Tropical–Midlatitude Interactions in the Indian and Pacific Sectors of the Southern Hemisphere*

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ABSTRACT

Observations and three General Circulation Model (GCM) simulations are analyzed to study interannual processes and interactions involved in the tropical Indian and Pacific sectors and higher southern latitudes. Major features involving the observed eastward progression of the tropical convective maximum from the Indian monsoon in northern summer to the Australian monsoon and Pacific in southern summer are represented in all three model simulations. This suggests that the distribution of land and sea accounts for the observed seasonal position of the convective maximum in these regions. However, the South Pacific convergence zone (SPCZ) is not present at any time of year in the slab-ocean models due to the anomalously warm sea-surface temperatures simulated in the equatorial Pacific.

Analyses of observed zonal mean 500-mb temperatures and sea-level pressure from 10° to 80°S for Warm-Event composites minus Cold-Event composites (representing two extremes of interannual variation) show the period of anomalies at high southern latitudes to be about half a year out of phase with the tropics. Poleward migration of observed anomalies is most apparent in the Pacific sector, suggesting a delayed communication between the tropics and high latitudes. A tentative physical explanation is postulated that involves the SPCZ as a conduit for communicating tropical anomalies to higher southern latitudes by changes in the dynamically coupled ocean–atmosphere system operating in the tropics and midlatitudes of the Pacific. If such processes are taking place in the real climate system, the tropics and high southern latitudes are not in equilibrium and immediate transmission of tropical anomalies to high southern latitudes may not occur. The GCM simulations under consideration here, which are run to near-equilibrium, therefore, are not capable of portraying such features of observed interannual variability. These conclusions point to the necessity of using a coupled ocean–atmosphere GCM capable of simulating realistic air–sea interactions to study such interannual and latitudinal linkages in the Indian and Pacific sectors of the Southern Hemisphere.

1. Introduction

A previous paper (Meehl, 1987) examined the observed annual cycle in the tropical Indian and Pacific sectors. It was noted that an area of intense convection, termed the convective maximum, moves from the northwest in the Indian summer monsoon to the southeast in the Australian monsoon and South Pacific convergence zone (SPCZ) during the mean annual cycle from May to the following April. Modulation of this convective maximum was observed to be associated with Warm and Cold Events in the Southern Oscillation with similar, albeit smaller, signals occurring in many other years. Warm Events are usually characterized by relatively higher sea-level pressure (SLP) west of the SPCZ and lower SLP to the east. With the lower SLP east of the SPCZ, higher sea-surface temperatures (SSTs) are associated with greater rainfall there. Opposite conditions prevail during Cold Events (e.g., van Loon and Shea, 1985). The entire ocean–atmosphere system in the tropical Indian and Pacific regions is involved in producing intraseasonal and interannual variations (Meehl, 1987).

Several studies have focused attention on tropical/midlatitude linkages in the Southern Hemisphere (SH). Simpson and Downey (1975) modeled the SH response to a midlatitude SST anomaly. They noted circulation changes both equatorward and poleward of the anomaly in different averaging periods. In analyzing the mechanisms in the SH midlatitudes that were responding to a tropical Pacific SST anomaly, Palmer and Mansfield (1986) concluded that the zonally symmetric changes in the SH were related to increased extratropical baroclinic instability associated with the high tropical SST anomalies in the model. Kaynay et al. (1986) postulated that mechanisms involved with the coupled ocean–atmosphere system were important in the SPCZ and South Atlantic convergence zone in regard to tropical/midlatitude circulation changes. In

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a case study, Huang and Vincent (1985) showed that the SPCZ played an important role in SH circulation. A GCM study by Von Storch et al. (1988) showed the sensitivity of SH circulation to SST anomalies in the vicinity of the SPCZ.

This paper examines the observations further for evidence of interannual variability in the extratropical Indian and Pacific sectors that may be associated with the modulations of the annual cycle. Analyses concentrate on the dynamically coupled air–sea system in these same sectors as described by Meehl (1987) for the tropics. In addition, three GCM simulations with different boundary forcings are examined for similar features to provide insight into processes that may be taking place in the real climate system.

In section 2, the three GCM simulations are described in more detail, and observed data sources are reviewed. In section 3, the annual cycle of the convective maximum in the tropics is examined in terms of precipitation from long-term mean observations and the three GCM experiments. In section 4, the annual evolution of zonal mean anomalies in the SH from 10° to 80°S is presented for the two extremes of the Southern Oscillation—observed Warm and Cold Events. Similar analyses are shown in section 5 for the GCM experiments. Discussion follows in section 6, and conclusions in section 7.

2. Observed data and GCM simulations

Long-term monthly means of precipitation are from Jaeger (1976). The 500-mb temperatures for the Southern Hemisphere (SH) are archived at the National Center for Atmospheric Research (NCAR). They are the Australian analyses on 5° × 5° grids covering the period of May 1972 to May 1984. The SH SLP data are also archived at NCAR and digitized on 5° × 5° grids. Two periods are represented—January 1951 to December 1958 from the South African analyses and May 1972 to May 1984 from the Australian analyses.

The three GCM experiments analyzed in this paper involve the same atmospheric GCM coupled to two different ocean surfaces. The atmospheric model, a version of the NCAR Community Climate Model (CCM), includes calculations for soil moisture, snow cover, cloudiness/radiation, convective adjustment and annual cycle. The model is a global, spectral GCM with realistic geography, nine layers in the vertical and rhomboidal 15 truncation which yields approximately a 4.5° latitude by 7.5° longitude resolution. Further description of the atmospheric model can be found in Washington and Meehl (1984).

The first experiment is run with the GCM driven by the annual cycle of observed SSTs and is denoted the specified sea-surface temperature or “SPEC SST” experiment. The model is run for five years, and averages are taken for the final three-year period. The second and third simulations involve the atmospheric GCM coupled to a simple mixed-layer ocean model. SSTs are computed by the model from surface-energy balance and crude seasonal heat storage. The ocean is a slab of water 50-m deep with no provision for currents, upwelling, or annual variation of depth. As described by Meehl and Washington (1985), the simple 50-m mixed layer roughly accounts for the annual cycle of ocean heat storage. But they also show that compensations in the surface-energy balance, which result in part from the lack of ocean heat transport and upwelling, produce SST anomalies as high as those observed during Pacific Warm Events. These are associated with other altered circulation features in the tropics, such as upper-level wind anomalies that are reminiscent of observed Warm Events (Meehl and Washington, 1986). The GCM, with the present amount of atmospheric carbon dioxide (CO₂) coupled to the simple slab-ocean model, is denoted MIX1. The third GCM experiment is run the same as the MIX1 case except that atmospheric CO₂ is doubled. This run is referred to as MIX2. Averages for MIX1 and MIX2 are taken over a 3-year period at the end of a 12-year integration, which was preceded by two previous periods with an accelerated annual cycle to enhance convergence toward equilibrium. This method is described in Washington and Meehl (1984). In the context of the present paper, these GCM simulations are interpreted as alternate climate states with different boundary forcings. As such, they are not direct analogues but are informative for what they tell us about physical processes in the climate system.

Basic features of the MIX1 and MIX2 runs have been documented by Washington and Meehl (1984). Further characteristics of the MIX1 run, compared to observations, are shown by Meehl and Washington (1985, 1986). Meehl (1984) describes some features of the SPEC SST simulation compared to observations.

3. Annual cycle of tropical precipitation

The successive southeastward positioning of the convective maximum, described by Meehl (1987), can be traced in monthly mean precipitation maps in Fig. 1 which shows observed precipitation from Jaeger (1976). The stippled areas highlight heavy precipitation, beginning in Fig. 1a with the Indian monsoon region in July around 80°–100°E north of the equator. In Fig. 1b, an area of regionally heavy precipitation is located to the southeast of the maximum in Fig. 1a. It covers much of Southeast Asia from about 15°N to 10°S and from roughly 85° to 120°E. During January (Fig. 1c), an area of heavy precipitation is farther southeast, covering parts of Indonesia and northern Australia associated with the Australian monsoon from about 90° to 150°E and from near 15°N to 25°S. In April (Fig. 1d), heavy precipitation occurs farther east in the tropical Pacific from about 130°E to the Dateline in the SPCZ and Intertropical Convergence Zone (ITCZ). The Solomon Islands region (near 10°S,
Fig. 1. Annual cycle of the progression of an area of observed heavy precipitation denoted by stippling for (a) July, (b) October, (c) January, (d) April. Note that some regions receive heavy precipitation year-round (e.g., southwest of Sumatra near 5°S, 95°E, and the Solomon Islands region near 10°S, 160°E). However, the stippled area highlights the annual maximum of precipitation in those regions.

160°E) experiences heavy precipitation year-round, but rainfall is greatest there at this time of year. The precipitation maximum is stronger in its eastward movement from July to January than in its westward return from January to July in the area west of about 160°E. This can be seen by comparing the area near Sumatra
(an island in the region of the equator and 100°E) in October (Fig. 2b) and April (Fig. 2d). Precipitation amounts exceed 10 mm day\(^{-1}\) in October, but are just above 8 mm day\(^{-1}\) in April. Meehl (1987) attributes this to the west–east circulations between the Indian and Pacific sectors. Stronger convection and precipitation in the Pacific sector are associated with suppressed activity in the Indian sector and vice versa. This affects convection and precipitation near 120°E during southern fall as the ITCZ returns northwestward from the northern Australian region to India. At this time of year, relatively greater rainfall associated with the convective maximum becomes established between about 140°E and the Dateline, and rainfall to the west is suppressed.

Washington and Meehl (1984) and Meehl (1984) show geographical distributions of precipitation averaged for December–February (DJF) and June–August (JJA) computed from the three GCM simulations as well as for observed values for the tropical Indian and Pacific regions. The seasonal maxima in the Asian monsoon regions are represented reasonably well by all three model simulations, but the position and magnitude of the precipitation in the tropical Pacific are significantly affected by the SST pattern. The warm SST anomalies in the tropical equatorial Pacific, documented by Meehl and Washington (1985, 1986) for the mixed-layer ocean model, are associated with a southward shift of the ITCZ and a northeastward displacement of the SPCZ. The result is a band of heavy precipitation across the near-equatorial Pacific in both of the GCM runs coupled to the slab mixed-layer ocean. To examine the seasonal evolution of selected areas of relatively intense precipitation for these model runs, isolaths of monthly mean precipitation for July, October, January and April are plotted in Fig. 2. The 8 mm day\(^{-1}\) isolaths for the observations (shaded areas in Fig. 1) are plotted for the four months in Fig. 2a. In Fig. 2b–d, 10 mm day\(^{-1}\) isolaths are shown in a similar manner for the three model simulations. Because precipitation is greater than observed in all three model runs, the 10 mm day\(^{-1}\) isolath is roughly comparable to the 8 mm day\(^{-1}\) contour from the observations. The successive eastward and southward positioning of these areas of regionally heavy precipitation is qualitatively represented by all three model simulations. In July, heavy rainfall occurs in the Indian monsoon region, in October over Southeast Asia, in January near the New Guinea/Indonesia area, and in April there is relatively heavy precipitation in the western Pacific. The simulation of the seasonal progression for this particular area of heavy rainfall is fairly well-represented in the Northern Hemisphere, but none of the model simulations captures the extent or intensity of precipitation in the northern Australia region or in the SPCZ during January or April.

The results in Figs. 1 and 2 suggest that the distribution of land and sea and associated differential heating accounts for much of the seasonal position of the convective maximum in the Asian monsoon regions. The warmer SSTs in the tropical eastern Pacific in the MIX1 and MIX2 (mainly a result of the slab ocean which does not include upwelling) are associated with greater precipitation there throughout the annual cycle, particularly when the convective maximum moves into the Pacific in February–April. However, the SPCZ is only weakly simulated by the SPEC SST case and is not represented in the simulations with the slab ocean.

The implications of this model deficiency in terms of interactions with the SH extratropics are discussed in the next two sections.

4. Interannual variation—observed Warm and Cold Events

To further examine the SH circulation in general and the linkages between the tropics and higher southern latitudes in particular, it is useful to study the two documented extremes of interannual variation associated with the Southern Oscillation (SO)—Warm and Cold Events. These events are well known for their effects in the tropics, and associations with higher-latitude circulation anomalies in the SH are being explored in more detail (e.g., van Loon and Shea, 1985, 1987). A number of investigators (e.g., Trenberth, 1981) have noted that the zonal nature of the SH circulation lends itself to analysis through the use of zonal averages. Here, zonal averages are computed globally and for two sectors—the Indian from 45° to 155°E and the Pacific from 160°E to 90°W. Since Warm and Cold Events represent opposite manifestations of a dominant mode of circulation—the SO—it can be assumed that composite differences (Warm minus Cold Events) should provide insights as to the functioning of interannual variability and tropical–midlatitude interaction.

In Figs. 3 and 4, composite differences are computed, Warm-Event years minus Cold-Event years, for 500-mb temperatures and SLP, respectively. Years zero for Warm and Cold Events are those listed by van Loon and Shea (1985). For the 500-mb temperatures, Warm-Event years zero are 1972, 1976 and 1982; Cold-Event years zero are 1973 and 1978. For SLP, Warm-Event years zero are 1951, 1953, 1957, 1972, 1976 and 1982; Cold-Event years zero are 1954, 1973 and 1978. Both Figs. 3 and 4 show the seasonal evolution of anomalies starting with year minus one through year zero and ending with year plus one. Three-month seasonal averages are computed based on the long-term mean annual cycle of 50°–65°S 500-mb temperature gradients and SLP in van Loon (1967). Seasonal averages are centered on the month most closely associated with a maximum or minimum of the second harmonic which is dominant at those latitudes. Therefore, seasonal averages are computed for May–July (MJJ), August–October (ASO), November–January (NDJ) and February–April (FMA). Aligning the seasons in this way also is
Fig. 2. The seasonal progression of areas of relatively heavy precipitation for the four months July, October, January and April. (a) Outlines of 8 mm day$^{-1}$ isopleths stippled in Fig. 1 for observed precipitation, (b) outlines of 10 mm day$^{-1}$ isopleths for the SPEC SST case, (c) same as (b) except for the MIX1 case, and (d) same as (b) except for the MIX2 case.

consistent with the annual cycle in the tropics depicted in Fig. 1 to run from May through the following April.

The 500-mb temperature differences, Warm minus Cold Events, are shown for global zonal averages in Fig. 3a. Zonal averages computed for the Indian sector, which includes the Indian Ocean and Australia, appear
FIG. 3. Observed annual cycle of seasonal zonal mean 500-mb temperature differences, Warm-Event composite minus Cold-Event composite, for 10°S to 80°S. Year zero is denoted as the year of a Warm or Cold Event as defined by van Loon and Shea (1985). Year minus one is the year before, year plus one the year after. The sign of the difference is that of a Warm Event. Differences greater than +0.5°C are stippled; differences less than −0.5°C are cross-hatched. Values are contoured for seasonal averages described in text. (a) Global zonal mean differences, (b) Indian sector zonal mean differences (45°S–155°E), (c) Pacific sector zonal mean differences (160°E–90°W).
FIG. 4. As in Fig. 3 except for SLP. Differences greater than +1 mb are stippled; differences less than −1 mb are cross-hatched.
in Fig. 3b, and for the Pacific sector in Fig. 3c. The sign of the differences is that of a Warm Event, and discussion will be in terms of Warm Events. Cold Events are assumed to have the opposite sign. Because of the small number of cases (three Warm Events and two Cold Events), the statistical significance of the differences will not be discussed. Standard deviations computed for the 500-mb monthly mean temperatures are typically less than 1°C and for Warm and Cold Events are often around 0.5°C. A more rigorous analysis of the significance of the differences awaits a longer period of data accumulation.

In Fig. 3 in the tropics, similar features emerge in all three sets of zonal averages for 500-mb temperatures. First, the cycle for anomalies of the same sign near 10°–20°S runs from MJJ to the following FMA, the same cycle in the tropics in other studies of the SO. Second, the anomalies globally in both sectors are of the same sign, with negative anomalies spanning MJJ of the year before a Warm Event (year minus one denoted in the format MJJ+1) to February–April of the year of an event (FMA0), and positive anomalies from MJJ0 through FMA of the year after an event (FMA+1). This conformity of the global tropical troposphere to the sense of SST anomalies in the equatorial Pacific has been noted by Horel and Wallace (1981) for southern summer (northern winter) at the end of year zero going into the beginning of year plus one. They compose an index of stations from all regions of the global tropics and find that periods of warm SSTs in the equatorial Pacific (during the Mature Phase of a Warm Event, in the terminology of Rasmusson and Carpenter, 1982) are coincident with above-normal temperatures throughout the global tropical troposphere. Further inspection of their Fig. 4 also shows the converse to be true—Cold-Event years occur in association with below-normal temperatures throughout the tropical troposphere. The seasonal evolution of these features in the tropics near 10°–20°S can be seen in Fig. 3. The largest anomalies of 500-mb temperature occur during southern summer, the time of year when the temperatures at 500 mb in the southern tropics are highest in the mean annual cycle. Therefore, these anomalies in the tropics involve a modulation of the annual cycle. From about MJJ+1 to FMA0 the annual cycle is suppressed, and from about MJJ0 to FMA+1 the annual cycle is enhanced. This goes along with results presented by Meehl (1987), who showed that the annual cycle of the convective maximum from the Indian monsoon region in northern summer to northern Australia in southern summer is often either strong or weak. This implies that in years when the convective maximum in the Indian sector is strong and convection in the Pacific is weak (e.g., for the period May+1 through Apr10 as noted by Meehl, 1987) the global tropical tropospheric temperatures are lower than in years when the convective maximum is weak in the Indian sector and strong in the Pacific.

Moving from the tropics to higher latitudes, it can be seen that an opposition of sign in the anomalies exists in three latitudinal zones: the first is north of about 30°S, the second is roughly 35° to 60°S and the third is poleward of 60°S. This is best seen in the Indian sector (Fig. 3b). It is also worth noting that the areas of opposite sign do not occur simultaneously. For example, in Fig. 3b negative anomalies north of about 30°S occur from about MJJ+1 to MJJ0. Positive anomalies are present from about 35° to 60°S during roughly ASO1 to ASO0. Negative anomalies poleward of about 60°S exist from shortly after NDJ+1 to about the beginning of NDJ0. Similarly, positive anomalies in the tropics from MJJ0 to MJJ+1 correspond to negative anomalies at midlatitudes from ASO0 to ASO+1 with positive anomalies from NDJ0 to almost NDJ+1. This suggests a poleward progression of anomalies with the tropics about half a year out of phase with the high latitudes. A more direct poleward progression of similar nature can readily be distinguished in the Pacific sector in Fig. 3c.

Previous studies, such as Trenberth (1981) and Rogers and van Loon (1982), have noted the zonal position in the SH such that circulation anomalies in the tropics and high latitudes are often opposite to the midlatitudes and vice versa. Yasunari (1987) shows a similar opposition of SST anomalies in filtered SST data, as well as an apparent poleward propagation of the SST anomalies in the Pacific sector. Trenberth (1981) has also noted poleward progression of geopotential height and westerly wind anomalies in the SH. He interprets these features to be most distinctive in the “Australasia” sector from nearly the west coast of Australia to almost halfway across the Pacific. Poleward progression of anomalies is evident but not nearly as pronounced in his “Pacific Ocean” sector (from halfway across the Pacific to the middle of South America) and “Indian Ocean” sector (from midway in Africa to about the west coast of Australia). Little poleward progression was evident in his “Atlantic Ocean” sector. The Indian and Pacific sectors in the present study cover most of Trenberth’s Indian Ocean and Pacific Ocean sectors, respectively, and share his Australasian sector. It is probable that the present results and Trenberth’s results are describing similar phenomena.

Clues to a physical explanation for poleward progression of anomalies, and that they appear to be most pronounced in the Pacific Ocean, may lie in the unique geography and resulting annual cycle of the coupled ocean–atmosphere system in the Indian and Pacific regions. As described by Meehl (1987), the annual cycle of the convective maximum follows the change of season and is positioned progressively southeastward from the Indian monsoon during northern summer to northern Australia and the SPCZ during southern summer. Meehl (1987) shows that high SSTs ahead of the path of the convective maximum are associated with a local enhancement of convection and precipi-
tation during the course of the annual cycle. The ocean is affected by passage of the stronger convection such that lower SSTs are left in its wake where higher SSTs existed before. The altered ocean surface then influences the atmospheric circulation the following year and so on, with an inherent biennial tendency. Trenberth (1975) and van Loon and Shea (1985) examine the SPCZ area and note similar relationships for that region. The SPCZ stretches to the southeast from the tropics into the subtropics. In the extratropics, it is a source of cyclonic activity as shown, for example, by van Loon (1965). By its nature, the SPCZ is a natural conduit for tropical effects to be passed to the midlatitudes. The Pacific Ocean is the only basin with such a dominant feature that directly links the tropics to the midlatitudes of the SH. A feature comparable to the SPCZ exists in the South Atlantic but is less extensive and located farther south (e.g., see van Loon, 1965). Therefore, it is probably no coincidence that the direct poleward migration of anomalies is most strongly seen in the Pacific sector (Fig. 3c). These results suggest that, because of the coupling of atmosphere and ocean, the anomalies do not instantaneously transform the SH circulation. Van Loon and Shea (1987) show warm SST anomalies ahead of the convective maximum in the SPCZ to 40°S the year before a Warm Event (their Fig. 4a and c). As increased cyclone activity migrates farther south during the course of the annual cycle, the SSTs would be lowered with the increased latent and sensible heat fluxes from greater cyclone activity. Lower SSTs at 40°S in the Pacific occur the year of a Warm Event (van Loon and Shea, 1987, Fig. 4b and d). Lower SSTs in the midlatitudes would have the effect of slackening the gradient between 50°S and 65°S and decreasing cyclonic activity there. The effects of tropical/subtropical conditions in the SPCZ could then be communicated to high latitudes via changes in air–sea interaction, but with a time delay as the coupled system adjusts at progressively more southern latitudes.

This line of reasoning can be followed in Fig. 4 for SLP compiled in a similar fashion to Fig. 3. Standard deviations of the monthly means are typically less than 2 mb in the tropics north of 30°S, 2–4 mb between 30° to 50°S, and 3–5 mb south of about 50°S. There is no consistent global anomaly sequence near 10°–20°S (Fig. 4a) as there is in the 500-mb temperature. Instead, as can be expected from the signal of the SO in SLP, the Indian and Pacific sectors are in opposition with a period nearly the same as the temperature in Fig. 3. From MJJ–1 to FMA0, there is relatively higher SLP in the tropical Pacific (Fig. 4c) and lower SLP in the tropical Indian sectors (Fig. 4b). Then, from roughly MJJ0 to FMA+1, the opposite pattern is seen. Due to the diagonal movement of the convective maximum from north of the equator in the Indian monsoon to south of the equator in the SPCZ, the SLP signal near 10°–20°S is most distinct in the Pacific. The increased cyclonic activity near 40°S (negative differences in the Pacific in Fig. 4c) moves south and intensifies from MJJ0 through FMA+1. This is in association with the cooler 500-mb temperatures there in Fig. 3c. At the same time, the slackened 500-mb temperature gradient between 50° and 65°S is associated with less cyclonic activity and a weakening of the circumpolar trough near 65°S. This is evidenced by the large positive SLP anomalies at that latitude in the Indian and Pacific sectors as well as in the global zonal means. As seen in Fig. 3 for the 500-mb temperatures, the tropical SLP anomalies are about half a year out of phase with the higher-latitude anomalies.

Another feature of interest is the opposition between the Indian and Pacific sectors during year minus one from about 50° to 70°S. Positive SLP differences appear in the Indian sector at those latitudes (Fig. 4b), while negative differences are seen in the Pacific sector (Fig. 4c). The negative SLP anomalies in the Pacific are associated with a steepened 500-mb temperature gradient between 50° and 65°S (cooler temperatures at 65°S than at 50°S in Fig. 3c). The positive SLP differences in the Indian sector occur in conjunction with slackened 500-mb temperature gradients between those latitudes in Fig. 3b. This opposition between the Indian and Pacific sectors during year minus one near 30°–40°S has been noted by van Loon and Shea (1985, 1987) and Kiladis and van Loon (1988) to be important in setting up SSTs critical to the Warm-Event sequence in the following year.

5. Seasonal evolution in the GCM experiments

Since observed Warm and Cold Events represent two realizations of the climate in these regions, the three model simulations will be compared as distinct alternate climate states to look for physical processes similar to those noted in the observations. This section addresses the evolution of the seasonal cycles of the three GCM experiments—SPEC SST, MIX1 and MIX2—for the SH. Seasonal averages and sector zonal means are computed in the same manner as for the observations in the previous section. Differences are calculated for MIX1 minus SPEC SST (Fig. 5) and MIX2 minus MIX1 (Fig. 6), for 500-mb temperature and SLP. Since the GCM averages are only for 3-year periods, conventional statistical tests are not performed here. Instead, it is useful to note that standard deviations of monthly mean 500-mb temperatures for all model runs are on the order of 2°C or less. Standard deviations of monthly mean SLP are similar to the observed values: about 2 mb or less in the tropics, 2–4 mb at midlatitudes and 4–6 mb at high latitudes. Differences greater than two standard deviations are the most noteworthy, but the general pattern and physical consistency of the results are discussed for all the differences.

The mixed-layer ocean generates SST anomalies in the equatorial Pacific similar to those observed during Warm Events (Meehl and Washington, 1986). There-
Fig. 5. Zonal mean differences, MIX1 minus SPEC SST, computed in a similar fashion to Fig. 3 and 4. (a) 500-mb temperature, global zonal mean, cross-hatched areas are differences less than −1 mb, (c) Same as (a) except for Indian sector zonal means (45°–155°E), (d) Same as (b) except for Indian sector zonal means. Stippled areas are differences greater than +1 mb, (e) Same as (a) except for Pacific sector zonal means (160°E–90°W), (f) Same as (b) except for Pacific sector zonal means.

Therefore, similar to the observed Warm Events, the tropical 500-mb temperatures are higher in MIX1 year-round for global, Indian and Pacific zonal averages (Fig. 5a, c, e). Greatest warming occurs in southern summer (NDJ–FMA), as noted for the observed values for MJJ0 to FMA11 in Fig. 3. SLP is lower in the Pacific sector and higher in the Indian, as seen in the observations in Fig. 4, with greatest values again occurring during southern summer. As noted by Washington and Meehl (1984), sea ice in MIX1 is more extensive than the observed values in SPEC SST. In addition, the 50-m ocean mixed layer produces SSTs that have an annual cycle at midlatitudes which is somewhat different from the observed SSTs used to drive SPEC SST (Meehl and Washington, 1986). These changes in surface forcing are reflected by the 500-mb temperatures as well. Meehl and Washington (1986) show that at 50°S the MIX1 SST during February–June is about 3°–5°C warmer than the SPEC SST, and slightly cooler during September–November. Similarly, Fig. 5a, c and e shows that the 500-mb temperatures at midlatitudes are highest in the MIX1, compared to SPEC SST, in FMA and MJJ. Smaller positive, and sometimes negative, differences occur during ASO. This change in the annual cycle of SSTs at 50°S has the effect of changing cyclonic activity in the circumpolar trough. In the global zonal mean (Fig. 5a), the warmer temperatures in the midlatitudes (50°–60°) are associated with lower SLP at higher latitudes for most of the year (Fig. 5b). The Indian and Pacific sectors reflect a similar pattern but reveal more seasonal detail. In particular, there appears to be an opposition in the sign of SLP differences at about 30°–40°S and 60°–65°S. Negative anomalies (lower SLP) in the Indian sector during MJJ and ASO at about 40°S are associated with small negative or positive anomalies (higher SLP) near 65°S and vice versa for NDJ–FMA. The opposite is the case in the Pacific. Larger negative anomalies occur around 30°–40°S in FMA and MJJ corresponding to positive differences near 60°S. The zonal opposition of circulation anomalies in the SH within each sector and the east–west opposition between Indian and Pacific sectors were noted for the observations in Figs. 3 and 4. However, it is apparent that the change in the annual cycle of mixed-layer SSTs (Meehl and Washington, 1986) and the significantly altered sea-ice distributions for the model mixed layer (Washington and Meehl, 1984) at midlatitudes in the MIX1 case make it difficult to draw more than general conclusions from a direct comparison with observations at high latitudes.

For MIX2 minus MIX1 (Fig. 6), similar features in the tropics (10°–20°S) are apparent. These include warmer temperatures year-round in both sectors and
globally (Fig. 6a, c, e) with greatest warming during southern summer. Lower SLP is seen in the Pacific, especially during FMA, with relatively higher SLP in the Indian sector near 10°–15°S at that time of year. The significant decrease of sea-ice extent in the MIX2 compared to MIX1 (Washington and Meehl, 1984) is manifest by the warmer SSTs at 50°S (Meehl and Washington 1985). This is also reflected by warmer 500-mb temperatures year-round near 50°–65°S in all zonal averages in Fig. 6a, c and e. These changes alter cyclonic activity at higher latitudes, as represented by the SLP differences in Fig. 6b, d and f. For the global and Indian averages, the poleward shift of cyclonic activity with the major retreat of the sea-ice line is represented by positive SLP anomalies near 50°–55°S and negative anomalies to the south during FMA and MJJ. This was also seen as a poleward shift of cloud amount shown by Washington and Meehl (1984, their Fig. 13). Positive SLP differences during NDJ, and especially ASO, can be traced to the retreat of sea ice. The result is a less “continental” and more “oceanic” annual cycle near 65°S. This effect, coupled with a decrease in the annual mean 500-mb temperature gradient, produces the result seen in Fig. 6—higher SLP during ASO and NDJ (late winter and fall) and lower SLP during FMA and MJJ (late summer and spring) at high latitudes in all sectors.

6. Discussion

From the preceding analyses and comparison of Figs. 3 and 4 (observations) with Figs. 5 and 6 (model results), it is apparent that the poleward propagation of anomalies connected with observed interannual extremes, Warm and Cold Events, is not reproduced in any of the GCM comparisons. This may not be too surprising given the deleterious effects at mid- and high latitudes of weaknesses in the simulation of the annual cycle of SSTs and sea ice in the two model runs with the simple mixed layer. In addition, the SPCZ was postulated to play a major role in the communication of anomalies via the coupled ocean–atmosphere system from tropics to higher southern latitudes in section 4. Even though the successive eastward positioning of the tropical convective maximum was noted to take place in the model simulations in Fig. 2, with a tropical annual cycle similar to the observed of May to April, examination of geographical maps of precipitation shows that the majority of the convection in the Pacific in both slab-ocean models occurs over the relatively warm SSTs in the equatorial region (Meehl and Washington, 1986). Therefore, the SPCZ is shifted northeastward and does not come into play in the evolution of the annual cycle in either of those model simulations.

When compared with one another, the model integrations, interpreted as alternate climate realizations, reproduce some of the observed features of variability in the tropics and subtropics, but do not capture the poleward migration of Warm-minus-Cold-Event anomalies in the extratropics. However, several significant results emerge with important implications for
understanding the coupled climate system. The slab-ocean model simulations produce SST anomaly patterns that resemble those observed during the mature phase of Warm Events in the tropics, in part, to the omission of certain ocean processes such as upwelling (Meehl and Washington, 1985). The atmosphere responds to those SST anomalies to produce direct circulation effects which, near-equilibrium for seasonal averages, resemble those observed during Warm Events (Meehl and Washington, 1986). However, to study the seasonal evolution of observed interannual events in the extratropics such as those noted in section 4, model simulations run near-equilibrium do not capture such features. By definition, the model runs are near-equilibrium with their various boundary forcings. The observed processes appear not to be at equilibrium, as discussed in section 4, neither from year to year nor from latitude band to latitude band. Since the extratropical anomalies in the observations appear to emanate from the tropics with a multiyear cycle and lags of about half a year at high latitudes, averages of the model runs for single annual cycles would not be able to replicate such features in the extratropics.

A dynamically coupled ocean–atmosphere GCM may be capable of reproducing the observed signals seen in section 4 if it is at least capable of simulating 1) a realistic response of ocean SSTs to changes in atmospheric circulation in the tropics and at higher latitudes; 2) a reasonable reproduction of the SPCZ; and 3) a plausible seasonal cycle of the successive eastward positioning of the tropical convective maximum with a tendency for the biennial modulation described by Meehl (1987), which results from the interannual fluctuation of the east–west circulation of the coupled ocean–atmosphere system.

Therefore, the present results not only point to the inadequacies of the present model simulations, but to the requirements for the performance of future coupled ocean–atmosphere models which must be capable of simulating the close coupling between ocean and atmosphere in Indian and Pacific sectors of the SH, as evidenced by the analysis of observed climate states represented here by Warm- and Cold-Event years.

7. Conclusions

Observations are analyzed to relate the annual cycle of convection in the tropical Indian and Pacific sectors to circulation in the SH extratropics. Three GCM simulations with different ocean surfaces are then analyzed and compared with observations. The GCM is a version of the NCAR CCM with computed soil moisture, snow cover, cloudiness/radiation and annual cycle. The GCM is first coupled to an ocean surface with the specified annual cycle of observed SSTs (SPEC SST). Then it is run with a 50-m deep, simple mixed-layer ocean whereby SSTs are computed from the surface-energy balance and heat storage in the 50-m thick slab of water.

There is no provision for ocean currents or upwelling. One experiment is performed with the model in this configuration with the present amount of atmospheric CO₂ (MIX1). Another is run with double the present amount of CO₂ (MIX2). The model simulations with various altered boundary forcings are interpreted as alternate climate states, not as direct climate analogues. Observed manifestations of alternate climate states are taken to be composites of Warm- and Cold-Event years. The following conclusions are reached:

1) The observed mean southeastward movement of the position of the convective maximum from the tropical Indian sector to the Pacific is evident in all three GCM experiments. This suggests that the distribution of land and sea in those regions can account for the observed seasonal movement of the convective maximum. Features during the tropical annual cycle, which appear to depend on the accurate simulation of SSTs, are captured better by the SPEC SST (run with observed SSTs) than either the MIX1 or MIX2 (run with the slab ocean). Similarly, the warmer-than-observed SSTs in the equatorial Pacific in both slab-ocean runs (MIX1 and MIX2) are associated with greater precipitation there throughout the annual cycle, particularly when the convective maximum moves into the Pacific in February–April. This greater-than-observed convective activity in the Pacific disrupts the east–west circulation between the tropical Indian and Pacific and is associated with suppressed precipitation in the Indian sector as noted in previous studies. However, the SSTs produced by the slab ocean preclude an accurate simulation of the SPCZ.

2) An analysis of zonal mean observed 500-mb temperature and SLP for Warm-Event year composites minus Cold-Event year composites is performed. Conditions in the tropical Pacific can influence the temperature of the entire global tropical troposphere during those years, as noted in other studies. However, the west–east circulation between the tropical Indian and Pacific sectors is represented as an opposition of sign of SLP anomalies in those regions.

3) The tropics and high southern latitudes appear to react about half a year out of phase. The tropical cycle runs from MJJ to FMA while the high-latitude period covers NDJ to ASO. This suggests a communication of information from tropics to high latitudes. Poleward progression of anomalies appears most prominently in the Pacific sector. This is likely to be related to the poleward progression of anomalies in this region noted by Trenberth (1981). It is postulated that the SPCZ could act as a conduit between the tropics and higher latitudes in the following manner. As strong convection and associated lower SLP move southeastward toward the SPCZ in year minus one of a Warm Event, the underlying SSTs are higher due to conditions in the previous year. This involves the entire annual cycle of the convective maximum as it moves
from the Indian monsoon to the Australian monsoon and SCZ as described by Meehl (1987), and the SPCZ region in particular as outlined by Trenberth (1975) and van Loon and Shea (1985, 1987). As lower SLP in the SPCZ moves farther southeastward with the annual cycle, the SSTs are lowered by increased cyclonic activity (greater sensible and latent heat loss, more cloudiness, etc.). The increased cyclonic activity then moves into midlatitudes in the Pacific, and SSTs are similarly altered. This, in turn, affects the annual cycle of the temperature gradient between middle and high southern latitudes and thus the semiannual oscillation of the atmospheric circulation at those latitudes which depends critically on the annual cycle of SSTs there (van Loon, 1967; van Loon and Rogers, 1984). Warm 500-mb temperatures in the tropics from MJJ_{0} to FMA_{+1} of a Warm Event could, therefore, be associated with a weakened high-latitude circulation a half-year later from NDJ_{+1} to ASO_{+1}. Those lower 500-mb temperatures at midlatitudes could effectively slacken the temperature gradient and weaken the high-latitude SH circulation from NDJ_{+1} to ASO_{+1}.

4) The observed dynamically coupled ocean-atmosphere system is not in a state of equilibrium between tropics and higher latitudes either seasonally or latitudinally if such poleward propagation is taking place. Therefore, the present GCM simulations, which are near equilibrium with their various boundary forcings, cannot, and indeed do not, capture the observed signals of poleward propagation as was noted in comparing the observed results in section 4 to the model results in section 5. Even though the brevity of the observed data record precludes definitive conclusions as to the observed processes postulated to be occurring in point 3 above, the present results suggest that a coupled ocean–atmosphere GCM would be necessary to simulate such observed effects.

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