The sensible heat of rainfall in the tropical ocean

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Abstract. Precipitation affects the buoyancy of the upper ocean by altering the salinity and temperature distribution. If precipitation exceeds evaporation, salinity decreases and the buoyancy of the upper ocean increases. If the temperature of the rain drops and the sea surface temperature are different, precipitation induces a sensible heat flux into the ocean. Generally, it is assumed that the rain drop temperature is close to the sea level wet-bulb temperature. However, because of the finite thermal relaxation time of large rain drops, the temperature of the rain drops could be colder than the sea-level wet-bulb temperature and the sensible heat flux enhanced. We test this possibility by developing a microphysical model to compute the temperature correction from first principles. For tropical warm pool conditions the temperature correction (i.e., difference between rain drop temperature and the wet-bulb temperature) has a maximum of only 0.2 K for normal conditions. The theoretical calculation thus corroborates the general assumption. The model is used to calculate sensible heat flux contributions with rain rate and surface meteorological data to estimate the rain cooling obtained during the Tropical Ocean-Global Atmosphere Coupled Ocean-Atmosphere Response Experiment (TOGA-COARE). The average heat flux from rainfall over the entire TOGA-COARE period for the rain was approximately 2.5 W m^{-2}. However, during intense rainfall events the rainfall sensible heat flux may be as high as 200 W m^{-2}. During these periods the magnitude of the rainfall sensible heat flux rain cooling matches that of the latent heat flux. Because of the uncertainties in measuring meteorological parameters and because the heat flux calculated from the wet-bulb temperature differed by only about 0.1 W m^{-2}, the wet-bulb temperature is still a reasonable value to use for the temperature of falling rain in sensible heat flux computations.

1. Introduction

The principal scientific objectives of TOGA Coupled Ocean-Atmosphere Response Experiment (TOGA-COARE) were designed to provide an understanding of the role of the warm pool regions of the tropics in the mean and transient state of the tropical ocean-atmosphere system [*Webster and Lukas*, 1992]. Specifically, the goals were to describe and understand (1) The principal processes responsible for the coupling of the ocean and the atmosphere in the western Pacific warm pool system; (2) The principal atmospheric processes that organize convection in the warm pool region; (3) The oceanic response to combined buoyancy and wind stress forcing in the western Pacific warm pool region; and (4) The multiple scale interactions that extend the oceanic and atmospheric influence of the western Pacific warm pool system to other regions and vice versa.

To be able to approach the scientific goals of TOGA-COARE, an elaborate overlapping observational network was set up in the western Pacific Ocean. The network allowed the evolving structures of the atmosphere and the ocean to be viewed simultaneously and in detail for the first time in the tropical warm pools.

The thermally driven part of the interaction of the upper ocean and the overlying atmosphere may be viewed in a convenient manner by considering the contribution of the surface heat and freshwater fluxes to the upper ocean buoyancy flux [*Gill*, 1982, *Kraus and Businger*, 1994];

$$F_B = C_w^{-1} \alpha g Q_T + \beta g S(E - P), \qquad (1)$$

where Q_T is the total upward heat flux from the ocean, E-P is the difference between the precipitation and evaporation rates (the freshwater flux), S is salinity, g is the gravitational acceleration, C_w is the specific heat of seawater, and α and β are the thermal and salinity expansion coefficients, respectively. F_B is written in such a manner so as to delineate thermal and haline effects. That is, (1) appears as an aggregate of the heat flux and the freshwater flux into the ocean. The total heating may be written as

$$Q_T = Q_{lh} + Q_s - Q_r, \qquad (2)$$

where the elements on the right-hand side signify the latent heating, the sensible heating, and the net surface radiation. There

are two components to the sensible heating, the turbulent transfer of heat from the ocean surface to the atmosphere (i.e., Q_{ts}) and the cooling of the ocean by the mixing of cooler precipitation into the upper ocean (Q_p):

$$Q_s = Q_{ts} + Q_p. \tag{3}$$

From (1)-(3) it can be seen that precipitation enters into the buoyancy flux as a component of the total heating and the freshwater flux. Because $(\alpha/\beta S) \sim 10^{-2}$ rainfall will virtually always increase the buoyancy of the upper ocean.

Although the contribution of precipitation to the cooling and turbulent mixing of the upper ocean has been of some interest [Katsaros, 1976; Katsaros and Buettner, 1969] it is usually ignored [e.g., Oberhuber, 1988]. The major components of the total heat flux are assumed to be the latent heat flux (80 to 125 W m^{-2}) and the net radiation (100 to 200 W m^{-2}) made up of net solar radiation (150 to 250 W m^{-2}) and the net surface long-wave heating (–40 to –60 W m^{-2}). The turbulent sensible heat flux is about an order of magnitude smaller than the latent heat flux. However, in the tropical warm pool regions, annual precipitation is generally between 3 and 5 m corresponding to an average precipitation rate of 0.34 to 0.50 mm hr^{-1}. It is important to determine the degree of cooling associated with Q_p.

The magnitude of the cooling of the upper ocean or, alternatively, the reduction of the ocean buoyancy will depend on the temperature of the rain when it enters the ocean and on the rain rate. It is not difficult to establish the maximum temperature of the rain. This must be the wet-bulb temperature of the ambient air which is the temperature a drop will equilibrate in a sufficiently long period in the atmosphere. Over the ocean the wet-bulb temperature rarely exceeds the water surface temperature. Thus, except in unusual circumstances, rainfall will always cool the ocean. The few oceanographic observations [e.g., Ostapoff and Worthem, 1974; Bahr and Paulson, 1991; Flament and Sawyer, 1995] support this claim.

Two basic groups of questions that concern us are the following. (1) What is the temperature of raindrops entering the ocean? Is it significantly below the near-surface wet-bulb temperature? We ask this question because the ensemble of droplets are usually falling through a vertical temperature gradient and their temperature may not lag the local wet-bulb temperature by an insignificant amount. This assumption has been questioned by oceanographic observations [e.g., Bahr and Paulson, 1991]. (2) What is the overall impact of the rain-cooling of the tropical ocean? The long-term impact is important, of course, but rainfall occurs episodically and, occasionally, violently in the tropics. The question may be extended to consider the temporal impacts of episodic events.

To answer the first question, we consider the temperature change of a droplet ensemble falling through an atmosphere from cloud base to the ocean surface. The problem is approached from first principles and standard analytic expressions applied to calculate the evolving temperature of the droplet ensemble. The temperature is calculated for a wide range of cloud heights and the temperature of the rain droplet ensemble compared with the surface wet-bulb temperature. Finally, using rain rates observed during TOGA- COARE, we calculate the sensible heat flux due to precipitation.

2. Simple Bulk Model

Because rain rate is, in fact, a mass flux, it can be simply related to an equivalent sensible heat flux by multiplying by the heat capacity and the rain-ocean temperature difference.

The rain rate R is defined by the expression

$$R = \frac{4\pi}{3}\rho_w \int a^3 v_f(a) n(a) da, \tag{4}$$

where ρ_w is the density of water, a is the droplet radius, $v_f(a)$ is the droplet fall velocity and $n(a)$ is the droplet size spectrum number density (i.e., number of drops per radius increment per volume). The effective sensible heat flux for this rain rate is

$$Q_p = c_w R(T_o - T_r), \tag{5}$$

where c_{wo} is the specific heat of water (4186 J kg^{-1}K^{-1}), T_o is the bulk sea surface temperature, and T_r is the mass-mean temperature of the rain as it reaches the ocean surface. A positive value of Q_p signifies that when the rain is cooler than the ocean, a loss of heat from the ocean to the atmosphere occurs.

For the simple bulk model we will assume that the bulk droplet temperature T_r is equal to the wet-bulb temperature T_w as determined from atmospheric measurements at a typical reference height of 10 m above the surface. We also assume that the droplet cannot respond to the thermodynamic gradients in the surface layer very close to the ocean. Between cloud base and the near-surface region, equilibrium occurs for small droplets, but the larger thermal equilibrium time of the large droplets causes them to top the local equilibrium values. Droplets with radii of the order of a few millimeters have fall velocities of order 10 m s^{-1} [Gunn and Kinzer, 1949] and their thermal equilibrium time is on the order of 1 s [Pruppacher and Klett, 1978]. Given a typical tropical boundary layer lapse rate of 0.01 K m^{-1}, the droplets are roughly 0.1 K cooler than T_w computed from near surface meteorological measurements. This rough estimate will be examined more closely in section 3. For the TOGA-COARE region T_w is typically 3 K cooler than the ambient air temperature T_{air}. This difference can be expressed more quantitatively with the standard psychrometer equation [e.g., Hess, 1959]:

$$T_w = T_{air} - \frac{L d_v}{d_h c_p}(q_s(T_w) - q), \tag{6}$$

where L is the latent heat of vaporization of water (2.5x10^6 J kg^{-1}), c_p is the specific heat of air (1004 J kg^{-1} K^{-1}), d_u and d_h are the diffusivities for water vapor and heat, respectively; q is the ambient specific humidity, and q_s is the saturation humidity at the relevant temperature. Because raindrops are large in size and are composed of fairly pure water, we ignore convective and salinity effects

Subtracting the T_o from both sides of (3) gives

$$T_o - T_w = T_o - T_{air} + \frac{L d_v}{d_h c_p}(q_s(T_w) - q). \tag{7}$$

Here $q_s(T_w)$ can be expanded to obtain an expression in terms of $q_s(T_o)$

$$q_s(T_w) = q_s(T_o) - \frac{dq_s}{dT}(T_o - T_w). \tag{8}$$

The term dq_s/dT is known from the Clausius-Clapeyron relation [e.g., Wallace and Hobbs, 1977]. With some manipulation we can write

$$T_o - T_w = \epsilon \left(\Delta T + \frac{L}{c_p}\Delta q \right), \tag{9}$$

where

$$\Delta T = T_o - T_{air} \text{ and } \Delta q = q_s(T_o) - q \tag{10}$$

are the normal bulk air-sea differences used to estimate the sensible and latent heat fluxes. Here ϵ is termed the wet-bulb factor and is given by

$$\epsilon = 1 / \left(1 + \frac{Ld_v}{d_h c_p}\frac{dq_s}{dT}\right).$$ (11)

Note that ϵ decreases with increasing temperature and has a value of approximately 0.2 for the TOGA-COARE region.

Using (2) and (6), the sensible heat flux due to rain can be written as

$$Q_p = -Rc_{wo}\Delta T\epsilon\left(1 + \frac{1}{B}\right),$$ (12)

where

$$B = \frac{c_p\Delta T}{L\Delta q}$$ (13)

is the bulk Bowen ratio [see Garratt, 1992 p130]. In the tropical oceans, B is roughly 0.1, so $\epsilon(1 + 1/B)$ can be replaced with the numerical factor 2.5 to give

$$Q_p = 1.05 \times 10^4 R\Delta T.$$ (14)

The average rainfall in the western Pacific warm pool region is about 1.2×10^{-4} kg m^{-2} s^{-1} (i.e., 0.42 mm hr^{-1} [e.g., Oberhuber, 1988, Webster and Lukas, 1992], and the average air-sea temperature difference is about 1.5 K. If the additional 0.1 K correction due to the thermal response lag is added to the air-sea temperature difference, the average sensible heat contribution of the rain is about 2.0 W m^{-2}. The largest one hour average rain rate observed by the R/V Moana Wave on the COARE Pilot cruise was 1.5×10^{-2} kg m^{-2} s^{-1} (i.e., 55 cm d^{-1} [Young, et al., 1992]), which is equivalent to about 250 W m^{-2}. This value is not negligible; it is roughly comparable to the reduction in solar radiation or the enhancement of the latent heat flux that occurs with these storms during the daytime.

The above discussion shows that the cooling due to rain may not be negligible as has been previously thought and may have very large intermittent values. To explore the effects of rain on the heat budget, a model was constructed to use observed meteorological data to calculate the thermal equilibrium response correction in the observed rain events. These corrections, along with the rain rate, are used to compute the actual rain cooling of the ocean surface from the R/V Moana Wave during three cruise legs of TOGA-COARE.

3. Numerical Model

We now describe a numerical model used to evaluate deviations of raindrop temperature from the near-surface wet-bulb temperature. The model is based both on an assumed initial lognormal rain droplet distribution and on microphysical rate equations for the evaporation and cooling of the droplets as they fall from cloud base to the surface. The model allows the size distribution of the rainfall ensemble to change as the drops fall from a given cloud base height through a subcloud layer with a dry adiabatic lapse rate to the ocean surface. The temperature of the falling rain and the amount of cooling at the ocean surface are calculated for each observation. This rain heat flux along with the sensible heat, latent heat and total energy fluxes is summed and averaged over the entire observation period.

The droplet model assumes an initial continuous lognormal spectral density function as a function of droplet diameter D given by

$$n(D) = \frac{N_t}{\sqrt{2\pi}D\ln\sigma}\ \exp\left[\frac{-\ln^2(D/D_g)}{2\ln^2\sigma}\right]$$ (15)

from Feingold and Levin, [1986]. A lognormal distribution is used because it generally provides a better fit to observed drop spectrum than some other distributions, plus it is analytically convenient because all of its higher moments are also lognormally distributed.

The parameters in (15) from Feingold and Levin [1986] are computed using

$$N_t = 172R^{0.22},$$ (16)
$$D_g = 0.75R^{0.21},$$ (17)
$$\sigma = 1.4300 - 3.1 \times 10^{-4}R,$$ (18)

where the rain rate R (millimeters per hour) is taken from the data set. Equations (16)-(18) give the number density, N_t (droplets per cubic meter), the geometric mean droplet diameter D_g (meters), and the standard geometric deviation σ, respectively.

Initial droplet size bins with diameters of 0.2 to 12.8 mm are defined. They are arranged such as to have a droplet's mass in the $i+1$ bin be twice that of a droplet's mass in the i th bin. This format yields 19 bins. Finally, the width of the i th bin times the probability density in the center of the i th bin gives the discrete number of drops per unit volume in each bin for a given rain rate ($n_i = n(D_i)\Delta D_i$). The sum of n_i yields N_t.

Once the initial cloud base droplet spectrum and atmospheric variables are set, the drops are released from the cloud base and changes in drop size and temperature with time are computed at 1-meter intervals. The number of drops in each bin is kept fixed but their radius (and bin width) evolve as the droplet evaporates. Pruppacher and Klett [1978] showed that the time rate of change of droplet size may be written as

$$\frac{da}{dt} = \frac{d_v f_v}{a\rho_w r_v}\left(\frac{e_\infty}{T_\infty} - \frac{e_a}{T_a}\right),$$ (19)

and the rate of change of the temperature difference between the environment and the droplet as

$$\frac{d(T_\infty - T_a)}{dt} = -X(T_\infty - T_a) + Y,$$ (20)

where a is the droplet radius in meters; ρ_w is the density of water (1000 kg m^{-3}), r_v is the gas constant for water vapor (461.5 J kg^{-1} K^{-1}), e_∞ is the vapor pressure (pascals) at the ambient air temperature T_∞, and e_a is the vapor pressure for the droplet surface temperature T_a.

The various factors in (16) and (17) are computed as follows [Pruppacher and Klett, 1978]:

$$d_v = 2.11 \times 10^{-5}\left(\frac{T_\infty}{271.15}\right)^{1.94}\left(\frac{101325}{P}\right),$$ (21)

$$f_v = 0.78 + 0.308R_s^{1/3}R_c^{1/2},$$ (22)

$$X = \frac{3}{a^2 c_{wo}\rho_w}\left[k_a f_h + Ld_v f_v\left(\frac{E_\infty - E_a}{T_\infty - T_a}\right)\right],$$ (23)

$$Y = \frac{3d_v L f_v}{\rho_w c_{wo}a^2}(1 - s)E_\infty,$$ (24)

where $E_\infty = e_{\infty,\text{sat}}/r_v T_\infty$ and $E_a = e_{a,\text{sat}}/r_v T_a$. P is the air pressure at the given level and f_v is the ventilation coefficient for water vapor diffusion and ρ_w is the density of liquid water.

R_s and R_e are the Schmidt and Reynolds numbers ($\mu/\rho d_v$ and $\rho D v_f/\mu$, respectively), μ is the dynamic viscosity of air, $e_{\infty,sat}$ and $e_{a,sat}$ are the saturated vapor pressure at T_∞ and T_a, respectively; and r is the relative humidity. The ventilation coefficient for heat diffusion is given by

$$f_h = \frac{d_v c_p \rho f_v}{k_a},\tag{25}$$

and the thermal conductivity for dry air by:

$$k_a = A + C(T_\infty - 273.15),\tag{26}$$

where A= 2.38x10^{-2} and C=7.12x10^{-5} yields k_a in W m^{-1} K^{-1}.

The model equations are calculated for atmospheric profile variables at 1–m intervals from the cloud base to the ocean surface. These atmospheric profiles vary through the integration. As stated above, the atmosphere is assumed to have an adiabatic lapse rate Γ. The vertical temperature profile is set by the near-surface ambient air temperature which has the form of a time series observed from the R/V *Moana Wave* [*Young, et al.*, 1992]. The air pressure and density are assumed to decrease exponentially with height, while the relative humidity increases linearly from the observed surface value to 100% at the cloud base. With the assumed lapse rate and the continually updated surface temperature, E_∞ and T_∞ evolve with time. In this manner the impact of the changes to the environment by the evaporating rain drops are taken into account. Clearly, this is an approximation as there may be other environmental changes that this simple procedure does not consider. However, most of these effects may be expected to be minimal due to the rapid falling rate of the droplets. Droplets falling at about 10 m s^{-1} will take less than 1–min to reach the ocean surface from the cloud base. It should be pointed out, though, that there is an assumption that the air below the cloud base is quiescent. It is possible that vertical velocities due to cloud dynamics will change the rate of fall. However, to consider these effects will require cloud resolving models which are beyond the level of sophistication considered here.

The fall speed of each bin of droplets is calculated using relationships derived by *Lin et al.* [1983] and *Gossard et al.* [1990];

$$v_f = 942 D^{0.8}\left(\frac{\rho_o}{\rho}\right)^{0.5}; \quad D < 8 \times 10^{-4}\text{m},\tag{27}$$

$$v_f = \left(9.65 - 10.3\exp^{-600D}\right)\left(\frac{\rho_o}{\rho}\right)^{0.4};$$
$$D \geq 8 \times 10^{-4}\text{m},\tag{28}$$

which give the best fit to the data of Gunn and Kinzer (1949) for small and large droplets, respectively. Here ρ_o is the air density at sea level and ρ is the air density at any level. D is given in meters and yields v_f in meters per second.

The incremental change of radius and temperature (i.e., 19 and 20) in a given layer is found by multiplying the rate of change by the time dt it takes a droplet to fall through a 1–m layer (i.e., $dt = 1/v_f$). From (20) the change of temperature through the layer can be calculated. If the residence time in a layer is very long, the drop is assumed to be in thermal equilibrium and the temperature of the drop is set equal to the wet-bulb temperature. Computationally, this is assumed to occur if $dt > 90\%$ of τ, where $\tau = 1/X$. From (20), $T_w = T_\infty - Y/X$. The error incurred by

this approximation is order 10^{-5} K or less. After the new radius and temperature of the droplets have been found in a given layer, the droplets are assumed to move down to the next layer and the whole process is repeated until the ensemble reaches the ocean surface. When the radius of a particular bin reaches zero, those droplets are considered to be completely evaporated and removed from further consideration.

Once the entire spectrum of droplets has reached the ocean surface, the final mass of the spectrum is compared with the starting mass and the evaporation mass to check mass continuity. The surface precipitation flux is calculated, along with the mass-weighted rain and wet-bulb temperatures. The heat flux is calculated using (2). As the drops fall, the total number of drops in each bin is conserved. Continuity requires that

$$(n_i v_f)_{z=z_c} = (n_i v_f)_{z=0}.\tag{29}$$

The left and right sides of the equation are the products of the number per unit volume and fall velocity calculated at the cloud base ($z=z_c$) and surface ($z=0$), respectively. As the drops fall, n_i will increase because v_f decreases due to the partial evaporation of the drops. Therefore the left-hand side of (29) can be used in computing, via (4), the precipitation flux at the surface. Of course, the mass flux at the surface will be different from that at cloud base because the droplet radius and bin increment will be different. Droplets that have evaporated will have an infinite number concentration but zero mass flux because $Q=da=0$.

4. Numerical Model Results

The model is used to assess the impact of rain rate on the temperature of the falling rain. Calculations of the correction ($T_w - T_a$) due to the effects of the thermal relaxation time on the temperature of the rain were made for rain rates varying from 1 to 120 mm hr^{-1} for cloud base heights of 250, 500, 1000, 2000 and 5000 m. The relative humidity was set to 80% at the surface and increased linearly up to 100% at the cloud base except in the 250–m case, where to keep the model stable, an 85% surface relative humidity was used. The ambient surface air temperature was set to 299 K. Figure 1 shows the results for the five cases.

The model is discrete in both radius space and the vertical coordinate. Our unsophisticated numerical approach has some problems where $(da/dt)dt$ is not small compared to a. This occurs for droplets of smaller radius (where da/dt is large) and when the thermodynamic vertical gradients are strong. The jagged appearance of the curves in Figure 1 is due to our crude method of overcoming this problem (i.e., labelling the droplets as evaporated). A better approach would be to increase the vertical resolution of the model. However, given the clear indication that the correction is small, we decided more sophisticated treatments would not be justified. The use of 85% relative humidity for the 250–m cloud base height is quite reasonable anyway because the convective boundary layer rarely produces low cloud bases simultaneously with dry conditions.

Both rain rate and cloud base height have a clear effect on the temperature of rain. As rain rate and cloud height increase, the correction (i.e., the difference between T_w and T_a) increases. However, even in the extreme case of high cloud bases and large rain rate, the temperature of the rain is no more than 0.2 K less than T_w. Thus the assumption that the temperature of the raindrops reaching the surface is given by the wet-bulb temperature appears to be very sound in the tropical atmosphere.

Next, the model was used to find the flux of sensible heat from actual rain events that occurred during TOGA-COARE. The
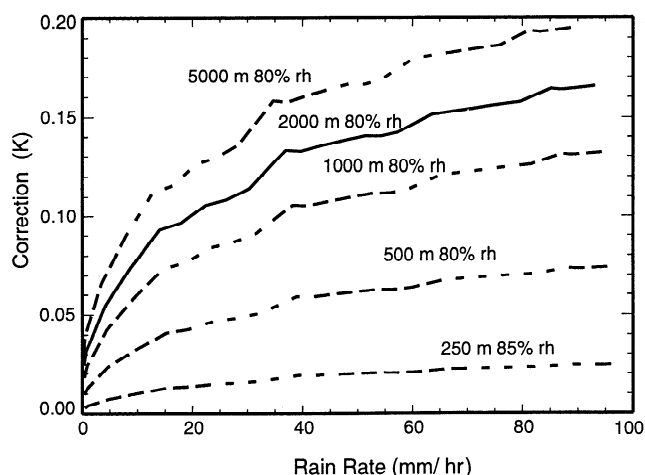
Figure 1. Difference (kelvins) between the wet-bulb temperature T_w and the raindrop temperature T_a as a function of rain rate (millimeters per hour) for various cloud base heights (meters) and surface relative humidities (percent).

sensible and latent heat fluxes were computed using the TOGA-COARE bulk flux algorithm (C. W. Fairall et al., Bulk parameterization of air-sea fluxes for TOGA-COARE, submitted to *Journal of Geophysical Research*, 1995). While the rain rates were available, the measurements of cloud base height are intermittent. To overcome this deficiency, the model was run twice with the cloud base height set to 500 and 4000 m to obtain lower and upper limits of the range of rain cooling experienced in TOGA-COARE. Data are from hourly meteorological data collected on the R/V *Moana Wave*. The data start at 0355 UT on November 11, 1992,
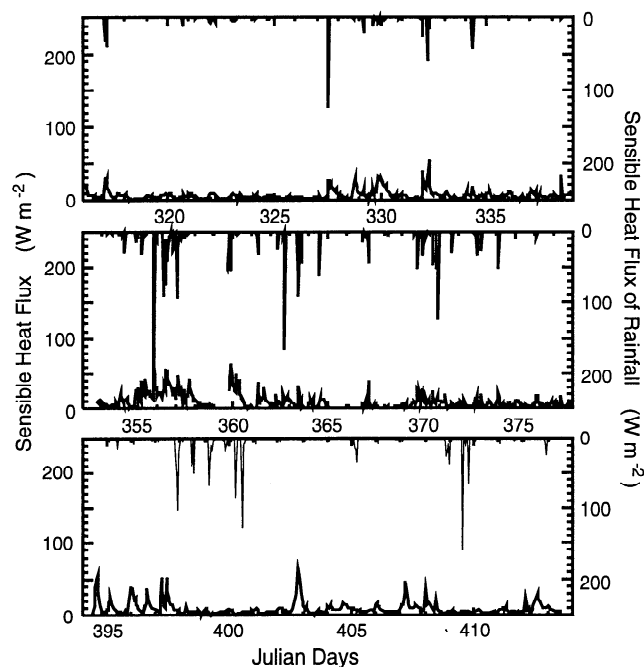
and goes to 1251 UT on February 16, 1993. A typical cloud base height in TOGA-COARE was about 650 m which is consistent with the local lifted condensation level measured from the surface. From the data that do exist a time series of the results for the 500-m run are shown in Figure 2 where the sensible heat flux is plotted, along with the rain heat flux. Figure 3 is a time series of observed latent heat flux for the same period. Over the entire TOGA-COARE period the average rain sensible heat flux for the 500-m cloud base height was 2.5 W m^{-2} and for the 4000-m case, 2.1 W m^{-2}. Thus the average value is about one fifth of the mean sensible heat flux value. The flux of rain cooling calculated using the wet-bulb temperature without the correction for thermal equilibrium time was 2.4 W m^{-2}. The flux obtained by using the average sea-rain temperature difference (1.55 K) and the average rain rate (1.34 10^{-4} kg m^{-2} s^{-1}) for the entire data set in (12) was 2.20 W m^{-2}.

While the average over all events (both rainy and dry) is small, the value of rain sensible heat for individual rain events may be very large. For the 241 hours of precipitation observed on the R/V *Moana Wave* during TOGA-COARE, the rain cooling averaged 17 W m^{-2} compared to a rain event sensible heat flux of 22 W m^{-2}. However, Figure 2 shows several events with cooling over 100 W m^{-2}. The maximum occurred on Julian day 356 with a value of 213 W m^{-2}. The rain rate for this hour was 10^{-2} kg s^{-1} (36 mm hr^{-1}). In other words, on certain days the sensible heat of rain can be substantial, indeed. Note that the peak rain heat flux computed from the time series is comparable to the value estimated in section 2 from the mean air-sea temperature difference despite a significantly smaller rain rate. This is because air temperature and wet-bulb temperatures tend to be depressed during rain events.

5. Summary

We have used theoretical models to calculate the temperature change of ensembles of rain droplets falling through the tropical boundary layer. These estimates were used to calculate the
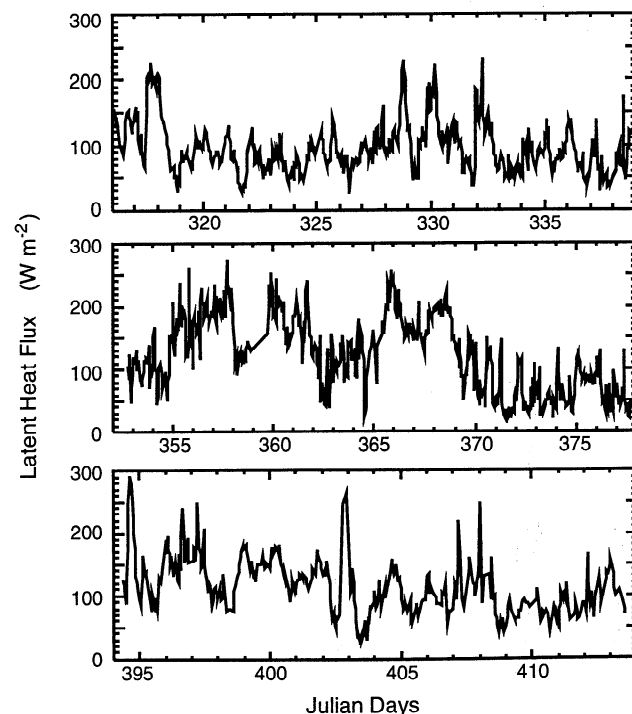

Figure 2. The sensible heat (watts per square meter) due to rain cooling and turbulent heat transfer calculated from observed data taken during the TOGA-COARE experiment. The rain cooling is plotted from the top down with the sign reversed for clarity. Data were taken from the R/V *Moana Wave* during the TOGA-COARE experiment in the western Pacific Ocean at 156°E, 2°S starting on November 11, 1992, and continuing to February 16, 1993. Julian day is referenced relative to January 1, 1992.

Figure 3. Same as Figure 2, except for the latent heat flux (watts per square meter).

sensible heat flux from the ocean due to rainfall. In general, it was found that the approximation that the temperature of the raindrops is given by the wet-bulb temperature is good. This is consistent with the recent findings by *Flament and Sawyer* [1994]. In fact, it was found that the difference between T_w and the temperature of the droplet T_a was minimal to the extent that the long-term average heat flux was within 0.3 W m^{-2}. However, this estimate considered many dry periods in the tropical ocean as well as precipitating periods. During convective periods the difference would be more substantial, and in periods of very heavy precipitation it might be important to use T_a rather than T_w. Because of the correlations of meteorological conditions with rain rate, the use of long-term average variables to compute Q_P leads to significant underestimations. Furthermore, we have ignored the possibly substantial effects of frozen precipitation and rain drops embedded in strong downdrafts. Heavy precipitation in the tropics generally occurs out of low cloud bases, so we assume that our near-surface thermodynamic measurements adequately sample the subcloud layer.

Although it is beyond the scope of this study, it is interesting to speculate on the importance of the rain sensible heat on the state of the ocean during intense convective events. Clearly, values of rain sensible heat of the order of the latent heat flux during heavy precipitation must cause considerable changes in the local buoyancy of the upper ocean. The freshwater flux from rain increases the buoyancy of the ocean. Rain sensible heat, on the other hand, decreases the buoyancy. In intensive precipitation events it is not clear which process would dominate.

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