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Net heat gain of the tropical Pacific Ocean computed from subsurface ocean data and wind stress data

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Abstract—The heat conservation equation in the mixed layer derived by BEHRINGER and STOMMEL (1981, *Journal of Physical Oceanography*, **11**, 1393–1398) is used to map the interseasonal and annual rates of heat gain in the tropical Pacific Ocean between 30°N and 30°S (except the 4°N–4°S band). The interseasonal rates of heat gain are mostly determined by the rate of local heating; horizontal heat advection is also determinant between 4° and 10° in each hemisphere. The annual rate of heat gain is several times smaller than the interseasonal ones and, is mainly dominated by the effects of oceanic heat transports.

Compared to earlier calculations based on bulk aerodynamical formulas, the patterns of the annual heat gain are similar to those given by ESBENSEN and KUSHNIR (1981) in the south tropical Pacific and north of 20°N and WEARE *et al.* (1981, *Journal of Physical Oceanography*, **11**, 705–717) between 4°N and 10°N west of 155°W. These results also depict a broad pattern of an annual net heat loss under the mean position of the Inter Tropical Convergence Zone (ITCZ) which was never reported before.

1. INTRODUCTION

EVIDENCE that anomalies of sea surface temperature (S.S.T.) in the tropical Pacific Ocean are related to several aspects of global climate has been found by many investigators (HOREL and WALLACE, 1981; PAN and DORT, 1983). Understanding the many physical processes which cause the low frequency variations of the tropical Pacific Ocean S.S.T. is a useful contribution to understanding the global climate. The most obvious processes that determine S.S.T. variations are net surface heat flux (Qn), vertical advection, vertical mixing and horizontal advection. Among these processes Qn is extremely important because it is one of the primary ways in which the ocean can be cooled or heated (MCPHADEN, 1982; STEVENSON and NIILER, 1983).

Since no direct measurements of Qn can be made, it has been calculated as a residual from

$$Qn = Qs - Ql - Qi - Qh, \tag{1}$$

where Qs, Ql, Qi and Qh refer to the net surface flux of solar energy, the latent heat, the infrared radiation and the sensible heat. These terms are only available from empirical relations, usually called the bulk formulas, using meteorological parameters (cloudiness, wind speed and air, sea and dew point temperatures). Therefore, the net surface heat flux (Qn) is a residual of quantities estimated by empirical formulas subject to different

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parametrizations (or calibrations). Estimates of errors in the bulk formulas have been made by various authors (TALLEY, 1984) and different values of Qn have been computed in the tropical Pacific Ocean by BUDYKO (1963), WYRTKI (1965), WEARE *et al.* (1980, 1981) and ESBENSEN and KUSHNIR (1981).

The present work is based on a different approach to compute the net heat gain of the tropical Pacific Ocean (30°N–30°S), using subsurface oceanographic data and wind stress data, in order to obtain an independent estimate. This indirect estimate was suggested by a similar calculation in the tropical Atlantic Ocean (BEHRINGER and STOMMEL, 1981).

In the following sections, the model of BEHRINGER and STOMMEL (1981) is reviewed and the data and processing techniques needed in the model are presented. The net rates of heat gain in the tropical Pacific Ocean, for the annual average and for four interseasonal periods, are then compared with some previous investigations.

2. THE SURFACE HEAT BUDGET IN THE MIXED LAYER

According to BEHRINGER and STOMMEL (1981), the conservation of heat in a mixed layer of depth (h) requires that

$$\underset{\text{(I)}}{Qn} = \rho \cdot Cp \cdot \Bigg[\underset{\text{(II)}}{h \cdot (\delta\bar{T}/\delta t)} + \underset{\text{(III)}}{(\delta\bar{T}/\delta x) \cdot \int_{-h}^{0} Ug \cdot \mathrm{d}z}$$

$$+ \underset{\text{(IV)}}{(\delta\bar{T}/\delta x) \cdot (\tau y/\rho \cdot f)} + \underset{\text{(V)}}{(\delta\bar{T}/\delta y) \cdot \int_{-h}^{0} Vg \cdot \mathrm{d}z} \tag{2}$$

$$- \underset{\text{(VI)}}{(\delta\bar{T}/\delta y) \cdot (\tau x/\rho \cdot f)} + \underset{\text{(VII)}}{(Wh \cdot Dt)/2} \Bigg]$$

where (Ug, Vg) are the components of the geostropic velocity vector, (τx,τy) those of the wind stress vector, and f is the Coriolis parameter. The temperature $T(x,y,z,t)$ in the mixed layer is represented by the sum of a term $\bar{T}(x,y,t)$ which is independent of z and a term which varies linearly in z by a small constant amount (Dt). The vertical velocity (W) has been approximated by a linear profile, and varies from (Wh) at the bottom of the mixed layer to zero at the surface, so that

$$W = -(Wh/h) \cdot z \tag{3}$$

where (Wh) is given by

$$Wh = (\delta/\delta x) \cdot (\tau y/\rho \cdot f) - (\tau x/\rho \cdot f) - (\beta/f) \cdot \int_{-h}^{0} Vg \cdot \mathrm{d}z \tag{4}$$

(β is the gradient of the Coriolis parameter with respect to the y direction).

In this formulation of Qn, the heat mixed down through the bottom of the mixed layer has been neglected so that we can interpret Qn as the rate of heat gain through a unit area of the sea surface. Each of the terms of equation (2) will be referred to in Section 4 as the net rate of heat gain (I), the local heat storage (II), the contributions of zonal geostrophic transport (III), zonal Ekman transport (IV), meridional geostrophic transport (V), meridional Ekman transport (VI) and vertical advection (VII).

3. DATA AND DATA PROCESSING

Two categories of basic data have been used in this study. The first ones are the subsurface data constituted by four seasonal means of temperature and salinity profiles obtained on a 1° square grid (i, j) of the tropical Pacific from the National Oceanographic Data Center (NODC). These profiles are parts of those derived by LEVITUS (1982) for the world ocean and constitute the most up-to-date files for the tropical Pacific Ocean. Unfortunately, the Levitus analysis did not include any estimate of the r.m.s. variability in the temperature and salinity profiles. Seasons were defined as follows: winter = February/March/April, spring = May/June/July, summer = August/September/October and autumn = November/December/January. It should be noticed that this definition of seasons makes the *a priori* assumption of a lag between ocean and atmosphere, and can probably induce smoother fields in the oceanic and atmospheric parameters derived further. Complementary details can be found in LEVITUS (1982). The second category of data is the climatological means of monthly wind stresses obtained on a 2° in latitude by 10° in longitude grid (I, J) from 30°N to 30°S and 120°E to 80°W (WYRTKI and MEYERS, 1976). A constant drag coefficient $(Cd) = 1.5 \times 10^{-3}$ was used to convert to stress units.

To compute the components of the geostrophic velocity vector (Ug, Vg) and the vertical velocity (W) involved in equation (2), we must specify the level of no motion (Zo) and a vertical variation of temperature in the mixed layer (Dt) which will implicitly define the mixed layer depth(h). The sensitivity of the calculation has been tested for four different combinations of these two parameters given by Zo = 500 and 1000 m and Dt = 0.5 and 1°C. The results were very similar, i.e. in-between the errors of the calculations (Figs 1b and 3e); the results presented in the following sections use Dt = 1°C and Zo = 1000 m.

To compute Qn from equation (2) the dynamic height was derived relative to Zo, on the (i, j) grid, at each standard level reported in LEVITUS (1982) and for the four seasons. The components of geostrophic velocity vectors (Ug, Vg) were then derived on the (I, J) grid.

The depth of the mixed layer (h), i.e. the depth at which the temperature is Dt = 1°C lower than at the sea surface, was estimated on the (i, j) grid for the four seasons and then averaged to give values on the (I, J) grid. The mean temperature of the mixed layer $(\bar{T})$ was derived similarly on the (I, J) grid. Finally monthly wind stress data were averaged over 3-month periods on the (I, J) grid points.

Table 1. Estimates of the basic quantities involved in equation (2) *at* 15°N, 155°W

	Winter (02/03/04)	Spring (05/06/07)	Summer (08/09/10)	Autumn (11/12/01)	Annual	Units
h	82.2 ± 2.5	79.1 ± 2.1	59.1 ± 5.7	66.5 ± 5.6	71.7 ± 2.2	m
$\bar{T}$	24.8 ± 0.2	25.7 ± 0.3	26.5 ± 0.2	26.6 ± 0.3	25.6 ± 0.2	°C
$\int_{-h}^{0} ug \cdot dz$	−6.5 ± 1.0	−6.2 ± 0.2	−6.9 ± 0.3	−7.4 ± 1.0	−6.7 ± 0.4	$m^2 s^{-1}$
$\int_{-h}^{0} vg \cdot dz$	−1.7 ± 0.7	−2.5 ± 1.4	−1.8 ± 0.5	−2.0 ± 0.3	−2.0 ± 0.4	$m^2 s^{-1}$
$\zeta_{y/pf}$	−0.95 ± 0.14	−0.84 ± 0.08	−0.82 ± 0.22	−0.89 ± 0.20	−0.87 ± 0.09	$m^2 s^{-1}$
$-\zeta_{x/pf}$	2.78 ± 0.46	2.80 ± 0.40	2.21 ± 0.24	2.49 ± 0.73	2.57 ± 0.25	$m^2 s^{-1}$

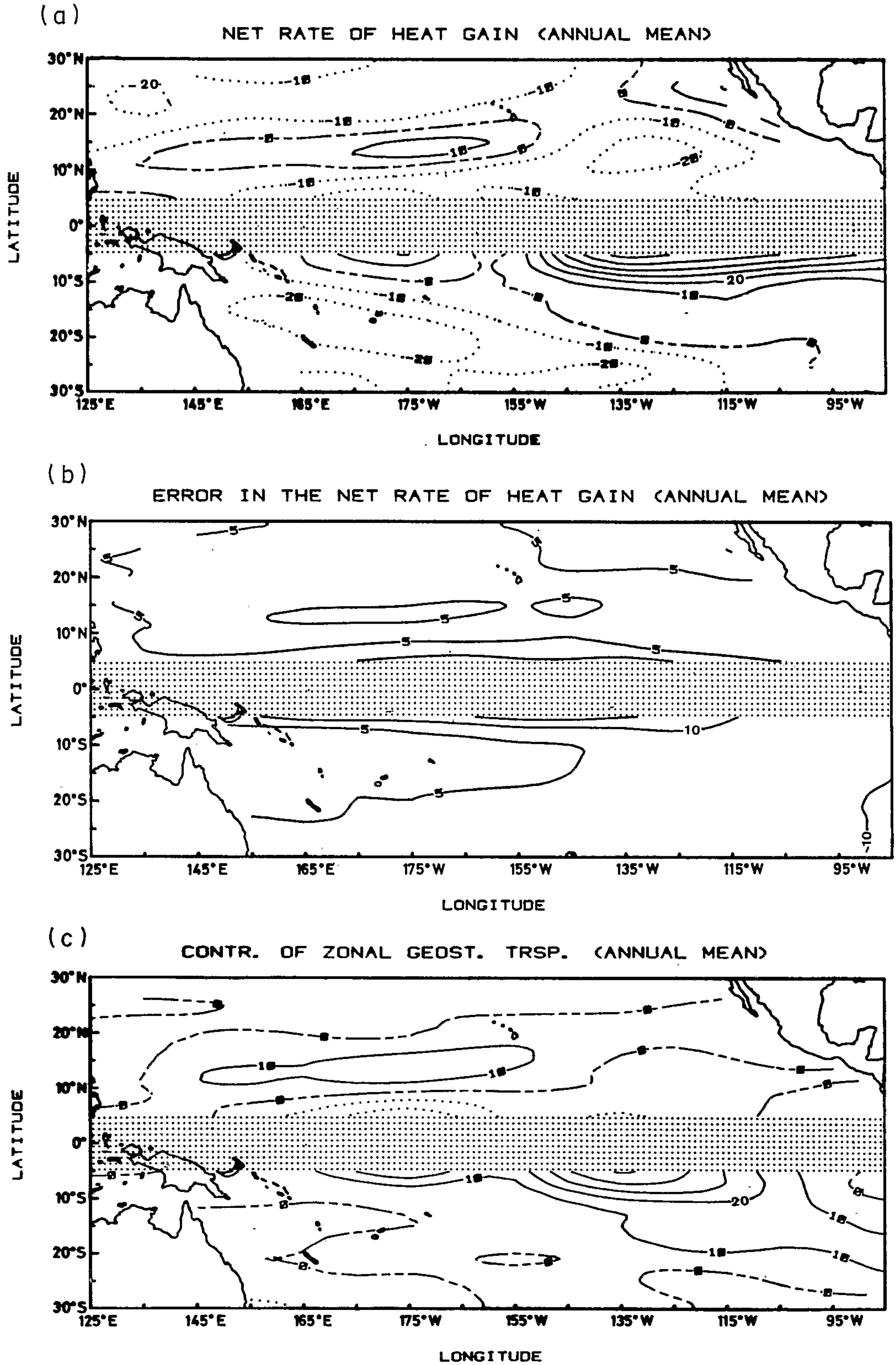
The standard deviation of every quantity relative to the average over the $2° \times 10°$ rectangles (subsurface data) or over 3-month periods (wind stress data) was used as a rough estimate of its respective error since the effects of noise and temporal or spatial variations can not be separated. Table 1 lists the values and the corresponding errors of the basic quantities used to evaluate equation (2) at 15°N–155°W. The error estimates for all these quantities have been used to obtain the error estimates for (Qn).

4. RESULTS AND COMPARISONS

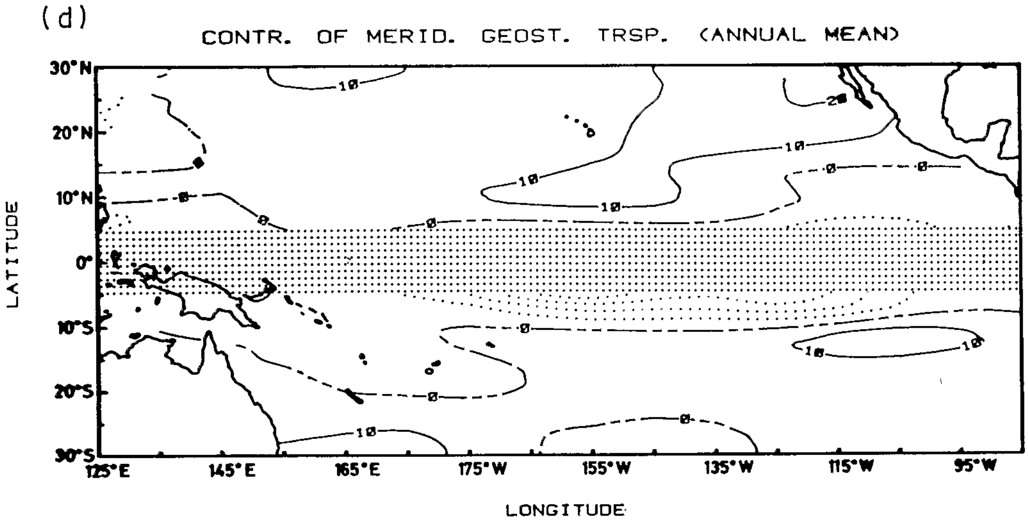
Due to equatorial singularity in the computations of vertical velocity and Ekman and geostrophic transports, BEHRINGER and STOMMEL (1981) had "no confidence" in their calculations of Qn within 10° of the equator. In fact, the terms involving the Coriolis parameter increase toward the equator inversely proportional to latitude and, in this study, unreported values close to the equator are very large and sometime unrealistic. For example, the annual mean vertical heat advection at 2–4°N ranges between 20 and 60 W m^{-2} east of the date line. This requires a vertical velocity of 0.8–2.5 m day^{-1}, whereas WYRTKI (1981) found a vertical upwelling velocity, averaged between 4°N and 4°S (i.e. including the equator where the highest value can be expected) equal to 1 m day^{-1}; BURKOV (1980) calculated upwelling speeds of 2–3 m day^{-1} for the equatorial Pacific. The value 0.8–2.5 m day^{-1} determined here is thus overestimated. However, similar to WYRTKI's (1981) calculations, the terms involving the Coriolis parameter can be considered to settle to reasonable values poleward of 4°. A cutoff latitude of 4° north and south of the equator was thus adopted, and Qn will not be presented in the 4°N–4°S latitudinal strip.

The annual mean net surface heating

The annual average of the net amount of heat (Qn) received (+) or lost (–) by the ocean and of its error is presented in Figs 1a,b. A weighted smoothing procedure involving the eight surrounding rectangles has been applied prior to contouring; this produces smoother fields and reduces the error estimates. We see two distinct regions of maximum heating, one between 10 and 16°N west of 155°W and the other in the southeastern part of the tropical Pacific. In the first region, the net heat gain is mainly determined by the balance between the contributions of meridional geostrophic and Ekman transports (Figs 1d,e) reinforced by the zonal geostrophic advections (Fig. 1c) associated with the North Equatorial Current (NEC). In the second region, the advection of cold water by the South Equatorial Current (SEC) (Fig. 1c) plus a large value of the meridional Ekman heat transport (Fig. 1e) due to a strong meridional gradient of the mean temperature in the mixed layer ($\bar{T}$), suggest an annual heat gain from the atmosphere. Spatial variations in the annual means of local heat storage and contribution of vertical advection are not significant, and they are not presented here. It is interesting to note that the locations of these two regions are very similar to the two broad maxima (i.e. larger than 220 W m^{-2}) of annual mean absorption of solar radiation (Qs) calculated by WEARE *et al.* (1981, Fig. 4). These two last regions are separated by the 4°N–4°S band (not investigated here) and by an area of negative heating situated between 4 and 20°N, east of 155°W, mostly due to the contribution of meridional Ekman transport (Fig. 1e). This area of net heat loss by the ocean is probably due to a low


(a)
NET RATE OF HEAT GAIN (ANNUAL MEAN)
LATITUDE
LONGITUDE
30°N
20°N
10°N
0°
10°S
20°S
30°S
125°E
145°E
165°E
175°W
155°W
135°W
115°W
95°W
(b)
ERROR IN THE NET RATE OF HEAT GAIN (ANNUAL MEAN)
LATITUDE
LONGITUDE
(c)
CONTR. OF ZONAL GEOST. TRSP. (ANNUAL MEAN)
LATITUDE
LONGITUDE

Figs 1a–c.



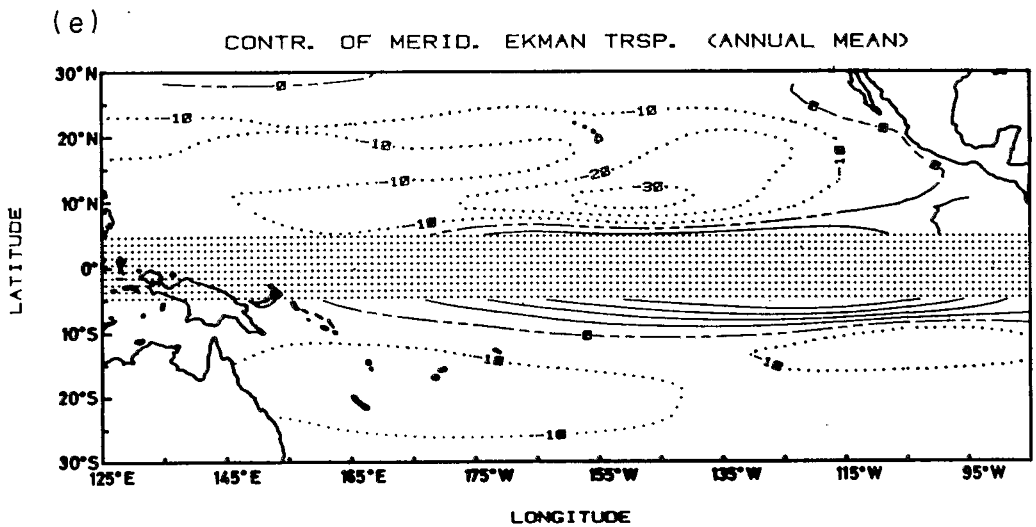


Fig. 1. (a) Net rate of heat gain (+) or loss (–) by the ocean (annual mean) (W m^{-2}). (b) Error in the net rate of heat gain or loss by the ocean (annual mean) (W m^{-2}). (c) Contributions of zonal geostrophic transport (annual mean) (W m^{-2}). (d) Contribution of meridional geostrophic transport (annual mean) (W m^{-2}). (e) Contributions of meridional Ekman transport (annual mean) (W m^{-2}).

incoming radiation associated with the mean annual position of the ITCZ. A meridional section along 150°–160°W (Fig. 2) showing the contributions of significant terms of equation (2) clearly indicates that the annual mean horizontal heat advection has a greater influence on Qn than the local heat storage. Contributions of meridional Ekman and geostrophic heat transports are indeed the dominant factors, although they have a tendency to balance each other. In conclusion, the annual mean surface heating is mostly accounted for by the horizontal oceanic heat advections.

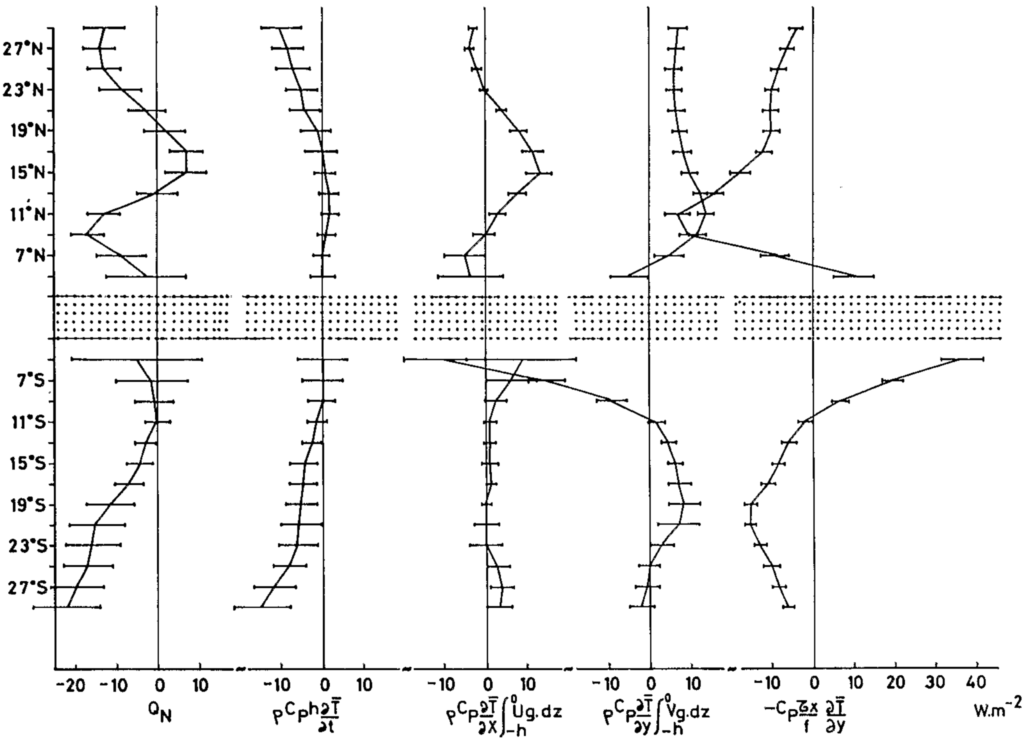
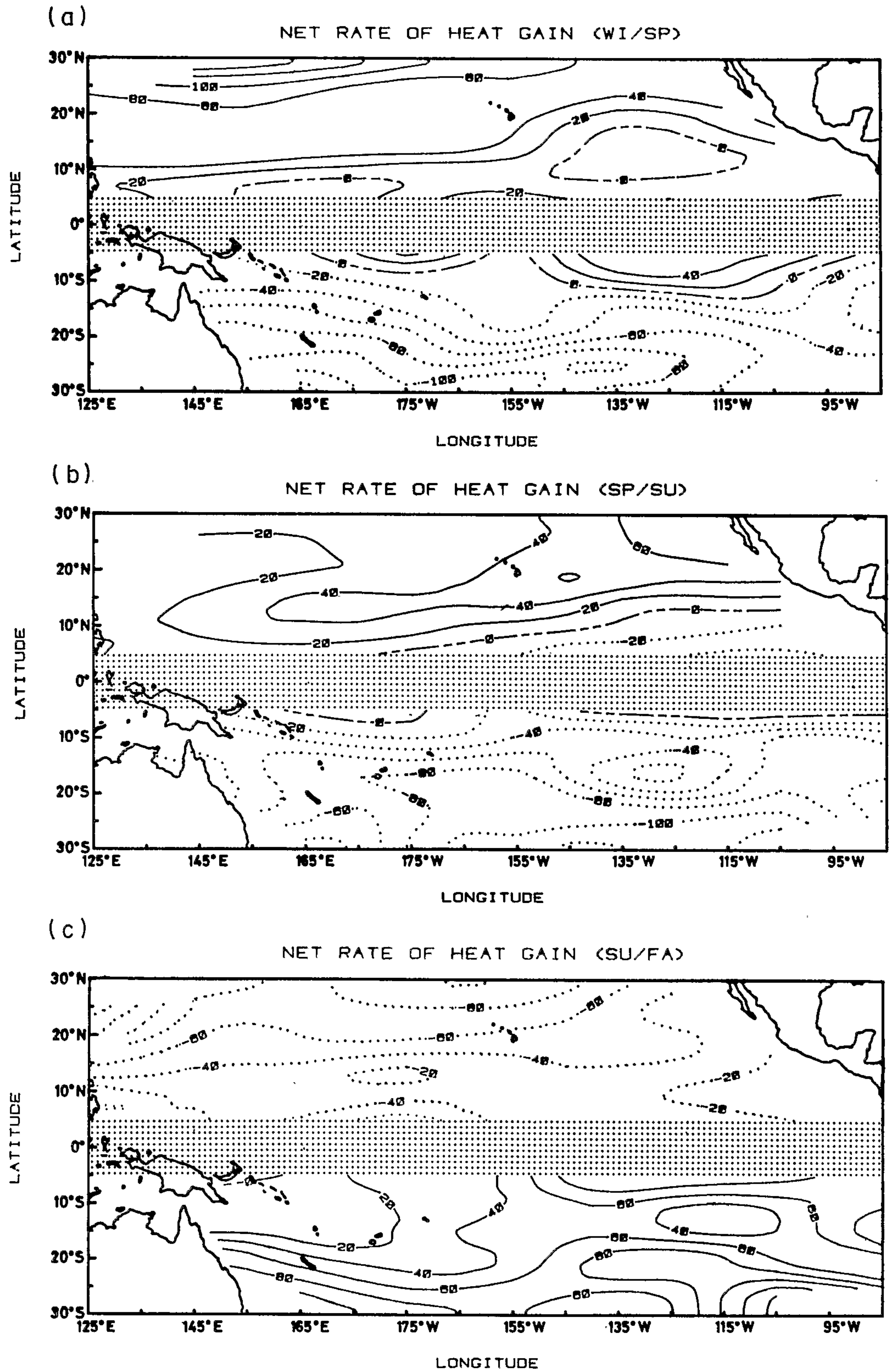


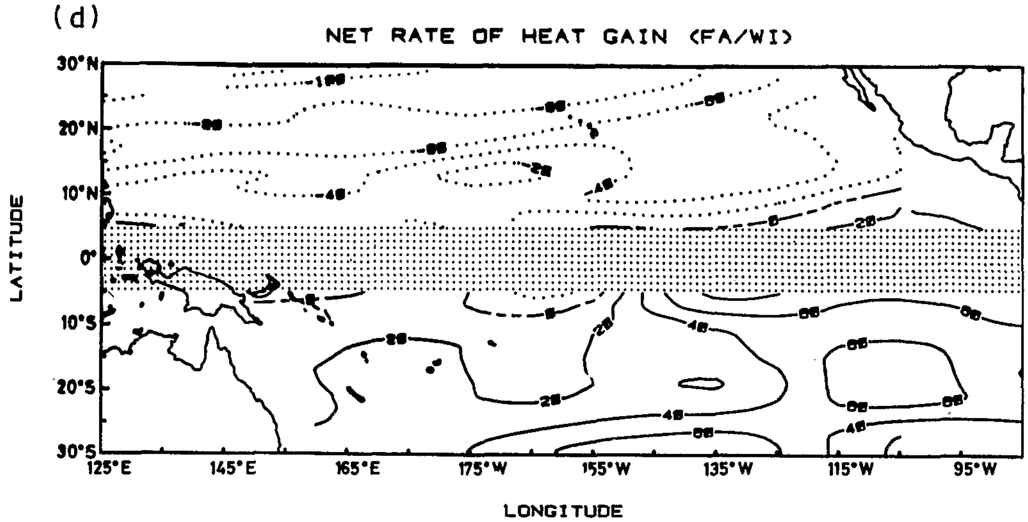
Fig. 2. Meridional sections along (150–160°W) of the significant terms of equation (2).

The patterns of net heating (Qn) computed by WEARE *et al.* (1981) and ESBENSEN and KUSHNIR (1981) are roughly in general agreement although they differ quantitatively near the equator in the eastern and western Pacific. They both differ quite significantly from those published by BUDYKO (1963), WYRTKI (1965) and HASTENRATH and LAMB (1979). Because a thorough comparison can be found in WEARE *et al.* (1981), the reader need only be reminded that the main difference is that the ocean gains heat in the north tropical Pacific according to WYRTKI (1965, Fig. 5) whereas it loses heat: (a) north of 10°N in the central Pacific according to WEARE *et al.* (1981) and (b) between about 2 and 10°N east of 110°W according to HASTENRATH and LAMB (1979). Considering the error (Fig. 1b) of the estimate of Qn, the present results are quite similar to those of Esbensen and Kushnir for the south tropical Pacific and north of 20°N. A poor agreement exists between 5 and 10°N west of 155°W where WEARE *et al.*'s (1981) calculations are more similar to the ones in this paper. Whether this difference is due to a real physical fact or from the approach or processing of the data field can not be said.

The most striking result of a comparison between the various estimations is that the present results show only a large region of net heat loss under the mean position of the ITCZ. Parts of this region present a good correspondence with Wyrtki's and Hastenrath and Lamb's calculations, the latter tending to show an even stronger cooling of the ocean east of 110°W. Therefore, this comparison suggests a difficult problem in the bulk formulas to determine the proper modification to take into account the effects of intense


(a)
NET RATE OF HEAT GAIN (WI/SP)
LATITUDE
LONGITUDE
(b)
NET RATE OF HEAT GAIN (SP/SU)
LATITUDE
LONGITUDE
(c)
NET RATE OF HEAT GAIN (SU/FA)
LATITUDE
LONGITUDE
30°N
20°N
10°N
0°
10°S
20°S
30°S
125°E
145°E
165°E
175°W
155°W
135°W
115°W
95°W

Figs 3a–c.



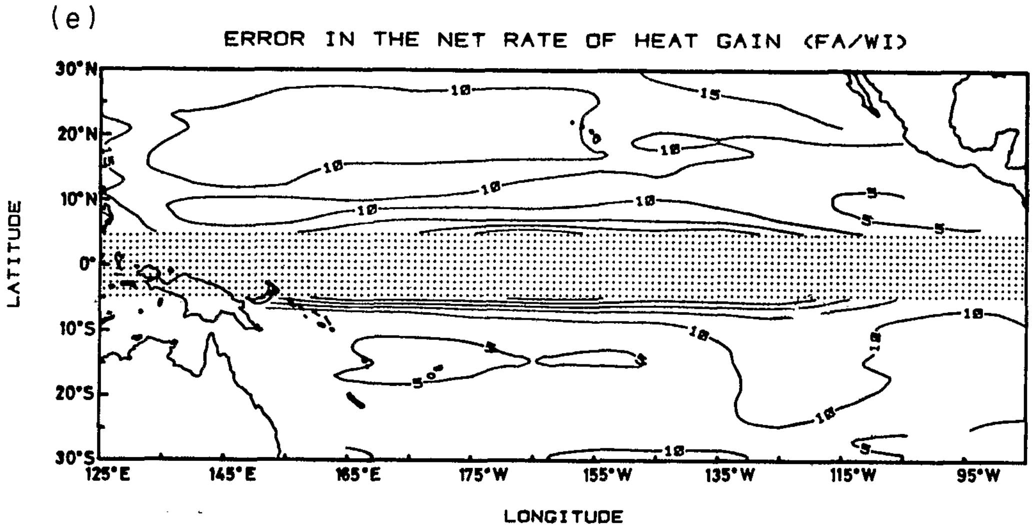


Fig. 3. (a) Net rate of heat gain by the ocean during the winter to spring (WI/SP) periods (W m^{-2}). (b) Net rate of heat gain by the ocean during the spring to summer (SP/SU) periods (W m^{-2}). (c) Net rate of heat gain by the ocean during the summer to autumn (SU/FA) periods (W m^{-2}). (d) Net rate of heat gain by the ocean during the autumn to winter (FA/WI) periods (W m^{-2}). (e) Error in the net rate of heat gain by the ocean during the autumn to winter (FA/WI) periods (W m^{-2}).

cloudy conditions associated with the ITCZ position. A thorough understanding of the effects of the ITCZ migrations upon the tropical Pacific Ocean circulation (the smoother of inequitable incoming solar radiation) might be attained by such considerations which, of course, require further heat budget studies.

The interseasonal net surface heating

The interseasonal net surface heat gains (Qn), several times larger than the annual one, are mapped on Figs 3a–d for the periods between the four seasons. Figure 3e shows the error estimates for the autumn to winter (FA/WI) period which is fairly representative of the error estimates for the other periods. *Qn* is maximum (minimum) during the interseasonal periods of maximum (minimum) incident solar radiation received in each hemisphere. This suggests that the interseasonal rates of heat gain or loss by the ocean are dominated by the effects of local heating or cooling. In other words, horizontal and vertical heat advections are not sufficient to account for the magnitude of the local heat storage that must be explained by local heat exchange with the atmosphere. Therefore, as previously demonstrated by GILL and NIILER (1972) for mid-latitudes, the interseasonal net surface heating seems predominantly local, but it is important to stress the oceanic heat transport contributions in particular areas. To this end, the various contributions of significant terms of equation (2) are presented in rectangles located in

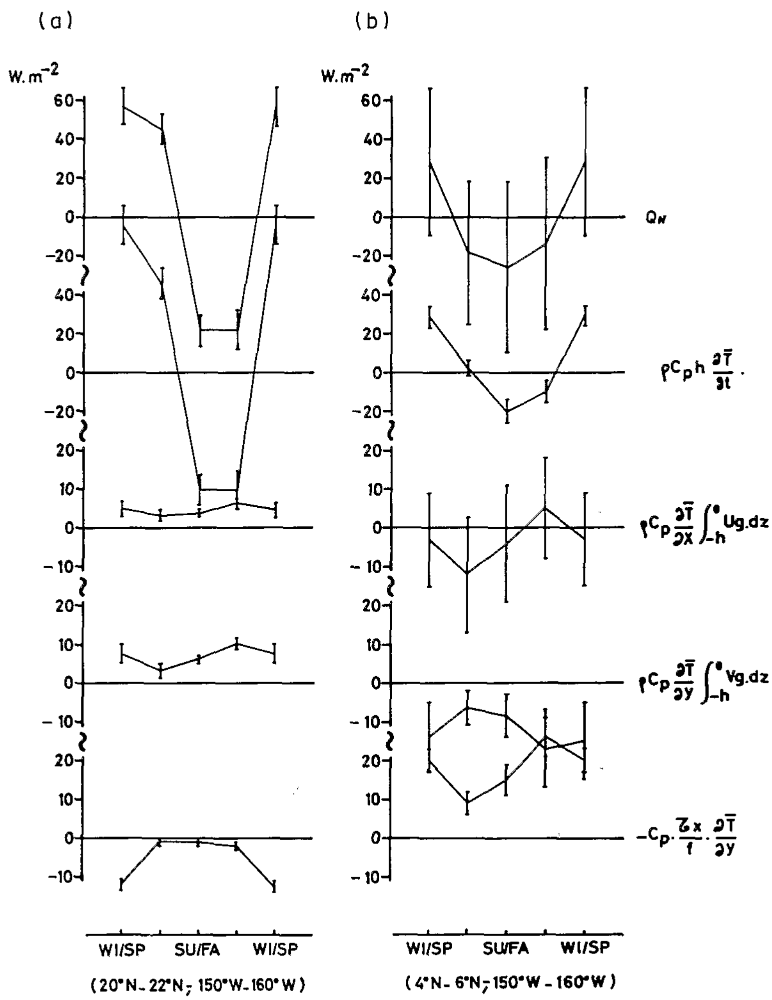


Fig. 4 (a) Contributions of terms of equation (2) in the (20–22°N; 150–160°W) rectangle during the four interseasonal periods. (b) Contributions of terms of equation (2) in the (4–6°N; 150–160°W) rectangle during the four interseasonal periods.

mid-latitude (20–22°N; 150–160°W) and close to the equator (4–6°N; 150–160°W). In each rectangle (Figs 4a, b), there is a strong consistency between the interseasonal variations in the local heat storage and in the net rate of heat gain. However, in mid-latitude (Fig. 4a), this similarity occurs because the oceanic heat transports are an order of magnitude less than the local heat storage, whereas, near the equator (Fig. 4b), these transports almost balance each other, and their contributions appear to be negligible. This is valid for the two 4° to 10° latitudinal bands (away from boundaries). The oceanic heat transport is thus crucial in determining the patterns of net heat gain by the ocean in these areas.

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