The Temperature at the Ocean-Air Interface[1]

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ABSTRACT

A simple theory is presented to account for the difference between the temperature at the ocean-air interface and that of the water at a depth of about one meter. Except in very light winds and intense solar radiation the mean temperature difference ΔT is expected to be of the form

$$\Delta T \propto \frac{q}{(\tau/\rho_w)^{\frac{1}{2}}},$$

where q is the sum of the sensible, latent, and long-wave radiative heat flux from ocean to atmosphere and τ/ρ_w is the kinematic stress. No data are available to test this prediction.

The influence of slicks and solar insolation on interface temperature is also briefly discussed.

1. Introduction

Recent years have witnessed an explosive increase in the use of radiometric techniques for the remote measurement of the surface temperature of the ocean (Clark, 1964). Successful efforts by the author (1967) to improve the absolute accuracy of measurements made with an airborne instrument, yielding a precision of ± 0.2C, have re-awakened interest in the connection between the temperature at the ocean-air interface and that in the uppermost meter of water. For some decades it has been known that the surface is generally cooler than the subsurface water with the major temperature variation concentrated in the uppermost millimeters [see the summary in Roll (1965)]. When used from a near surface position, radiometers afford a unique tool for determining the magnitude of this surface anomaly, and, when their idiosyncrasies are more widely understood, it seems likely that they will become indispensable in field experiments where the transfer of heat, moisture, and momentum are measured. Already they reveal what perhaps might be guessed from the patterns of sea surface streaming observed by Woodcock (1941), *viz.*, that the surface of even a well-mixed ocean can exhibit considerable fluctuation in temperature, >0.5C (Fig. 1b), and can possess a mean value which differs by 0.3 C from the conventional dip-bucket value. Only when the exchange of energy between ocean and atmosphere is vanishingly small (Fig. 1a) is the surface the uniform, noise-free boundary that it is convenient to assume.

It is in the area of sea-air exchange that the temperature assumed by the interface is of prime concern; yet apart from Hasse (1963), researchers have generally assumed that the surface and subsurface temperature are identical. The magnitude of the temperature anomaly at the ocean surface is commonly a non-negligible fraction of the air-sea temperature difference as conventionally defined, and over the tropical oceans the anomaly is such that even the *direction* of the heat flux may be misjudged.

In this short paper we shall attempt to establish what factors determine the difference between the temperature at the ocean-air interface and that at a depth of some tens of centimeters, including the two more important in a simple theoretical discussion. Comments will also be made on two others whose contribution is less easy to assess.

2. Principal factors determining the interface temperature

At the interface continuity requires that q, the flux of sensible heat within the ocean, balance $H+LE+R$, the total flux of heat within the atmosphere, where H is the sensible heat flux, LE the latent heat flux, and R the net long-wave radiation. At night these are the only terms in the energy equation and we infer that within the uppermost meter of the ocean q changes only slowly with depth. By day, because of the absorption of direct and diffuse solar radiation, we can no longer consider q independent of depth even in the topmost millimeter. Some discussion of this difficulty is presented in Section 3; in what follows we shall assume solar radiation small enough to be ignored.

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Our discussion is based on a recognition that except in very light winds (perhaps less than 2 m sec^{-1}) the Richardson number for the upper meter of the ocean is very small. As in the atmosphere immediately above the surface, heat transfer is *forced* or passive rather than *free*.

The stress of wind on water is commonly considered to have two components, a wave drag in which mo-

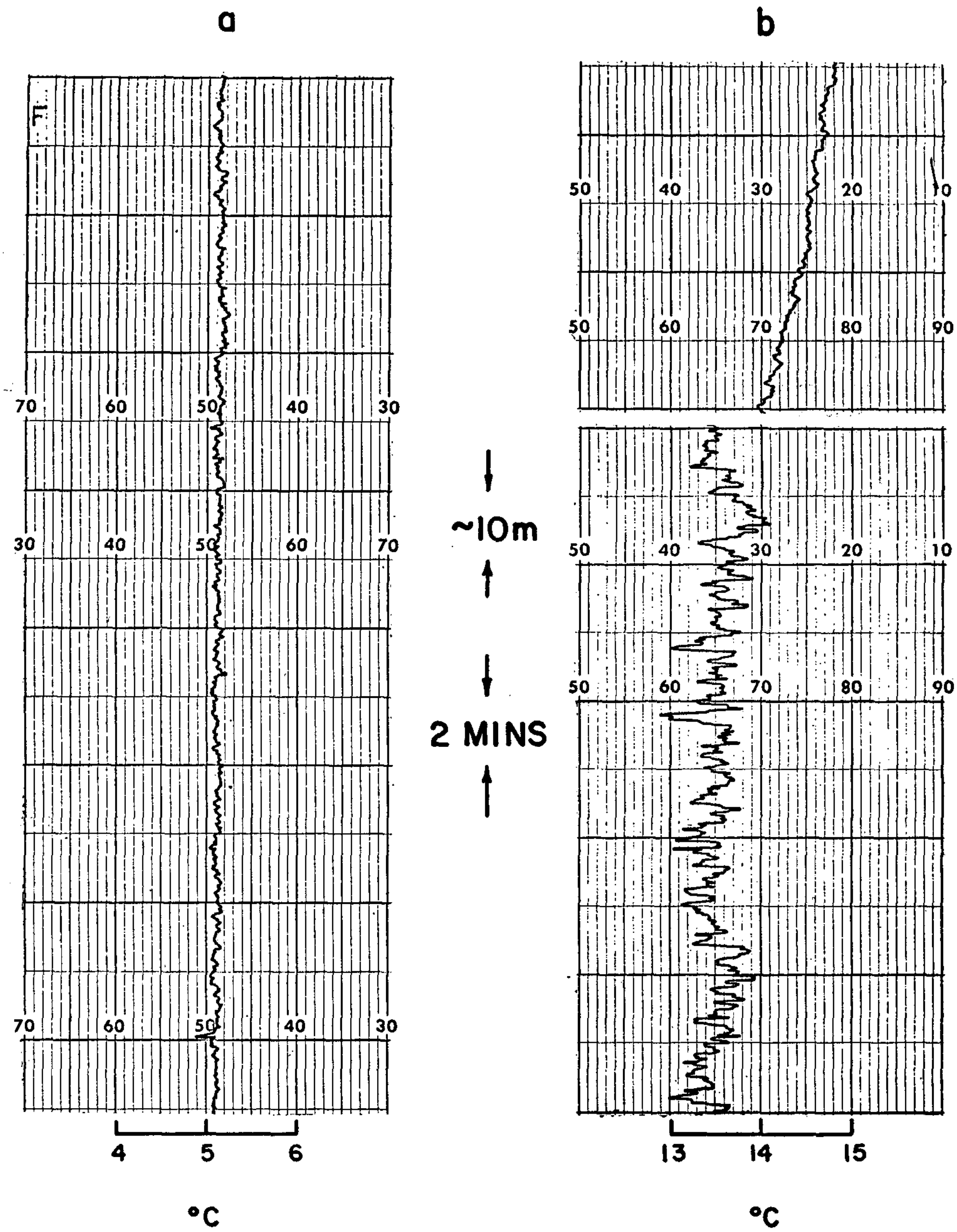

FIG. 1. Time series record of radiometric observations of sea surface temperature (corrected for non-blackness of surface). Corresponding meteorological observations were as follows:

	(a)	lower (b)
Air temperature (°C)	5.2	12.0
Dew point (°C)	3.9	7.8
Dip-bucket temperature (°C)	5.2	13.9
Wind (m sec^{-1})	2.5	10
Sky	low overcast	clear
Current, tidal (cm sec^{-1})	...	10

The record shown in upper (b) is from a well-stirred open water bath and indicates instrument noise in conditions of lower (b).

mentum is transmitted to the water because of the correlation between local pressure and local surface slope, and a frictional drag which acts tangentially to the local surface. Although in recent discussion Stewart (1961) has championed the role of wave drag, we know of no evidence which establishes that the frictional component is negligible. Accordingly, we hypothesize the existence of a thin quasi-laminar region in the ocean and adjacent to its surface within which the frictional component of the wind stress is communicated viscously. If the mean thickness of this region δ is determined only by the viscous stress τ', the kinematic viscosity ν, and the water density ρ_w, a dimensional argument leads to the relation

$$\delta \sim \nu \Big/ \left(\frac{\tau'}{\rho_w}\right)^{\frac{1}{2}}. \tag{1}$$

Co-existing with the viscous layer is supposed a region within which the flux of heat is transmitted principally by molecular conduction. Since the Prandtl number ν/κ, where κ is the thermal diffusivity, is larger than unity (6–13), the conduction region is probably somewhat thinner than the viscous region. If we accept the analogy with a rigid boundary, the ratio of their thicknesses is $(\nu/\kappa)^{-\frac{1}{3}}$ or approximately one half; however, the assumption is not essential to this part of our discussion. Taking the major part of the temperature variation to occur within the conduction region we obtain

$$q \sim \frac{k\Delta T}{\delta}, \tag{2}$$

where k is the thermal conductivity of sea water and ΔT is the mean temperature difference between the surface and the interior.

Combining these relations and introducing a numerical coefficient λ to take account of the order of magnitude character of our argument[3] we obtain

$$\Delta T = \frac{\lambda q \nu}{k(\tau/\rho_w)^{\frac{1}{2}}}. \tag{3}$$

Whether the surface is cooler or warmer than the interior depends only on the direction of total heat flow; since this is generally from ocean to atmosphere we conclude that the ocean is usually covered with a "cool skin." Because the fluxes of sensible and latent heat and the square root of the stress are approximately proportional to the surface wind speed, a dependence not shared by the net flux of long-wave radiation, it is readily seen that for a given thermal/moisture contrast between ocean and atmosphere the temperature anomaly ΔT is most pronounced in light winds and declines with increasing wind.

Although there is no comprehensive body of data available in a form suitable to test Eq. (3), a few observations by the author of ΔT between 0.2 and 0.3C (Saunders, 1967), together with corresponding estimates for q and τ, suggest that λ is from 5 to 10. A temperature difference of 0.2–0.3C is probably typical of much of the sub-tropical and tropical areas [see also the observations of Hasse (1963)], but in middle-latitude winters and in the western parts of the oceans we compute with $\lambda = 7$ that in extreme circumstances temperature differences may reach 1C.

Not only does the temperature at the ocean-air interface differ from that of the subsurface water, but also the salinity. Surface waters are enriched in salts because of their separation from water in the evaporation process. Corresponding to the region of viscous dominance we suppose a region within which salts are returned to the ocean interior by molecular diffusion down a concentration gradient. We can compute the salt enrichment (or gas enrichment if we are concerned with the air-sea transfer of gases) by an argument entirely analagous to that used in determining the temperature anomaly. We shall make explicit use of the one-third power law mentioned earlier for the ratio of the thickness of the diffusion region to that of the viscous region. It is then readily shown that the fractional enrichment of the sea surface concentration of salt is given by the expression

$$\frac{\Delta c}{c} = \lambda \left(\frac{\kappa}{D}\right)^{\frac{1}{3}} \left(\frac{\nu C}{k}\right) \frac{E}{(\tau/\rho_w)^{\frac{1}{2}}}, \tag{4}$$

where D is the diffusivity of salt of concentration c, C is the specific heat of water, E the evaporation rate, and other terms are as defined previously.[4] Since both the evaporation rate and the square root of the wind stress are approximately proportional to the wind speed we see that the salt enrichment depends principally on the moisture contrast between ocean surface and atmosphere. Even when the difference in specific humidity is as large as 10 gm kg^{-1}, we find $\Delta c/c$ only about 2%. It can be concluded that in computing the equilibrium vapor pressure for air in contact with the ocean it is a sufficiently good approximation to assume the surface salinity equal to the conventional near-surface value.

In a dead calm, when the Richardson number is very large, the surface temperature of the ocean may also differ significantly from its interior value. Studies of *free* convective loss from a horizontal surface or between parallel horizontal surfaces demonstrate a rela-

[3] For example, we shall make τ the total wind stress, as conventionally measured, and absorb the factor expressing the frictional component in λ.

[4] If we assume that the viscous, conduction, and diffusion regions are of equal thickness, the only change in (4) is to make the power of κ/D unity.

tion of the form

$$q=Ak\left(\frac{g\alpha}{\kappa\nu}\right)^{\frac{1}{3}}\Delta T^{\frac{4}{3}}, \tag{5}$$

or

$$q=\beta\Delta T^{\frac{4}{3}},$$

where g is the acceleration of gravity, α is the coefficient of thermal expansion of the fluid and other terms are previously defined. From the work of Globe and Dropkin (1959) the numerical coefficient A was determined as 0.20.[5] Values of β for sea water determined from the tabulations of Fofonoff (1962) are shown in Table 1.

TABLE 1. Magnitude of cool skin from Eq. (5).

Temperature, [°C]	0	10	20	30
β [cal cm^{-2} min^{-1} (°C)$^{-4/3}$]	2.3×10^{-2}	3.4×10^{-2}	4.3×10^{-2}	5.1×10^{-2}
ΔT, [°C]	0.6	0.5	0.4	0.35

If we suppose that long-wave radiation to clear skies is the major agency for heat exchange between ocean and atmosphere and take a value of $q=0.12$ cal cm^{-2} min^{-1}, then the temperature difference derived from Eq. (4) is about 0.5C; see also the bottom line of Table 1.

3. Other factors which influence the interface temperature

In this concluding section we consider the role played by slicks and by solar radiation in influencing the temperature at the ocean-air interface.[6]

From a near surface position our radiometer records show that the surface temperature anomaly commonly increases when a natural oil film or slick moves into the field of view of the instrument. (We can show that the film must be thick enough to exhibit interference colors before its influence on the infrared emissivity of the surface can account for the observed effect; most natural slicks are believed monomolecular.) Such a surface temperature variation is sometimes a manifestation of the organized overturning to be found in the uppermost meters of the ocean in light and moderate winds [see the thermal maps of McAlister and McLeish (1966)], where surface convergence and descent concentrate slicks in regions of cold surface water. Slicks may also locally change both the stress and heat transfer between atmosphere and ocean but we believe that these effects are less important than that resulting from the change in hydrodynamic boundary condition. A film can rigidify a water surface [see, for example, Van Doorn (1966)] converting the boundary from a mobile state, in which spatially varying surface motions are permitted, to an immobile state, in which relative motions vanish. It is hypothesized that by arresting surface fluid a slick causes a local thickening of the viscous-conduction region within the ocean and hence causes an increased temperature drop between the surface and the interior. Essentially the same idea has been advanced by Jarvis (1963).

In our discussion we have assumed that vertical temperature gradients can exist only immediately adjacent to the ocean surface. Such an assumption is justified when there is sufficient wind mixing or, in the absence of wind, when the ocean is losing heat to the atmosphere. In light winds and strong insolation, a measurable temperature stratification may arise throughout the entire near surface layer and caution must be exercised in applying the arguments of the previous section.

As was mentioned earlier, even in the presence of wind mixing, solar radiation must be reckoned with. Schmidt [see Defant (1961)] has shown that 15% of the energy in the direct solar beam, mainly in the near infrared beyond 1.5 μ, is absorbed within the topmost millimeter of the ocean. This quantity is comparable with q. The assumption that q is invariant with depth has therefore to be replaced by the invariance of the sum of the conductive and radiative fluxes. If the resulting equation is integrated over the depth δ (determined externally by the stress), we obtain

$$q-\frac{1}{\delta}\int_0^\delta (F_0-F)dz\sim\frac{k\Delta T'}{\delta}, \tag{6}$$

where F is the solar insolation at depth z, F_0 the surface value, q the total heat loss across the surface as defined previously, and $\Delta T'$ the new temperature difference between the surface and the interior. At depths greater than δ it is supposed that turbulent processes redistribute absorbed solar energy and lead to uniformity of temperature. We conclude that to account for the effect of solar radiation q in Eq. (3) must be replaced by $q-\frac{1}{\delta}\int_0^\delta (F_0-F)dz$. For zenith sun the integral, which is approximately one half of the energy absorbed in depth δ, has values of 0.07 cal cm^{-2} min^{-1} when $\delta=0.1$ cm and 0.03 cal cm^{-2} min^{-1} when $\delta=0.05$ cm. Because q is usually larger than these values we deduce that the ocean is commonly covered by a cool skin by day as well as by night.

Because of the tentative nature of this discussion it is anticipated that a fair test of Eq. (3) would be made in circumstances where the absorption of solar energy is small enough to be neglected.

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[5] Globe and Dropkin's definition of ΔT differs by a factor of 2 from that used here.

[6] Ewing and McAlister (1960) have commented on the importance of wave breaking: we have nothing further to add to their views.

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