THE COOL SKIN AND THE WARM LAYER IN BULK FLUX CALCULATIONS

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1. Background

Bulk flux routines make use of the sea surface temperature as the lower thermal boundary condition. Logically, the proper temperature is taken at the air-sea interface, T_s; this is essentially that temperature determined by an infrared radiometer. However, the measurement of T_s by infrared radiometer requires careful correction for reflected atmospheric radiance and places extreme demands on the absolute calibration accuracy and stability of the device. Inexpensive instruments that can routinely do this job are not available, so we must make do with *in situ* temperatures placed in the water. These so-called bulk temperature sensors may use ship intakes at 2 m depth or floating thermistors at a few cm depth. To obtain the correct interfacial temperature, these bulk temperatures must be corrected for the "warm layer" and the "cool skin". The warm layer is a region in the upper few meters of the ocean where solar radiation has caused significant warming relative to the deeper mixed-layer temperature. Warm layers occur during the day when temperature stratification caused by absorption of the solar flux is sufficiently strong to overcome shear-induced mixing. The cool skin is a layer in the upper few millimeters of the ocean caused by the combined cooling effects of the net longwave radiation, R_{nl}, the sensible heat flux, H_s, and the latent heat flux, H_l. The cool skin is on the order of 0.1 to 0.5 K and is almost always present (although its total effect may be cancelled by the presence of a warm layer).

In the interest of improving the accuracy of the COARE bulk flux algorithm, we have investigated various methods for correcting bulk temperature measurements for the cool skin and the warm layer. In section 2 we examine the physical principles, experimental basis, and the implementation of the cool skin correction, including the development of a wind speed dependent form of the Saunders coefficient. In section 3 we discuss the warm layer and show how it is described by the Price Weller Pinkel surface layer model.

2. The Cool Skin

The basic physics of the cool skin was described completely by Saunders (1967). The total cooling at the interface, Q, is given by

$$-Q = R_{nl} - H_s - H_l \qquad \textbf{(1)}$$

where we use the normal meteorological sign conventions (e.g., $R_{nl} = R_{\downarrow l} - R_{\uparrow l}$) and a positive value for Q represents cooling of the water. Because this cooling is realized at the interface (the

longwave penetration depth is about 10 μm), the temperature gradient at the interface is defined by molecular thermal conductive processes:

$$k\frac{\partial T}{\partial z}|_0 = Q \qquad \textbf{(2)}$$

where k is the thermal conductivity of water, z the vertical ordinate (+ downward), and T the temperature profile. Away from the interface, the temperature gradient is quickly destroyed by turbulent mixing . Thus, the cool-skin temperature change, ΔT, is confined to a region of thickness δ , which is referred to as the molecular sublayer. From (2) we can see that

$$\Delta T \simeq Q\delta/k \qquad \textbf{(3)}$$

where we use the convention that ΔT is positive if the surface is cooler than the bulk. We can estimate δ by assuming that it is proportional to the Kolmogorov microscale

$$\delta \simeq (\nu^3/\epsilon)^{1/4} \qquad \textbf{(4)}$$

where ν is the kinematic viscosity and ϵ the rate of dissipation of turbulent kinetic energy. For turbulence generated by surface shear ϵ scales as u_{*w}^3/δ where u_{*w} is the friction velocity (in the water) ; for turbulence generated by convection ϵ scales as $\alpha g Q_b/(\rho c_p)$ where α is the thermal expansion coefficient, Q_b the virtual surface cooling that includes the buoyancy effects of salinity due to evaporation, g the acceleration of gravity, ρ the density and c_p the specific heat of water.

For the shear generated turbulence regime, the friction velocity leads to

$$\delta_s = \frac{\lambda \nu}{u_{*w}}$$
$$\Delta T_s = \frac{\lambda Q \nu}{k u_{*w}} \qquad \textbf{(5)}$$

where λ is an empirical coefficient determined by simultaneous measurements of ΔT, Q and u_{*w} . For convectively generated turbulence (4) leads to a minus one-fourth power dependence of δ on Q_b; the form that is usually used follows from Rayleigh number scaling of free convection and

is

$$\delta_c = \left(\frac{k^2 \nu}{A^3 g \alpha \rho c_p Q_b}\right)^{1/4}$$
$$\Delta T_c = \left(\frac{\nu}{A^3 g \alpha \rho c_p k^2}\right)^{1/4} Q/Q_b^{1/4} \quad \textbf{(6)}$$

where A is an empirical constant. The form of the virtual cooling follows from the relation of the density flux to the thermal and salinity fluxes (Paulson and Simpson 1981)

$$Q_b = Q + \left(\frac{S\beta c_p}{\alpha L_e}\right) H_l \quad \textbf{(7)}$$

where S is the mean salinity, β the salinity expansion coefficient, and L_e the latent heat of vaporization. Note that $S\beta$ is relatively constant in the ocean ($\approx$0.026) but α decreases from about 3 x 10^{-4} K^{-1} in the tropics to near zero in polar regions. Thus, in the COARE region the parenthetical factor in the second term of (7) has a value of about 0.14 and , in the light wind conditions where (6) is applicable, Q_b will differ from Q by no more than 10% .

Numerous studies of λ have been performed over the ocean (see Robinson et al. 1984 or Fedorov and Ginsburg 1992 for reviews). Fedorov and Ginsburg list 12 different field studies of λ - a typical value being 6. There are reports of λ increasing with wind speed and there are reports of λ decreasing with windspeed. Because many of the studies ignore the buoyancy driven regime, their smaller values of λ in light winds are quite consistent with (6). Most of the studies suffer from either crude determinations of at least one of the key terms in computing Q or inadequate accuracy for T_s. Consider that a typical value for ΔT is about 0.3 K but an accuracy of $\pm$0.1 K for T_s has yet to be demonstrated. Even with careful longwave radiative flux measurements, uncertainties in bulk flux computations (particularly in the earlier studies) suggest that Q is uncertain by at least 10%. Thus, 30% systematic errors in estimates of λ are difficult to avoid. For example, Hasse (1971) used empirical formulae to estimate Q and determined T_s by equating it to the measured air temperature when the gradient was small. Paulson and Simpson (1981) used bulk formulae for the turbulent fluxes and direct measurements of R_{nl} but T_s was determined from a narrow band IR thermometer corrected for sky reflection with broad-band , global IR flux data. Furthermore, the absolute calibration of the IR thermometer was adjusted with a polynomial of order 5 generated by minimizing the rms difference between the measured values of ΔT and Saunders' formula. Two recent experiments (Schluessel et al. 1990 and Coppin et al. 1991) have featured improved IR thermometers and flux measurements, although their daytime results are confused by the effects of the warm layer. Typical night time values of ΔT are about 0.3 K and are consistent with λ between 4 and 8. However, these new measurements have not really cleared up the picture because they give conflicting results; Schluessel et al.'s largest ΔT's are at high wind speeds whilst Coppin et al.'s are at low wind speeds.

Fedorov and Ginsburg (1992) list three laboratory and one field study of the convective coefficient A; values vary from 0.16 to 0.26. We are going to use 0.23 (i.e., $A^{-3/4} = \lambda/2 = 3.0$), which is midway between the value 0.2 suggested by Saunders (1967) and the value 0.25 recommended by Fedorov and Ginsburg (from Shigayeva et al.'s 1982 field program in the Sea

of Japan). Combined with λ=6 this will lead to a shear-driven to convective driven transition beginning at about 4 ms^{-1} for typical values of Q; this is consistent with Paulson and Simpson's results (see their figure 8).

The convective and shear forms for δ can be smoothly blended by approximating their relative contributions to the dissipation rate. For example, the Kansas experiment (Wyngaard and Cote, 1971) yields

$$\epsilon = (\frac{u_*^3}{\kappa z})(1+(z/L)^{2/3})^{3/2} \quad \textbf{(8)}$$

where L is the Monin-Obukhov stability length. We can approximate (8) as

$$\epsilon \approx \frac{u_{*w}^3}{\delta \lambda^3} + constant \frac{g\alpha \rho c_p Q_b \nu^2}{k^2} \quad \textbf{(9)}$$

where the constant will be chosen to be consistent with (6). Dividing by $(\nu)^3$ yields

$$\frac{1}{\delta^3} = \frac{1}{\delta_s^3} + \frac{1}{\delta_c^3}\frac{\delta}{\delta_c} \quad \textbf{(10)}$$

We assume that the δ/δ_c term on the right hand side of (10) is about 1 and we use (10) to incorporate these effects into a λ coefficient that is not constant but depends on conditions. Thus

$$\delta = \frac{\lambda \nu}{(\rho_a/\rho)^{1/2} u_{*a}} \quad \textbf{(11)}$$

where ρ_a is the density of air and u_{*a} is the atmospheric friction velocity. The cool skin is given by

$$\Delta T_{cool} = \frac{Q\delta}{k} \quad \textbf{(12)}$$

and

$$\lambda = 6[1+(\frac{Q_b 2^4 g\alpha \rho c_p \nu^3}{u_{*a}^4 (\rho_a/\rho)^2 k^2})^{3/4}]^{-1/3} \quad \textbf{(13)}$$

Shear and convective effects are comparable when the term inside the three-quarter power is on the order of one (yielding a value for λ of 4.8). This occurs (using parameters appropriate for COARE) when

$$u_{*a}^4 = 4.4x10^{-7} Q_b \quad \textbf{(14)}$$

which implies $u_{*a} \approx$ 0.08 ms^{-1} and u $\approx$ 2.5 ms^{-1}.

The final point we must consider is the effect of solar heating. Saunders (1967) pointed out that about 5% of the net solar flux will be absorbed in the nearest 1 mm of the water surface and that this heat acts to reduce Q. If we define δS as the fraction of the net solar flux absorbed within the sublayer, then we simply replace Q in (12) with Q-δS. Following Saunders (1967) and Paulson and Simpson (1981), we represent δS as

$$\delta S = R_{ns} \sum_{i=1}^{N} F_i [1 - \frac{\gamma_i}{\delta}(1 - \exp(-\delta/\gamma_i)] \qquad \textbf{(15)}$$

which describes the absorption in 9 wavelength bands of amplitude F_i and absorption length γ_i (see Table 1). The first four bands have absorption scales much greater than 1 mm and the last four have absorption scales much less than 1 mm. Thus, we can approximate the absorption in the sublayer (δ < 1 cm) as

$$\delta S = S_{ns}[0.137 + 11\,\delta - \frac{6.6x10^{-5}}{\delta}(1 - \exp(-\delta/8x10^{-4}))] \qquad \textbf{(16)}$$

where 0.137 is the sum of the last five F_i , and 11 is one-half the sum of the first four values of F_i/γ_i . For δ = 1 mm , δS is 0.091*S_{ns} and, for even quite thin sublayers, at least 5% of net solar will be absorbed.

In is not clear that such a complicated scheme for adjusting this correction depending on the actual value of δ is justified by the present accuracy of λ and the specification of (15). Paulson and Simpson (1981) obtained better results with (15) compared to the simple approximation $\delta S = 0.19*R_{ns}$ suggested by Hasse (1971), but a reduction of the coefficient from 0.19 to 0.10 would give reasonable agreement. However, for the very light winds often encountered in COARE, we will probably need to use (16) because δ will tend to be large. These results imply that for winds around the 1 ms^{-1} , the will be no cool skin for most of the day. On those occasions when Q is negative, then there is no cool skin and ΔT_{cool} is set to zero.

Using expressions (11), (12), (13), and (16) we have computed estimates of the cool skin for the Moana Wave COARE data. The average diurnal cycles of ΔT and Q-δS are shown in Fig. 1 for the entire data set (1622 50-min samples) and for the cases where the 10-m wind speed is less than 4.0 ms^{-1}. The average night time value is about 0.3 K and the average day time value is about 0.2 K. In Fig. 2 we show the same quantities averaged as a function of wind speed broken down into day and night cases. Note that the ocean convective effect causes the night time cool skin to reach a maximum value of about 0.35 at 1.5 ms^{-1}. In Fig. 3 we compare our parameterization with results from the two recent measurement studies described earlier. Schluessel et al. (1990) give a simple equation in terms of the bulk parameters for the night time case which is indicated by the circles. Coppin et al. (1991) state that for wind speed greater than 1.5 ms^{-1} the simple Saunders formula with λ=6.5 fits the average of their measurements (shown in Fig. 3 by the X's). Notice the substantial disagreement of these two field programs at low wind speeds. This disagreement is in fact due to the use of Schluessel et al.'s (1990) parameterization outside the range of conditions upon which it is based. This points out the necessity of using a more general, physically based model.

3. The Warm Layer

On a clear day the sun deposits an average of about 400 Wm^{-2} of heat into the ocean over the twelve daylight hours. Roughly half of this heat is absorbed in the upper two meters (the details are dependent on water clarity). In the absence of mixing, this is sufficient heat input to warm this 2-m deep layer uniformly by 2.0 K. Observations (Price et al. 1986; Fedorov and Ginsburg 1992) show that the region of significant warming begins at the surface and propagates downward as it intensifies with increasing solar intensity. Under light wind conditions, the near surface warming can be quite abrupt in the morning and even peak before local noon. Measurable warming occurs as deep as 20 m and may persist well past sundown. Because the majority of the solar heat is deposited in the upper two meters and the warming leads to a stably stratified surface layer, the turbulence necessary to mix this heat deeper into the mixed layer can only come from shear driven by wind. Thus, the characteristic depth of the warm layer, D_T , is smaller in light winds and the temperature increase across the warm layer is greater.

For the purposes of computing bulk air-sea fluxes we are interested in the difference between our subsurface *in situ* temperature measurements (at depth z) and the surface temperature,

$$T_s = T_m(z) - \Delta T_{cool} + \Delta T_{warm} \qquad \textbf{(17)}$$

Whereas the cool skin is unambiguously above any realistic *in situ* sensor we are likely to use, this is not true for the warm layer. Thus, the warm layer correction depends on the depth of our sensor - the shallower the sensor, the smaller the correction. Because the warm layer may be only a few meters deep on some days, we anticipate that an *in situ* sensor at one tenth this depth (say 10 cm) may require warm layer corrections on the order of tenths of degrees; a sensor at 1 m may require corrections on the order of degrees. This scaling analysis suggests that a sensor at 1 cm depth will not require a significant correction for the warm layer.

The physics of the cool skin are dominated by molecular transport processes with very rapid time scales and small spatial scales that are amenable to laboratory investigation. The same is usually not true for the warm layer. The vertical distribution of the warm layer is determined by a combination of the radiation absorption profile plus molecular and turbulent mixing processes on vertical scales of centimeters to tens of meters. The absorption profile is straightforwardly handled by expressions such as (15), although more attention must be paid to the clarity of the water. This is typically done by using the Jerlov classifications (Fedorov and Ginsburg 1992).

Whereas we believe that turbulent mixing is usually important in describing the warm layer, it is instructive to examine some analytical solutions for purely molecular heat conduction (taken from Carslaw and Jaeger 1971). For a semi-infinite medium initially at temperature T_b with a constant surface flux F_0 (equal to $F_i R_{ns}$ for each appropriate component in Table 1) imposed at time t=0, the surface temperature increases as

$$T_s - T_b = \delta T_0 = \frac{2F_0}{\rho c_p \sqrt{\pi \kappa}} t^{1/2} \quad \textbf{(18)}$$

where $\kappa = k/(\rho c_p)$ is the thermal diffusivity. The time scale for the surface warming to penetrate to some depth, τ_0, is related to D_k, the molecular conduction penetration depth

$$z = D_k = 2\sqrt{\kappa \tau_0} \quad \textbf{(19)}$$

For times much greater than τ_0 ,the difference in the surface temperature and the temperature at depth z is

$$T_s - T(z) = \Delta T_0(z) = \frac{F_0 z}{\rho c_p \kappa} \quad \textbf{(20)}$$

For the case of exponentially absorbed radiative heating (the analytical solution is left to devotees of Carslaw and Jaeger) described as per (15), the characteristic molecular conduction time scale becomes

$$\tau_e = \gamma_i^2 / \kappa \quad \textbf{(21)}$$

For times small compared to τ_e the surface warms linearly with time at a rate given by

$$\delta T_e(t<\tau_e) = \frac{F_0}{\rho c_p \gamma_i} t \quad \textbf{(22)}$$

For large times, the surface follows the same time to the one-half power as above, but with a constant thermal lag

$$\delta T_e(t>\tau_e) = \delta T_0 - \frac{F_0 \gamma_i}{\rho c_p \kappa} \quad \textbf{(23)}$$

For large times the difference between the surface temperature and the temperature at depth z, ΔT_e , is negligible if $z<\gamma_i$ and is the same as (20) for $z>\gamma_i$.

These results are quite informative, because they put an upper limit on the temperature errors we can expect in the absence of wind mixing. For example, for a sensor at 1 cm depth the last four components in Table 1 are effectively surface sources and the fifth contributes modestly to ΔT. Therefore, we expect that

$$\Delta T(z=0.01) = \frac{[(0.13+0.06)R_{ns} - Q]}{\rho c_p \kappa}(0.01) \quad \textbf{(24)}$$

describes the temperature error of a sensor at 1 cm depth. A net solar flux of 800 Wm^{-2} and Q of 100 Wm^{-2} gives a temperature error of 2.5 K! Of course these large surface warmings cannot actually occur because increases in T_s will cause increases in Q which will quickly limit the warming process. However, these results clearly imply that we must not be cavalier in assuming that a sensor at 1 cm depth provides an unambiguous measurement of the surface temperature in very light winds.

Price, Weller, and Pinkel (1986 - hereafter PWP) describe the essential turbulent dynamical processes in the development of the PWP mixed-layer model. These can be crudely summarized by assuming that, once the stratified warm layer forms shortly after δS exceeds Q, then the temporal integrals of the surface fluxes induce changes in the near-surface temperature and current that are confined to a depth D_T :

$$\Delta T = \frac{2\int(\delta S - Q)dt}{\rho c_p D_T} = \frac{2I_S}{\rho c_p D_T}$$
$$\Delta u = \frac{2\int u_{*w}^2 dt}{D_T} = \frac{2I_\tau}{D_T} \qquad \textbf{(25)}$$

where u denotes the magnitude of the current. The factor of two appears because we assume that the profile is linear from the bottom of the layer to the surface. The depth is determined by requiring the bulk Richardson number to be no greater than a critical value, $R_{ic} = 0.65$:

$$D_T = (2R_{ic})^{1/2}\frac{I_\tau}{(\alpha g I_S/\rho c_p)^{1/2}} \qquad \textbf{(26)}$$

This depth can be combined with (25) to yield an estimate of the temperature gradient

$$\frac{\Delta T}{\Delta z} = \frac{\alpha g(I_s/\rho c_p)^2}{R_{ic} I_\tau^2} \qquad \textbf{(27)}$$

If we assume that the time integrals over some time interval t can be approximated as $<\delta S\text{-}Q>t$, then we can write the temperature change between the surface and the depth z as

$$\Delta T_{warm} = \frac{\alpha g(<\delta S - Q>/\rho_a c_p)^2}{R_{ic}<u_{*a}^4>} z \qquad \textbf{(28)}$$

Furthermore, if we wish to select a threshold value for an acceptable temperature difference, say ΔT_{th}, then we can estimate the minimum wind speed to achieve the necessary wind mixing:

$$U_{th} = \left[\frac{\alpha g z}{R_{ic}\Delta T_{th}}\right]^{1/4}\left[\frac{<\delta S - Q>}{\rho_a c_p c_d}\right]^{1/2} \qquad \textbf{(29)}$$

For example if we use a net solar heat flux of 800 Wm^{-2} and Q=100, then a threshold of 0.1 K in

(29) gives a wind speed threshold of 0.4 ms^{-1} for z = 1 cm and 3.1 ms^{-1} for z = 1 m. We can also compute the increase in the surface temperature in the upper 2-m by combining (25) and (26).

A simple preliminary investigation has been done with the Moana Wave COARE data to look at the warm layer effects. We picked out about 40 days that had a complete diurnal cycle, no big rainstorms, and reasonably constant wind speeds (Table 2). There were additional days of strong winds with 0 warm layer that we did not bother to include. We took the surface temperature just before sunup as the mixed layer temperature and then found the peak T_s in the afternoon; the difference (ΔT_m) is the solar heating peak. We then computed average solar heat flux (S), average Q, and the mean wind speed. We then estimated mean u_*^2 as .0013*u^2, and computed the Price Weller Pinkel model estimate of the peak warming:

$$\Delta T_c = \left[\frac{2\alpha g t}{R_{ic}\rho_w}\right]^{1/2} \frac{1}{\rho_a c_{pw}^{3/2}} \frac{[f*S-Q]^{3/2}}{u_{*a}^2} \qquad \textbf{(30)}$$

where t=7 hr to cover the period from just after sunup to just past noon. The factor f represents the fraction of solar radiation absorbed above the PRP depth . This is computed using (15), but we chose a form of the absorption profile from Soloviev (1982) that is deemed appropriate for the moderately clean waters of the COARE region [N=3; F_i=(.45,.27,.28); γ_i=(12.8,0.357,0.014) m]:

$$f = 1-\left[\sum_{i=1}^{N} F_i*\gamma_i*(1-\exp(-D_T/\gamma_i))\right]/D_T \qquad \textbf{(31)}$$

Note that (26) and (31) must be solved iteratively before we can calculate (30). The comparison between this equation and the data from COARE is shown in Fig. 4. It is fairly good except for the obvious tendency for the model to underpredict the measured values. In Fig. 5 we show a comparison of ΔT_m normalized by $(f*S-Q)^{3/2}$ versus $1/u_{*a}^2$, which from (30) we expect to be linear. Note that the slope through this data is about 1.1 x 10^{-6} implied by (30) with R_{ic} = 0.65 as per PWP. For light winds, when the warm layer is most important, this result is fairly insensitive to the choice of R_{ic} Using this fit to the data we have constructed typical total warm layer afternoon peaks as a function of wind speed for essentially cloudfree skies (see Table 3). At a wind speed of 5 ms^{-1} the correction for a sensor at 1-m depth is less than 0.1 k (i.e., 1/7*0.4) but at 1 ms^{-1} the correction is the full 3.7 K (because $z>D_T$).

On the assumption that the Moana Wave water temperature sensor may not be measuring the true surface temperature during very light winds, we computed the correction for a sensor at depth z using (28); we used a value for z of 5 cm, which is an upper limit for a realistic number. It turns out that this makes little difference. Clearly this result is not perfect, but we are encouraged that there is some merit. When the winds are near 1 m/s, (28) indicates that the sensor is a few tenths of a K too low. However, this conclusion is highly dependent on the values of u_*, which are quite uncertain at these low wind speeds. We have obtained u_* rather crudely, by averaging the wind speed and using a drag coefficient. Probably we should be doing a vector average of u_*^2, which could give significantly smaller values as the wind direction wanders around. Even modestly lower values of u* will change things (including the ΔT comparison) considerably.

The analysis here is just for the peak warming. On the grounds that it would be amusing to see if this simple model reproduces the hourly evolution of the diurnal cycle with any accuracy, we again processed the Moana Wave 50-min data. Starting at midnight (LST) of each measurement day, we initialized D_T at 19 m and set I_S and I_τ to zero. Beginning at the time when $f*R_{ns} - Q$ first exceeds zero, we computed the time integrals as described by (25). Measured values of ΔT were defined as the difference in measured surface temperature at the subsequent times and the surface temperature at this initial time. Note that (26) and (31) were solved iteratively to give self consistent values for D_T and f. The average diurnal cycle of measured and computed ΔT and computed D_T is shown in Fig. 6a for the 51 measurement days where the average daytime value of u_*^2 was less than 0.05 m^2s^{-2} (i.e., wind speed less than about 6.5 ms^{-1}). To illustrate the effects of wind speed, we show the same diurnal averages for $0.02 < u_*^2 < 0.05$ (Fig. 6b) and $u_*^2 < 0.008$ (Fig. 6c). Notice the asymmetry of the signatures for the lighter wind case. The warm layer becomes quite shallow (1 m) almost immediately and deepens steadily through the day; the surface temperature rises very quickly and peaks very close to noon with some residual warming still present at midnight.

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FIGURE CAPTIONS

Fig. 1. Average diurnal behavior from Moana Wave COARE data computed with the cool-skin model - all data (solid line) and wind speed less than 4.0 ms^{-1} (dashed line). Upper panel is the cool skin temperature; lower panel is the net interfacial heat flux.

Fig. 2. Average cool skin behavior from the Moana Wave COARE data computed with the cool-skin model (11)-(16) versus the mean 10-m wind speed.- day time (dashed line) and night time (solid line). Upper panel is the cool skin temperature; lower panel is the net interfacial heat flux.

Fig. 3. As in Fig. 2 except the X's are the Saunders model with λ=6.5 as described by Coppin et al. (1991), and the O's are the Schluessel et al. (1990)

Fig. 4. Measured peak afternoon surface temperature warming from the Moana Wave COARE data versus computations with (30).

Fig. 5. Measured peak afternoon surface temperature warming normalized by the 3/2 power of δS-Q versus u_*^{-2}.

Fig. 6. Average diurnal behavior from Moana Wave COARE data computed with the PWP warm-layer model. The upper panel is the warm layer surface temperature: measured (solid line) and model computed (circles). The lower panel is the computed warm layer depth.

a) All days with daytime mean $u_*^2 < 0.05$ m^2s^{-2} (n=51).
b) All days with daytime mean $0.02 < u_*^2 < 0.05$ m^2s^{-2} (n=20).
c) All days with daytime mean $u_*^2 < 0.008$ m^2s^{-2} (n=10).

Table 1. Exponential coefficient for 9-wavelength fit to absorption of solar radiation in sea water (Paulson and Simpson, 1981)

i	wavelength, μm	F_i	γ_i ,m	F_i / γ_i
1	0.2-0.6	0.237	34.8	6.8×10^{-3}
2	0.6-0.9	0.360	2.27	1.6×10^{-1}
3	0.9-1.2	0.179	3.15×10^{-2}	5.7
4	1.2-1.5	0.087	5.48×10^{-3}	1.6×10^{1}
5	1.5-1.8	0.080	8.32×10^{-4}	9.7×10^{1}
6	1.8-2.1	0.0246	1.26×10^{-4}	1.9×10^{2}
7	2.1-2.4	0.025	3.13×10^{-4}	8.0×10^{1}
8	2.4-2.7	0.007	7.82×10^{-5}	9.0×10^{1}
9	2.7-3.0	0.0004	1.44×10^{-5}	2.8×10^{1}

Table 2. Some selected days from the Moana Wave data during COARE showing the peak solar warming of the surface temperature and the background meteorological information averaged over seven hours after sunup.

Q	u	u_*^2	ΔT_m	S	f	$\delta S-Q$	ΔT_c	ΔT_w	D_T
Wm^{-2}	ms^{-1}	$(ms^{-1})^2$	K	Wm^{-2}		Wm^{-2}	K	K	m
280	6.0	.05	.2	610	.76	180	0	0	17
120	1.3	.003	1.9	690	.52	240	1.4	.1	.7
160	4.0	.02	.3	625	.72	290	.3	0	5
160	4.0	.02	.4	780	.72	400	.5	0	5
150	3.0	.011	.8	640	.65	270	.5	0	2.8
140	2.5	.008	1.2	700	.61	290	.7	0	2.1
115	2.0	.005	.9	560	.58	210	.7	0	1.5
165	3.2	.013	.5	650	.68	280	.4	0	3.5
145	3.4	.015	.9	670	.70	320	.5	0	3.8
135	3.5	.016	.7	700	.70	355	.5	0	3.8
155	5.0	.032	0	415	.76	160	.07	0	11
205	5.8	.043	.2	450	.76	140	.04	0	17
260	6.0	.057	0	230	.76	-10	0	0	100
180	4.6	.021	.5	680	.73	320	.3	0	5
200	5.3	.036	0	350	.76	60	.02	0	21
130	1.6	.004	2.0	730	.54	265	1.2	.1	1.1
135	1.8	.004	1.5	590	.56	195	.8	.0	1.3
150	2.9	.01	.6	660	.65	280	.5	.0	2.7
185	3.6	.017	.6	690	.71	305	.4	0	4.3
115	1.2	.002	2.2	710	.49	230	2.0	.1	.65
140	2.2	.006	.7	680	.58	255	.8	.0	1.5
175	5.6	.042	.1	360	.76	10	0	0	100
190	5.0	.032	.2	480	.76	175	.1	0	11
165	4.1	.022	.2	430	.75	160	.1	0	8
260	9.4	.11	0	560	.76	170	.05	0	38
200	9.7	.12	0	640	.76	290	.04	0	32
190	5.6	.042	.2	680	.76	330	.15	0	11
160	3.5	.016	0	245	.76	25	0	0	15
100	1.9	.005	.3	350	.61	115	.3	0	2.0
130	2.9	.011	.2	310	.71	90	.1	0	4.0
110	1.3	.002	2.5	610	.50	195	1.5	.1	.65
120	3.0	.012	.9	610	.67	290	.5	0	3.2
130	3.6	.018	.6	615	.72	315	.4	0	4.5
130	2.2	.007	1.1	730	.60	305	.85	0	1.7
160	5.0	.032	0	275	.76	50	0	0	21
175	4.0	.02	.6	560	.74	240	.2	0	6
200	5.0	.032	.5	730	.75	350	.25	0	8
130	3.4	.015	.7	700	.69	350	.5	0	3.6
145	3.2	.013	.5	590	.69	260	.35	0	3.6
175	4.2	.022	.5	750	.73	375	.35	0	5
105	3.0	.012	.3	350	.71	145	.15	0	4.5
130	2.9	.011	.4	500	.68	210	.3	0	3.3

Table 3. Warm layer afternoon peak as a function of wind speed for cloud-free COARE conditions.

u, ms^{-1}	S, Wm^{-2}	Q, Wm^{-2}	f	f*S-Q	u_*^2, m^2s^{-2}	D_T , m	ΔT, K
1	710	100	0.48	240	0.0015	0.6	3.7
2	710	110	0.59	310	0.0052	1.5	1.6
3	710	125	0.67	350	0.011	2.4	0.9
4	710	165	0.72	345	0.016	4	0.6
5	710	190	0.75	345	0.025	7	0.4
6	710	230	0.76	310	0.034	9	0.2
7	710	260	0.76	280	0.049	13	0.15