



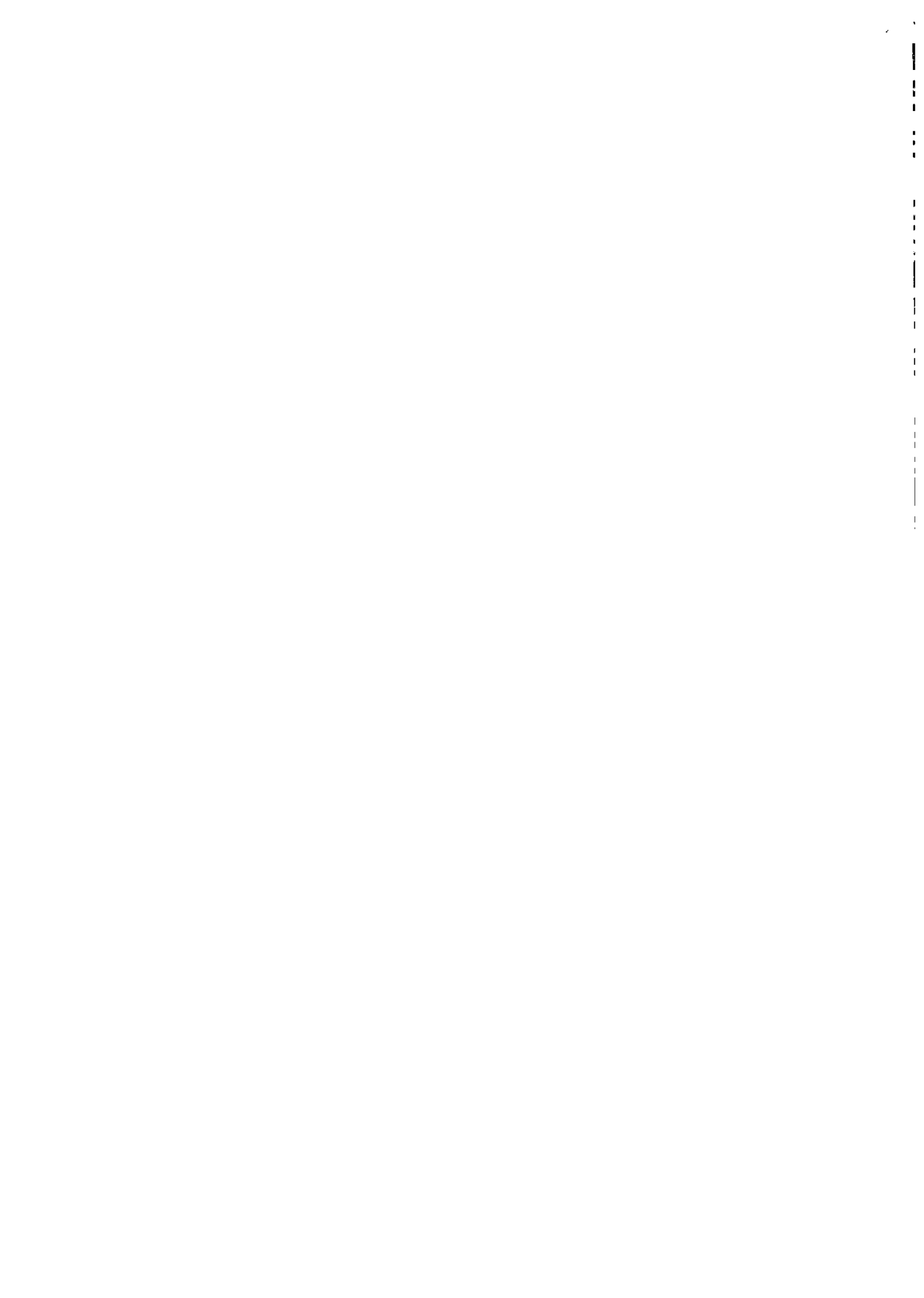
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***"A Coupled Air-Sea Biennial Mechanism in the Tropical Indian
& Pacific Regions: Role of the Ocean"***

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A Coupled Air–Sea Biennial Mechanism in the Tropical Indian and Pacific Regions: Role of the Ocean

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ABSTRACT

Variations in the upper-ocean heat content are part of a mechanism to explain biennial signals in the tropical Indian and Pacific ocean regions. The mechanism involves modulations of the annual cycle of convection and processes of the dynamically coupled ocean and atmosphere system in the tropics. A critical element of that mechanism is persistent sea surface temperature anomalies on the time scale of one seasonal cycle. Analyses of composite vertical temperature profiles from hydrographic station data for various near-equatorial areas in the Indian and Pacific oceans show that variations in the ocean heat content depend on the depth of the thermocline in the warm-pool region (both eastern Indian and western Pacific) and temperatures in the upper-ocean mixed layer away from the warm pool (western Indian and eastern Pacific). The variations in thermocline depth in the Pacific are similar to those for El Niño–Southern Oscillation (ENSO) events and are present in the biennial mode as well. Apparently, similar sets of mechanisms operate on both ENSO and biennial time scales. These results suggest that changes in upper-ocean heat content contribute to the persistence of sea surface temperature anomalies important to the biennial mechanism, that both the Indian and Pacific oceans are actively involved in ENSO, and that ENSO could be an amplification of the biennial cycle.

1. Introduction

Meehl (1987) identified a biennial signal in the coupled ocean–atmosphere system in the tropical Indian and Pacific regions and proposed a mechanism to explain those signals that involved large-scale interaction between the ocean and atmosphere. Trenberth (1975) had postulated this type of interaction as important for biennial signals in the area north of New Zealand. Nicholls (1978) formulated a set of equations that produced a biennial oscillation dependent on air–sea interaction. He suggested that these types of processes could explain biennial signals in the Indonesia–North Australia region. Brier (1978) conceptualized these ideas to explain how biennial signals could arise at a given location in an interactive ocean–atmosphere system. Subsequently, Nicholls (1979, 1984) has shown further evidence for the importance of air–sea coupling and the annual cycle for a biennial oscillation and how such an oscillation could be a part of the Southern Oscillation.

The mechanism proposed by Meehl (1987) extended the earlier work and expanded the space scales to in-

volve circulation anomalies over the entire expanse of the tropical Indian and Pacific oceans. Of critical importance were the interaction of the annual cycle with the large-scale east–west atmospheric circulation, the annual movement of a convective maximum that was strongest at any given location during one season, and persistent sea surface temperature (SST) anomalies on annual time scales. Since then, a number of other studies have further documented biennial signals in these regions. For example, Kiladis and van Loon (1988) documented large anomalies in the tropical Indian and Pacific regions associated with extremes in the Southern Oscillation that had biennial characteristics. Lau and Sheu (1988) and Rasmusson et al. (1990) noted biennial and lower-frequency signals associated with the Southern Oscillation in these same regions. Both papers also documented a tendency for phase locking to the annual cycle for the biennial signals. Meehl (1987) provided a plausible physical explanation for this linkage of the biennial oscillation to the annual cycle, but Barnett (1991) questioned whether this could explain observed quasi-biennial variations. However, Barnett (1991) did not rule out that possibility and suggested that the relationship of the biennial oscillation with the annual cycle could be intermittent. As will be seen shortly, this is a possibility since the years used by Meehl (1987) to document biennial signals are a subset of the total number of years available for analysis. However, the years in the composites to be analyzed shortly involve roughly two-thirds of the available years

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and represent a biennial mechanism linked to the seasonal cycle that is in evidence in some form or other a significant part of the time. In the present study, Meehl's (1987) mechanism is reviewed to provide a context for an analysis of vertical ocean temperature profiles from hydrographic station data to document the role of heat storage in the ocean.

2. The biennial mechanism

The possible role of upper-ocean heat content in the biennial mechanism is illustrated schematically in Fig. 1. The basic premise is that at any given location in the tropical Indian and Pacific regions air-sea coupling is strongest during one season of the year. The rest of the year is relatively quiescent for ocean-atmosphere interaction on the annual time scale. The season of greatest air-sea coupling coincides with the time of year when local convection is strongest. The passage of this convective maximum corresponds to regional monsoon regimes (India, Australia) or when the intertropical convergence zone (ITCZ) is otherwise overhead. Meehl (1987) showed that at any location in this vast area, the local convection is relatively strong during one season, even in near-equatorial locations where the ITCZ passes twice a year. (At these locations, it is possible that there are two times of year—the equinoxial seasons as the ITCZ crosses the equator—when

air-sea coupling could be strong enough to alter the anomalies in the Indian and Pacific regions, but this prospect will be left to a subsequent study.)

To follow the biennial cycle at a given location, begin in Fig. 1 at the top frame during year 1 (stippled area). Assume that the SSTs start out relatively high at this location. When the convective maximum comes overhead, the warm water is associated with greater evaporation and low-level moisture convergence and, consequently, stronger convection. As this season progresses, convection intensifies with stronger low-level winds and greater heat loss and mixing in the ocean, and the upper ocean cools. At the end of this season, the convective maximum moves on and relatively cool water remains in its wake. Because of the strong air-sea coupling during this season, the vertical temperature structure of the ocean is significantly altered, thus reducing heat stored in the upper ocean. This then contributes to the persistence of cool water through the rest of the year (Fig. 1, left frame).

A year later (year 2 in Fig. 1, bottom frame), the convective maximum again comes overhead at this location. During this season, however, relatively cool water remains from the previous year and results in less evaporation and low-level moisture convergence and weaker convection. As the season progresses, the weaker convection is associated with weak low-level winds and less heat loss and mixing in the ocean. By

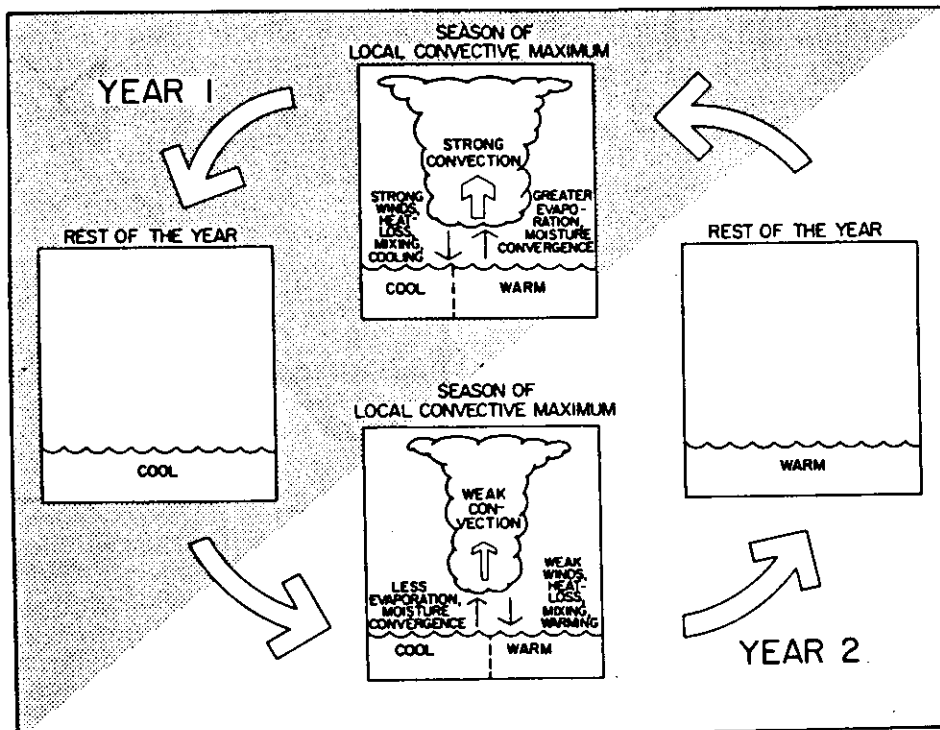


FIG. 1. Schematic illustration of the biennial mechanism of Meehl (1987) at a given location in the tropical Indian and Pacific regions.

the season's end, as the convective maximum moves on, relatively warm SSTs are left in its wake with associated increases of heat storage in the ocean. This warmer water then persists through the rest of the year (Fig. 1, right panel) until the convective maximum returns one year later and the biennial cycle starts over.

The interactions between ocean and atmosphere act as a simple switch—with the ocean first acting on the atmosphere via reinforcing or weakening convection via SST anomalies, and then the atmosphere acting on the ocean as the processes associated with convection alter the state of the ocean. Conceptually, the coupled system can be thought of as acting on SST as follows:

$$T'_i = -aT'_{i-1}, \quad (1)$$

where T' is the SST anomaly, i is the present year, and a is an air-sea coupling coefficient that is either zero or one. On the annual time scale in this scheme, $a = 0$ for most of the year and $a = 1$ during the one season when air-sea coupling is strongest. The SST of the present year T'_i is determined by the SST left over (or set up) from the previous year T'_{i-1} , acted on by the sign switch involved with the coupling coefficient a . The coefficient a is nonzero only during the season when the convective maximum is overhead at any given location. This conceptually implies that, for a perfectly biennial system, SSTs would always be either relatively warm or relatively cool compared to previous and/or following years. Therefore, the actual long-term mean SST would simply be the average of the warm and cool states.

To study this proposed mechanism, Meehl (1987) chose a long-term time series of Indian monsoon precipitation as an index tied to processes taking place

over a very large area in the tropical Indian and Pacific regions (due to well-known linkages associated with the Southern Oscillation). The time series of monsoon precipitation from Verma et al. (1984) was used to identify relatively "strong" and "weak" monsoon years. This was done by choosing years when monsoon rainfall was less than the year before and the year after (a weak year) and vice versa for a strong year. In effect, this is equivalent to running a high-pass filter over the annual data to capture the year-to-year variability characteristic of the Indian monsoon. Figure 2 shows the strong and weak years (S and W above the years) and the warm and cold events (WE and CE below the years), as defined by van Loon (1984) and van Loon and Shea (1985). Indicative of the biennial signal in the system, a number of years have consecutive strong and weak monsoons, and very few strong and weak events occur singly. The weak years tend to be warm events and strong years cold events, but there are many other strong and weak years that are neither warm nor cold events. There is also a tendency for some warm and cold event sequences to be biennial. This suggests that the Southern Oscillation is operating with the same set of mechanisms as the biennial oscillation and that extremes in the Southern Oscillation—the warm and cold events—could be exaggerations of the biennial oscillation.

Meehl (1987) showed that the mechanism outlined in Fig. 1 can plausibly explain some aspects of signals of SST and sea level pressure (SLP) in the tropical Indian and Pacific. In an analysis of results from a coupled GCM, Meehl (1990) notes that a somewhat different set of processes may be involved in the far eastern Pacific, but the basic premise of strong air-sea

"STRONG" OR "WEAK" BASED ON RELATIVE INDIAN MONSOON RAINFALL

S=STRONG		W=WEAK				CE=COLD EVENT				WE=WARM EVENT																				
S	W	S	W	S	W	S	W	S	W	S	W	S	W	S	W															
1900	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30
		WE	CE	WE		CE		CE		WE		WE		CE		WE		CE		WE		CE		WE		CE		WE		WE

S	W	S	W	S	W	S	W	S	W	S	W	S	W	S	W	S	W													
1930	31	32	33	34	35	36	37	38	39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60
		WE	CE	WE			CE	WE		CE			CE		WE		WE		CE		WE		WE		CE		WE			WE

W	S	W	S	W	S	W	S	W	S	W	S	W	S	W									
1960	61	62	63	64	65	66	67	68	69	70	71	72	73	74	75	76	77	78	79	80	81	82	83
		WE	CE	WE	CE		WE	CE		WE	CE		WE	CE		WE	CE		WE			WE	

29 STRONG, 17 PRECEDED BY WEAK, 10/16 COLD EVENTS (ONE WE)
 30 WEAK, 23 PRECEDED BY STRONG, 14/19 WARM EVENTS (TWO CE)

FIG. 2. Relatively strong and weak Indian monsoon years and cold and warm events, 1900-83.

coupling during one season of the year is the same. Coupled processes in the eastern Pacific are under study and will be the subject of a subsequent paper.

One aspect of the mechanism critical at all locations that Meehl (1987) did not address was ocean-heat storage. Certainly, for the extremes of the Southern Oscillation (warm and cold events), changes of slope of the thermocline in the Pacific, with attendant alterations of heat storage, are important (e.g., Wyrtki 1985). During warm events, the thermocline shallows in the western Pacific and deepens in the east and vice versa for cold events. This is associated with an attendant decrease and increase of ocean heat storage. Meehl (1987) suggests that similar heat-storage changes could be at work for the biennial mechanism in the Pacific, as well as in the Indian Ocean. This possibility will be explored in the analyses to follow.

3. Data

Vertical ocean temperature profiles are compiled from observed hydrographic station data [a reliable subset of the observations used to produce the Levitus (1982) ocean atlas] for three separate areas in the equatorial Pacific, as well as one in the Banda and Arafura seas north of Australia and two in the Indian Ocean (Fig. 3). Most of the hydrographic stations were Nansen casts taken from the National Oceanographic Data Center station-data archive. The station data were screened for quality control by Levitus for his atlas.

The hydrographic station data are not evenly distributed in time. Figure 4 shows the number of available observations for each area in Fig. 3 as a function of season. Observations in the eastern Indian and Pacific oceans tend to be clustered mostly in the 1950s, 1960s, and 1970s, while the data from the other three areas in the western Pacific are more scattered throughout the period. North Indonesia has the most available hydrographic station data, with observations spread from the late 1920s to the mid-1970s.

To compare the characteristics of the upper-ocean structure, annual mean temperature profiles are formed for all available data from each area. Four representative long-term mean area-averaged profiles are shown in Fig. 5. The equatorial eastern Pacific is characterized by low SSTs (about 23.5°C), a very shallow mixed

layer (order 30 m), and a correspondingly shallow thermocline (from about 30 to 75 m). In the western Pacific, the north New Guinea profile is representative of the other area-averaged profiles in the far western Pacific. Mean SSTs are greater than 29°C with a well-mixed layer in the upper 50–75 m. The thermocline is deep and extends from about 75 to 250 m. The eastern Indian profile from an area in the eastern Indian warm pool is similar to the one in the western Pacific except that the thermocline is not quite as deep and extends from about 50 to 200 m. The western Indian Ocean area lies outside the warm pool and has a shallower mixed layer than the eastern Indian Ocean area.

4. Interannual variability

Figure 6 shows composite strong-minus-weak SST differences from Meehl (1987) through the annual cycle from July, October, and January to April. The SST data are from the Comprehensive Ocean–Atmosphere Data Sets (COADS). The sign of the anomalies is that of a strong annual cycle (strong Indian monsoon, cold water in the eastern Pacific, warm water in the western Pacific; in the extreme, a cold event). Superimposed on the SST differences are hatched areas indicative of strong convection. In this case, the hatching highlights areas of long-term mean outgoing longwave radiation (OLR) less than 220 W m⁻² (from Janowiak et al. 1985).

The maximum of convection in Fig. 6 moves from India in July, to southeast Asia in October, to northern Australia in January, and back north of the equator and into the ITCZ in the Pacific in April. In a strong year, the convection is more intense through the entire year along its track west of the South Pacific convergence zone (SPCZ), extending diagonally from northwest in the equatorial western Pacific to southeast in the subtropical Pacific (Meehl 1987). The convection feeds on energy from the warm water in its path and is positively reinforced through increased latent and sensible heat fluxes. As noted in Fig. 1, the stronger convection at a location is associated with stronger heat fluxes, enhanced mixing in the ocean, and cooling of the water. As the convective maximum moves on through the seasonal cycle, it leaves SSTs of opposite sign (negative in this case) in its wake. By northern

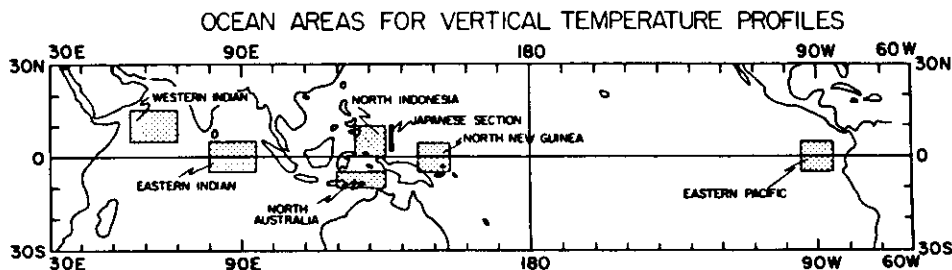


FIG. 3. Areas chosen for hydrographic station data composites (stippled areas). Japanese section along 137°E from Yasunari (1990) is also shown.

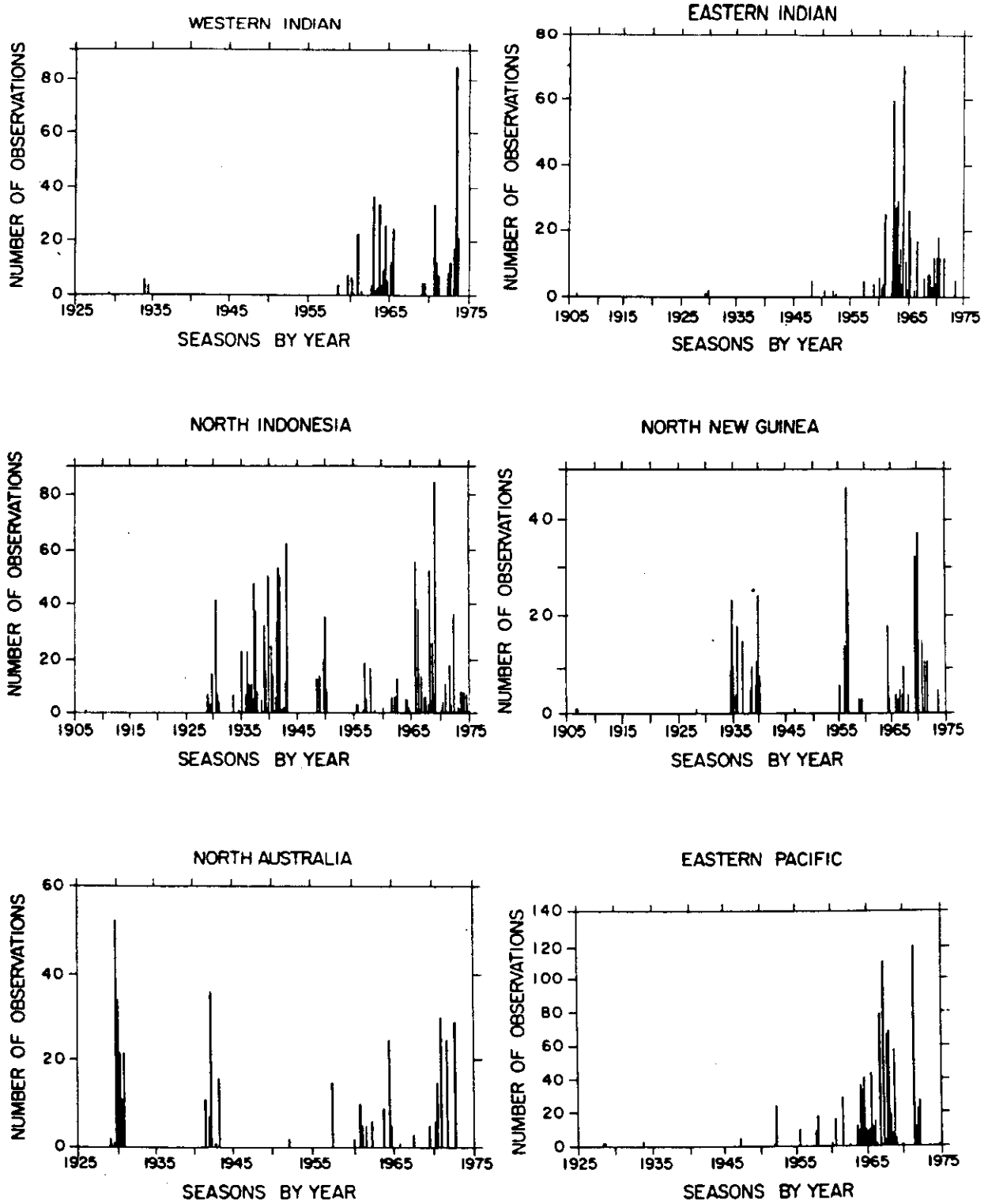


FIG. 4. Number of observations by season from the Levitus (1982) hydrographic station data for each area in Fig. 3.

spring, the entire system west of the SPCZ has been "reset" (cool water), as the SST anomalies in the eastern Pacific make the transition from negative to positive.

The subsequent seasonal cycle (in a perfectly biennial system) then has SST anomalies of opposite sign to those in Fig. 6 with weakened convection west

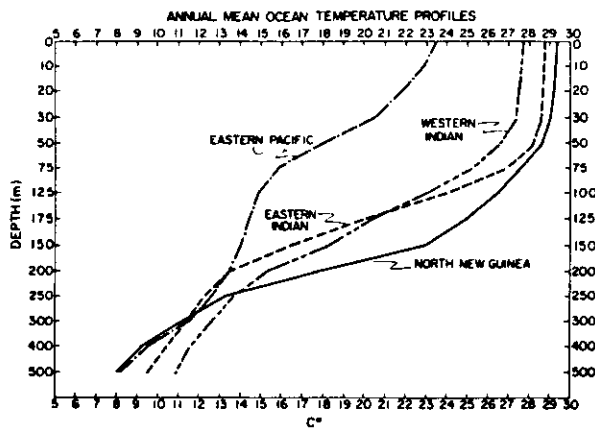


FIG. 5. Long-term annual mean ocean vertical temperature profiles for three representative areas—eastern Pacific, north of New Guinea, and eastern Indian Ocean (refer to Fig. 3 for location).

of the SPCZ. These anomalies are negatively reinforced along its path during the seasonal cycle by the residual cool water from the previous year, leaving SSTs of the opposite sign (positive in this case) in its wake. Concurrently, positive SST anomalies are in evidence in the eastern Pacific (a warm event in the extreme). By the end of that seasonal cycle, the system would then be reset for a strong annual cycle the following year (warm SSTs west of the SPCZ, cool water to the east of the SPCZ) and so on.

In the present study, annual means of the vertical ocean temperature profiles for each area in Fig. 3 are based on the duration of the 12-month period when SST anomalies of one sign are persistent at a given location. This depends on either the season when the convective maximum passes overhead (Indian and Pacific ocean regions west of the SPCZ), or when the transition of sign of SST anomalies occurs due to processes in the eastern Pacific (Fig. 6). Since the patterns of SST are seasonally evolving at each location (Fig. 6), it is useful to choose one time of year as a reference and formulate differences in relation to it. The averaging periods here are arbitrarily chosen to correspond to the SST anomaly pattern during a strong seasonal cycle in October (i.e., Fig. 6b). At this time in a strong seasonal cycle, the relatively strong convective maximum is over western Indonesia and there is cool water in the Indian Ocean, warm in the western Pacific, and cool in the eastern Pacific. The duration of the annual mean averaging period to produce the correct phase of the seasonal evolution to key on this time of year is given for the areas at the top of each plot in Fig. 7.

All available years are composited on the basis of strong and weak monsoon years in Fig. 2 and averaged for each area. Because of the disparities in the spatial and temporal distribution of the hydrographic station data in Fig. 4, none of the areas is quantitatively comparable, but all provide semiautonomous checks on the others for qualitative consistency of the signals. The

hydrographic stations also provide an independent check on the SST results from the COADS data in Fig. 6. A data cutoff of 70 hydrographic stations per composite is used. Less than this number and uneven seasonal distributions begin to contaminate the results. The exception is the area north of Australia for the composites without warm and cold events. A fortuitously good seasonal distribution of the station data in this area allows inclusion of these results, but it should be noted that fewer hydrographic stations were used for that one composite.

Assessing statistical significance of the composite differences is difficult for the same reasons listed above. For example, if standard deviations are computed for all members of a given space-time composite, the composite differences are usually greater than one standard deviation at each level. However, such standard deviations include both spatial and temporal variability. Therefore, for the general purposes of this paper, comparison to COADS SST data and to the other observed features of vertical temperature structure will be noted where possible. As will be seen, the signals are consistently large enough to compare favorably with other sources in spite of the coarse spatial and temporal resolution of the data. A more quantitative evaluation must await better time-series data from oceanographic moorings. Such data are now being collected for the tropical Pacific (e.g., McPhaden and Hayes 1991).

Figure 7 shows temperature differences in the upper 500 m of the ocean, strong-minus-weak composites, as a function of depth for the six areas. For all strong-minus-weak composites (solid lines, including the extremes of the Southern Oscillation, the warm and cold events), the two areas in the western Pacific (North Indonesia and North New Guinea, Figs. 7b and 7e) show small positive SST differences consistent with the sign of the COADS SST data for a strong year in those locations (Fig. 6). But a larger positive signal is evident at the depth of the thermocline near 100–150 m (Fig. 5). In the ocean north of Australia (Fig. 7c), the thermocline also seems to be affected in the same sense as in the western Pacific but with somewhat larger signals of SST.

As expected from El Niño–Southern Oscillation (ENSO) observations, the thermocline in the western Pacific moves up and down, deeper during strong years (often cold events), as evidenced by the larger positive temperature differences at thermocline depth, and shallower during weak years (often warm events). Figure 8 shows this phenomenon for the actual composite profiles during strong and weak years for the north New Guinea area. However, if the extremes of the Southern Oscillation (the warm and cold events) are removed, the same signal appears (dashed lines in Fig. 7), indicating, as with SST and SLP documented by Meehl (1987), that similar processes are taking place for the biennial oscillation. This suggests that the extremes in the Southern Oscillation are an exaggeration

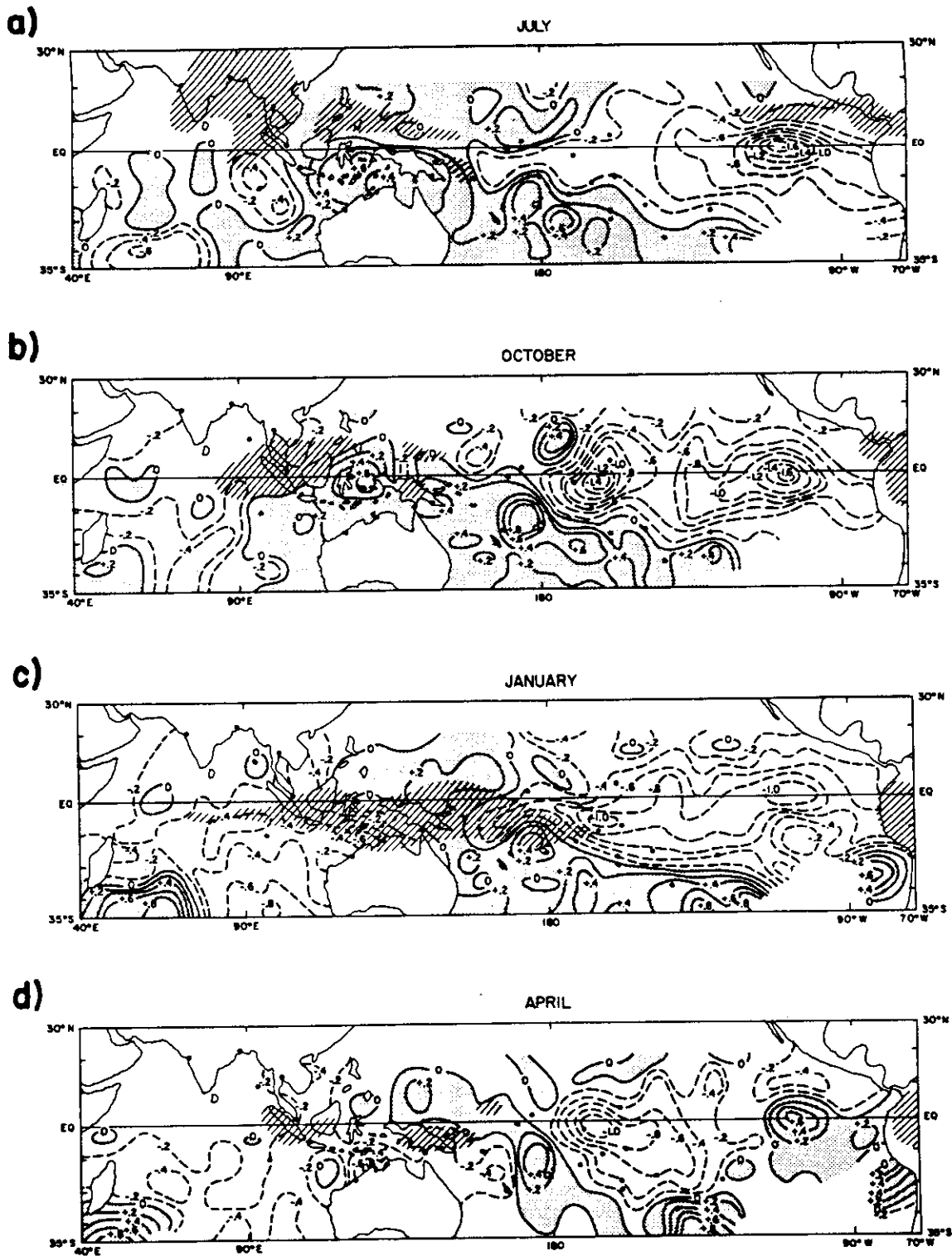


FIG. 6. Long-term mean OLR values less than 220 W m^{-2} indicative of areas of heaviest convection (hatching) and SST differences, strong years minus previous years for (a) July, (b) October, (c) January, and (d) April (SST values after Meehl 1987). OLR values from Janowiak et al. (1985).

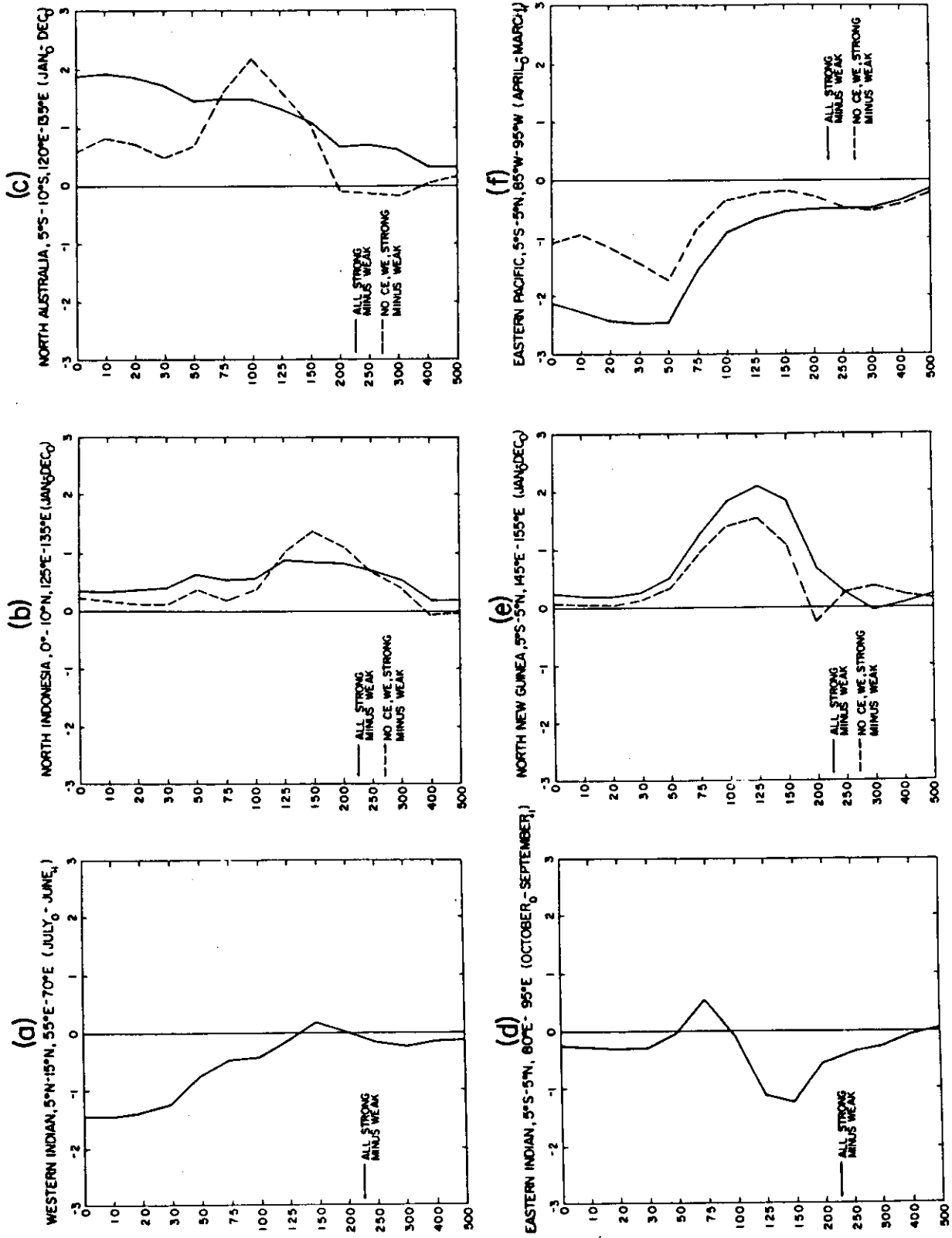


FIG. 7. Strong-minus-weak composite ocean vertical temperature profiles for all strong-minus-weak years (solid lines) and strong-minus-weak years without warm and cold events (dashed lines) for areas shown in Fig. 3: (a) western Indian, (b) north India, (c) north Australia, (d) eastern Indian, (e) north New Guinea, and (f) eastern Pacific. Latitudinal and longitudinal extents of the areas, as well as the duration of the annual mean averaging intervals, are given at the top of each frame.

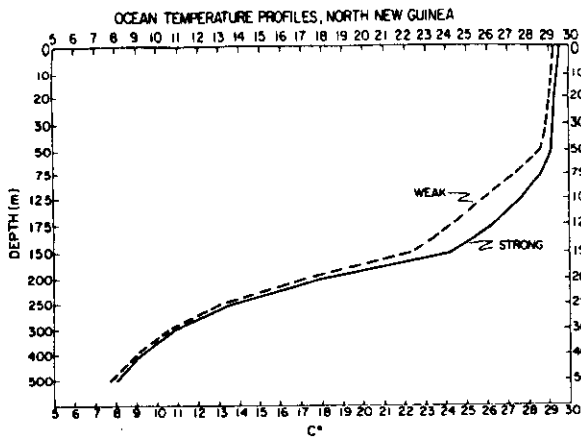


FIG. 8. Average composite temperature profiles ($^{\circ}\text{C}$) for strong (solid) and weak (dashed) years for the north New Guinea area. The long-term mean profile (not shown) lies roughly midway between the two lines.

of the biennial oscillation operating with the same mechanism.

For the eastern Pacific (Fig. 7f), the thermocline is shallow (Fig. 5), and the signal for the strong-minus-weak years is evidenced by lower SSTs extending through the depth of the mixed layer to around 50–75 m. This is also consistent with the COADS SST data and could be expected from observations of the extremes in the Southern Oscillation (e.g., McPhaden and Hayes 1990) since strong years are often cold events. As in the western Pacific, if the warm and cold events are removed from the composites, a similar signal still appears with smaller amplitude.

The area in the eastern Indian Ocean (Fig. 7d) shows a large signal in the variation of thermocline depth, in the same sense as the smaller-amplitude SST anomalies that are negative during a strong annual cycle in this region for the time of year in the seasonal evolution (again consistent with the COADS SST data in Fig. 6). Because of insufficient numbers of hydrographic stations, it is not possible to show results for composites of years without warm and cold events for this area. This is also the case for the area in the western Indian Ocean (Fig. 7a). Most of the temperature change in the profile of composite differences for all years for that area occurs in the mixed layer in the upper 100 m, similar to that seen for the eastern Pacific (Fig. 7f).

The results from the vertical ocean temperature profiles in Fig. 7 show that the sign of the SST change for all the areas is consistent with the COADS SST data from Fig. 6. Also, there are large changes in heat storage in the upper ocean that could contribute to the persistence of ocean temperature anomalies on annual time scales. For example, the change in the vertical integral of temperature in the upper 500 m for all strong-minus-weak years for the north New Guinea area is $+270.35^{\circ}\text{C m}^{-1}$. For the strong-minus-weak years ex-

cluding warm and cold events, the change is $+181.30^{\circ}\text{C m}^{-1}$ for that area. A large contributor to these changes in heat storage is the depth of the thermocline. If, for example, only the average temperature change in the upper 50 m of the ocean is used to calculate the change in the vertical integral of temperature (presumed to be in the upper 500-m ocean column), the change for all strong-minus-weak years is $+135.00^{\circ}\text{C m}^{-1}$ for north New Guinea, about half the value of $+270.35^{\circ}\text{C m}^{-1}$ that includes the temperature change in the thermocline.

To look at the changes in upper-ocean heat storage and SST in more detail, we compute the rate of change of heat storage Q as a function of depth h and time t by

$$Q(h, t) = \frac{\partial}{\partial t} \int_0^h c_p \rho T(z) dz,$$

where c_p is specific heat of seawater, ρ is density, and T is temperature. It must be kept in mind that this equation accurately describes Q for short time-averaging intervals. But for longer-term composited time averages, the magnitude of Q is likely to be underestimated. Therefore, the present calculations represent qualitative changes in the mixed layer. Other processes are likely to come into play at depths below the mixed layer to 500 m. Additionally, Stevenson and Niiler (1983) showed that heat storage calculated between an isotherm in the upper ocean and the surface more closely corresponds to the variation of SST. We will return to this aspect shortly.

The rate of change of heat storage Q for $h = 500$ m for the strong-minus-weak composites over the period of one year for the north New Guinea area is 35.9 W m^{-2} . The strong and weak years in Fig. 8 roughly straddle the long-term annual mean (not shown) so that the change of heat storage for only the strong years in relation to the long-term mean is about 18 W m^{-2} . If it is assumed that most of this change is taking place during one season, this number is about 72 W m^{-2} . This same quantity for the years without warm and cold events is about 48 W m^{-2} . As noted above, these quantities are likely to be underestimated.

However, considering the upper mixed layer as noted above by Stevenson and Niiler (1983), fluctuations of heat storage from the mean have been calculated for the ocean layer above the depth of the 26°C isotherm to be as high as about 70 W m^{-2} for 1 month (Meyers et al. 1986, their Fig. 2f). In Fig. 5, the depth of the 26° isotherm in the mean is about 110 m. If all of the annual rate of change of heat storage calculated for the north New Guinea area for the depth of the upper ocean above 110 m took place in 1 month, that value would be 84 W m^{-2} . This is near the extreme range of observed fluctuations of heat storage for this area of the western Pacific, as noted above. More typical observed fluctuations near 40 W m^{-2} are closer to the

present computed monthly value of 56 W m^{-2} for strong years without cold events. Meyers et al. (1986) conclude that the surface latent heat flux due to anomalous wind forcing is the primary contributor to the observed fluctuations of SST and heat storage to the depth of the 26° isotherm in the upper ocean. McPhaden and Hayes (1991) came to a similar conclusion for the same region. They noted typical heat-content changes of 35 W m^{-2} in the upper 75 m, with evaporative cooling being the most important process contributing to these fluctuations. This is consistent with the processes proposed earlier to take place in association with intense atmospheric convection during a strong annual cycle that would act to remove heat from the mixed layer and lower SSTs.

5. Conclusions

The role of ocean-heat content in the biennial mechanism of Meehl (1987) is explored with vertical temperature profiles in the upper 500 m of the ocean from hydrographic station data. The thermocline moves up and down in the western Pacific, as could be expected from ENSO studies, and a similar thermocline movement exists for the biennial time scale. That is, a strong Indian monsoon is associated with a deeper thermocline and warmer water in the western Pacific and a shallower thermocline and colder water in the eastern Pacific (vice versa for a weak Indian monsoon with warm water in the eastern Pacific). The eastern Indian Ocean exhibits changes in thermocline depth comparable to the western Pacific. If the extremes in the system—the warm and cold events associated with the Southern Oscillation—are removed, similar signals are still evident in the ocean temperature profile data for the Pacific areas. The data are insufficient to form similar composites for the Indian Ocean areas consistent with the signals of SST documented by Meehl (1987). The differences in upper-ocean temperature structure are associated with changes in upper-ocean heat content. These changes in the western Pacific are consistent with observed fluctuations of ocean heat storage in the mixed layer forced by evaporative cooling from anomalously strong winds. However, these processes that affect the mixed layer cannot explain all of the changes in temperature noted in the observed temperature profiles down to 500 m. The areas outside the warm-pool region (western Indian and eastern Pacific) have shallower thermoclines and show the largest temperature-change signal in the mixed layer above 100 m. Changes in ocean-heat content could contribute to the persistence of SST anomalies in these regions on annual time scales. This points to the important role of the ocean in the biennial mechanism of which the extremes of the Southern Oscillation appear to be an amplification.

Since these results are derived from different temporal distributions of hydrographic station data, the

quantitative aspects must be treated with some caution. Qualitatively and in spite of notably different temporal distributions, the results are consistent with the COADS SST data and certainly deserve further examination with better time series of ocean temperature profiles. Such consistent results from the sparse observations suggest that the signals noted here are fairly robust. Vertical ocean temperature profile data from the Japanese section along 137°E (location in relation to areas in the present paper is shown in Fig. 3) are characterized by a similar strong biennial signal of thermocline depth and heat content for the period 1972–85 (Yasunari 1990). Masumoto and Yamagata (1991) postulate that coupled processes involving Ekman pumping and vertical thermocline movement combine to produce the biennial signal in that region. Lukas and Lindstrom (1991) have also noted important coupled processes involving westerly winds that can deepen the mixed layer in the western equatorial Pacific. The present study shows that coupled processes extend over much larger areas to produce the biennial signal. The results from the Indian Ocean are especially intriguing and suggest that the entire tropical Indian and Pacific regions are involved in the tropical biennial and Southern oscillations.

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