Observations of Air-Sea Interactions in the Western Pacific Warm Pool During WEPOCS

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I. Introduction

The meteorological observations made during the Western Equatorial Pacific Ocean Circulation Study (WEPOCS) show the complex nature of air-sea interaction occurring in the western equatorial Pacific warm pool. The winds in the core of the western Pacific warm pool are usually light (cf. Sadler *et al.*, 1987). During these low mean wind speed conditions, a radiative-convective equilibrium (cf. Sarachik, 1978; Betts and Ridgeway, 1988) seems to hold in which the strong coupling between atmosphere and ocean, controlled by local convective processes, provides a negative feedback that stabilizes the system with near-zero net heat flux. Recently, the importance of the hydrological cycle in this coupling has become apparent, both in observations (Lukas, 1989) and in simple models (Garwood and Chu, 1989; Stevens *et al.*, 1989).

Strong winds, associated with synoptic disturbances such as monsoon surges and westerly wind bursts, appear to be responsible for most of the transfer of heat from the western Pacific Ocean to the atmosphere. The character of the convection and air-sea interaction during these disturbances is different from the low mean wind speed conditions (Keen, 1988; Lander and Morrissey, 1988; Lau *et al.*, 1988; Lukas, 1988), involving spatial scales of thousands of kilometers, versus the 10-100 km scales seen during the low mean wind speed conditions. The transition from the local 1-dimensional balance to the fully 3-dimensional interaction of the ocean and atmosphere in the western equatorial Pacific is of particular interest, as the latter thermodynamics is relevant to ENSO onset.

II. Data

A. WEPOCS

WEPOCS is a joint U.S.-Australian program to study the low-latitude western boundary currents of the Pacific Ocean, and to study the response of the warm pool of the western equatorial Pacific to the Northwest Monsoon (Lindstrom *et al.*, 1987). Figure 1 shows the cruise tracks from WEPOCS I, II, and III, conducted in June-August 1985, January-February 1986, and June-July 1988 respectively.

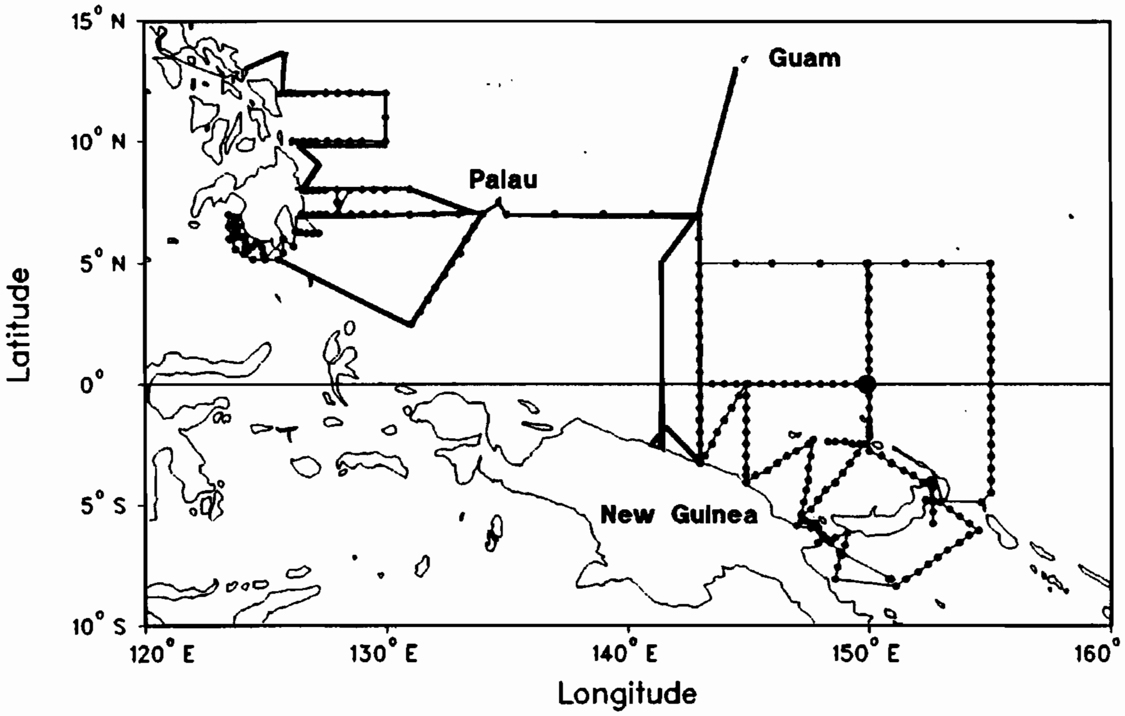

Figure 1. Cruise tracks from the three WEPOCS expeditions. The heavy line is the track of R/V MOANA WAVE during WEPOCS III (19 June - 30 July, 1988). Leg 1 is that portion to the east of Palau and north of New Guinea; the portion around Mindanao and west of Palau is Leg 2. The large sold circle indicates the location of the equatorial surface mooring during WEPOCS I, and the small sold circles indicate hydrographic station positions.

B. Observations

During WEPOCS III, hourly shipboard observations were made of sea surface temperature (SST) at 0.5 m depth, wet- and dry-bulb air temperature (from sling psychrometer at 3 m height), and wind speed and direction at 10 m (Fig. 2). Cloudiness observations were made every 3 hours and during hydrographic stations.

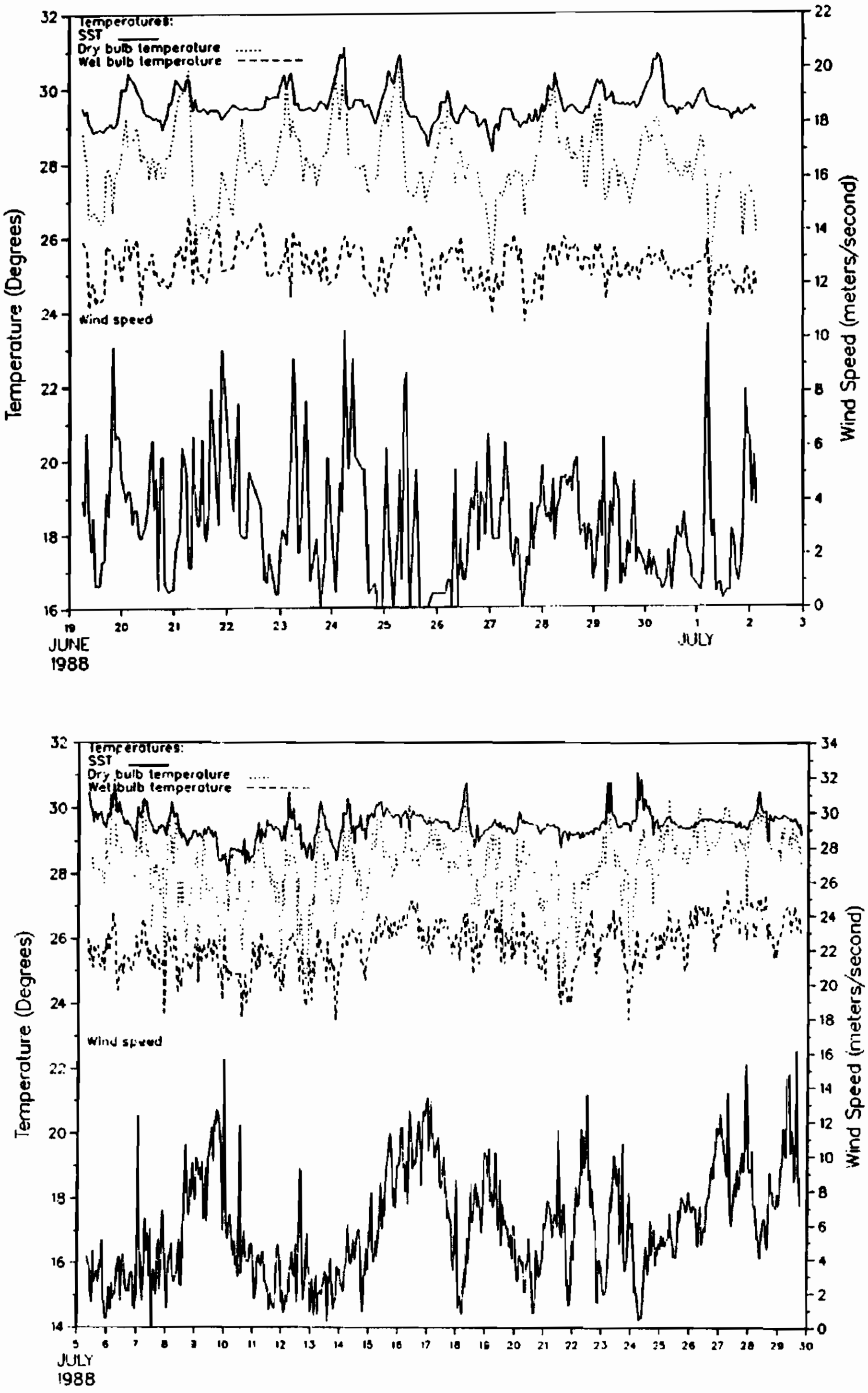

Figure 2. Hourly shipboard meteorological observations made during WEPOCS III, Leg 1 (top) and Leg 2 (bottom).

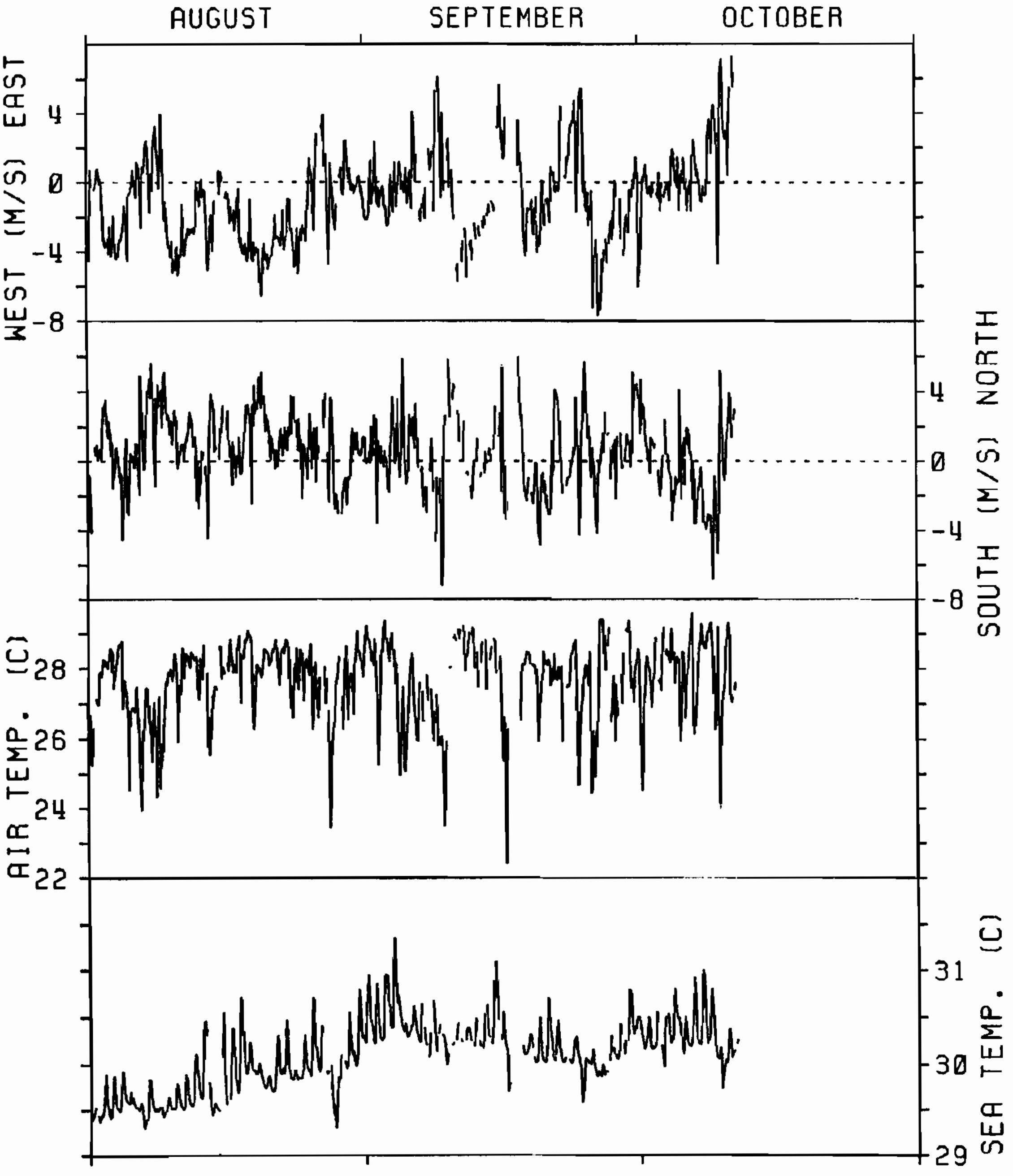

Figure 3. Time series of wind, air temperature, and sea surface temperature from the WEPOCS equatorial surface mooring of the equator at 150°E.

An Eppley full-sky radiometer was mounted on the ship's stern A-frame, and data were recorded at 1 minute intervals, but only a partial record was obtained because of data logging problems.

Wind and temperature measurements were also made from the surface buoy on the equatorial current meter mooring deployed during WEPOCS I in 1985 (Fig. 3). Winds and air temperature are from a height of 3 m, and SST is from 1 m below the surface.

During WEPOCS II and III, R/V MOANA WAVE was equipped with a thermosalinograph in the intake to the cooling pump for the hydrographic winch. The water was drawn continuously from a depth of 4 m. Seabird temperature and conductivity sensors were used, and the system sampled every 20 seconds. After editing, and calibration of salinity against water samples, the data were smoothed and subsampled at 1 minute intervals.

C. Heat flux computations

The components of the net heat flux were computed from the hourly meteorological observations using the bulk formulae as per Stevenson and Niiler (1982) and Large and Pond (1982), and with direct measurements of solar radiation. When daily averages of these fluxes were compared to fluxes calculated using daily averaged variables (Table 1), the differences were generally small. Thus, heat fluxes from daily means are presented below, using cloudiness observations to estimate insolation because the radiometer records were not complete. Note that the net insolation parameterization used in these estimates was checked against the radiometer data and the calculated insolation is biased high by -26 W/m^2 for the 24 days of data. The insolation and net heat flux were not corrected for this bias. Also, low wind-speed modifications (Liu *et al.*, 1979) were not made to the exchange coefficients. These computations will be made in the near future.

Table 1.

Difference of heat flux components and net heat flux for daily averages calculated as the flux from daily-averaged meteorological variables minus the daily-averaged fluxes calculated from hourly observations.

day	ΔQ_h	ΔQ_e	ΔQ_b	$\Delta Q(Wm^{-2})$
172	-4.7	6.3	.6	-2.2
173	-2.3	4.4	.2	-2.3
174	-.9	11.8	2.0	-12.8
176	-4.9	8.8	.4	-4.2
177	-.5	4.7	.0	-4.2
179	-.3	2.1	.2	-2.0
180	.1	1.2	.2	-1.6
181	-3.9	1.3	.1	2.5
182	2.9	2.0	.1	-5.0
183	-3.1	5.4	-.4	-1.8
189	.1	-.5	-.9	1.3
190	.8	-2.0	-2.8	4.0
192	-1.1	7.7	.9	-7.5
193	4.6	4.4	-.2	-8.8
194	-6.1	1.9	-1.9	6.1
195	2.7	2.0	.2	-4.9
196	5.4	1.4	.6	-7.4
197	-2.5	1.3	-.6	1.8
198	-3.1	1.1	-.3	2.3
199	-7.2	1.3	1.1	4.9
200	2.0	6.6	-.7	-7.9
201	-8.5	1.0	.9	6.5
203	.4	5.6	-.2	-5.8
204	2.4	6.7	.4	-9.5
205	.9	8.8	.4	-10.1
206	-1.5	4.5	-.6	-2.5
207	-1.2	-.1	-.1	1.5
208	-1.8	-4.6	-1.0	7.4
210	-.2	1.2	.3	-1.3
avg	**-1.08**	**3.33**	**-.05**	**-2.20**
std	**3.31**	**3.57**	**.91**	**5.38**

III. Results

A. Heat Fluxes

During Leg 1, the winds were usually less than 4 m/s, except during convective downdrafts and squalls (Fig. 2). Note the very high frequency variations in these hourly data, associated with convective activity, that can not be resolved with twice or even four times per day sampling. The mean air-sea temperature difference was 1.6°C during this period, and was as large as 3-4°C during the downdrafts. This large mean air-sea temperature difference is three times larger than climatological estimates for this region (Weare *et al.*, 1981; Esbensen and Kushnir, 1981). This is probably due to biases in merchant ship temperature observations; large air-sea temperature differences have been observed on all three WEPOCS expeditions and from the WEPOCS moored buoy in the warm pool.

The time series from the WEPOCS mooring shows the light mean winds (speed less than 4 m/s most of the time, and variable direction) that are characteristic of the warm pool (Fig. 3). The air temperature is usually in the range 27-29°C, with negative excursions (attributed to convective downdrafts) below 24°C on occasion. The SST shows a strong (0.5°C amplitude) diurnal cycle during light winds (cf. Ostapoff and Worthem, 1974), and SST is usually warmer than the air temperature by 1.5-2.0°C (Lukas, 1987).

During Leg 2, the calm conditions of the warm pool region were disturbed by several surges of the Southwest Monsoon (Fig. 2). During these periods, the large air-sea temperature difference is reduced (cf. 25-30 July), and the diurnal cycle is supressed.

The insolation and latent heat are the largest terms in the net heat flux, and they also have the largest variability (Fig. 4). The sensible heat flux is proportional to the air-sea temperature difference, which varied diurnally but not as much from day to day. Large negative excursions of dry air temperature are associated with convective gust fronts (squall lines) in which the air is cold and saturated. This air is brought rapidly to the surface in downdrafts inside the convective towers. Because of the large air-sea temperature difference at these times, a considerable sensible heat flux from the ocean to the atmosphere is occurring in the region of such downdrafts (eg. Gautier, 1978; Bean and Reinking, 1978).

Both wind and humidity changes were responsible for modulating the daily mean latent heat fluxes during Leg 1 of WEPOCS III, where the mean wind speed was less than 4 m/s. The net heat flux over 11 days was 40 W/m^2, with excursions between +130 and -30 W/m^2 for the daily values. During Leg 2, several surges of the Southwest Monsoon resulted in winds of 8-14 m/s for 2-3 days at a time, and an average wind speed of about 6 m/s. The large latent heat losses (200-270 W/m^2) and reduced insolation during the monsoon surges resulted in net heat losses

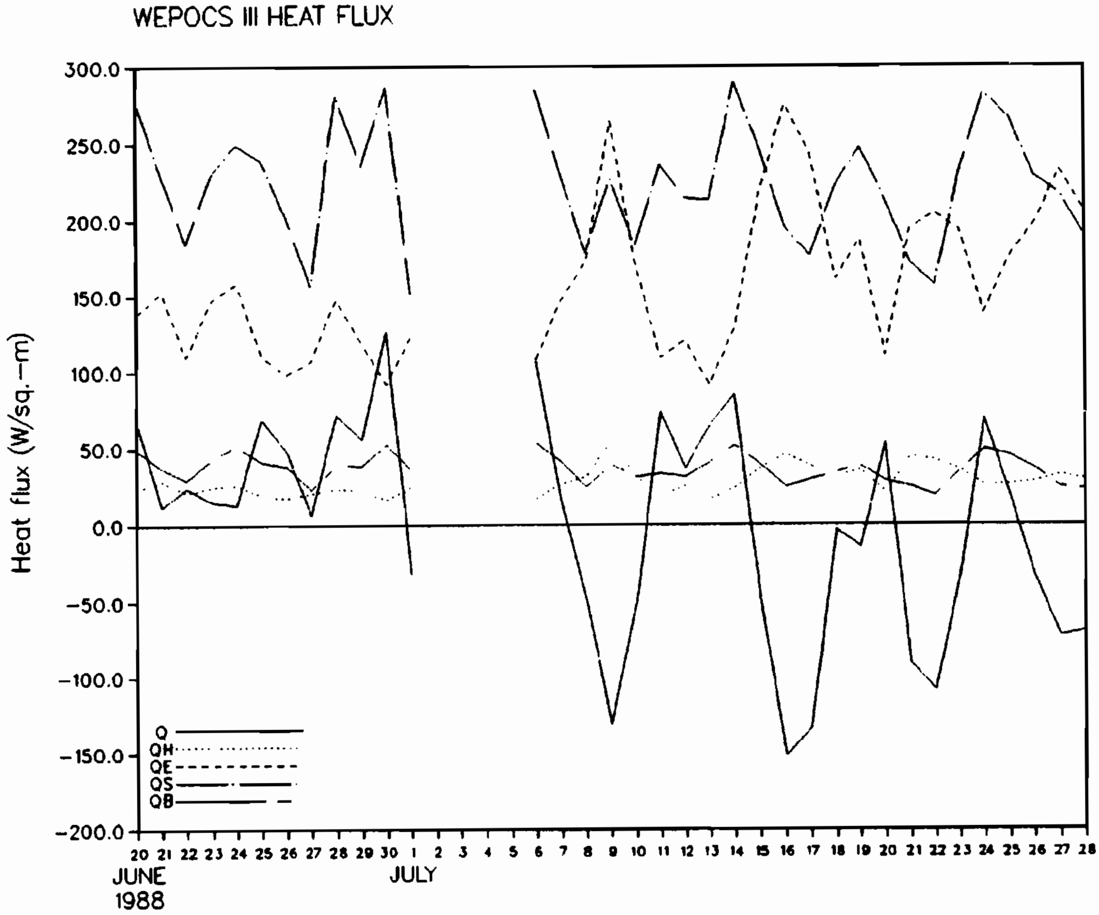

Figure 4. Heat fluxes computed from daily mean meteorological variables observed during WEPOCS III. See text for description.

from the ocean of 100-150 W/m^2 during these periods, and an average over 23 days of -20 W/m^2. Note that these net fluxes may be biased high by 20-30 W/m^2 because the algorithm used to calculate insolation from daily cloud coverage is probably not applicable to the cloud types found in the western tropical Pacific.

B. Precipitation signatures

The thermosalinograph record from 21 July 1988 (near 8°N, 128.5°E), shows the effect of the strong precipitation and convective downdrafts, which is seen in the freshening of sea surface salinity, and the cooling of SST (Fig. 5). The width of this cold, fresh puddle is about 20 km, comparable to the scale of the convection cell. It is not clear whether the cooler water is due to cold rain, or to heat being extracted from the ocean by cold downdrafts, although previous observations (Greenhut, 1978) suggest the latter, which is consistent with the 1-dimensional radiative-convective equilibrium theory.

Temperature and salinity from the thermosalinograph on R/V MOANA WAVE during WEPOCS III Leg 1 show the aggregate effect of the mesoscale convection on the upper ocean temperature and salinity fields (Fig. 6). Individual convective cells freshen and cool the upper ocean as indicated by the heavy vertical lines. The

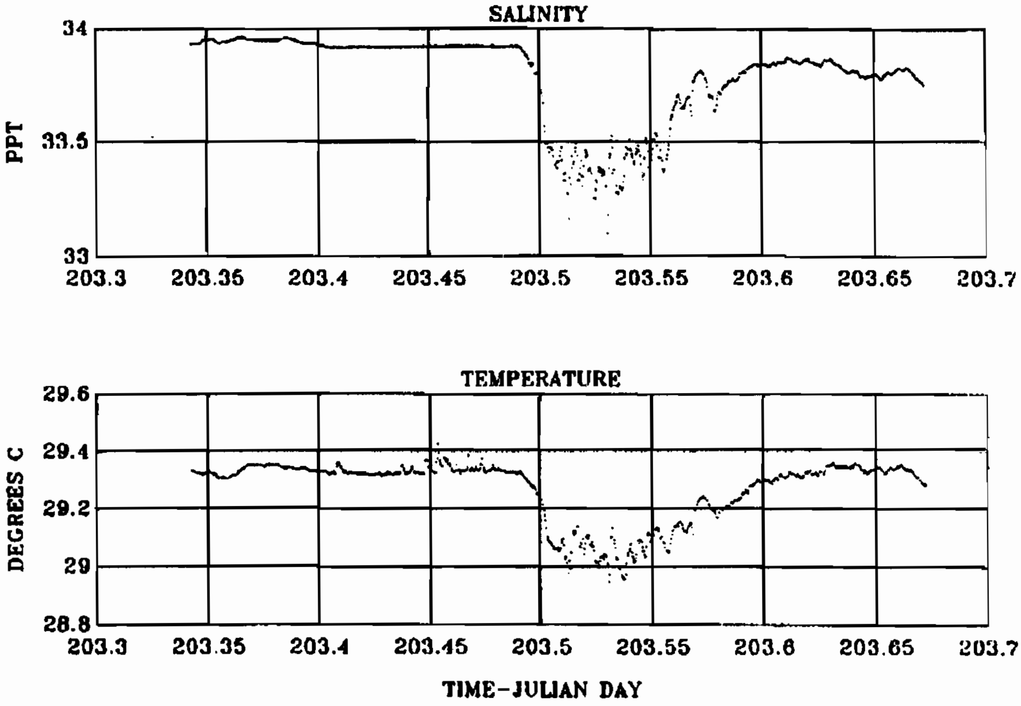

Figure 5. Segment of WEPOCS III thermosalinograph record (salinity at top, temperature at bottom) showing the depression of salinity and temperature as R/V MOANA WAVE passed through a convective cell. Approximate length scale associated with this large freshwater input is 20 km.

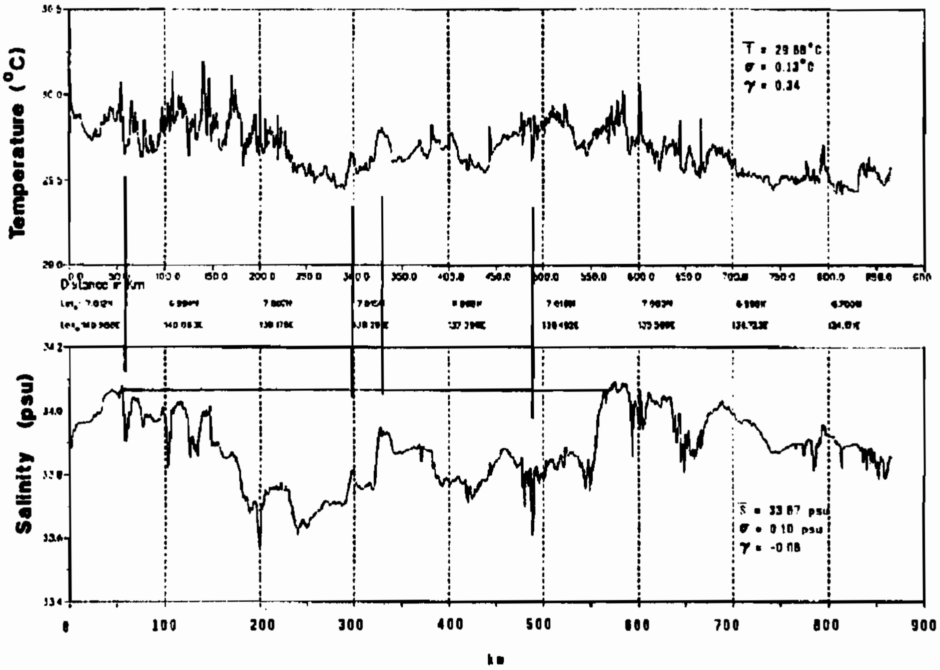

Figure 6. A two-day segment of the thermosalinograph record from R/V MOANA WAVE during the WEPOCS III expedition showing temperature (top) and salinity (bottom) at a depth of 4 m versus distance (km) along track. Note the general depression of salinity between 50 km and 550 km, as well as the more localized depressions (eg. 250-300 km).

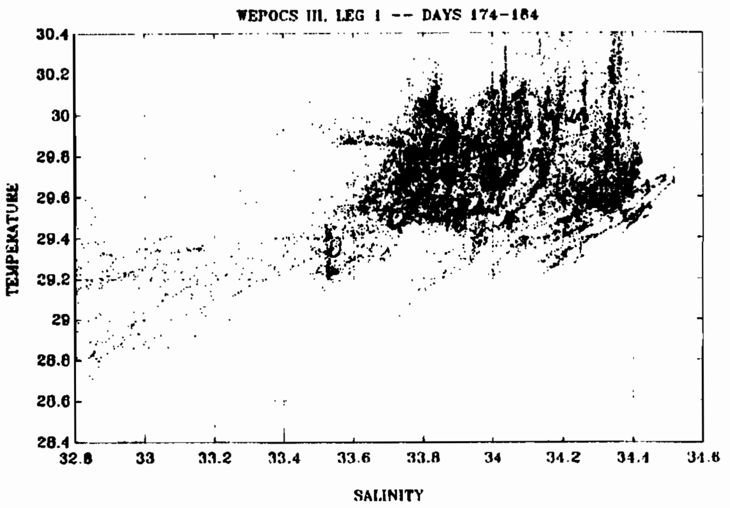

Figure 7. Temperature-salinity diagram for thermosalinograph data from WEPOCS III for the same period as the observations in Fig. 6. Note the general tendency for low salinity to be associated with lower temperature.

freshening of the sea surface between 100 and 550 km along track is attributable to a mesoscale convective complex of this spatial extent. The diurnal cycle in SST is seen in the apparent long wavelength variation.

A surface temperature-salinity diagram (Fig. 7) shows the strong correlation of SST and surface salinity variations, with cold, fresh "puddles" from mesoscale convection mixing with warmer, saltier waters. This good correlation suggests that the buoyancy forcing of the warm pool is dominated by mesoscale convection during the time of these observations. Elliot (1974) did not find such a good correlation during BOMEX, probably because the relative influence of wind and buoyancy forcing was so different from the western equatorial Pacific.

IV. Discussion

From the WEPOCS observations, it is clear that air-sea fluxes in the warm pool are dominated by processes associated with mesoscale convection. Thus, it is appropriate to review these processes here. The view of the tropical convective cell that arose from the intensive GATE observations is summarized by Bean and Reinking (1978). The complexity of the circulation and the vertical fluxes is substantial; a complete description is given by Leary and Houze (1979).

The heaviest rain occurs in the convective tower, but when weighted by area, the stratiform anvil contributes as much as 50% of the precipitation. Within the tower, strong downdrafts occur leading to the formation of gust fronts, and because of the rapid sinking, this air remains cold. Because of the very thick and dark cloud in the tower, incoming solar radiation is severely attenuated. In the regions outside of the convective cell, the cloud cover is generally thin stratocirrus cloud, permitting solar radiation to reach the sea surface without much attenuation. This region is where cooler, drier air from the outflow regions of the convective cells returns to the near-surface layers (Betts and Ridgway, 1988).

The incoming solar radiation is primarily modulated by the diurnal cycle, and by the clouds associated with mesoscale convection. When the winds are light, and the skies are clear, the SST responds rapidly to the heating (Ostapoff and Worthem, 1974), as does the overlying atmosphere. This can be seen in Figs. 2 and 3, where the diurnal cycle of SST is as large as 1°C, and the diurnal cycle of air temperature is about 2°C.

According to bulk parameterizations, the latent heat flux depends on the vapor pressure difference between air at the sea surface temperature and that of the air above the sea surface, and on the wind speed. For the case of no wind, the bulk formula predicts no moisture flux. Liu *et al.* (1979) suggest that this is not the case, and that the exchange coefficient in this formula varies strongly with wind speed at low wind speeds, especially when the air-sea temperature difference is large. The subsiding air outside the convective towers is relatively dry (see the large difference between dry bulb and wet bulb air temperatures in Fig. 2), and is thus capable of picking up substantial moisture.

The heat flux depends on SST, but changes of SST depend strongly on mixed layer thermodynamics. The momentum and buoyancy fluxes at the sea surface, combined with the initial buoyancy profile, determine the evolution of the mixed layer and its heat budget. Thus, precipitation and salinity effects, through their influence on the buoyancy flux and the buoyancy profile, may exert a control on the mixed layer heat budget (Ostapoff *et al.*, 1973; Miller, 1976), and thus on the flux of heat to the atmosphere.

V. Conclusions

The fluxes of heat and moisture in the western Pacific warm pool are generally controlled by mesoscale convection, with a quasi-one-dimensional radiative-convective equilibrium in which there is near-zero net heat flux on time scales longer than about one day (Godfrey and Lindstrom, 1989). This equilibrium is occasionally disturbed by synoptic scale atmospheric events; it is during these times that heat is effectively extracted from the warm pool by the atmosphere (Meyers *et al.*, 1986). The horizontal scales of SST and sea surface salinity variability are controlled by the convective processes and the synoptic scale forcing.

Scale interactions occur when processes on at a particular time or space scale modulate the energy of processes on different time or space scales. This can arise only from nonlinear thermodynamics. The turbulent fluxes of heat, moisture, and momentum between the upper ocean and the lower atmosphere are nonlinear, and are dependent on one another. Synoptic scale forcing modulates the atmospheric convection, and thus substantially modulates the air-sea fluxes (eg. Seguin and Kidwell, 1980). It is this synoptic atmospheric forcing which can change the warm pool system from a nearly one-dimensional radiative-convective equilibrium into a fully three-dimensional system.

Acknowledgements

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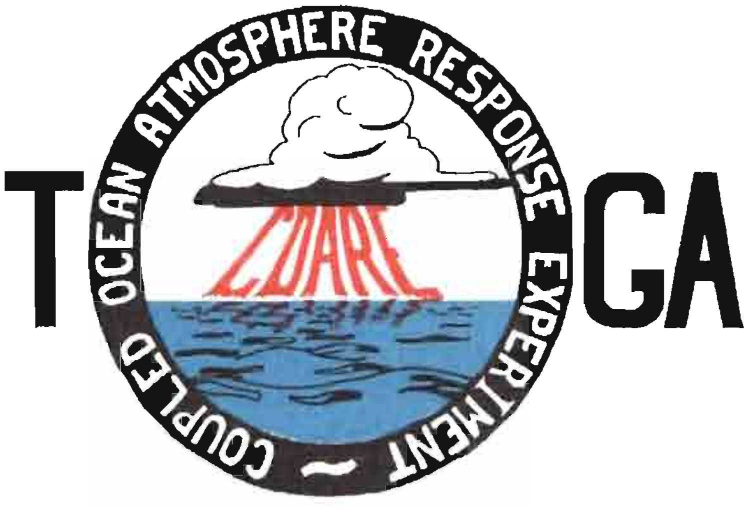

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