1960; Katsaros & Liu, 1974) and from field observations (e.g., Woodcock, 1941; Woodcock & Stommel, 1947; Ball, 1954; Ewing & McAlister, 1960; Hasse, 1963; Saunders, 1967a; McAlister, McLeish & Corduan, 1971; Clauss et al., 1970; Katsaros, 1973).

Observational techniques for obtaining the surface temperature have varied from specialized mercury thermometers, use of infrared radiometers or fine resistance wires to extrapolated atmospheric temperature profiles.

In the article by Hasse (1963) most of the work previous to 1963 was summarized in a figure (his Fig. 2). This figure shows an intriguing limit of about -1 °C of the surface temperature deviation, ΔT, as the negative air to water temperature difference increases beyond -5 °C. ΔT is defined as $T_0 - T_w$, where T_0 is the water temperature at the interface, and T_w is the bulk water temperature measured between depths of 0.2 and 2 m. The data for air to bulk water temperature difference <-5 °C are from two pioneering papers by Woodcock (1941) and Woodcock & Stommel (1947). (See Fig. 1 where these data are included as open and filled circles.)

Since in this paper we will be looking for extremes of negative $T_0 - T_w$ the following discussion will be limited to the case of net upward heat flux with no insolation. The case of calm conditions and heating from above is trivial in that the boundary layer is not limited but grows through the whole fluid. Very large positive values of $T_0 - T_w$ are therefore possible. Stable stratification also occurs with negative $T_0 - T_w$ in salt water of salinity $<24.7\%$ at temperatures less than those of maximum density (4 °C for fresh water and decreasing with salinity). $T_0 - T_w$ of -2.3 °C was observed before freezing in a laboratory vessel of fresh water (Katsaros & Liu, 1974). In deeper stably stratified water even larger temperature deviations are probable. The limit in these cases depends on the upward heat flux, the depth of the fluid and the amount of supercooling allowed by the nucleating agents. The following discussion will, therefore, also be restricted to cases of gravitationally *unstable* stratification.

In this paper three mechanisms which impose limits on ΔT in unstable stratification will be discussed, namely, Reynolds stresses in the water, gravitational instability and surface tension effects. Both laboratory and field observations will be presented which show ΔT's beyond -1 °C when the water is in free convection. An explanation for the discrepancy with the earlier data will be given in terms of the characteristics of the thermal boundary layer on water in free convection and the

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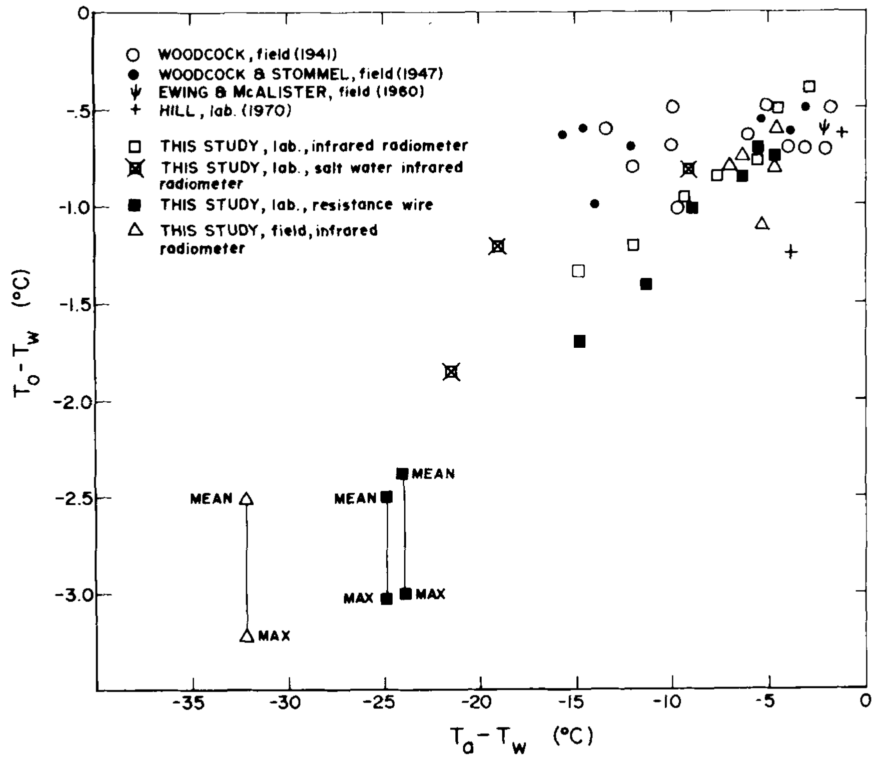

Fig. 1. Measured values of negative temperature deviations of the free surface boundary layer in water at low wind speeds versus air–water temperature difference from several investigators.

limited vertical resolution of the mercury thermometers used in obtaining these data.

2. The boundary layer influenced by turbulent stresses

In deep water with heat flux out of the top surface the temperature drop across the boundary layer is determined by the net heat loss and the thickness of that layer. In discussions of the "cool film" it has often been assumed that the gradient immediately below the interface is approximately linear in a thin layer where molecular conduction dominates which borders on a turbulent well-mixed layer (e.g., McAlister & McLeish, 1969). However, it is understood that where there is divergence or convergence due to varying Reynolds stresses or waves, the mean gradient will be of error function shape (Katsaros & Businger, 1973). The same is true for a time varying boundary layer (Howard, 1966; Liu [0] Businger, 1975; Katsaros et al., 1977), but for the sake of discussion we define an *equivalent* boundary layer depth, δ. δ is then a depth through which a linear temperature gradient with a slope equal to the one at the surface gives rise to the correct mean heat flux, Q. Saunders (1967b) gave a dimensional argument for this boundary layer thickness being inversely proportional to the $\frac{1}{2}$ power of the turbulent Reynolds stress in the water, τ, namely:

$$\delta \propto \nu \left(\frac{\tau}{\rho}\right)^{-1/2} \tag{1}$$

where ρ is water density and ν is kinematic viscosity.

Using the heat conduction equation in the form:

$$\Delta T = \frac{\lambda Q \delta}{(\kappa \rho c_p)} \tag{2}$$

where λ is a constant of proportionality, κ is thermometric diffusivity and c_p specific heat at constant pressure, he also predicted that since sensible and latent heat flux as well as the square root of the turbulent Reynolds stress are proportional to wind speed, but infrared heat loss is not, the surface temperature deviation will be only weakly affected by windspeed but will *decrease* somewhat with windspeed when the infrared heat flux is relatively large. Hasse (1971) confirmed this general relationship for windspeeds >2 m sec^{-1} using atmospheric profiles with vanishing potential temperature gradient. Hill's (1970) laboratory measurements of ΔT obtained with an infrared radiometer are also in agreement, but Paulson & Parker's (1972) study gave two different values of the constant λ in Saunders' equation depending on whether waves were being generated or not.

Other turbulence producing processes in the water even in the total absence of wind would also tend to limit the value of ΔT. Practical examples are a running stream or tidal and gradient currents in the sea producing velocity gradients near the surface. Internal waves or the propeller of a ship, etc. also cause turbulent stresses in the water which would similarly limit the region of conductive heat flux and, therefore, ΔT.

From considerations of the effects of turbulent stresses on the boundary layer one can conclude that the largest values of negative ΔT should be expected at low wind and current speeds. Saunders (1967b) suggested from an estimate of λ and (1) and (2) that $\Delta T = -1$ °C may be reached in extreme circumstances.

3. The boundary layer in free convection

In free convection at high Rayleigh number, mean flow and shear are absent, but the convective motions are none-the-less disorganized and turbulent, the Rayleigh number being defined through:

$$R = \frac{\alpha g}{\kappa \nu} \frac{\partial T}{\partial z} \cdot d^4 \quad \text{or} \quad R = \frac{\alpha g}{\kappa \nu} \Delta T \cdot d^3 \tag{3}$$

where α is the thermal expansion coefficient, g the acceleration due to gravity, d the depth of the fluid, and z the vertical coordinate. This regime begins already at Rayleigh numbers of about 10^5 (Willis & Deardorff, 1967; Krishnamurti, 1970) which in fresh water at 27 °C with a $\Delta T = -0.5$ °C corresponds to a depth of only 0.02 m. Thus, most natural water bodies of interest are in turbulent free convection in the absence of Reynolds stresses. Already when the wind speed drops below 2 m sec^{-1}, if there are no mean water currents, it is likely that the free convection regime dominates (Hasse, 1971). The wind speeds accompanying the data of Woodcock & Stommel (1947) for $T_a - T_w < -5$ °C are in all but one case less than 1.2 m sec^{-1}.

The established theoretical and experimental relationship between heat flux, ΔT, and depth for high Rayleigh number free convection is contained in:

$$N = AR^{1/3} \tag{4}$$

where the Nusselt number is given by $N = Q/[(\kappa\rho c_p)(\Delta T/d)]$ and A is a constant. In recent studies of convection between parallel rigid plates (e.g., Rossby, 1969; Chu & Goldstein, 1973) the power in this relationship was found to be slightly less than $\frac{1}{3}$. However, laboratory experiments at $R \simeq 10^9$ of convection below a free water surface (Katsaros et al., 1977) verify the $\frac{1}{3}$ power law almost exactly with $A = 0.156$.

When the power in (4) is $\frac{1}{3}$ the depth dependence in the Nusselt and Rayleigh numbers cancel. This implies that the fluid is deep enough that the upper boundary layer is unaffected by the existence of a lower one. There is then only one independent variable, namely the externally applied heat flux. Rewriting (4) using (3) we obtain:

$$\Delta T \propto \left(\frac{\kappa\nu}{\alpha g}\right)^{1/4} \left(\frac{Q}{(\kappa\rho c_p)}\right)^{3/4} \tag{5}$$

This relationship between ΔT and Q was also discussed by Saunders (1967b). Defining an equivalent depth, δ_c, for the boundary layer in a convective regime as in the stressed case by eq. (2) with δ_c replacing δ, and substituting (2) into (5) we obtain:

$$\delta_c \propto Q^{-1/4} \tag{6}$$

In free convection, as in the case of convergence and divergence in the boundary layer, the temperature profile was predicted to be highly nonlinear (Howard, 1966, in a theory based on Rayleigh instability). Howard references observations by Thomas & Townsend (1957) and Elder (unpublished) which verify the theoretical shape of the temperature profile. Recent laboratory measurements (Katsaros et al., op. cit.) verify this theory in greater detail, and also show some modifications. δ_c is therefore best thought of as a scaling length, and is given the subscript c to distinguish it from δ,

Table 1. *Corresponding values of heat flux, surface temperature deviation and equivalent boundary layer thickness from the relationship* $N = 0.156 R^{0.33}$ *for fresh water at 27 °C*

Q		ΔT	δ_c
cal/cm² sec	W/m²	(°C)	(mm)
2.0×10^{-3}	0.84×10^2	0.44	9.6
4.0	1.67	0.73	8.1
6.0	2.51	1.0	7.3
8.0	3.35	1.23	6.8
10.0	4.18	1.46	6.4
12.0	5.02	1.68	6.1
14.0	5.86	1.88	5.9
16.0	6.69	2.08	5.7
18.0	7.53	2.28	5.5

since the two depths would differ in general for the same heat flux.

From (5) it is evident that ΔT should increase as Q becomes larger without any asymptotic limit. From (6) we also see that the layer across which the temperature drop occurs becomes thinner as the heat flux increases. Table 1 gives corresponding values for Q, ΔT and δ obtained from (4), (5) and (6) with the constant $A = 0.156$ and the physical constants evaluated for fresh water at 27 °C (Chemical Rubber Company Handbook, 1972). Fig. 1 includes the data from laboratory measurements where the mean air temperature is controlled at 21.5 ± 0.1 °C and the wet bulb temperature between 12 and 14 °C. $T_a - T_w$ has been used in Fig. 1 to conform with Hasse's (1963) Fig. 2, although $T_a - T_w$ does not have a unique proportionality to Q, since evaporative and radiative heat fluxes may vary independently of temperature. (See further discussion of this figure below.)

Thus, it is evident from (6) that, if the water was in free convection, when Woodcock and Stommel obtained their data, their 1.5 mm diameter mercury thermometer misses more and more of the steep gradient in temperature near the interface as the heat flux increases; the contention being that the indicated limit on ΔT is artificial, introduced by the measuring technique.

There may, of course, have been additional reasons for a low ΔT in the particular circumstances of those measurements. No data on the turbulence in the water were given, but the water bodies are referred to as ponds, so it is unlikely that strong currents and Reynolds stresses were present. Also, according to Jones (1974) the eddy exchange coefficient is smaller near the surface in cases where tides create currents than when wind stress is the driving force. Supporting evidence that water currents have weak effects on the thermal boundary layer is Ewing & McAlister's (1960) finding that unless their underwater pump ruptured the surface in "the manner of a bubbling spring", it did not produce measurable effects on the radiation temperature, which agrees with my own experience. Since the winds were weak during Woodcock and Stommel's measurements, it is likely that the free convection regime applies.

Another force which could conceivably affect the boundary layer will be considered for completeness, namely the force due to surface tension.

4. Surface tension effects on the thermal boundary layer

When looking at the millimeter-thick boundary layer on a kilometer deep ocean it is perhaps not obvious that the length scale to be used when evaluating the Rayleigh number should be the total depth of the ocean, or the depth of the oceanic mixed layer, rather than the thickness of the boundary layer itself.

Let us for a moment consider a boundary layer of 1.5 mm thickness as a separate entity. Across such a layer it would require temperature differences in excess of 33 °C to exceed the Rayleigh criterion (for a critical Rayleigh number of 650; Katsaros, 1969). Thus, if 1.5 mm were the relevant depth, the Rayleigh mechanism could not be responsible for the observed convective motions. Another candidate is a mechanism due to horizontal variations in surface tension, which was discussed in relation to the oceanic thermal boundary layer by Lee (1972), and Webb (1974). (A comprehensive historical review of this subject is given by Berg et al., 1966a.) In regions of divergence on the surface, the water is somewhat warmer than in the convergence lines (the two-dimensionality of the sinking cold water has been photographed by Spangenberg & Rowland, 1960; and Katsaros, 1975). Surface tension of the warmer water is less than that of the colder water. Therefore, the horizontal surface tension gradient tends to amplify the convection. The critical value of the temperature gradient in surface tension driven convection is obtained from the Marangoni number, M, which relates surface tension variation with temperature, $\partial\sigma/\partial T$, and the vertical temperature gradient, $\partial T/\partial z$, to the diffusivity for heat, κ, and kinematic viscosity, ν, viz.:

$$M = \frac{-\dfrac{\partial\sigma}{\partial T}\dfrac{\partial T}{\partial z}\cdot d_{st}^2}{\rho\nu\kappa} \tag{7}$$

σ, the surface tension of water is both a function of temperature and concentration of solutes. A critical value of 80 for the Marangoni number was determined by Pearson (1958). For a water layer of 1.5 mm thickness Lee (op. cit.) gives the limiting temperature drop of only 0.054 °C. However, Lee also shows, as did Pearson, that in a deep convecting water layer the Rayleigh number is the critical one, while for shallower layers the Marangoni mechanism dominates. His argument is as follows.[1]

Finding the critical thicknesses for water layers in convection driven by surface tension and buoyancy respectively, he obtains from (3) and (7)

$$d_c^4 = R_c\,\nu\,\kappa\bigg/\left(g\alpha\frac{\partial T}{\partial z}\right) \tag{8}$$

$$d_{st}^2 = M_c\,\rho\cdot\nu\cdot\kappa\bigg/\left(-\frac{\partial\sigma}{\partial T}\cdot\frac{\partial T}{\partial z}\right) \tag{9}$$

where R_c and M_c are the critical Rayleigh and Marangoni numbers; exact values for the critical condition are not important for the present argument. Since d_{st} enters to the second power in the Marangoni number, while d_c enters to the fourth in the Rayleigh number, one expects surface tension effects to be more important for thin layers of water. Lee shows that for typical values of the constants in water, surface tension dominates for $d <$ 21.6 mm, the two forced being of equal magnitude for d = 21.6 mm. However, both Lee and Webb consider the relevant depth to be the boundary layer thickness and contend that the reason we observe negative ΔT's greater than -0.054°C is due to surfactants which reduce $\partial\sigma/\partial T$ drastically even for very small amounts of contaminant (Palmer & Berg, 1972). The divergent regions have less surface solutes and the surface tension is therefore higher, while in the convergent regions σ is not as high as it would be without the solutes. The result is that the Marangoni effect is diminished. This was advanced by both Lee and Webb as an indication of the omnipresence of water pollutants, natural or man-made.

However, the discussion above rests on the assumption that the boundary layer can be considered independent of the rest of the fluid, implying that the depths in (8) or (9) should be of the order of millimeters. This is in opposition to Pearson's (1958) conclusion. There exists, in addition, experimental evidence for the dominance of the buoyancy force for water depths > about 10 mm. Berg & Morig (1969) used the schlieren technique to observe the effect of stabilizing or destabilizing density gradients on convection driven by diffusion of a solute across the boundary of a two-layer liquid system. Transfer of the solute across a boundary can be designed to cause instability in either the upper or the lower layer or both. If the density gradient becomes gravitationally stable in either layer, the motions, in that layer, which are driven by surface tension forces are confined to a narrow region just below the interface. If the density gradient becomes unstable, deep streamers penetrate far from the interface. This latter situation is the typical one on oceans and lakes for upward net heat loss. Since even for depths of only a few centimeters both M_c and R_c would usually be exceeded in clean water, one can conclude that the buoyancy force dominates in determining the motion and one should look to eq. (5) for predicting ΔT.

It has been the experience of researchers on liquid pools (Berg et al., 1966b) that water does not show the surface tension driven convection, even in shallow layers (<10 mm), because it is *always* contaminated enough (even after distillation) to increase the critical Marangoni number by orders of magnitude. Thus, Webb's (1974) assertion of the omnipresence of surfactants on water is verified. It is, however, not the primary reason we observe large ΔT on the oceans. We have, in fact, two theoretical and experimentally tested reasons why the buoyancy force dominates the motion in turbulent free convection in water: (a) the large depth of the fluid makes the critical Rayleigh number exceeded more strongly than the critical Marangoni number, and (b) the high surface tension of pure water makes it so attractive to surfactants that natural water is always contaminated and the Marangoni mechanism becomes inoperative.

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Table 2. *Environmental conditions during measurements of the data obtained for this study*

Location	T_b (°C)	T_a (°C)	T_{WB} (°C)	Wind speed	Currents, waves	Water depth (m)	Sky condition	Technique: Rad, infrared radiometer: Res, resistance wire	Average ΔT °C	Number of samples	Comment
Laboratory	46.0	22.0	16.0	none	none	0.45	ceiling	Res	2.4	3	—
University of	43.6	22.4	16.0	none	none	0.45	ceiling	Rad	1.9	3	8.4% NaCl
Washington	36.3	21.4	12.2	none	none	0.45	ceiling	Res	1.7	100	—
	33.6	21.4	12.2	none	none	0.45	ceiling	Res	1.2	100	—
	32.6	21.4	12.2	none	none	0.45	ceiling	Res	1.4	100	—
	30.8	22.4	16.0	none	none	0.45	ceiling	Res	1.0	100	—
	30.7	21.4	12.2	none	none	0.45	ceiling	Rad	0.95	3	—
	30.3	21.4	12.2	none	none	0.45	ceiling	Res	1.0	100	—
	28.8	21.4	12.2	none	none	0.45	ceiling	Res	0.9	100	—
	28.1	21.4	12.2	none	none	0.45	ceiling	Rad	0.8	3	—
	26.9	21.4	12.2	none	none	0.45	ceiling	Res	0.7	100	—
	26.3	21.4	12.2	none	none	0.45	ceiling	Res	0.75	100	—
	26.0	21.4	12.2	none	none	0.45	ceiling	Rad	0.5	3	—
	24.2	21.4	12.2	none	none	0.45	ceiling	Rad	0.4	3	—
	1.0	−10.0	$\simeq -11.0$	none	none	0.45	ceiling	Rad	0.8	2	3.5% NaCl
	1.0	−20.0	$\simeq -21.0$	none	none	0.45	ceiling	Rad	1.2	2	3.5% NaCl

Field											
North Sea, 54.9° N 8.2° E, Sept. 26, 1973	15.2	11.0	9.0	< 3 m/sec variable	0.4 m/sec tidal current, < 0.5 m swell, no wind waves	13.0	variable strato-cumulus	Rad	0.6 0.8	8 9	radiometer 4 m above water surface
Lake Washington, Jan. 27, 1975	7.4	−0.8		< 2 m/sec variable	no mean current, low amplitude, waves occasionally	4.7	variable stratus	Rad	0.8	16	max. $\Delta T = 1$ °C
Lake Washington, March 5, 1975	6.8	1.4 0.5	0.8	calm < 1 m/sec	no mean current, low amplitude, gravity waves, < 10 cm amplitude	4.7	mostly clear, a few altocumulus	Rad	1.1 0.75	7 5	radiometer 0.5 m above water surface
Alpental, Washington, April 1, 1975	34.0	3.0	1.6	shielded < 1 m/sec	none	1.3	stratocumulus, stratus	Rad	2.5	31	max. $\Delta T =$ 3.4 °C

We will return to the effect of surface films below. First, measurements of ΔT in the laboratory by the techniques which led to the relationship $\Delta T \propto Q^{\frac{3}{4}}$ will be presented, then field data obtained to show that this relationship carries over to nature, when winds are weak.

5. Laboratory observations

ΔT and heat flux were measured in a tank of dimensions 0.50×0.75 m cross-section with water depths of 0.45 m for room to bulk water temperature differences up to 18 °C.

The ΔT across the boundary was determined by two methods. For the first technique, which was also used in the field, a Barnes Engineering Company PRT5 infrared radiometer with a 2° angular opening was positioned 2 m above the water surface. Periodically the boundary later was destroyed by vigorously stirring the water. This increases the turbulent Reynolds stress which decreases the boundary layer depth by (1). If the boundary layer is made infinitesimally thin, the increase in observed radiation temperature is a measure of ΔT.

Alternatively, the temperature profile in the water was measured with a 25 μm diameter resistance wire. The difference between the mean temperature from a horizontal traverse at the level where the wire just begins to show a distortion of the surface and at the depth in the interior where no vertical gradient can be discerned gives a value of ΔT. In some cases the ΔT is taken from the average profile obtained from 100 individual ones.

Heat flux was determined calorimetrically with carefully measured corrections for losses through the walls (see Katsaros et al., 1977). The observations are summarized in Table 2 and the ΔT values plotted in Fig. 1.

Maximum ΔT observed was 3.2 °C with a heat flux of 860 wm^{-2} and $T_a - T_w = -25$ °C. (In evaporating 4.6 or 8.4% sodium chloride solutions the ΔT for the same heat flux is reduced by 10 to 20%, because of the larger value of the "thermohaline" Rayleigh number compared to the purely thermal one (e.g., Stern, 1960; or Turner, 1973).)

6. Field observations at low wind speeds

On September 26, 1973, during the Joint North Sea Wave Project II (JONSWAP II), the same radiometric technique was used to evaluate ΔT, except that in this case the boundary layer was disrupted by throwing buckets of water drawn from about 0.5 m depth into the field of view. This rather primitive technique was used because the radiometer had to be mounted on the railing about 4 m above the water on the F/S *Gauss*. The ship was 4 km west of the Island of Sylt on this day. ΔT was measured during a period of low winds and waves and -4.2 °C air–water temperatures difference. Currents in the water were weak at the time (Krall, personal communication). An average value of 0.7 °C $\pm$ 0.1 °C results from 17 observations (Table 2). Because the radiometer measures both T_0 and T_w, corrections due to reflection from sky and ship, which would be very difficult to assess, are changed very little between the two measurements. Specular reflection from water at 8–12 μm wavelengths is about 2% at zenith angles from 0–30° increasing rapidly beyond (McSwain & Bernstein, 1961). With r.m.s. slopes on water typically $<16°$ (Cox & Munk, 1954) about the same fraction of the radiance entering the radiometer would be due to reflection in the disrupted and undisrupted conditions and the reflected signal would come from approximately the same cone of the sky. If any correction needs to be applied it is likely that there would be more reflected zenith radiance from the roughened surface than from the calm one, observed at an angle of about 20°, so that T_w would be too cold and ΔT would be slightly underestimated. A possible larger error may occur if the boundary layer of the total field of view was not completely disrupted. This is difficult to ascertain from ship-board. If such an error occurred, 0.7 °C would again be an underestimate of ΔT.

Measurements of ΔT under low wind conditions were also obtained in Lake Washington, Seattle, Washington, and in a heated swimming pool at Alpental, Washington. The experimental instrumentation and procedure were the same as described above with hand-stirred water to reveal T_w, and on the spot calibrations of the sensing and recording system. The results and environmental conditions are listed in Table 2 and plotted in Fig. 1.

A new boundary layer is quickly re-established after a stirring. This takes only about 10 sec., but 10 min was often required before the convection pattern in the signal was fully re-established. This is indicative of the lack of advection.

One data point in Fig. 1 at $T_a - T_w$ of -2 °C is taken from Ewing & McAlister (1960) who also used an infrared radiometer and obtained this measurement from Scripp's pier in 7 m of water at a wind speed of 0.5 m sec^{-1}. Two values also at low air–water temperature difference are from Hill (1970).

A large range between mean and maximum ΔT values at large negative $T_a - T_w$ occurs because the boundary layer in turbulent free convection is an intermittent phenomenon (Katsaros et al., 1977) with space and time scales larger than the temporal and spatial resolution of the radiometer (0.3 sec and 0.25 to 1.0×10^{-2} m^2 respectively). The field data of $T_0 - T_w$ is probably also affected by occasional weak wind gusts and waves.

Although there is some scatter of the data from both field and laboratory, the recent data obtained with fine resistance wires or infrared radiometers, show the general trend of increasing negative $T_0 - T_w$ for negatively increasing $T_a - T_w$.

$T_a - T_w$ is, of course, only a crude indicator of the total heat flux. However, since saturation vapor pressure at the surface is a function of T_0, sensible and latent heat flux tend to increase together. This is especially true when $T_0 - T_a$ is large, since the atmospheric vapor pressure is then a smaller fraction of the vapor pressure at the surface. However, variations in relative humidity and the non-linear variation of saturation vapor pressure with absolute temperature are not included. The net infrared heat flux may also vary independently of $T_0 - T_a$, and is usually a larger fraction of the total heat flux in the field compared to the laboratory.

Keeping these limitations in mind Fig. 1 shows none the less clearly that negative ΔT's in excess of -1 °C are possible for water in free convection.

7. Concluding remarks

It has been shown that the surface temperature deviation from the bulk value in evaporating and cooling water bodies does not have a -1 °C limit as suggested by earlier data obtained with a thin bulb mercury thermometer. At low wind speeds much greater values of ΔT have been observed. It was also argued, using relation (6), that if these earlier

Table 3. *Locations where the air–sea temperature difference is climatologically < −2.5 °C for at least two of the three months in a season*
The locations where the winds have force Beaufort 2 or less (i.e. < 2–3 m sec^{-}) at least 10% of the time are marked by an asterisk. (From *U.S. Navy Marine Climatic Atlas of the World*, 1955–1959)

Location	Season: Dec., Jan., Feb.	Mar., Apr., May.,	June, July, Aug.	Sept., Oct., Nov.
Northern Hemisphere	*China Sea up to (−4.5 °C)			*China Sea
	East Coast U.S.A.	*East Coast U.S.A. Gulf Stream		*East Coast U.S.A. Gulf Stream
	*Gulf Stream East of Florida			
	Davis Strait–Baffin Bay (−7.5 °)			*Davis Strait–Baffin Bay
	*Barents Sea	*Barents Sea		*Barents Sea
Southern Hemisphere			*Indian Ocean S.E. of Madagascar	Indian Ocean S.E. of Madagascar
		*S. of Cape of Good Hope	S. of Cape of Good Hope	
		E. of Australia	*E. of Australia extending to 160 °W, just W. of Australian coast	
			*E. of New Zealand	
			*E. of S. America 40–50 °S, 40–50 °W	

measurements were made in water which was in turbulent free convection, then more and more of the steepest part of the temperature gradient would be excluded by a thin bulb mercury thermometer as the heat flux increased. It has been established in earlier studies that the largest values of the surface temperature deviation occur when Reynolds stresses are small (Saunders, 1967b; Hasse, 1971) and there is furthermore no theoretical limit of ΔT in buoyancy driven convection. Consequently, a real limit of ΔT on the earth is only given by the maximum net upward heat flux possible in earth's most extreme climates under low wind conditions. Such extreme values of ΔT can be expected to occur just off-shore where cold dry continental air slowly advances over a warm ocean. Table 3 lists some examples of oceanic locations and seasons where and when extreme ΔT's might be observed. How often weak wind conditions are correlated with extreme $T_a - T_w$ cannot be deduced from the existing data. Conditions of extreme heat flux with weak winds might more readily be met on shallow lakes in the fall or on industrial cooling ponds.

It was also argued that the limit on ΔT and the boundary layer thickness imposed by surface tension gradients which develop as a consequence of temperature gradients does not apply to deep water bodies. Both theoretical justification and experimental proof of this exist in the engineering literature. This seems also to be borne out by the laboratory data of ΔT reported here.

The effect of surface films in changing ΔT should, however, not be overlooked. The presence of an organic film on the surface, usually does not inhibit evaporation significantly (Barger & Garrett, 1968), but will rather give rise to *increased* ΔT. With a film on the surface, which does not take part in the convective motions, the temperature gradient retains a steep slope through this layer as well as in the top layer of the water. If the boundary layer and the surface film are of comparable thicknesses and conductivities, the surface temperature deviation can double. Hill (1970) found in laboratory observations that a maximum ΔT of $-2.0\,°C$ occurred for low wind speeds at the downwind end of his tank where a surface "slick" collected, while without the slick the ΔT was $-1.2\,°C$.

8. Acknowledgements

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ТЕМПЕРАТУРНЫБ ОТКЛОНЕНИЯ НА ПОВЕРХНОСТИ МОРЯ ПРИ ОЧЕНЬ СЛАБОМ ВЕТРЕ. ЕСТЬ ЛИ ГРАНИЦА?

Лабораторные и полевые измерения спада температуры вдоль пограничного слоя верхней поверхности испаряющейся и охлаждающейся воды при турбулентной и свободной конвекции указывают на то, что возможны более высокие значения, чем зарегистрированные прежде. Верхняя граница отклонения температуры поверхности обусловливается только максимальной теплоотдачей имеющей место в чрезвычайных климатических условиях.