The South Asian Monsoon and the Tropospheric Biennial Oscillation

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ABSTRACT

A mechanism is described that involves the south Asian monsoon as an active part of the tropospheric biennial oscillation (TBO) described in previous studies. This mechanism depends on coupled land–atmosphere–ocean interactions in the Indian sector, large-scale atmospheric east–west circulations in the Tropics, convective heating anomalies over Africa and the Pacific, and tropical–midlatitude interactions in the Northern Hemisphere. A key element for the monsoon role in the TBO is land–sea or meridional tropospheric temperature contrast, with area-averaged surface temperature anomalies over south Asia that are able to persist on a 1-yr timescale without the heat storage characteristics that contribute to this memory mechanism in the ocean. Results from a global coupled general circulation model show that soil moisture anomalies contribute to land-surface temperature anomalies (through latent heat flux anomalies) for only one season after the summer monsoon. A global atmospheric GCM in perpetual January mode is run with observed SSTs with specified convective heating anomalies to demonstrate that convective heating anomalies elsewhere in the Tropics associated with the coupled ocean–atmosphere biennial mechanism can contribute to altering seasonal midlatitude circulation. These changes in the midlatitude longwave pattern, forced by a combination of tropical convective heating anomalies over East Africa, Southeast Asia, and the western Pacific (in association with SST anomalies), are then able to maintain temperature anomalies over south Asia via advection through winter and spring to set up the land–sea meridional tropospheric temperature contrast for the subsequent monsoon. The role of the Pacific is to produce shifts in regionally coupled convection–SST anomalies. These regions are tied together and mutually interact via the large-scale east–west circulation in the atmosphere and contribute to altering midlatitude circulations as well. The coupled model results, and experiments with an atmospheric GCM that includes specified convective heating anomalies, suggest that the influence of south Asian snow cover in the monsoon is not a driving force by itself, but is symptomatic of the larger-scale shift in the midlatitude longwave pattern associated with tropical SST and convective heating anomalies.

1. Introduction

Coupled interactions between the ocean and atmosphere have previously been identified (Meehl 1987, 1993, 1994a) as contributing to a mechanism that produces a biennial component of interannual variability in the troposphere in the tropical Indian and Pacific regions [the tropical or tropospheric biennial oscillation (TBO)]. Biennial signals in various parts of those areas have been noted by a number of studies (e.g., Trenberth 1975; Nicholls 1978, 1979, 1984; Tomita and Yasunari 1996; Clarke et al. 1997, manuscript submitted to J. Climate, hereafter referred to as CLV97). The role of air–sea coupling (e.g., Brier 1978) has been central to all studies of such biennial oscillations, and several of those studies have linked the biennial component of variability to the Southern Oscillation. In particular, Meehl (1987) noted that this linkage involved the seasonal cycle and postulated that peaks in the biennial cycle were made manifest via a similar set of processes as the tropical Pacific warm and cold extremes in the Southern Oscillation. Subsequent studies have confirmed these earlier results (e.g., Terray 1995; Shen and Lau 1996; Tomita and Yasunari 1996) and have also further quantified the biennial component of tropospheric variability in the tropical Pacific and/or Indian regions of the SST and surface wind (Lau and Sheu 1988; Ras- musson et al. 1990; Barnett 1991; Kutsuwada 1991; Ropelewski et al. 1992; Jiang et al. 1995; CLV97). Biennial tendencies have also been noted to involve ocean dynamics in the western Pacific (e.g., Lukas 1988; Masumoto and Yamagata 1991). The active role of the east Asian summer monsoon in the TBO has been emphasized by Lau (1992) and Shen and Lau (1995), as has the associated coupled air–sea interaction in the South China Sea–Maritime Continent region (Tomita and Yasunari 1996; Ju and Slingo 1995; Soman and Slingo 1997; CLV97). A covarying biennial component of sea level pressure and surface temperature over Asia has
also been shown to have connections to North Atlantic atmospheric circulation (Mann and Park 1996).

It has been suggested that the TBO may arise from interaction between the annual cycle and the El Niño–Southern Oscillation cycle (Goswami 1995). That study assumed that the ENSO cycle is a fixed mode of the climate system, with a set period of 4 yr. However, if the TBO and ENSO operate with similar mechanisms, the TBO could be a determining factor in setting up the ENSO cycle and monsoon interactions, and not vice versa (Shen and Lau 1995). More recently, modeling results have suggested that the Indian summer monsoon could generate an internal biennial oscillation without any interactions with other components of the climate system, or possibly through the interaction of intraseasonal oscillations with the annual cycle (Goswami 1995). Yet other modeling studies have demonstrated the fundamental role of SST anomalies (and thus the role of interannual atmosphere–ocean interactions) in the Indo–Australian monsoons (Kitoh 1992; Palmer et al. 1992; Chen and Yen 1994) and in the TBO (Alexander and Weickmann 1995). Thus the possibility exists that the south Asian monsoon could generate its own biennial oscillation independent of other interactions or in concert with irregular ENSO variability (e.g., Jin et al. 1994). But it would appear equally plausible that the timing, strength, and nature of the TBO involves multiscale land–atmosphere–ocean interactions such that the south Asian monsoon plays an interactive role in larger-scale tropical interannual variability (Yasunari 1990) that could provide deterministic links to assist in forecasting monsoon rainfall (Lau and Yang 1996).

Meehl (1987) used Indian monsoon rainfall as an index of coupled air–sea interactions over all of the regions of the Indian and Pacific Oceans that produce a tropospheric biennial signal in about two-thirds of the years considered. It was also noted that SST anomalies in the tropical Indian and Pacific Oceans varied in phase in the biennial oscillation and the Southern Oscillation, and that Indian monsoon rainfall was inherently tied to the TBO. There have been attempts to link the stratospheric quasi-biennial oscillation (QBO) to interannual variability in the troposphere involved with the Southern Oscillation (e.g., Gray et al. 1992). However, the work cited above dealing with the TBO has not yet been linked to the QBO a priori. Such linkages may indeed exist, but are beyond the scope of the present paper.

Since the TBO is postulated to be made manifest as extremes in the Southern Oscillation (SO) and the Indian monsoon has been shown to be closely linked to the SO via the large-scale east–west circulation in the atmosphere (e.g., Webster and Yang 1992; Ju and Slingo 1995), it is clear that the Indian monsoon is closely connected with the TBO. However, since air–sea interaction is postulated as the central mechanism for the TBO in those regions, it is unclear exactly what role the Indian monsoon plays either directly or indirectly in the TBO. The purpose of this paper is to specifically address the role of the Indian monsoon (or more generally the south Asian summer monsoon) in the TBO. This paper will also address the role of tropical–midlatitude interactions in greater detail since the TBO and monsoon interactions most likely involve large time and space scale interactions (Webster and Yang 1992; Meehl 1987, 1994a; Tomita and Yasunari 1996; Nigam 1994; Yang et al. 1996; Yasunari and Seki 1992; Chen and Yen 1993, 1994).

2. The coupled ocean–atmosphere biennial mechanism

As first described by Meehl (1987) and later elaborated on by Meehl (1993), the biennial mechanism involving air–sea coupling can be described at a given location schematically by the diagram in Fig. 1a. For this mechanism to work, there are several premises that must be satisfied. First, the seasonal cycle must play such a role that convection is locally strong during one season per year. Second, the air–sea coupling must be strong only during that one season when the convection is locally strong. Third, upper-ocean heat content and SSTs may be affected during that season via vigorous air–sea coupling such that anomalies set up during that season persist for 1 yr. Meehl (1987, his Fig. 3a) shows a Hovmöller diagram of the long-term mean seasonal cycle of outgoing longwave radiation (OLR) averaged from 30°N to 30°S. Convection is locally strong once per year at all locations in the Indian–Pacific region as the convective maximum translates from west to east and north to south going from northern summer to southern summer (Meehl 1987; Murakami and Wang 1993), even in areas where the ITCZ passes overhead twice per year. This demonstrates that east–west atmospheric circulations are a dominant factor in the seasonal cycle, as well as in their more familiar role in interannual variability.

To summarize the biennial mechanism involving air–sea coupling (Fig. 1a), if the SST at a given location starts out relatively warm (as in the right panel), convective activity associated with the seasonal cycle encounters this relatively warm water at the time of year of the local convective maximum (or local “rainy season”). The relatively warm water enhances the convection with increased evaporation and moisture convergence (top panel). This produces relatively strong winds, high latent heat flux, and enhanced mixing that acts to cool the upper ocean. At the end of that season, the convective maximum moves on with the seasonal cycle, leaving the upper ocean relatively cool. Via alterations of the upper-ocean heat content (Meehl 1993), the ocean provides the memory to retain the relatively cool conditions (left panel in Fig. 1a) until a year later, when the convective maximum associated with the seasonal cycle again arrives at that location (bottom panel of Fig. 1a). The cool SSTs, associated decreases of evaporation, and reduced low-level moisture convergence all
contribute to weaker convective activity. Accordingly, the low-level winds are weaker and there is proportionately less latent heat flux. Consequently, the SSTs are left relatively warm at the end of that season in conjunction with higher heat content in the upper ocean (Meehl 1993). This provides the memory for the relatively warm SSTs to persist (right panel in Fig. 1a) until a year later, when the convective maximum is again relatively strong, and so on.

In this conceptual view, the coupled system exists in two states relative to each other—one warm and one cool. In the idealized case, the average (or zero) state never occurs. The mean climate is simply the arithmetic average of the relatively warm and cool states. However,
as shown by Rasmusson et al. (1990) and Ropelewski et al. (1992), what actually seems to happen is that the TBO is superimposed on a low-frequency component of variability such that the interactions between the biennial and low-frequency components could be made manifest as extremes in the SO. Yet the conceptual model of Fig. 1a is useful for visualizing the coupled processes that could actually produce the TBO. In this view, dynamic coupling between the ocean and atmosphere in conjunction with the seasonal cycle of convection over the tropical Indian and Pacific regions and the large-scale east–west circulation in the atmosphere combine to produce the TBO. This precludes the TBO from simply occurring as an incidental interaction between the annual cycle and the ENSO cycle. However, the interaction of the TBO with the seasonal cycle could be intermittent (Barnett 1991) and the amplitude in any given year could vary (Terray 1995). This could partly explain why the TBO does not have higher total amplitude in the studies of Rasmusson et al. (1990) and Ropelewski et al. (1992).

Since Meehl (1987) and others have shown that the south Asian monsoon has links to the TBO, it is possible to pose a set of processes that could make the monsoon an active participant in working with large-scale atmospheric circulations and air–sea interaction mechanisms to produce the TBO. For this to happen, there should be (at least) coupled land–ocean–atmosphere interactions that directly involve the seasonal cycle, as was noted for the air–sea mechanism described above. Figure 1b shows a schematic of possible processes that could actively involve the south Asian monsoon in the TBO. Here, there are similar premises to those posed for the air–sea mechanism. First, convection must be strong during only one season per year. This premise is easy to satisfy since the south Asian monsoon has a well-known once-per-year precipitation maximum during northern summer. Second, land–atmosphere coupling must be strong during that season. Third, there must be a set of processes that provide a 1-yr timescale memory for land-surface temperature anomalies. This function was performed by heat storage in the ocean for the air–sea mechanism. This last premise is more difficult to satisfy for land since there is much less capability for heat storage. This issue will be addressed in more detail later.

In addition to the fundamental difference of surface type between land and ocean, the atmosphere–land biennial mechanism must work somewhat differently than the atmosphere–ocean mechanism. The latter relies mainly on local air–sea interaction (along with low-level moisture convergence and large-scale linkages associated with the atmospheric east–west circulation) for most of the coupled processes that contribute to either a strong or weak convective maximum. For the south Asian monsoon, the fundamental concept that is important for producing a biennial signal is land–sea temperature contrast or, more accurately, the north–south meridional tropospheric temperature gradient. This is the basic forcing of the monsoon (e.g., Webster 1987), such that enhanced meridional temperature contrast is associated, in the observations, with strong Asian summer monsoons and vice versa for weak monsoons (e.g., Li and Yanai 1996). This basic phenomenon also appears as a fundamental element of model simulations of the monsoon. Stronger land–sea or meridional tropospheric temperature contrast at the beginning of the monsoonal season is associated with the greater monsoon precipitation over south Asia in a variety of GCMs with the same fixed SSTs (Meehl 1994b).

The role of land–sea or meridional tropospheric temperature contrast in tropical rainfall regimes has been well studied. For example, Charney et al. (1977) showed that increased land albedo produced cooler land temperatures, reduced land–sea temperature contrast, and reduced precipitation over land in the West African monsoon. A recent GCM study has demonstrated this relationship, which includes the associated increase in soil moisture with reduced tropical land albedo and increased precipitation (Lofgren 1995). Other GCM studies of the paleoclimate have shown that changes in solar parameters can enhance summer heating of south Asia, increase land–sea temperature contrast, and produce a stronger south Asian monsoon (e.g., Kutzbach et al. 1993). The role of snow cover in altering surface temperature and land–sea temperature contrast was first put forth by Blanford (1884) and was shown in more recent observations by Hahn and Shukla (1976). These studies (and others, see Shukla 1987) noted a tendency for enhanced winter and spring snow cover over south Asia to be followed by below-normal monsoonal rainfall. The mechanism involves proportionately more solar radiation being reflected by the excessive snow cover and a subsequently wetter land surface as the excessive snow cover melts. Both of these processes act to produce colder land temperatures, reduced land–sea temperature contrast, a weakened meridional tropospheric temperature gradient, and a weaker monsoon. These mechanisms have been studied more recently and shown to work in GCM studies (e.g., Barnett et al. 1989; Yasunari et al. 1991; Vernekar et al. 1994; Douville and Royer 1996). Similar linkages between snow cover and ENSO have also been noted (Groisman et al. 1994; Yang 1996; Sankar Rao et al. 1996).

To follow the biennial sequence for the south Asian monsoon in Fig. 1b, if the south Asian land area is relatively warm to begin with (right panel of Fig. 1b), this produces enhanced land–sea temperature contrast and a relatively strong monsoon (or a strong convective maximum, to use the terminology of the atmosphere–ocean biennial discussion above). The strong monsoon is associated with greater precipitation, higher evaporation, stronger winds from the Indian Ocean, and elevated land-surface moisture (top panel of Fig. 1a). At the end of the strong monsoon, the saturated land surface produces greater latent heat flux and contributes to rel-
Fig. 2. General regions of anomalously wet (hatched) and dry (stippled) areas from station data for composites based on seasons the year before (year $-1$) and the year during (year 0) warm events in the SO. (a) MJJ year $-1$, (b) November–January (NDJ) year $-1$, (c) MJJ year 0, and (d) NDJ year 0 (after Kiladis and van Loon 1988).

At the relatively cool temperatures over land. If these cool land temperatures can persist for 1 yr (left panel of Fig. 1b), reduced land–sea temperature contrast is set up to contribute to a relatively weak monsoon the following year (bottom panel of Fig. 1b). The weak monsoon is associated with less precipitation, lower evaporation from the Indian Ocean, and weaker low-level inflow. The reduced precipitation does not saturate the land surface, latent heat flux is not as high, and the land surface is left relatively warm at the end of the weak monsoon season. If these relatively warm land-surface temperatures can persist for 1 yr (right panel of Fig. 1b), enhanced land–sea temperature contrast sets up a relatively strong monsoon during the next northern summer, and so on.

3. Observations

As noted above, there are well-known links between the south Asian monsoon and the SO (e.g., Webster and Yang 1992; Yasunari and Seki 1992; Ju and Slingo 1995). Rasmusson and Carpenter (1983) showed that for roughly a 100-yr period, there were 25 warm events in the Pacific and 21 of those had Indian monsoon rainfall below normal. Meehl (1987) used relatively strong and weak Indian monsoons to show associated signals in the Pacific Ocean that resembled those of cold and warm events, respectively, and to show that SST anomalies in the tropical Indian and Pacific Oceans varied in phase. These links between the Indian monsoon, Indian Ocean, and the tropical eastern Pacific were shown to have a biennial tendency in that study, and those results have been reconfirmed in subsequent analyses.

For example, Figs. 2 and 3 show precipitation and surface temperature anomalies for warm minus cold event composites (after Kiladis and van Loon 1988). In the season MJJ-1 in Fig. 2a (the May–July season prior to a warm event or, for the monsoon, prior to a weak monsoon), many areas in the south Asian monsoon region show excess precipitation (as in the top panel of Fig. 1b). Associated with the strong monsoon, the land temperatures decrease in Fig. 3a. By the end of the strong monsoon season (Fig. 3b), the land temperatures over most of south Asia are below normal. These conditions over south Asia then persist through winter (Fig. 3c) into spring (Fig. 3d). This corresponds to the left panel in Fig. 1b. These land temperature anomalies vary in phase with tropical Indian Ocean SSTs and eastern Pacific SSTs (Meehl 1987).

During the northern winter, precipitation anomalies in the tropical Pacific and Indian Ocean areas follow the conceptual model in Fig. 1a. That is, a relatively strong convective maximum moves over the relatively warm waters in the Australasian region (Figs. 3b and 3c) with the seasonal cycle. The strong convective maximum in Fig. 2b is made manifest as positive precipitation anomalies in the seasonal rainfall areas of Australasia. In response, SSTs cool and are left relatively low to the north and west of the convective maximum,
as also described by Meehl (1987, 1993). The large- scale east–west circulation is playing a role in the Aus- tralasian sector, with increased vertical motion and up- per-level outflow, and suppressed convection over the western Indian Ocean and eastern Pacific in association with cool SSTs in those regions. Just out of the picture in Fig. 2c are positive precipitation anomalies over cen- tral Africa, which are also involved with the large-scale east–west circulation in the atmosphere (see also Kiladis and Diaz 1989, Halpert and Ropelewski 1992; Rope- lewski and Halpert 1987; Lau 1992).

The relatively cold land temperatures over south Asia during northern winter and spring set up decreased land– sea temperature contrast and contribute to a weak Indian
monsoon the following summer, as in Fig. 2c (and also as noted in Fig. 1b, bottom panel). The reduced precipitation amounts are then associated with a warmer, relatively drier land surface, and positive surface temperature anomalies are left at the end of the monsoon season (Fig. 3f). These conditions then persist through the rest of the year (Figs. 3g and 3h, corresponding to the right panel of Fig. 1b) to set up enhanced land–sea contrast for the following strong monsoon season, and so on. Note in Fig. 2d that the weak convective maximum that has moved over Australasia with the seasonal cycle is associated with reduced precipitation amounts in that region, and relatively warm SSTs are left to the north and west of the convective maximum. Also note in Fig. 2d that the weak convective maximum in Australasia is part of the large-scale east–west circulation, such that elevated precipitation amounts are occurring to the west over the eastern Indian Ocean and over the eastern Pacific. Kiladis and Diaz (1989) also note enhanced precipitation over continental Africa in association with these alterations of the east–west circulation.

A further documentation of observed variability of precipitation in the global Tropics has been performed by Lau and Sheu (1988). Figure 4a shows the pattern of the first EOF of the monthly mean precipitation from their analysis of monthly station precipitation data. This EOF explains 9% of the monthly variance of the station precipitation data. The pattern reflects the features of precipitation variability discussed above. When precipitation is above normal in the Indian monsoonal region, it is also above normal over the rest of Australasia and below normal over the eastern Pacific and central Africa. Spectra of this leading EOF show a biennial peak at a little more than 2 yr, as well as a peak associated with the SO at about 5–6 yr (Fig. 4b). Also note the opposition of sign between India and parts of east and South Asia. This suggests that the “monsoon indices” used in this paper for area-averaged monsoon rainfall mostly over India may not be representative of other parts of south Asia.

4. Coupled model analysis

In this section, we use a coarse-grid global coupled ocean–atmosphere GCM as a heuristic tool to study the processes postulated above for the observed climate system. Details of the model are described elsewhere (e.g., Washington and Meehl 1989; Meehl 1989). Briefly, the model uses a global spectral rhombooidal 15 (R15, roughly 4.5° lat × 7.5° long) nine-level spectral atmospheric GCM. The atmospheric model is coupled to a global coarse-grid (5° lat × 5° long) four-layer ocean GCM. The ocean is a simple thermodynamic formulation. The atmospheric model was spun up for 25 yr with a 50-m deep nondynamic slab mixed layer. The ocean was initialized with observed cross sections of temperature and salinity and then run with observed atmospheric surface temperatures and wind stress for 50 yr. The models were then coupled together, and an integration was performed without flux correction for present day climate for 70 yr. The analyses presented here will use the last 50 yr of that integration.

In spite of systematic errors in the non-flux-adjusted integration, this model has been shown to simulate many of the aspects of the observed large-scale mean climate (Washington and Meehl 1989) and the climate variability compared to satellite observations (Meehl et al. 1994), as well as some aspects of ENSO (Meehl 1990). Large-scale features of the east–west atmospheric circulation associated with interannual variability of tropical Pacific SSTs and teleconnections to Africa, Asia, and South America comparable to observations are also represented in the coupled model (Meehl et al. 1993, Fig. 4b). Variability of the simulated south Asian monsoon has been shown to resemble the observed monsoon as well in Meehl and Washington (1993), although that study noted that the colder-than-observed tropical SSTs in the coupled model contribute to an enhanced land–sea temperature contrast in the model and, consequently, somewhat more land-based precipitation relative to Indian Ocean precipitation in the model compared to the observations. Figures 5a and 5b and Table 1 show mean and DJF precipitation from the coupled model and observations (also shown in Meehl 1989; Washington and Meehl 1989). As noted in these earlier studies, the major monsoon precipitation centers are represented, but with enhanced precipitation in DJF over central Africa and weaker-than-observed precipitation over the Australian monsoon region (in association with the colder-than-observed tropical SSTs). In JJA in Figs. 5c and 5d, precipitation over Africa is also enhanced, with relatively higher precipitation over the south Asian monsoon region and somewhat less Indian Ocean precipitation, as noted above. Surface winds in DJF and JJA compare favorably with observations of flow in the Somalial jet over the Arabian Sea reversing direction from JJA to DJF (Meehl 1989, Figs. 7 and 8). Weak minus strong monsoonal differences (Meehl and Washington 1993, Fig. 2a) show, for weak monsoon years, decreased precipitation over most of India, Bangladesh, Burma, Thailand, Laos, and Vietnam.

To examine timescales of variability in monsoon observations, spectra are computed for 50 yr of an observed Indian monsoon index (Parthasarathy et al. 1991) and compared to similar spectra from the coupled model for JJA monsoonal rainfall averaged over India [land points only from 5°–30°N, 65°–95°E; time series is shown in Meehl and Washington (1993), Fig. 1a]. Over the past 100-yr record of the Indian monsoon index, the 1901–50 period had a particularly active biennial cycle, and the observed time series shows a peak significant at the 95% level at the quasi-biennial timescale of 2.3 yr in Fig. 5e. Similar spectra for the 50-yr time series of JJA rainfall from the coupled model also show a biennial peak at 2.3 yr. Both the observed and model monsoon time series also show peaks associated with...
ENSO. For the observed, there is a peak at 3.3 yr, and for the model, there is a peak at 4.5 yr. The model biennial peak in Fig. 5f is below the 95% significance level, indicating that the biennial amplitude is less than that of the observations. This lower-amplitude variability in the coupled model compared to observations is consistent with other aspects of tropical interannual variability that have amplitudes smaller than those observed (e.g., Meehl 1990; Meehl and Washington 1993; Meehl et al. 1994) and with other coupled models of this class (e.g., Latif et al. 1994).

It was noted earlier that the south Asian monsoon has a strong link to the ENSO cycle. For the coupled model, during the 50-yr period of analysis, there were eight warm events in the Pacific (defined as the Niño3, 10°N–10°S and 90°–150°W, area-averaged SST anomalies greater than 0.25°C, or roughly one standard deviation) and five of those had Indian monsoon precipitation below normal. For nine cold events during that 50-yr period in the model (Niño3 SST anomalies less than −0.25°C), there were six Indian monsoons above normal. When the definition basis was the monsoon, for 10 Indian monsoons with area-averaged precipitation above one standard deviation, the SSTs in the Niño3 area were below normal seven times. For 11 Indian monsoons with precipitation less than one standard deviation, Niño3 SSTs were above normal seven times. As another measure of the connections between the tropical Pacific and Indian sectors in the coupled model, sea level pressure over the Indian Ocean and Niño3 SSTs have a correlation coefficient of $0.54$, which is significant at the 5% level.

To compare the patterns of precipitation variability in the model with the various indices discussed above, as well as with the observations in Figs. 4a and 4b, an EOF analysis is performed on the detrended monthly mean anomaly time series of precipitation from the coupled model. The first EOF is plotted in Fig. 4c, with the spectra of the time series of the principal component in Fig. 4d. As in the observations, the percent variance explained is quite low (2.3% for EOF1 in the coupled model and 9% for the observations in Fig. 4a). This is
thought to be due to the noisy nature of precipitation data. The lower number for the model is also an indication of the lower-amplitude tropical interannual variability in the model compared to the observations [documented, e.g., for ENSO-related variability by Meehl (1990), for monsoon variability by Meehl and Washington (1993), and for lower-tropospheric temperature by Meehl et al. (1994)]. However, what is of interest here is a comparison of the patterns of variability represented by EOF1 between observations in Fig. 4a and the coupled model in Fig. 4c, and there are some striking similarities. Same-signed areas (dashed lines) cover India, Southeast Asia, and the Australian monsoon region, while opposite-signed areas (solid lines) occur over central and eastern equatorial Africa and the central equatorial Pacific. This demonstrates the out-of-phase relationship between precipitation in the tropical Pacific–eastern Africa and the Indian–Australian monsoons. As in the observations in Fig. 4a, there is an opposition of sign between India and parts of east and Southeast Asia, with the similar suggestion that the monsoon index used in this paper, computed mostly over India, may not represent monsoonal variability in all parts of south Asia.

The spectra of the time series of the principal component from the coupled model (Fig. 4d) have similar timescales to those observed in Fig. 4b, with peaks in the TBO (roughly 2.3 yr) and ENSO (about 5.3 yr) timescales. This illustrates for the model, as well as for the observations, the linkages between TBO and ENSO in terms of similar patterns of precipitation anomalies occurring on different timescales.

In Fig. 6a, the evolution of various elements of the monsoon system in relation to a strong monsoon is shown schematically in composite form, as gleaned from the observed sources cited in the introduction, and configured to conform to the conceptual model shown in Fig. 1b. As noted above, for the northern winter and spring seasons prior to a strong monsoon, there is a tendency for Asian land temperatures (refer to Fig. 3) to be above normal (solid line in Fig. 6a), contributing to enhanced land–sea temperature contrast. The Indian Ocean SSTs (dash–double-dot line) are also above nor-
mal, but the land temperatures are proportionately higher. This maintains enhanced land–sea temperature contrast with the positive SST anomalies over the Indian Ocean contributing to elevated evaporation levels. The warm south Asian land temperatures are then associated with less snow cover over south Asia (bars in Fig. 6a). During the strong monsoon season with monsoon precipitation above normal (dashed line), the south Asian land temperatures drop, along with Indian Ocean SSTs. These conditions persist through the next northern fall, winter, and spring, contributing to increased snow cover, decreased land–sea temperature contrast, and a subsequent weak monsoon the following northern summer. The connection to the tropical Pacific is shown in terms of Niño3 SSTs that are below normal during the northern spring before the strong monsoon and then transition to above normal during the northern spring before a weak monsoon.

To examine the proposed processes involved with the coupled land–atmosphere–ocean mechanism in Fig. 1b, various area averages are computed from the coupled model. Asian monsoon precipitation and soil moisture are computed as the average of land points in the area 5°–40°N, 60°–100°E (see Figs. 5a–d for a geographic plot of precipitation). South Asian land temperatures are calculated for land points in the area 2°–30°N, 70°–120°E (see Washington and Meehl 1989, Fig. 8 for a geographic plot of temperatures over Asia). Asian snow cover is liquid water equivalent over the area 20°–70°N, 50°–130°E (refer to Washington and Meehl 1989). Indian Ocean SST is from ocean points in the area 5°S–30°N, 50°–100°E, and Niño3 SST is computed for the area 10°S–10°N, 150°–90°W (see Washington and Meehl 1989, Fig. 10 for plot of SSTs). Australian monsoon precipitation is computed for all points in the area 25°S–0°, 100°–150°E; African precipitation for land points 5°N–5°S, 15°–35°E; and Pacific precipitation for all points 0°–15°N, 150°–170°W.

These area averages are chosen for their relevant anomalies in the TBO cycle and composited as in Meehl (1987, 1993), based on an index of area-averaged Indian monsoon rainfall. This index is formed by averaging all land points over the Indian region in the model as noted above. From this time series (plotted in Meehl and Washington 1993, Fig. 1a), extreme wet or dry monsoons (in excess of one standard deviation) for the JJA season from the model are used to compute area averages for various other parameters, and differences are computed for strong versus weak monsoons. These differences are plotted as a function of season in Fig. 6. Though significance testing cannot be rigorous for the relatively small sample size analyzed here, an idea of the relative size of the anomalies is given by the maximum magnitudes of these differences being about 5% to 20% of the strong or weak monsoon averages, and in most cases they are around one standard deviation of the monsoon seasonal values. The sign of the differences is that of a strong monsoon, the calendar year of the strong monsoon is designated year 0, and the seasons preceding and following the strong monsoon follow that convention.

For the coupled model in Fig. 6b, there are a number of similarities to the schematic representation of the observed system in Fig. 6a. In particular, south Asian land temperatures are elevated in the northern winter and spring before a strong monsoon in relation to Indian Ocean SSTs to produce decreased land–sea temperature contrast. This is associated with reduced Asian snow cover. During the JJA season of a strong monsoon, the south Asian land temperatures decrease as soil moisture amounts increase due to the heavy rainfall. The cool south Asian land temperatures persist through the fol-
TABLE 1. Area-averaged coupled model precipitation anomalies (mm day
1) for East Africa (5°N–5°S, 15°–35°E), the Australian monsoon (5°–20°S, 130°–150°E), and the tropical Pacific (0°–15°N, 150°–170°E), and observed area-averaged OLR anomalies (W m
2) for East Africa (0°–10°N, 25°–40°E), the Australian monsoon (0°–
20°S, 120°–140°E), and the tropical Pacific (10°S–10°N, 140°–165°E). Areas are chosen to coincide with key areas of variability in reference to the first EOF of monthly anomaly precipitation from the coupled model and observations (Fig. 4). Area outlines are shown in respect to seasonal mean precipitation in Fig. 5. Italicized values are consistent with the signs of specified heating anomalies in the experiments described in text. Note that negative OLR anomalies are interpreted as analogs for positive precipitation anomalies.

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<th>Area</th>
<th>East Africa</th>
<th>Australian monsoon</th>
<th>Tropical Pacific</th>
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<td>Coupled model</td>
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<tr>
<td>DJF precipitation</td>
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</tbody>
</table>

5. Tropical forcing of midlatitude circulation

The question before us now is, what is causing the persistence of south Asian land-surface temperature anomalies with a 1-yr timescale that contributes to setting up land–sea temperature contrast in the TBO? It would appear that some other forcing, possibly from remote sources, could be important.

Evidence has been presented in previous studies of the role of tropical convective heating anomalies in forcing anomalous midlatitude circulation via remote Ross-
by wave response (e.g., Branstator 1983, 1990; Ting 1994) and that such forcing could affect tropical and extratropical circulation having to do with the monsoon (Yasunari and Seki 1992; Nigam 1994; Ting 1994). These studies and others (e.g., Barnett 1988; Chen and Yen 1993, 1994) have shown the possibility of connections between the Tropics and midlatitudes involving the south Asian monsoon. Perhaps heating from anomalous convection associated with the coupled air–sea biennial mechanism summarized in Fig. 1a and documented by Meehl (1987, 1993) could be altering the midlatitude circulation in such a way as to contribute to the persistence of the surface temperature anomalies over south Asia.

To investigate this hypothesis, surface temperature and 500-mb height anomalies from the coupled model, strong minus weak monsoons as in Fig. 6, are calculated and shown in Figs. 7a and 7b for the DJF season prior to a strong monsoon (DJF0 in Fig. 7). Area-averaged interannual standard deviations from the model over south Asia are roughly 1.0°C for surface temperature and about 20 m for 500-mb height. As noted in Fig. 6, temperature differences are positive over south Asia. These are separated from the positive differences over north Asia by small negative values over central Asia. The 500-mb height differences are shown in Fig. 7b. There is a well-organized large-scale pattern with negative differences over the Middle East, positive differences of over 35 m over north Asia, and negative differences over northeast Asia. This is indicative of a midlatitude longwave pattern, such that there is anomalous ridging over Asia. Such a pattern would contribute to maintaining relatively warm surface temperatures over south Asia via southerly advection of warm air under the ridge, as illustrated, for example, by van Loon and Williams (1977).

To compare these model results to recent observations, surface temperature differences from the European Center for Medium-Range Weather Forecasts (ECMWF) analyses (for a description, see Trenberth 1992) for the northern winter of 1988 (i.e., December 1987, and January and February 1988) minus 1987 are shown in Fig. 7c. The 1988 monsoon season was strong, and 1987 had weak monsoon (e.g., Krishnamurti et al. 1989; Krishnamurti et al. 1990). Therefore, the 1988 minus 1987 northern winter differences have the sign of conditions prior to a strong monsoon. Similar to the coupled model composites in Fig. 7a, positive differences of +2°C to +3°C occur over most of south Asia and north Asia, with negative differences over central Asia. Most of Africa is covered by positive differences as well. Associated with these surface temperature differences are the 1988 minus 1987 DJF 500-mb height differences in Fig. 7d. Also similar to the coupled mod-
Fig. 7. (a) Surface temperature anomalies representative of DJF0 (before a strong monsoon) for the coupled model formed from strong minus weak monsoon composites. Positive contours are solid, and negative contours are dashed. (b) Same as (a) except for 500-mb height anomalies. (c) Same as (a) except for observed surface temperature differences DJF 1988 ± 1987. (d) Same as (c) except for 500-mb height differences from the specified heating anomaly model. The anomalous convective heating centers are designated by + and cooling by − in (e) and (f). Major warm surface temperature areas in (a), (c), and (e) are denoted by W. Key negative or anomalously low 500-mb height centers are labeled with an H.

...el, negative differences near 30 m are located over the Middle East, with positive differences of over 100 m over north Asia and negative differences of about 40 m over east Asia, signifying anomalous ridging over Asia in the winter prior to the strong monsoon year (1988) compared to the winter prior to the weak monsoon year (1987).

To test the hypothesis that anomalous heat sources associated with tropical convective anomalies are associated with the anomalous 500-mb circulation and sur-
face temperature anomalies, a global atmospheric GCM is run in perpetual January mode with specified climatological SSTs and anomalous heat sources and sinks. The heat sources corresponding to positive precipitation anomalies are centered at 30°E, 0° over continental Africa and at 160°E, 0° over the Pacific, with an anomalous heat sink (corresponding to negative precipitation anomalies) at 130°E, 0°. These heating anomalies roughly correspond to centers of precipitation maxima and minima in EOF1 from the observed precipitation analysis of Lau and Sheu shown in Fig. 4a, and also match observed northern winter precipitation anomalies from the Kiladis and van Loon analysis in Fig. 2d, as well as similar anomalous precipitation maxima shown by Meehl (1987) for the tropical Pacific. The coupled model EOF1 suggests similar anomalous heating regions (see Fig. 4c).

The anomalous heating in the GCM is specified as in Branstator (1990) and Meehl et al. (1993). The heating is sinusoidal in the vertical, with a midtropospheric maximum of 5°C. The heating decreases linearly away from the central point in all directions to a radius of 1500 km. Figures 7e and 7f show differences for the perpetual January integration, the 795-day average (days 100±2800) from the control integration. Standard deviations are similar in magnitude to those of the coupled model and OLR (taken as a proxy for precipitation EOFs). In general, the African heating anomalies in EOF1 from the observed precipitation analysis of Lau and Sheu shown in Fig. 4a, and also match observed northern winter precipitation anomalies from the Kiladis and van Loon analysis in Fig. 2d, as well as similar anomalous precipitation maxima shown by Meehl (1987) for the tropical Pacific. The coupled model EOF1 suggests similar anomalous heating regions (see Fig. 4c).

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The perpetual January GCM experiment is next configured to have the opposite anomalies to those for the experiment in Figs. 7e and 7f to represent precipitation anomalies for the northern winter following a strong monsoon. Negative heating anomalies (representative of suppressed convection and thus a heating deficit) are centered at 30°E, 0° and at 160°E, 0°, with a positive heating anomaly at 130°E, 0°. Surface temperature differences computed as described above for Fig. 7e appear in Fig. 8e and show that surface temperatures are cooler over much of south Asia and Africa, with values of about −2°C to −6°C. The 500-mb height anomaly pattern in Fig. 8f also shows agreement with Figs. 8b and 8d in that there are positive values over the eastern Mediterranean and the Middle East, negative values of about −70 m over north Asia, and positive values over east Asia, indicative of a shift in the longwave pattern with an anomalous Asian trough with associated cold air advection over south Asia.

A number of other sensitivity experiments were carried out with anomalous equatorial heating and cooling centers shifted to various other longitudes to represent other features noted in the observed and model-simulated precipitation EOFs. In general, the African heating anomaly in concert with either the Indonesian and/or the western Pacific anomaly can produce the appropriate circulation anomalies over Asia. However, the combinations shown here involving heating—cooling anomalies over the Pacific—Australasia and Africa produced the best match to the coupled model and the observed midlatitude response over Asia. This points to the role of African precipitation anomalies (see Hastenrath et al. 1993 for a discussion of East African precipitation regimes) acting in combination with convective heating anomalies over the western Pacific in contributing to midlatitude conditions conducive to subsequent Indian monsoon development.

This analysis shows that by placing tropical convective heating anomalies in a perpetual January version of a GCM, patterns similar to those seen in the 1987–89 sequence of years prior to and following the strong monsoon of 1988 can be reproduced. The question is, are these precipitation anomalies in the coupled model or in the observations that correspond to these convective heating anomalies? To address that question, we compute area-averaged precipitation anomalies from the coupled model and OLR (taken as a proxy for precip-
Fig. 8. Same as Fig. 7 except for the DJF season after a strong monsoon (and before a weak monsoon).

[Diagram of coupled model, DJF+1, and 500 mb z differences with labeled areas and fields.]

(continued)

uation) for the northern winter of 1988 minus 1987 to illustrate conditions prior to a strong monsoon and for 1989 minus 1988 for conditions after a strong monsoon (Table 1). These areas are outlined in Fig. 5 and were chosen based on key regions of interannual variability from the EOF1 in Fig. 4a (observations) and the coupled model (Fig. 4c). In the coupled model for DJF0, the positive precipitation anomalies over East Africa and the western Pacific are +0.35 and +0.24 mm day\(^{-1}\), respectively, and are consistent with positive convective heating anomalies in those regions. The Australian monsoon precipitation anomalies are near zero. Thus two of the three areas (the two positive heating anomalies) are consistent with the optimum specified heating anomaly pattern in Figs. 7e and 7f. For the coupled model DJF+1 season, negative precipitation anomalies exist over East Africa and the western Pacific (−0.24 and −0.08 mm day\(^{-1}\), respectively), with positive anomalies over the Australian monsoon region (+0.36 mm day\(^{-1}\)). These anomalies are consistent with convective heating anom-
alias in the DJF+1 case in Figs. 8e and 8f. For the OLR observations for DJF0 (1988 minus 1987), the East Africa and Australian monsoon areas are consistent with the heat sources and sinks in the atmospheric GCM specified heating experiments. Negative OLR anomalies (indicating intensified convection and precipitation) of $-3.46$ W m$^{-2}$ in DJF0 are present over East Africa, with positive OLR anomalies (indicative of suppressed convection and less precipitation) over the Australian monsoon ($+9.95$ W m$^{-2}$). Anomalies of opposite sign are present in those areas for the DJF+1 season following the strong monsoon ($+5.67$ and $-6.67$ W m$^{-2}$, respectively). The signal is not consistent in the Pacific for the DJF0 season ($+1.38$ W m$^{-2}$) but does go along with the composite picture in DJF+1, with an area-averaged anomaly of $+2.61$ W m$^{-2}$ indicative of suppressed convection. These results are consistent with the specified convective heating anomaly experiments in that some combination of convective heating anomalies over Africa and the western Pacific is likely to be important for the appropriate response of the midlatitude longwave pattern over Asia.

6. The March–May (MAM) season

So far we have focused on the DJF SST and convective heating anomalies and their association with anomalous circulation over the Asian continent. This circulation, if persistent, could set up conditions that would affect the subsequent monsoon season. Ju and Slingo (1995), Soman and Slingo (1997), and Tomita and Yasunari (1996) have pointed to the importance of SST and associated convective heating anomalies in the northern spring season over the southeast Asian region in setting up conditions for either a strong or weak south Asian monsoon the following summer. Meehl (1987, 1993) noted that the northern spring season is a time of transition for SST anomalies in the tropical Pacific region in the TBO cycle. As shown in Fig. 6a, observed Niño3 SST anomalies tend to change sign going from DJF0 to MAM0, thus also changing the pattern of associated convective heating anomalies in the western Pacific. Could the midlatitude circulation patterns over Asia, set up by convective heating anomalies in the DJF0 season, be maintained through the MAM0 season with a different set of SST and associated convective heating anomalies, as noted, for example, by Ju and Slingo (1995) and Soman and Slingo (1997)?

To give a preliminary idea as to whether this is a possibility, the specified convective heating anomaly version of the atmospheric GCM is run to provide an analog to the MAM season in two additional experiments. As indicated by Meehl (1987, 1993) and noted by Ju and Slingo (1995) and Soman and Slingo (1997), conditions in the Pacific region in the MAM0 season (prior to a strong south Asian monsoon) involve relatively warm SSTs over Indonesia and north of Australia (left that way by the weak Australian monsoon during DJF0), while the tropical Pacific SST anomalies transition from relatively warm to relatively cool (see also Fig. 6a). Therefore, in association with those SST anomalies, during MAM0 there should be somewhat stronger convection over the Indonesia area than over the western tropical Pacific. To represent these conditions, an anomalous heat source is placed at $120^\circ$E, $0^\circ$ in the atmospheric GCM. The results in Figs. 7a and 7b show an anomalous ridge over Asia and positive land-surface temperature anomalies similar to those seen for DJF0 in Figs. 7e and 7f. This pattern would be associated with weakened tropospheric westerlies over south Asia that have been shown in observations to be associated with a strong monsoon the following summer (e.g., Yang and Webster 1990; Webster and Yang 1992; Yasunari and Seki 1992; Ju and Slingo 1995; Soman and Slingo 1997; Tomita and Yasunari 1996).

Conversely, for the MAM+1 season following a strong monsoon and preceding a weak monsoon, Niño3 SSTs transition from anomalously cool to anomalously warm (Fig. 6a), while SSTs over Indonesia and north of Australia have been left anomalously cool after the relatively strong Australian monsoon. Thus, a convective heating anomaly placed at $150^\circ$W, $0^\circ$ in the region of anomalously warm SSTs in the tropical Pacific is associated with an anomalous trough over Asia and with relatively cold land-surface temperatures there as well (Figs. 9c and 9d). This would result in anomalously strong midtropospheric westerlies over south Asia prior to a weak monsoon, as noted in observations prior to a weak monsoon (Yang and Webster 1990; Webster and Yang 1992; Yasunari and Seki 1992; Ju and Slingo 1995; Soman and Slingo 1997; Tomita and Yasunari 1996).

Similar patterns for the coupled model and the 1987–89 sequence for the MAM seasons (not shown) seem to suggest that even though the SST anomaly patterns and associated convective heating anomalies in the Tropics transition to different locations going from DJF to MAM, the contribution of the convective heating anomalies to the midlatitude circulation anomalies is still such that the pattern from the preceding DJF season could persist through MAM and thereby maintain the DJF conditions to set up the following monsoon season. A more rigorous examination of the transitions that occur during the MAM season is currently being formulated.

7. Links to other regions

Several key points have emerged from the previous analysis concerning mechanisms involved with the TBO and the Indian monsoon. First, a large part of the global Tropics appear to be involved (see also Ward 1995), and, in particular, African precipitation anomalies may play a role in helping to set up the midlatitude circulation in conjunction with convective heating anomalies in the western Pacific and Australasia. Second, the patterns of precipitation anomalies involved with the TBO
also appear to be involved in a similar way with the Southern Oscillation. This was essentially what was postulated by Meehl (1987), such that excursions of the TBO are made manifest as extremes in the SO. Consequently, anomalies in the climate system in various other regions that are associated with the SO (and by extension the TBO) can be examined to look for similar connections.

It has been noted above that anomalously warm SSTs in the tropical Pacific are often associated with a weak south Asian monsoon. Barnett et al. (1989) used model results to suggest that the remote response from a weak south Asian monsoon could be to enhance warm SST anomalies in the tropical Pacific via surface wind stress anomalies directly forced by the weak monsoon. To further address this issue here, we show surface vector wind anomalies (Fig. 10) from a weak south Asian monsoon sensitivity experiment [induced by increasing land albedos in the atmospheric GCM—see the full description of the experiment in Meehl (1994b)]. SSTs in this sensitivity experiment do not change. The surface wind response in the Indian region shows anomalous northeasterlies, as expected, in association with the weakened monsoon southwestery flow (for the full wind field, see Meehl 1989). However, in the tropical Pacific, there are westerly surface wind anomalies of around 1 m s\(^{-1}\) (about 15% of the magnitude in the control case). Such westerly surface wind anomalies would be conducive to weakened upwelling, reduced latent heat flux, and the advection of western Pacific warm pool surface water to the east. All of these processes would contribute to positive SST anomalies in the central equatorial Pacific. Thus, the model results indicate that a weak monsoon can independently produce surface westerly wind stress anomalies in the equatorial Pacific via the large-scale east–west atmospheric circulation. These wind anomalies can contribute to SST anomalies in the tropical Pacific, thus confirming the earlier results of Barnett et al. (1989) and emphasizing the active role of the south Asian monsoon in the TBO.

Due to the linkage of the Indian monsoon to African precipitation, several additional area averages relating to precipitation anomalies over regions of Africa are calculated. Nicholson (1997), Lamb and Peppler (1991), and others have documented factors that are associated with African precipitation anomalies. Lamb and Peppler (1991) note that strong Sahel rainfall is often associated with a weak subtropical high in the North Atlantic, a
strong subtropical high in the South Atlantic, and a northward shift and intensification of precipitation over West Africa and the Sahel. The two contrasting West African monsoon seasons of the northern summers of 1987 (weak) and 1988 (strong) suggest a link to the 1987 Indian monsoon (weak) and the 1988 Indian monsoon (strong). Consistent with these associations, area averages from the coupled model show that, on average, anomalously high Sahel precipitation (+1.02 mm day$^{-1}$ over the area $10^\circ-20^\circ$N, $20^\circ$W$-10^\circ$E) is associated with strong Indian monsoons along with a weakened subtropical high in the North Atlantic ($-0.3$ mb over the area $20^\circ-40^\circ$N, $70^\circ-10^\circ$W). The opposite is the case for the year following a weak Indian monsoon (Sahel rainfall is anomalously low at $-0.45$ mm day$^{-1}$, and the North Atlantic subtropical high is stronger at +0.7 mb), Lamb and Peppler (1991) and others have also shown that the South Atlantic high is relatively weaker during years of deficient Sahel rainfall and stronger during years of abundant Sahel rainfall. In the coupled model the South Atlantic high shows negative anomalies for both strong and weak Sahel rainfall years, suggesting that the signal is not consistent in this region of the model.

Nicholson (1997) shows that anomalously warm SSTs in the tropical eastern Pacific are often associated with warm SSTs in the Indian Ocean (also shown by Meehl 1987), as well as warm SSTs in the equatorial Atlantic south of about $10^\circ$N. Such an SST pattern has been shown by Lamb and Peppler (1991) and others to be associated with a reduction in Sahel rainfall. The coupled model shows that when Sahel precipitation is strong in the year of a strong Indian monsoon, equatorial Atlantic SSTs are lower by $-0.20^\circ$C over the area $20^\circ$S$-10^\circ$N, $35^\circ-10^\circ$W, while the following year, when Sahel precipitation is weaker, the Atlantic SSTs are warmer at $+0.07^\circ$C.

Previous associations of the monsoon with Southern Hemisphere circulation have shown that a strong Indian monsoon is often associated with stronger subtropical highs in the southern Indian and Pacific Oceans, while the opposite is the case for a weak Indian monsoon (e.g., Meehl, 1987, 1988; Clemens and Oglesby 1992). This is also the case in the coupled model with area-averaged sea level pressure in the subtropical high regions of the south Indian ($20^\circ-40^\circ$S, $60^\circ-100^\circ$E) and the South Pacific ($20^\circ-40^\circ$S, $90^\circ-150^\circ$W) Oceans during the JJA season at $+0.4$ and $+0.5$ mb, respectively. In the year following a relatively weak Indian monsoon, these two subtropical highs are also weakened at $-0.1$ and $-0.4$ mb, respectively.

South African precipitation during southern summer has also been shown to be related to SSTs in the tropical Pacific. Kiladis and Diaz (1989) show deficient South African precipitation during years of warm SSTs in the tropical eastern Pacific prior to a strong Indian monsoon. This is also the case in the coupled model, as was demonstrated first by Meehl et al. (1993) and also shown here. In the southern summer prior to a strong Indian monsoon, South African precipitation is deficient in the coupled model ($-0.60$ mm day$^{-1}$ over the area $40^\circ-20^\circ$S, $10^\circ-50^\circ$E), while for southern summers following a strong Indian monsoon, South African precipitation anomalies are positive ($+0.49$ mm day$^{-1}$).

Other studies have also linked Pacific SSTs to precipitation over northern South America (e.g., Kiladis and Diaz 1989; Ropelewski and Halpert 1987, 1989; Marengo and Hastenrath 1993). Meehl et al. (1993) showed this linkage for ENSO-like events in the coupled model. But for the monsoon definition base used here,
that association is not consistent with deficient South American rainfall both prior to and after a strong Indian monsoon.

8. Conclusions

The tropospheric biennial mechanism (TBO) involving atmosphere–ocean coupled processes postulated by Meehl (1987, 1993, 1994a) is reviewed. Premises involving dependence on the seasonal cycle of convection being strong during one season of the year at any given location in the Indian and Pacific sector, air–sea coupling being strongest during that season, and upper-ocean heat content providing the memory to maintain SST anomalies for 1 yr are all shown to have some basis in observations. The sign of South Asian land temperature anomalies, and tropical Indian and eastern Pacific SST anomalies all vary roughly in phase within about a season over the duration of the seasonal cycle beginning in about June.

It is noted that earlier studies of the TBO showed a connection with the south Asian or Indian monsoon, but it was unclear whether the monsoon was an active or passive participant. A biennial mechanism involving coupled atmosphere–land–ocean interactions is postulated that involves the monsoon as an active participant in the TBO.

A 50-yr time series from a global coupled ocean–atmosphere GCM is analyzed to look for evidence of this proposed mechanism. The simulated monsoon is shown to have links to the model’s version of ENSO in the Pacific not unlike the observed system, but the amplitude of the variability, and thus the strength of those linkages, is weaker than in the observations.

Results from the coupled model for strong minus weak monsoon years demonstrate similarity to the observed relationships in terms of the biennial mechanism. However, since the soil moisture anomalies that occur in conjunction with heavy rainfall during a strong monsoon only last for one additional season, it is suggested that remote processes, not local hydrology, must be contributing to maintaining anomalous land temperatures over south Asia on a 1-yr timescale. It is hypothesized that anomalous heat sources and sinks associated with SST and convective anomalies produced by air–sea interaction and east–west atmospheric circulation over the tropical Indian and Pacific sectors involved with the atmosphere–ocean biennial mechanism may be forcing circulation anomalies in the midlatitudes via remote Rossby wave response. Such forcing could alter the mid-latitude circulation in such a way as to maintain the surface temperature anomalies over south Asia (which are essential to the atmosphere–land–ocean biennial mechanism).

To test this hypothesis, a global GCM with specified SSTs, as well as static soil moisture and snow cover, is run in perpetual January mode with heat sources and sinks specified to correspond to key precipitation anomalies in the coupled model, as well as to conform to heating anomalies suggested by the first EOF of observed tropical precipitation and similar patterns from other observed studies. It is found that by specifying a combination of heat sources and sinks over Africa and the western Pacific Ocean, anomalous surface temperature signals over Asia resemble those from the coupled model, as well as from the ECMWF analyses from a recent weak to strong monsoon evolution (1987 to 1988). An additional GCM simulation with specified climatological SSTs shows that westerly surface wind anomalies in the equatorial Pacific develop as a direct consequence of a weak south Asian monsoon induced in a sensitivity experiment. This link between south Asian monsoon strength and conditions that could intensify SST anomalies in the tropical Pacific agrees with previous studies and further points to the active role of the south Asian monsoon in the TBO and ENSO.

Additional precipitation and sea level pressure anomalies are noted from previous observational studies and identified in the coupled model to show the probable near-global nature of the mechanisms involved with the TBO. However, of crucial importance for the monsoon biennial mechanism is the combination of convective heating anomalies over Africa and the western Pacific that involves alterations of the midlatitude circulation over Asia such that subsequent monsoon development is affected.

To summarize the mechanisms involved with the TBO, a schematic evolution of the features noted in Table 1 and in Figs. 6–9 is shown in Fig. 11. Though associations with rainfall in other parts of Africa, SST anomalies in the Atlantic, and subtropical SLP anomalies in the Southern Hemisphere have been noted above for the coupled model, we limit this schematic to the strongest associations in the East Africa–India–Pacific sectors. It is likely that the TBO could be characterized as a Tropics-wide or even larger-scale oscillation (e.g., see Ward 1995), but in this paper we have chosen to focus on the sectors illustrated in Fig. 11. Year 0 and year +1 indicate calendar years of strong and weak monsoons, respectively. The monsoon seasonal cycle [and also the TBO seasonal cycle, as noted by Meehl (1987)] is associated with transitions late in northern spring with the beginning of the south Asian monsoon season. That is the starting point of a set of seasonally linked anomalies until late the following northern spring (e.g., Meehl 1987; Yasunari 1991), which is arbitrarily denoted in Fig. 11 by stippling for TBO seasonal cycle 1 and nonstippled for TBO seasonal cycle 2.

Since Fig. 11 depicts an idealized TBO cycle, the starting point for discussion is arbitrary. We choose to begin during the northern winter DJF0 before a strong monsoon (at the right side of Fig. 11), when there is relatively heavy rainfall over East Africa and the tropical Pacific, and a weak Australian monsoon. The SST anomaly pattern involves positive values in the Indian
Fig. 11. Schematic of idealized TBO seasonal evolution. Year 0 refers to the calendar year of the strong monsoon; year -1 refers to calendar year of following year with a weak monsoon. Stippled area indicates year 1 of the TBO seasonal cycle, going from late northern spring of year 0 to late northern spring of year +1. An idealized TBO seasonal cycle 2 would follow TBO seasonal cycle 1 and proceed from the late northern spring of year +1 to the late northern spring the following year.
and Pacific, with negative anomalies north of Australia. The SST and associated convection anomalies have been set up in the previous seasons in the TBO cycle. The tropical convective heating anomalies during DJF are associated with a midlatitude longwave pattern with an anomalous ridge over Asia. This ridge is associated with warmer and drier conditions over Asia and thus decreased snow cover.

By the following northern spring season (MAM at the upper right in Fig. 11), the weak Australian monsoon has left the SSTs over Indonesia and north of Australia relatively warm, while in the tropical Pacific, the SSTs have become relatively cool. Thus, as the ITCZ heads north, there is relatively weaker convection over the Pacific, with somewhat stronger convection over the Indonesian region. This anomalous convective heating pattern maintains the midlatitude circulation pattern from the prior DJF season in Fig. 11a, with an anomalous ridge, weakened midtropospheric westerlies over south Asia, and relatively warm land-surface temperatures.

The following monsoon season (JJA in Fig. 11c), the anomalously warm south Asian landmass has set up enhanced land–sea or meridional tropospheric temperature contrast. This contributes to a strong monsoon, while air–sea interaction produces an anomalous SST pattern, with cool SSTs over the tropical Indian and Pacific regions and with anomalously warm SSTs over Indonesia and north of Australia. As the seasonal cycle of convection proceeds to the following northern fall (SON at the upper left in Fig. 11), the relatively strong convective maximum moves south and east with the seasonal cycle over the tropical Pacific, with somewhat stronger convection over the Indonesian region. The seasonal cycle of convection over Asia from the prior DJF season in Fig. 11a, a strong East Asian monsoon has set up over the anomalously warm SSTs in that region, while the large-scale east–west circulation in the atmosphere contributes to suppressed convection over Asia and the central tropical Pacific (where SSTs remain anomalously cool). This combination of convective heating anomalies in the Tropics is then associated with a midlatitude circulation that includes an anomalous trough over Asia. The associated cold air advection from higher latitudes maintains a cold south Asian landmass with increased snow cover.

In the subsequent northern spring season (MAM in the lower left in Fig. 11), the SSTs in the tropical Pacific transition to positive anomalies and the convective maximum is somewhat stronger over that region, with suppressed convection over the Indonesian region. These convective heating anomalies contribute to maintaining the midlatitude circulation pattern from the preceding northern winter, with an anomalous trough over south Asia and increased midtropospheric westerlies. The anomalously cool south Asian land temperatures maintain a decreased land–sea or meridional tropospheric temperature gradient that contributes to a weak monsoon the following northern summer (JJA+1 at the bottom in Fig. 11). As this weak convective maximum moves south and east with the seasonal cycle over the relatively cool SSTs near Indonesia in the following northern fall (SON+1 in the lower right in Fig. 11), the weak monsoon the preceding season has left the south Asian landmass relatively dry and warm, and this idealized version of the TBO cycle continues with conditions the following northern winter, as shown at the right in Fig. 11.

Of course, the TBO cycle depicted in Fig. 11 is not perfectly biennial, as noted in other studies, and only appears to operate intermittently or in a subset of years (Meehl 1987; Barnett 1991; Terray 1995). However, recent studies have shown that these associations involving midlatitude circulation patterns and monsoon strength are strong enough to be evident in composites of a number of observed strong and weak monsoon years (e.g., Yang et al. 1996).

Asian snow cover appears to contribute positively to the land temperature anomalies. However, the specified heating experiments demonstrate that tropical convective heating anomalies alone can produce circulation patterns over Asia that in turn produce anomalously warm or cold land-surface conditions that contribute to land–sea temperature contrast without any feedbacks from snow cover (snow areas and surface moisture are specified and cannot change in that model). For example, for the two contrasting observed monsoon years considered in this paper, spring snow cover over Eurasia was above normal prior to the weak 1987 monsoon but near normal before the strong 1988 monsoon, even though Eurasian surface temperature anomalies were below normal before the 1987 monsoon and above normal prior to the 1988 monsoon (Halpert and Ropelewski 1991). Therefore, changes of snow cover appear to be in part an artifact of the midlatitude circulation pattern associated with convective heating anomalies, rather than an independent forcing.

The northern spring season is important for a number of reasons. This is the beginning of the time of transition for SST anomalies in the tropical Pacific, as shown in Figs. 6a and 11. This transition appears to be necessary for the establishment of convective heating anomalies that could help maintain the midlatitude circulation pattern over Asia and thus contribute to subsequent monsoon development. Perhaps this is why the northern spring season is problematic for forecasting many features in the tropical Indian–Pacific region and is sometimes referred to as the “spring predictability barrier” (e.g., Webster and Yang 1992; Lau and Yang 1996). The mechanisms of this transition involve ocean dynamics and coupled air–sea interaction in the tropical Pacific, as well as features of the large-scale east–west circulation in the atmosphere involving the TBO. Future analyses will aim to describe this transition more fully, as well as the role of tropical forcing in the extratropical
circulation and the interaction of processes over various other regions of the globe that produce the TBO.

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