Annual Mean Surface Heat Fluxes in the Tropical Pacific Ocean

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ABSTRACT

The four components of the long-term annual mean net surface heating of the tropical Pacific Ocean between 30°N and 40°S are calculated and portrayed. These flux elements were derived by using the bulk formulas and about 5 million marine weather reports for the years 1957–76. In addition to illustrating the mean solar, latent heat, infrared radiation and sensible heat fluxes, annual mean values of the atmospheric variables which contribute to those fluxes also are illustrated. A simple error analysis is carried out from which it is concluded that the 95% confidence bands for solar heating, latent heat loss and net oceanic heating are ±29, ±39 and ±49 W m^{-2}, from the respective mean values. The validity of the results for the net heating is partially tested by comparing the horizontal heat transports required by the pattern of heating with independent estimates of those dynamical transports.

1. Introduction

In recent years there has been growing interest in the meteorology and oceanography of tropical ocean regions. This interest has developed in part because of the recognition of the importance of air-sea interaction in determining the weather and climate of the extratropics as well as the tropics. Coincident with this recognition has been a strong suggestion that some aspects of the circulation of the tropical oceans and atmosphere may be predictable one or more months in advance.

One of the most basic observational questions in this region is that of the pattern of the interchange of heat between the sea surface and the lower atmosphere. This interchange is the result of heating by absorption of solar radiation reduced by the losses through longwave emission, the loss of latent heat associated with evaporation, and the loss (or gain) of sensible heat. The pattern of oceanic surface heating has recently been identified for the tropical Atlantic by Bunker and Goldsmith (1979),[1] Bunker (1976), Bunker and Worthington (1976) and Hastenrath and Lamb (1978). A recent analysis of the heat fluxes in the Indian Ocean has been carried out by Hastenrath and Lamb (1979). For the bulk of the tropical Pacific Ocean the most recent such analysis is that of Wyrtki (1965b) based upon data prior to 1962. Because of the considerable increase in data available since that time and slightly improved parameterizations of the basic heating terms, we calculated all of the surface heating components for the tropical Pacific between 30°N and 40°S for each month from January 1957 through December 1976. At the same time we derived means and other statistics of the basic meteorological parameters in this region, some of which will be described in a later paper. These fluxes and meteorological statistics were derived from a total of ~5 × 10^6 marine weather report sets and were defined for grid regions of 5° latitude × 5° longitude. This report concentrates upon displaying the 20-year annual means of the heating terms and the basic meteorological parameters used in the formulas to calculate those means.

2. Basic energy equations

The exchange of energy between the ocean and the atmosphere may be written to a very good approximation as

$$Q_N = Q_S - Q_I - Q_H - Q_L, \tag{1}$$

where N, S, I, H and L refer to net heating, absorption of solar radiation by the ocean, net loss of infrared radiative energy from the ocean surface, loss of sensible heat through turbulent fluxes at the ocean surface, and net loss of latent heat through the evaporation of water from the surface, respectively. In this work, as with its predecessors, the terms on the right-hand side of Eq. (1) are calculated using the observed oceanic conditions and the so-called bulk flux formulas; the net heating is found as a residual.

Bunker (1976), Hastenrath and Lamb (1978) and others have discussed the difficulties in choosing the formulas most applicable over oceanic regions. The

[1] Bunker, A. F., and R. A. Goldsmith, 1979: Archive time-series of Atlantic Ocean meteorological variables and surface fluxes. WHOI-79-3, 28 pp.

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proper choices for the determination of absorbed solar and net emitted infrared radiation under generally cloudy skies are the most difficult. This is true, in part, because of the scarcity of radiation data over the open ocean and in part because these fluxes are affected by the structure of the atmosphere far above the surface for which no data is available. We have chosen to prescribe the solar radiation incident on the ocean under clear skies by the harmonic representation of the *Smithsonian Meteorological Tables* (List, 1958) suggested by Seckel (1970):[2]

$$Q_{S0} = (A_0 + A_1 \cos\phi + B_1 \sin\phi + A_2 \cos 2\phi + B_2 \sin 2\phi). \quad (2)$$

Here ϕ is $2\pi/[365(t - 21)]$, where t is the Julian day and the A's and B's are latitude-dependent coefficients, which are given by Reed (1977) with constants appropriate to calculating Q_{S0} in SI units for tropical latitudes. Seckel (1970)[2] found that this formula gave results which compared well with observations taken between 10 and 25°N near 150°W. Reed (1975) and Reed and Halpern (1975) also found good agreement between the predictions of this formula and data taken at several coastal stations and aboard the research vessel *Oceanographer*.

While there is general agreement about the utility of the Smithsonian formula [Eq. (2)], for clear sky insolation, there is a much greater problem in determining the proper modification to take into account the effect of clouds. We have chosen the formula derived by Reed (1977)

$$Q_C = Q_{S0}(1 - AC + 0.0019\beta), \quad (3)$$

where Q_C is the solar radiation incident on the ocean under cloudy conditions, A is a constant taken as 0.62, C is the fractional cloud cover and β is the noon solar altitude in degrees. Reed (1977) found good agreement between fluxes predicted by this formula and those measured from islands or coastal stations. Simpson and Paulson (1979) compared the results of several suggested cloud adjustment formulas with measurements taken aboard research platform FLIP and found that this formula was in best agreement with observations. However, they concluded that errors of $\sim$10 W m^{-2} may still exist.

The solar radiation absorbed by the ocean is the result of Eq. (3) multiplied times the factor $1 - \alpha$, where α is the ocean surface albedo. We used the values of α computed by Payne (1972) which vary with latitude and calendar month. Simpson and Paulson (1979) found generally excellent agreement of these values with their measurements.

This formulation is similar to that being used by Clark (personal communication) for the North Pacific but differs from that chosen by Bunker (1976) for the Atlantic. The latter chose a cloud dependence given by $Q_C = Q_{S0}(1 - aC - bC^2)$, where C is the fractional cloudiness and a and b are constant coefficients. Simpson and Paulson (1979) and Weare and Remer (unpublished manuscript) suggested that this latter formula tends to underpredict insolation by $\sim$10%.

The choice of formulas to predict net infrared loss from the ocean surface using marine meteorological reports is as difficult as the choice of the appropriate solar formulas. For net longwave radiation emitted from the surface under clear sky conditions (Q_{I0}) we have applied the formula suggested by Berliand (Budyko, 1963), i.e.,

$$Q_{I0} = \epsilon\sigma T_s^4(0.39 - 0.05e^{1/2}) + \epsilon\sigma T_s^3(T_s - T_a), \quad (4)$$

where ϵ is the emissivity of the ocean (0.97), σ the Stefan-Boltzmann constant, T_s and T_a the sea and air temperatures and e the atmospheric vapor pressure. Reed and Halpern (1975) and Simpson and Paulson (1979) found generally good agreement between the results of this formula and their rather limited data.

As in the case of solar radiation, cloudiness modifies the longwave flux considerably. Again, there is some uncertainty as to the proper form for the cloud correction. Following Reed (1976) we use

$$Q_I = Q_{I0}(1 - BC), \quad (5)$$

where C is fractional cloud cover and B is a factor that depends on cloud type ($B = 0.9$ for stratocumulus, $B = 0.6$ for altostratus and $B = 0.25$ for cirrus). Simpson and Paulson (1979) also suggested the use of this formula with $B = 0.8$ for the 35°N, 155°W location in which they made measurements. Generally our choice of longwave formulas are in agreement with those used for the Atlantic (Bunker, 1976) and in the North Pacific (Clarke, personal communication).

There is a greater consensus about the best choices of formulas to predict the sensible and latent heat fluxes. The formulas used are

$$Q_H = \rho_a C_p C_E |\mathbf{V}| (T_s - T_a), \quad (6)$$

$$Q_L = \rho_a L C_E |\mathbf{V}| (q_s - q_a), \quad (7)$$

where ρ_a is the surface atmospheric density, C_p the heat capacity of dry air, L the latent heat of vaporization, $|\mathbf{V}|$ the magnitude of the surface wind, T_s and T_a the sea and air temperatures, q_s and q_a the saturation specific humidity for the sea surface temperature and the measured atmospheric specific humidity, and C_E the turbulent exchange coefficient; ρ_a, C_p and L are assumed constant. C_E is assumed

[2] Seckel, G. R., 1970: The trade wind zone oceanography pilot study Part VIII: Sea-level meteorological properties and heat exchange processes July 1963 to June 1965. Report-Fisheries No. 612, 129 pp. [Available from the authors].

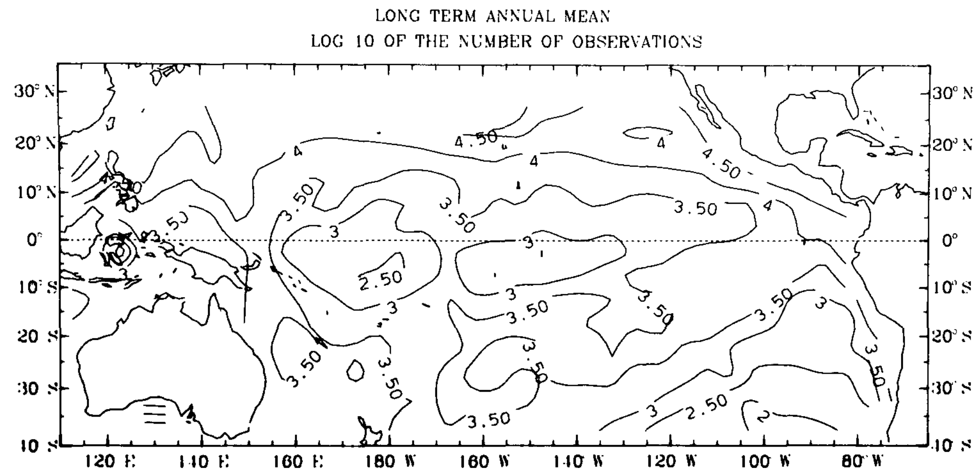

FIG. 1. Logarithm (base ten) of the total number of marine reports per 5° × 5° grid for the period 1957–76.

to depend upon the bulk Richardson number (Deardorff, 1968). This dependence was tabulated by Bunker (1976), who adjusted for apparent biases in ship-derived temperatures and humidities.

3. Data processing

The data used in this study are individual marine weather reports for the tropical Pacific Ocean between 30°N and 40°S and between the South American coast and 110°E for the period 1957–76. Observations for the period 1957–74 were derived from the archives of ships logs maintained at the National Climatic Center and obtained from the Fleet Numerical Weather Central. These were supplemented by archived radioed weather reports for the period 1972–76 obtained from the Southwest Fisheries Center. These data were reformatted, checked for gross errors and merged. After duplicates (adjudged by having identical date, place, air and sea temperatures and winds) were eliminated, long-term monthly means and standard derivations of eleven variables were calculated for each 5° × 5° grid in the tropical Pacific. The eleven variables were longitude, latitude, air temperature, sea surface temperature, specific humidity, pressure, fractional total and low cloud amount, zonal and meridional wind velocities and the magnitude of the wind speed.

Several possible schemes were tested for deriving the desired 5° × 5° monthly mean values of the heat fluxes and other parameters (see Weare and Strub, 1981). We chose to calculate the mean fluxes by calculating flux estimates for every observation set (if necessary data were available) and simply averaging all such estimates for each individual month in each 5° grid. Before these final flux estimates and means were derived, a final error check was carried out. We rejected all data in a report if the air temperature, sea temperature or humidity differed from the previously calculated long-term mean for that calendar month by more than three standard deviations (as calculated previously). In addition, any observations with wind speeds > 35 m s^{-1} were eliminated. These procedures resulted in the loss of about 2% of the observations. Spot checks of the observation sets, which were eliminated, indicated that many of the errors were due to much too low air and sea temperatures suggestive of the fact that many were high latitude reports with improper location codes.

The number of observation sets used to calculate the mean fluxes and other parameters for each 5° grid is indicated in Fig. 1. The data density is poorest just south of the equator to the west of the date line and in the eastern Pacific between about 80 and 140°W and south of 20°S. Similar density maps plotted for each year between 1957 and 1976 (not shown) show that the number of observations increases, especially south of the equator, through 1967–68 and then declines slightly through 1972–73. After 1973, when almost total reliance was placed on the radioed reports, the number of observations was considerably less than for the 1967–68 period nearly everywhere.

4. Results

In order to put into perspective the surface energy fluxes, we present in Fig. 2 a plot of the 20-year mean annual average sea surface temperature for the tropical Pacific Ocean. This mean, as with all the means which will be presented, is based on averages for grids which are 5° latitude × 5° longitude. Furthermore, as with most of the subsequent figures, the isolines in this figure are a simple contouring of the 5° grid means and are not an objective analysis but only the result of a very simple smoothing for plotting purposes. These sea temperatures differ

LONG TERM ANNUAL MEAN
T_s SEA TEMPERATURE (°C)

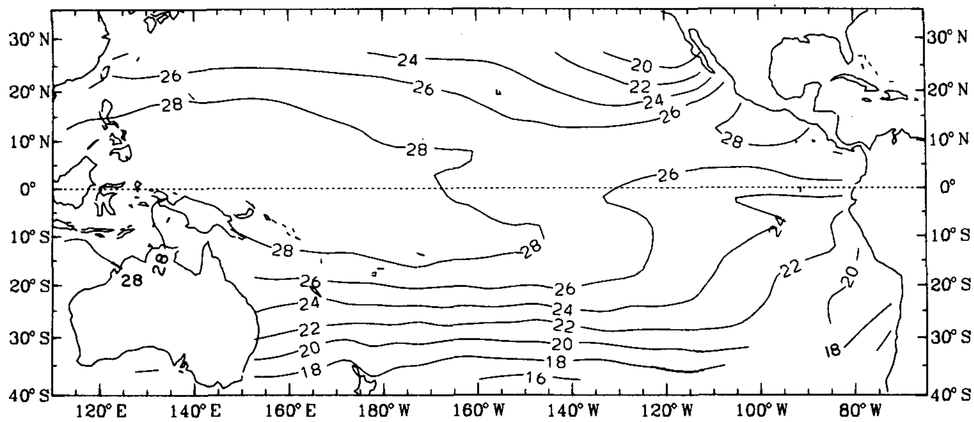

FIG. 2. Annual mean sea surface temperature (°C) for the period 1957–76. Values are based upon 5° × 5° grid averages of the marine data described in the text.

slightly, but we believe significantly, from that published by Wyrtki (1965a) based upon means for 2° grids for the period 1947–60. A qualitative comparison of our Fig. 2 with Wyrtki's Fig. 1 suggests that his means are 0.2–0.6°C warmer in the region 30°N to 20°S. Since the two data sets only overlap for four years, this suggests a generally cooling of the sea surface, which has also been noted by Kukla *et al.* (1977) for the North Pacific. We believe it is impossible to say whether this difference can be attributed to real geophysical phenomena or differences in measurement or processing of the data.

An important feature of Fig. 2 is the relatively low temperatures along the South American coast and the equator in the eastern Pacific associated with often intense upwelling. This is in sharp contrast with the equatorial western Pacific which has some of the highest sea temperatures observed. Other important features include the relatively cool water near Baja California associated with upwelling and the cold California current and the relatively warm water near 30°N between 120°E and 180°, probably associated with the southern edge of the warm North Pacific current. A general question of interest in viewing Fig. 2 is whether or not surface heating differences for various oceanic regions are at least partly responsible for the temperature patterns which we and others usually attribute primarily to oceanic dynamics.

Fig. 3 illustrates the annual mean net heating of the ocean through surface processes Q_N. This figure shows two distinct regions of rather intense heating near the equator in the eastern and western Pacific. In addition, there is a large region of moderate heating in the vicinity of the South Pacific high and smaller regions of moderate heating along the Mexican and South American coasts. This pattern of net heating differs quite significantly from those previously published by Hastenrath and Lamb (1978), Wyrtki (1965b) and Budyko (1963). Although the differences be-

LONG TERM ANNUAL MEAN
$Q_S-Q_H-Q_L-Q_I$ NET HEATING (W/m^2)

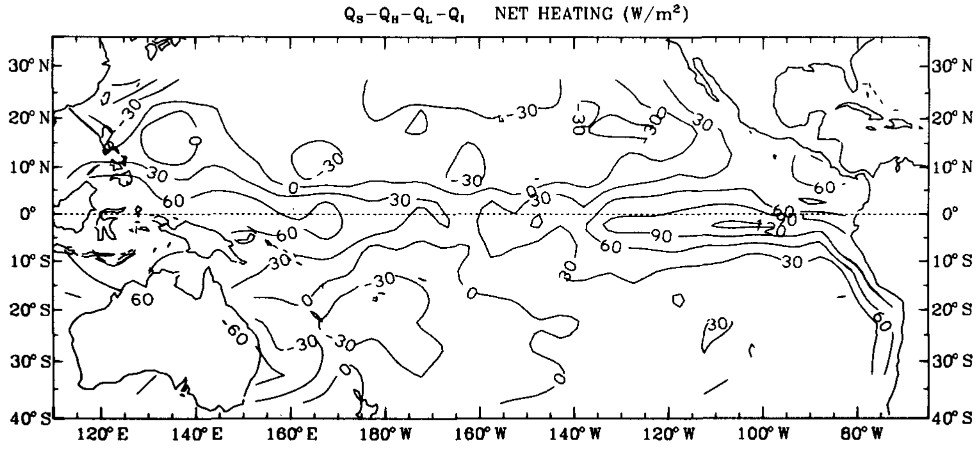

FIG. 3. Annual mean net oceanic surface heating (W m^{-2}) for 1957–76.

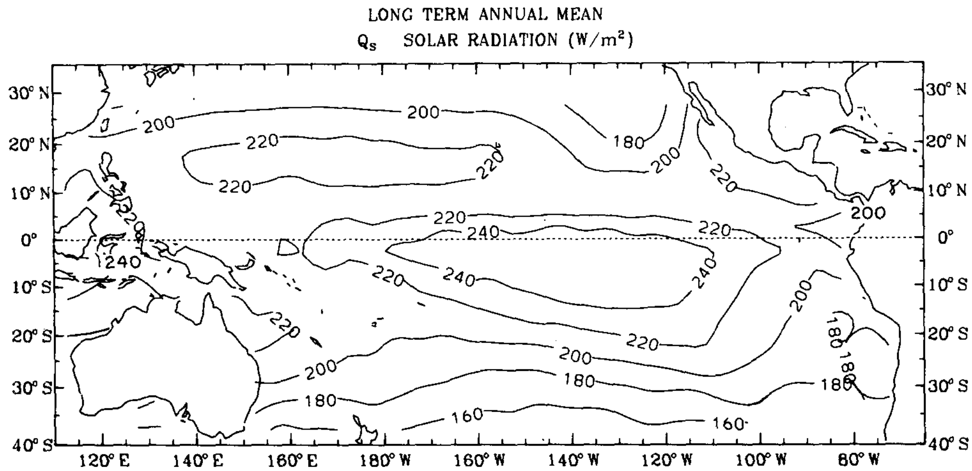

FIG. 4. Annual mean absorption of solar radiation by the ocean (W m^{-2}) for 1957–76.

tween Fig. 3 and the analysis of Budyko are great, they probably can be attributed primarily to Budyko's coarse resolution (ostensibly 10° latitude-longitude) and to his lack of oceanic data. The especially large differences with Wyrtki's analysis cannot be so easily reconciled. Wyrtki (1965b, his Fig. 5) shows net heating in the entire North Pacific between 30°N and 20°S except west of 150°E near 30°N and west of the 160°W near 20°S. This is in sharp contrast to the slight cooling in the central Pacific shown in Fig. 3, poleward of about 10°N and 10°S. Hastenrath and Lamb (1978) tend to show an even greater region of cooling than the present work for their rather limited coverage east of 110°W. In a later discussion we will show the differences with Wyrtki's (1965b) results are largely attributable to large differences in the calculated latent heat flux.

The principal energy input into the tropical ocean is the absorption of solar radiation which is shown in Fig. 4. This pattern is dominated by two rather broad regions of absorption > 220 W m^{-2} in the North and South Pacific divided by a slight local minima associated with the mean position of the Intertropical Convergence Zone. There is also a region of fluxes > 220 W m^{-2} in the western Pacific which is apparently separated from that of the central South Pacific by the Southern Convergence Zone (Trenberth, 1976). Finally, there is a fourth local maximum near the coast of Central America.

The solar heating pattern illustrated in Fig. 4 is similar to that published by Wyrtki (1965b), except that the previous analysis shows heating values which are ~10% less everywhere. Hastenrath and Lamb (1978) give results for part of the Pacific which are almost 20% less. The recent analyses of Reed (1977) and Simpson and Paulson (1979) suggest that the Berliand cloudiness formula used by Wyrtki overestimates the dimunition of solar radiation under cloudy conditions. Fig. 5 shows our analysis for the annual mean total cloud cover. Observer-estimated

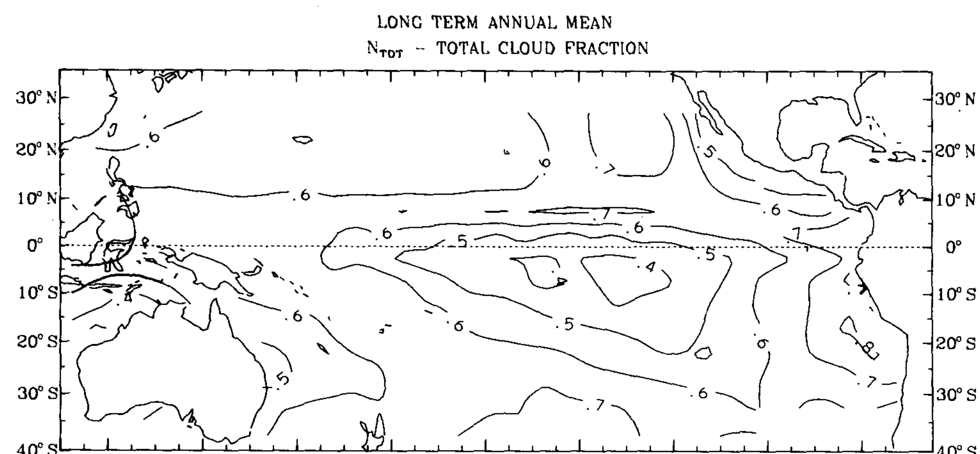

FIG. 5. Annual mean fraction of sky covered with cloud for 1957–76.

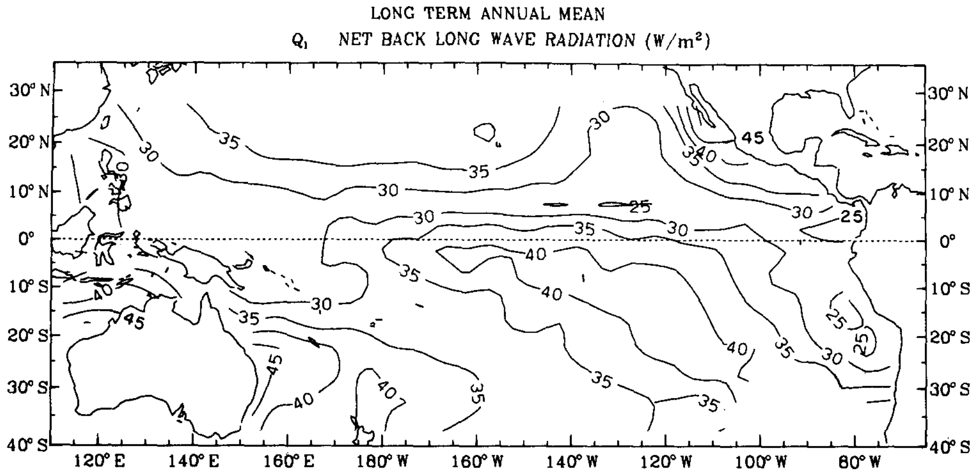

FIG. 6. Annual mean net infrared radiation emitted from the ocean surface (W m^{-2}) for 1957–76.

total cloudiness does not in general average less than 0.4 fraction of the sky covered. This implies that the formula used by Wyrtki will tend to underestimate absorbed solar radiation nearly everywhere. There remains some uncertainty, however, since none of the formulas distinguish cloud type and since it is well known that stratus dominate the eastern Pacific, whereas cumulus dominate much of the remainder of the basin.

The other radiative flux contributing to the oceanic surface net heating is the longwave radiation emitted to space minus that emitted from the sky back to the ocean, i.e., the net infrared. The spatial pattern of this flux (Fig. 6) is very azonal through the South Pacific and in the far eastern and western North Pacific. This is due in large part to the cloudiness pattern shown in Fig. 5 but also in part to the temperature and humidity. Wyrtki (1965b) shows a similar pattern with slightly smaller fluxes. It is possible that some of the differences between the present radiative analyses and those of Wyrtki is an increase in observed cloudiness from the 1946–60 period to the 1957–76 period. Unfortunately, Wyrtki (1965b) does not show an analysis of cloud cover in order for us to make the comparison.

The second most important heat flux contributing to oceanic heating or cooling is the transfer of energy during evaporation. This latent heat flux is shown in Fig. 7. The greatest losses are in the central North Pacific and a smaller region in the central South Pacific. An important minimum extends up the South American coast and along the equator. It is quite apparent that most of the features of the net oceanic heating shown in Fig. 2 may be suggested by subtracting the latent loss (Fig. 7) from the solar heating (Fig. 3). For instance, the large net heating in the eastern and western equatorial regions is primarily due to a minimum in latent heat loss. Also, the slightly negative net heating exists near the local maximum of solar heating near 15°N, 170°E largely because the latent losses are a maximum there.

As may be seen from Eq. (7), the latent heat loss

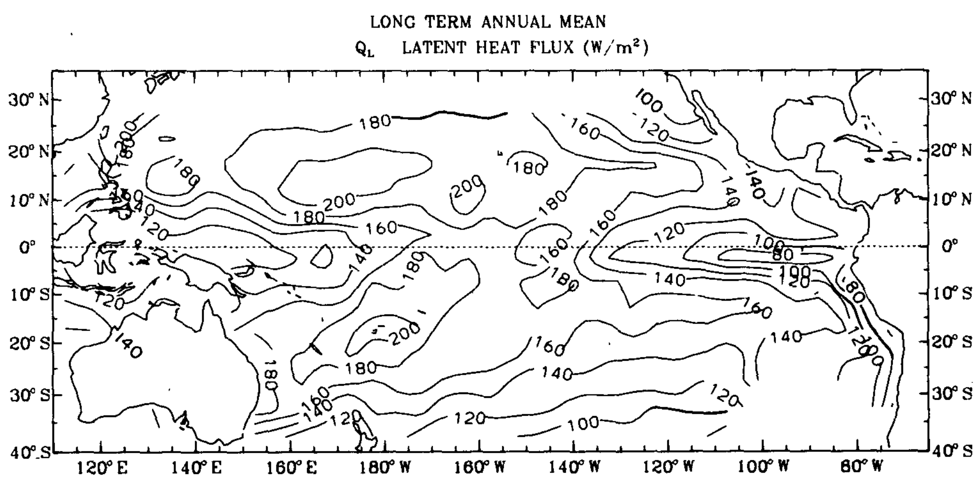

FIG. 7. Annual mean loss of latent heat through evaporation (W m^{-2}) for 1957–76.

LONG TERM ANNUAL MEAN
M WIND SPEED (m/s)

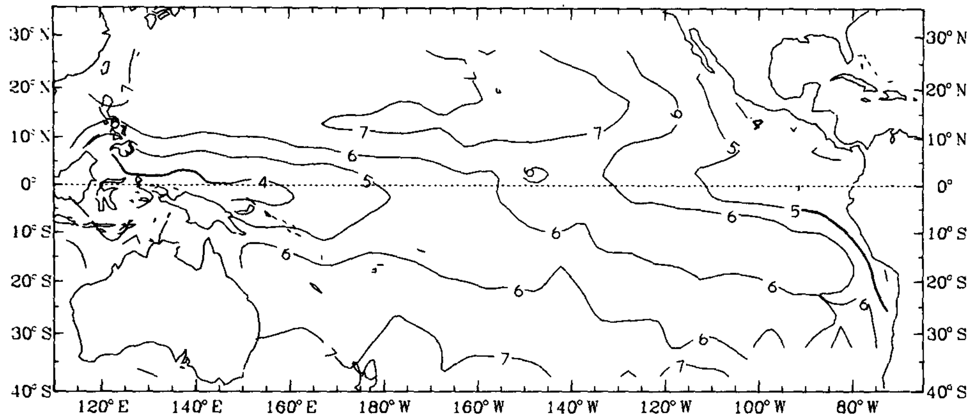

FIG. 8. Annual mean wind speed (m s^{-1}) for 1957–76.

from the ocean is proportional to both the average wind speed and the difference between the saturation specific humidity for the sea surface temperature and the measured humidity. We will call this latter parameter the humidity difference. The spatial patterns of annual mean values of these two parameters are illustrated in Figs. 8 and 9. The wind speed is greatest adjacent to the Chinese coast, near Hawaii and south of 30°S. The first two regions correspond to locales of large latent losses; the third does not. Minimum wind speeds are near the equator in the far eastern and far western Pacific where latent losses are generally relatively small. The humidity difference pattern (Fig. 9) shows a broad maximum west of the dateline between approximately 20°N and 20°S. There is a local minimum in the eastern tropical Pacific and generally lower values in the higher latitude of the Southern Hemisphere. This latter feature is undoubtedly related to the exponential relation of saturation humidity with temperature, although perhaps it also is related to the strong vertical mixing associated with the relatively high wind speeds in this area. In fact, over much of the Pacific it seems to generally hold that high wind regions are low humidity difference regions.

A third factor which weakly affects the magnitude of latent loss is variability of the latent heat exchange coefficient C_E with wind speed and air-sea temperature difference as shown by Bunker (1976). Fig. 10 illustrates the mean annual values of this coefficient. While it remains generally between 1.30 and 1.55, possibly important spatial differences do exist. The coefficient is especially low in the very stable region of minimum latent loss running up the South American coast and along the equator to ~140°W.

Whereas the latent heat losses from the ocean surface illustrated in Fig. 7 are generally comparable to those calculated by Hastenrath and Lamb (1978) they are far greater than those shown by Wyrtki (1965b). Wyrtki's values are as much as 60 W m^{-2}

LONG TERM ANNUAL MEAN
$q_s - q_a$ AIR-SEA HUMIDITY DIFFERENCE (g/kg)

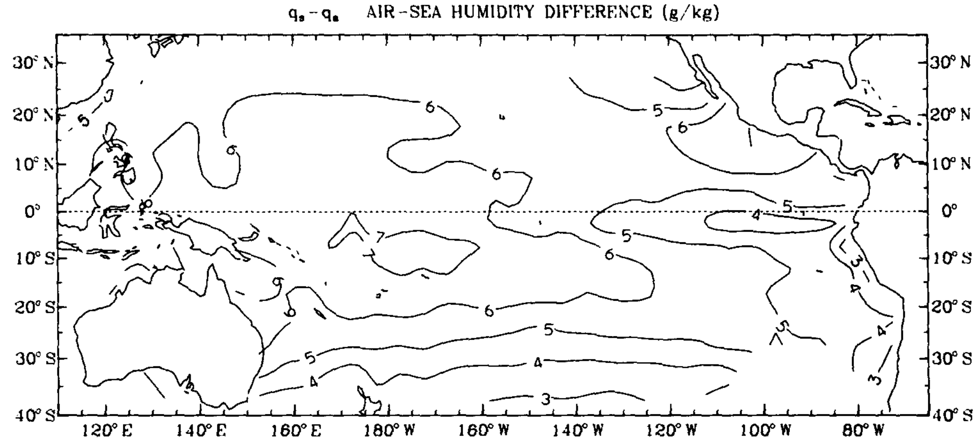

FIG. 9. Annual mean humidity difference (g kg^{-1}) for 1957–76.

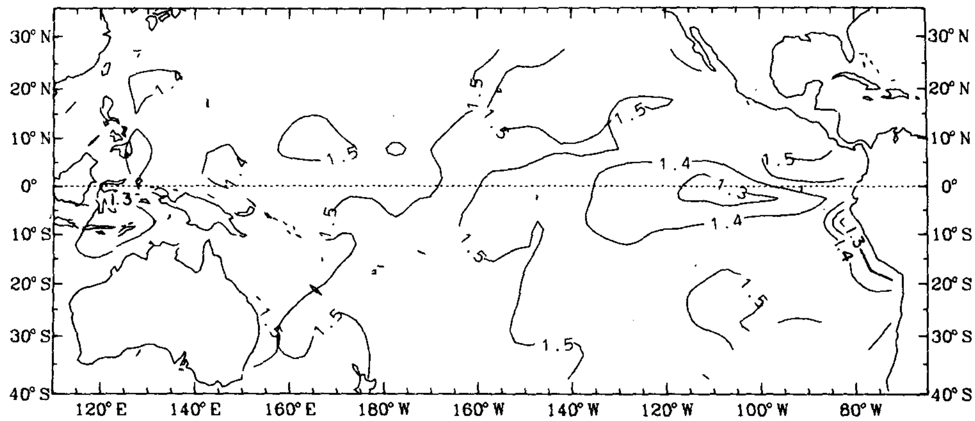

FIG. 10. Annual mean latent heat exchange coefficient (×1000) for 1957–76.

less than the current estimates. This difference is clearly significant. The primary reason Wyrtki's values are so much lower is his treatment of the sea surface temperatures. He chose to follow the suggestion of Saur (1963) and reduced all sea temperatures by 0.7°C. This substantially reduces the surface saturation humidity and hence the humidity difference. Bunker (1975)[3] has discussed in considerable detail the possible biases in sea surface temperatures *and* air and dew-point temperatures. He concluded that sea temperature differences are insignificant. Clarke (personal communication) has arrived at a similar conclusion based on comparing observations at Ship N (30°N, 140°W) and commercial ship observations in the same vicinity. Bunker concluded that the most significant bias is in dew point. This was based upon a comparison of special humidity measurements taken from a boom far from the ship superstructure and those taken in the standard manner. We corrected for this bias by using a slightly adjusted exchange coefficient as suggested by Bunker (1976).

[3] Bunker, A. F., 1975: Energy exchange at the surface of the western North Atlantic Ocean. WHOI-75-3, 107 pp.

The smallest surface heat flux in the tropical Pacific Ocean is the loss or gain of heat from the ocean surface by turbulent exchange of sensible heat whose annual average is shown in Fig. 11. Note that the magnitudes are generally more than an order of magnitude less than the solar heating and latent loss and considerably smaller than the infrared radiative loss. Furthermore, the spatial character is quite uniform with the most important maximum near the Asian coast. This flux term is primarily related to the windspeed pattern shown in Fig. 8 and the air-sea temperature difference shown in Fig. 12. This figure shows that the ocean is generally warmer than the air by less than an average of 1°C. The only region in which the sea is colder than the air and hence

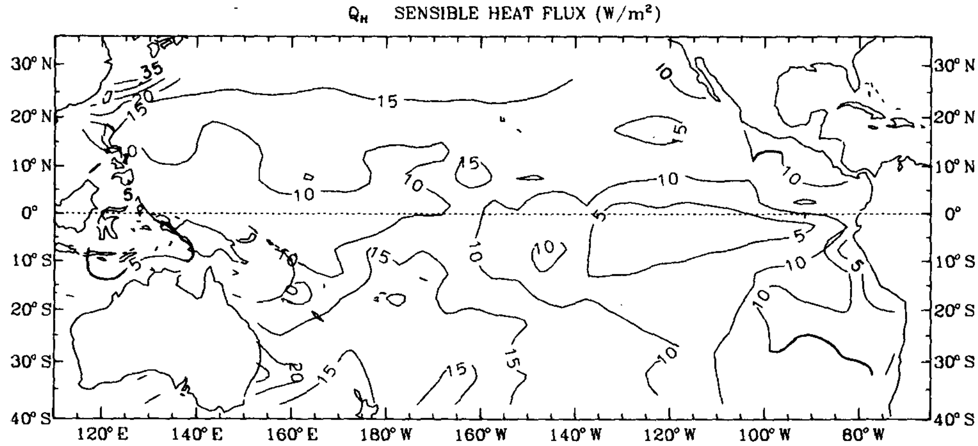

FIG. 11. Annual mean loss of sensible heat from the ocean (W m⁻²) for 1957–76.

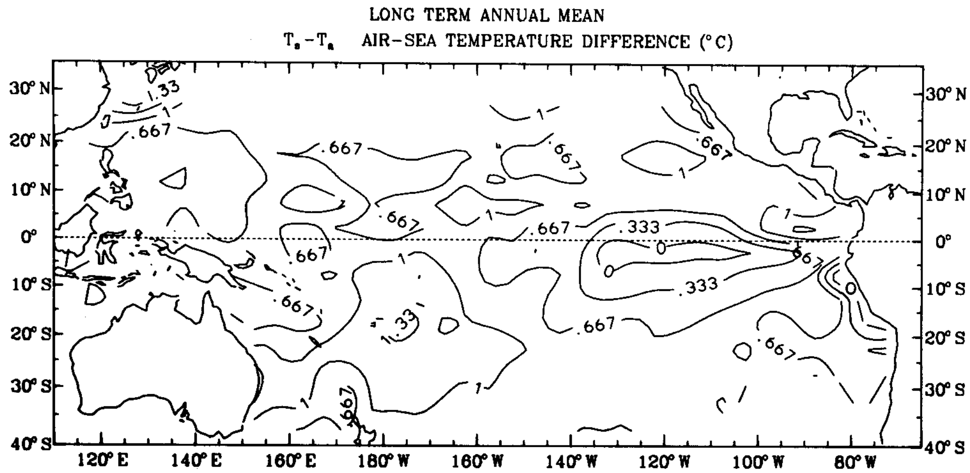

FIG. 12. Annual mean sea surface minus surface air temperature difference (°C) for 1957–76.

heated by the air through the sensible heat exchange is in the most intense upwelling region in the eastern equatorial Pacific. Conversely, the largest positive differences are near the Asian coast in which area the sea is up to 1.7°C warmer than the air.

5. Error analysis

Our analysis has shown that the annual average net heating of the ocean through the surface may be as high as +125 W m^{-2} and as low as −95 W m^{-2} in the Pacific Ocean between 30°N and 40°S. However, much of the ocean has heat fluxes which are between ±30 W m^{-2}. Important meteorological and oceanographic implications ensue from these net heating calculations and therefore require that we evaluate carefully the certainty of our flux estimates. Uncertainties in these values are due to two different sources. The first problem is defining the "bulk" formulas which best relate ship-derived meteorological variables with radiative and turbulent heat fluxes. The second arises from errors and sampling biases which are inherent in such a heterogeneous data set as the archive of marine weather observations.

We have analyzed the errors in the calculated net heating using the formalism outlined in Beers (1957). If that net heating is given by Eq. (1), then the standard deviation of the estimates is given by

$$s_{Q_N} = (s^2_{Q_S} + s^2_{Q_I} + s^2_{Q_L} + s^2_{Q_H})^{1/2}, \tag{8}$$

where s is the standard deviation and the other symbols conform to Section 2. Eq. (8) assumes that errors in the four fluxes are independent. The standard deviations of the individual components may be ascertained using an equation of the form

$$s_V = \left[\left(\frac{\partial V}{\partial X}\right)^2 s_X{}^2 + \left(\frac{\partial V}{\partial Y}\right)^2 s_Y{}^2 + \left(\frac{\partial V}{\partial Z}\right)^2 s_Z{}^2\right]^{1/2}, \tag{9}$$

where V is the dependent variable and X, Y and Z are three possible independent variables and, again, it has been assumed that errors in X, Y and Z are uncorrelated.

This method of error analysis was used to estimate the impact of the uncertainties in the specification of the bulk formulas as well as the contribution of sampling and measurement errors on our estimates of the solar, infrared and latent fluxes. In determining the effect of the uncertainties in the formulas themselves, it was assumed that the general forms were correct but that the "constants" Q_{S0}, A, Q_{I0}, B and C_E are known imprecisely. This was done by assuming that these parameters have the "standard deviations" illustrated in Table 1. To account for errors in the flux estimates due to measurement and sampling problems, standard deviations were assigned to the measured variables of cloudiness (C), wind speed ($|\mathbf{V}|$) and humidity difference ($q_s - q_a$). In general, it was assumed that the standard devia-

TABLE 1. Parameters used in and the results of the error analysis of the surface fluxes of downward solar radiation (Q_C), net infrared radiation (Q_I) and the loss of latent heat through evaporation (Q_L).

Parameter	Mean	Standard deviation
Q_{S0} (W m^{-2})	275	14
A	0.62	0.05
C	0.5	0.05
β	15°	0
Q_C (W m^{-2})	229	15
Q_{I0} (W m^{-2})	55	8
B	0.7	0.1
C	0.5	0.05
Q_I (W m^{-2})	36	7
C_E	1.5	0.05
$\lvert\mathbf{V}\rvert$ (m s^{-1})	6	0.6
$(q_s - q_a)$ (kg kg^{-1})	0.006	0.0006
Q_L (W m^{-2})	135	20

tions are ~10% of the mean. Those of the radiative fluxes were inferred from Simpson and Paulson (1979).

The results of applying Eq. (9) using the data in Table 1 are also indicated in Table 1. As may be seen, this analysis shows that the possible errors in the estimated absorbed solar and latent heat flux are similar with that of the latent slightly larger. If these results are applied to Eq. (8) ignoring the sensible flux, then the standard deviation of the net oceanic surface heating is ~25 W m^{-2}. This corresponds to a 95% confidence interval of ±49 W m^{-2} if one assumes that Q_N is normally distributed. Given this confidence level, many of the features in the map of net heating (Fig. 3) are not statistically distinguishable. The two intense heating regions along the equator do, however, remain significantly different.

While this result is somewhat discouraging in regard to the net heating, many more of the features of the maps of the solar, infrared and latent heat fluxes are statistically significant at the 95% confidence level. Those confidence levels for these fluxes are ~29, 14 and 39 W m^{-2}, respectively. Thus the principal features of Figs. 4, 6 and 7 are significant. It is very difficult, however, to say whether or not the differences between these analyses and those of previous authors are significantly different since the formulas and data of the other authors have different means and standard deviations.

6. Discussion

The mean annual net heating of the ocean surface shown in Fig. 3 is the most important parameter which we have discussed. Given the rather large 95% confidence intervals for this variable derived in the last section, it would seem necessary to test the validity of our estimates by some independent means. One such way is to evaluate the physical reasonableness of the annual average ocean dynamics which is required to balance the heating or cooling of a region through its surface in order to have a steady temperature. If one assumes that surface advection and/or upwelling are the dynamical processes which must balance the surface processes, then

$$Q_N = \int_Z^0 \rho C_W \left(V \frac{\partial T}{\partial x} + W \frac{\partial T}{\partial z} \right) dz, \qquad (10)$$

where Z is the mixed-layer depth, V the current speed along direction x, and W the vertical velocity. Recent estimates by Wyrtki (1977) of the horizontal velocities of the Peru current allow a calculation of the advective cooling in the eastern tropical Pacific where surface heating rates are of the order of 100 W m^{-2}. Given Wyrtki's estimates of an upper limit of velocity of ~0.3 m s^{-1} and gradients of -3×10^{-6} °C m^{-1} and given an approximate mixed layer depth of 25 m then the horizontal advection would lead to a cooling rate of ~90 W m^{-2}, which is in general agreement with our results. Much uncertainty remains, however, due to our lack of knowledge of typical advective velocities and the degree that upwelling must contribute.

Another region which may be tested in a similar manner is the region of weak cooling in the central North Pacific near 20°N, 160°W. This cooling requires warm advection or some higher order heating mechanism. Since this is in the region of the easterly Equatorial Current and since sea temperatures to the east are cooler, this zonal flow will not account for the required heating. Northward Ekman drift could result in the necessary warm advection, but the results of Wyrtki and Myers (1975)[4] suggests that the annual mean northward drift reaches only to ~18°N in this region. Thus it would seem that we have overestimated the net cooling in this region (and perhaps elsewhere). The magnitude of this overestimate is difficult to establish, but an increase in the net heating of this region of 30 W m^{-2} would probably be sufficient to make our results consistent with these simple oceanographic requirements. Such an adjustment is well within the estimated uncertainty of our values of 49 W m^{-2}.

As a final test of the overall validity of our estimates of net surface heating, we have computed the total oceanic surface heating for the region 30°N to 40°S from the American coasts to 110°E. Energy enters this region at an annual average rate of ~1.9 × 10^{15} W. If one assumes that to maintain a steady annual average temperature half of this energy must be transported northward across 30°N, an oceanic northward transport of heat of ~0.95 × 10^{15} W is required. Oort and Vonder Haar (1976) have estimated from global energy balance considerations that the Atlantic and Pacific Oceans contribute a northward transport across 30°N of 1.9 × 10^{15} W. Bryan (1978) has calculated from the annual average net heating estimates of Bunker and Worthington (1976) that the Atlantic transports ~0.75 × 10^{15} W across this latitude. Subtracting the estimate for the Atlantic from the total gives an estimate of 1.5 × 10^{15} W which must be transported in the Pacific Ocean northward across 30°N. This is in quite good agreement with our estimate, especially if one assumes that northward transport across 30°N is likely to be considerably greater than southward transport across 40°S in the Pacific Ocean due to the lack of a strong western boundary current in the Southern Hemisphere.

If our estimates of net surface heating are too low in the central Pacific as is suggested in a previous paragraph, then it seems that the solar radiation at the surface has been underestimated or the latent

[4] Wyrtki, K., and G. Meyers, 1975: The trade wind field over the Pacific Ocean. Rep. HIG-75-1, University of Hawaii, 26 pp. [Available from the first author].

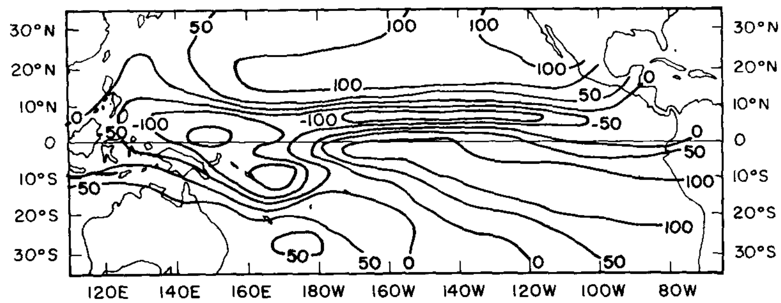

FIG. 13. Annual mean difference between evaporation and precipitation expressed in flux units (W m^{-2}). Evaporation is determined from Fig. 3 and precipitation is taken from interpolated 5° × 5° grid values of the map of Taylor (1973) multiplied by the latent heat of evaporation.

losses overestimated in this region. The earlier work of Wyrtki (1965b) gave net annual surface heating in the central Pacific in the range of +10 to +40 W m^{-2}, which could be consistent with the maintenance of steady-state temperatures by cold easterly advection. The primary reason Wyrtki's net heating is larger than ours is that his latent losses are ~60 W m^{-2} less. His latent losses are smaller primarily because he chose to reduce sea surface temperatures by 0.7°C to adjust for the bias Saur (1963) found between commercial ship-derived sea temperatures (those used by both Wyrtki and ourselves) and selected "bucket" measurements. However, more recently Husby and Seckel (1975),[5] Bunker (1976) and Clarke (personal communication) have carried out comparisons similar to those of Saur and have not found the regularly derived sea temperatures to be too warm. Furthermore it may be seen from Fig. 12 that, if the sea temperature were reduced by 0.7°C, then the sea and air would have nearly identical temperatures over most of the tropical Pacific, which is contradictory to usual observations.

In the same sense that our estimates of net heating may be "tested" through the use of conservation of surface ocean energy, the estimate of the loss of latent heat may be partially tested through the application of conservation of water in the atmosphere. For each 5° grid north of 30°S in Fig. 7 we have subtracted from our estimate of the annual average latent heat loss, the latent heat gain through precipitation, estimated from the annual precipitation map published by Taylor (1973).[6] The difference (expressed in W m^{-2}) is shown in Fig. 13. Unlike all of the previous contoured maps, the isolines on Fig. 13 are the result of a subjective analysis of the 5° grid differences. The differences shown in Fig. 13 must be balanced by dynamical convergence or divergence of moisture. For instance, Fig. 13 suggests that convergence must occur in most of the western Pacific between 15°N and 15°S and in the eastern Pacific near 8°N. This would be in agreement with the observations of low-level atmospheric convergence in these regions of the Intertropical Convergence Zone. From Fig. 13 we have estimated that near 20°N excess evaporation over precipitation requires an annual average divergence of moisture in the atmosphere comparable to 95 W m^{-2}. Peixoto and Crisi (1965; Plate 141)[7] give annual average vertically integrated divergence for 1958 from which we estimate a divergence for this region equivalent to 71 W m^{-2}. Given the many difficulties in both estimates, this would seem to be quite good agreement. We also calculate for this region the divergence implied by using Wyrtki's (1965b) annual average latent loss results and the Taylor (1973) rainfall estimates. This calculation gives a mean at 20°N of about 38 W m^{-2}. This is less than the estimated zonally and vertically averaged moisture divergence equivalent to 44 W m^{-2} calculated from five years of data by Rasmusson (1972).

Recently Hastenrath (1980) has used satellite-derived net radiation at the top of the atmosphere and previous estimates of net surface heating [a combination of the results of Hastenrath and Lamb (1978), Wyrtki (1965b) and Budyko (1963)], to estimate the net heating of the atmosphere for the tropical Pacific Ocean and other tropical oceanic regions. We have carried out a similar calculation using our net heating (Fig. 3) and the unadjusted satellite data used by Hastenrath, published by Ellis *et al.* (1978), kindly provided by Dr. Ellis. Fig. 14 shows this net heating of the atmosphere. It should be noted that so much energy goes into heating the eastern tropical Pacific Ocean that the atmosphere has a large net energy deficit which must be balanced by atmospheric processes. A small energy deficit also is apparent in the

[5] Husby, D. M., and G. R. Seckel, 1975: Large-scale air-sea interactions at ocean weather station V, 1951–71. NOAA Tech. Rep. NMFS SSRF-696, 44 pp. [Available from the authors].

[6] Taylor, R. C., 1973: *An Atlas of Pacific Islands Rainfall.* Hawaii Inst. of Geophysics, HIG-73-9, University of Hawaii, 29 pp. [Available from the author].

[7] Peixoto, J. P., and A. R. Crisi, 1965: Hemispheric humidity conditions during the IGY. Dept. of Meteorology, MIT, 169 pp.

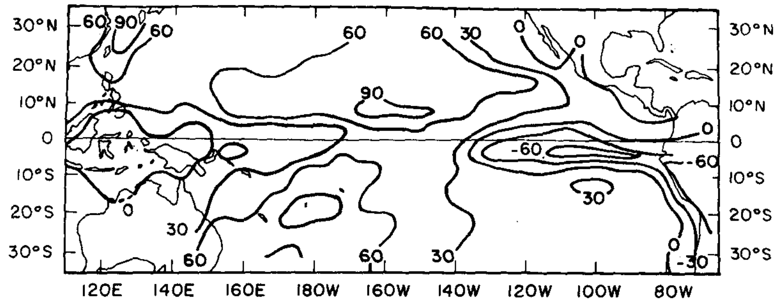

FIG. 14. Annual mean net heating of the atmosphere (W m^{-2}) determined as the difference between the net radiation at the top of the atmosphere (Ellis *et al.*, 1978) and the net surface heating of the ocean (Fig. 3).

western equatorial region. This latter deficit is less certain because of our assumption that the net heating of the ocean occurs everywhere (including on such islands as New Guinea). Hastenrath's (1980, Fig. 4c) results indicate an overall flatter pattern with less atmospheric cooling in the eastern equatorial region and less heating elsewhere. Thus, our results require more atmospheric transport of energy and an overall more vigorous circulation.

In this presentation no mention has been made of the nature and possible impact of the seasonal cycle of meteorological variables on the calculated vertical heat flux fields. Weare *et al.* (1980)[8] have prepared an atlas of monthly mean maps of basic and derived meteorological fields which partially rectifies this problem. In that work they show the 20-year monthly average of means, intramonth variances and covariances and interannual standard deviations for a number of variables.

Although we have briefly discussed the many possible factors which contribute to uncertainties in the surface fluxes, and we have made quantitative estimates of those uncertainties, it seems impossible at this time to pinpoint the principal causes of error. Nevertheless, it seems likely that the large and uncertain solar and latent fluxes should be most carefully studied as to the realism of the bulk formulas using regular marine data. We are convinced that it is these formulas rather than the questionable quality of a small part of the marine weather reports which have the largest influence on long-term mean fluxes, such as we have presented. Our conclusion suggests that there is an urgent need to obtain carefully calibrated radiation measurements and directly measured vapor flux measurements during all seasons and over a broad range of oceanic locales. Of secondary importance, with respect to deriving accurate long-term mean fluxes, would be attempts to broaden the data base for those formulas by using satellite and other remotely sensed data.

[8] Weare, B. C., P. T. Strub and M. D. Samuel, 1980: *Marine Climate Atlas of the Tropical Pacific Ocean*. Dept. of Land, Air and Water Resources, University of California, Davis, 147 pp. [Available from the first author].

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