Sensitivity of the Asian summer monsoon to aspects of sea-surface-temperature anomalies in the tropical Pacific Ocean

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SUMMARY

The response of both the time of onset, and the strength of the Asian summer monsoon, to regional aspects of the sea surface temperature (SST) anomalies in the tropical Pacific Ocean associated with El Niño/Southern Oscillation (ENSO) has been investigated through a series of sensitivity experiments with the Universities’ Global Atmospheric Modelling Programme (UGAMP) General Circulation Model (UGCM). This paper builds on the work of Ju and Slingo (1995) and on their hypothesis that the relationship between the Asian summer monsoon and ENSO involved the latitudinal position and strength of the tropical convective maximum (TCM) over Indonesia and the west Pacific in the preceding spring. The inference from their results was that the modulation of the TCM might be associated either with changes in the Walker circulation through the influence of the east Pacific SST anomalies, or with changes in the local Hadley circulation associated with the in situ SST anomalies in the west Pacific. The investigation has focused on the particular contrasting years of 1983 and 1984. The experiments described in this paper are designed to isolate the effects of the principal SST anomalies in the east and central Pacific, associated with El Niño/La Niña, from those of opposite sign which develop in the west Pacific as a complementary pattern during the mature phase of El Niño/La Niña.

The results of the experimentation suggest that, at least for the test cases of 1983 and 1984, the modulation of the Walker circulation, with implied additional subsidence over the eastern hemisphere, is the dominant mechanism whereby the Asian summer monsoon is weakened during El Niño years. However, the late onset during El Niño years may also be associated with the complementary cold SST anomalies in the west Pacific which delay the northwards transition of the TCM. During La Niña, the modulation of the Walker circulation appears not to be the controlling factor which determines the stronger monsoons. The UGCM results suggest that the complementary warm SST anomalies in the west Pacific enhance the TCM, and it is this in situ response by the TCM that leads to an early onset and stronger monsoon. The importance of warm anomalies in the west Pacific in the development of a strong monsoon has been investigated further through a case-study of the 1994 season. The year 1994 was an El Niño year in which the monsoon was unexpectedly active, but which was also marked by warmer than normal SSTs in the west Pacific.

The sensitivity experiments have also elucidated the role of El Niño in influencing the precursory signature of stronger subtropical westerlies over India and south-east Asia during the winter and spring preceding weak monsoons. The results suggested that the equatorwards shift of the subtropical jet is a remote response to the warm SST anomalies in the central and east Pacific associated with El Niño.

KEYWORDS: Asian monsoon El Niño General circulation model Walker circulation

1. INTRODUCTION

The Asian summer monsoon is a major component of the general circulation. It influences over one third of the tropical region for several months during the northern summer, and involves a substantial exchange of air between the southern and northern hemispheres. From June to September, the rainfall associated with the monsoon provides the main source of fresh water for millions of people in India and south-east Asia, and the economies and livelihood of these countries depend heavily on the reliability of the return of the summer rains. However, the monsoon exhibits substantial interannual variability in its time of arrival and in its seasonal mean rainfall. For example, in 1995 the onset of the monsoon over India was later than normal by about one week and its subsequent progression northwards was slower than usual. This, combined with an unprecedented heatwave, led to severe water and electricity shortages over much of India, and to considerable hardship for many people.

The interannual variability in the seasonal mean rainfall over India has been related to the phase of the El Niño/Southern Oscillation (ENSO) (e.g. Rasmusson and Carpenter

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1983; Webster and Yang 1992). In essence, drought years over India are often, but not exclusively, related to warm sea surface temperature (SST) anomalies in the equatorial central and east Pacific (El Niño), and wet years with abnormally low SSTs (La Niña). However, there are notable exceptions. For example, in 1994 the southern oscillation index (SOI) was negative and warm SST anomalies prevailed in the central and east Pacific. On the basis of these El Niño conditions, a slightly weaker than normal monsoon was predicted for India. In fact the monsoon proved to be one of the strongest in recent years with a seasonal mean rainfall that was over 10% higher than normal.

As well as finding that the monsoon strength was sensitive to the phase of ENSO, Joseph et al. (1994), on the basis of the rains over Kerala, southern India, found that the date of monsoon onset tended to be delayed significantly in El Niño years. The ability to predict the timing as well as the overall strength of the monsoon is important for local agriculture since the planting of crops depends crucially on when the rains occur.

Despite the clear need for prediction of the interannual variability of the monsoon, the results of a comprehensive study of the behaviour of thirty-three General Circulation Models (GCM) of the atmosphere which participated in the Atmospheric Model Intercomparison Project (AMIP; Sperber and Palmer 1996) have shown that, so far as Indian rainfall is concerned, the predictive skill of the GCMs was very limited. However, the models showed much more predictive skill when assessed in terms of the interannual variation of the broad-scale monsoon circulation. A clear relationship between the strength of the large-scale monsoon circulation and the phase of ENSO has been suggested from the results with the models (Ju and Slingo 1995; hereafter referred to as JS) and from the NWP analyses (Webster and Yang 1992; hereafter referred to as WY). JS concluded that those years with a delayed onset and weaker monsoon circulation tended to have warm SST anomalies in the central and east Pacific during the preceding spring, whereas those years with an earlier onset and a stronger monsoon circulation tended to have cold SST anomalies. In other words, weak monsoons were related to El Niño conditions and strong monsoons to La Niña conditions.

Using the results of an AMIP integration with the Universities’ Global Atmospheric Modelling Program (UGAMP) General Circulation Model (UGCM), JS compared the interannual variability of the simulated large-scale monsoon circulation with that from the European Centre for Medium-Range Weather Forecasts (ECMWF) analyses (their Fig. 7) and concluded that, not only did the model display a fair degree of skill, but also that the reproducibility of the monsoon index suggested a dominant role for the imposed boundary forcing of the tropical SSTs. JS also noted, in agreement with WY, that the greatest difference in the strength of the monsoon flow occurred during the onset of the monsoon but did not persist throughout the whole season. As Fig. 1 demonstrates, the dynamical monsoon index is even more reproducible for the months of May and June alone, than for the whole season (May to August) as shown by JS (their Fig. 7). JS therefore concluded that the main influence of ENSO on the Asian summer monsoon was associated with the strength and timing of the dynamical onset of the monsoon. This implied that changes in circulation patterns associated with the equatorial Pacific SST anomalies, in the preceding winter and spring, were possibly responsible.

JS argued that the developing phase of the monsoon was dependent on the strength and position of the heating maximum over Indo-China and the maritime continent. They proposed that the influence of El Niño on the large-scale monsoon could be explained in terms of the modulation of the latitudinal position and strength of the tropical convective maximum (TCM) over Indonesia and the west Pacific in the preceding spring. Figure 2 shows the composite outgoing longwave radiation (OLR) anomalies from the National
Figure 1. Bar chart of the monsoon strength during the onset phase, May and June, from ECMWF analyses and as simulated by the UGCM for the AMIP decade, 1979–1988. (Note that no analyses were available for 1979, and that the UGCM index for that year was zero.) The monsoon strength is inferred from a dynamical index which considers the anomalous vertical shear of the zonal wind over south-east Asia (0°–20°N, 40°E–110°E).

Figure 2. Composites of the mean outgoing longwave radiation anomalies (W m⁻²) observed by the NOAA AVHRR satellite for March, April, May preceding (a) weak monsoon years, and (b) strong monsoon years. The contour interval is 5 W m⁻² with values less than −5 W m⁻² shaded. No zero contour is drawn and positive contours are dashed.
Oceanographic and Atmospheric Administration (NOAA) AVHRR* satellite data for the springs preceding both weak and strong monsoons; these clearly demonstrate a substantial modulation of the TCM over the west Pacific and Indonesia. This variation in the position of the TCM may arise indirectly from changes in the Walker circulation in response to SST anomalies in the central and east Pacific. This has generally been accepted as a mechanism to explain the weaker monsoons in El Niño years (e.g. Palmer et al. 1992). However, JS found much less perturbation in the divergent circulation during La Niña years, which led them to postulate that the modulation of the strength and location of the TCM may also be related to changes in the local Hadley circulation as a direct response to in situ SST anomalies in the Indonesian and west Pacific region.

The SST anomaly patterns associated with El Niño are highly variable and depend on whether the El Niño is in its growing, mature or decaying phase. Although El Niño is traditionally thought of as a phenomenon that affects the east Pacific, it is now recognized that there are associated coherent patterns in the SST anomalies over the global tropical and subtropical oceans. Although the largest anomalies are seen in the equatorial central and eastern Pacific, a well-defined and coherent pattern of SST anomalies, with opposite sign, prevails over the western and subtropical Pacific (see JS; their Fig. 16). The patterns of the response by the convection, implied by the OLR anomalies (Fig. 2), suggest that these local SST anomalies may be instrumental in determining the latitudinal position and strength of the TCM.

To investigate the relative roles of these remote and direct responses in determining the behaviour of the monsoon, a series of sensitivity experiments with the UGCM were performed in which elements of the SST anomaly patterns in the tropical Pacific were removed. These experiments were intended to isolate the effects either of the warm anomalies or of the cold anomalies, within the El Niño or La Niña patterns, on the onset and strength of the monsoon flow. The approach taken has been to use two particular case studies, in this instance for the years 1983 and 1984, and to understand the mechanisms operating in those years rather than to use a composite or ensemble approach. Each El Niño event develops differently and, as will be shown at the end of the paper, it is details in the SST anomaly fields which can have a strong influence on the monsoon for a particular year. It is the intention that the increased understanding that is gained from these case-studies can be applied to the wider problem of seasonal monsoon prediction.

In section 2, the details of the UGCM used in the study are given. A detailed description of the design of the SST experiments is presented in section 3. The results from various integrations of the model are described in section 4 and, in section 5, the implications of the results are discussed and related to the anomalous monsoon of 1994.

2. Description of the model

A full description of the UGCM has been given by Slingo et al. (1994), and its application to the AMIP is described by JS. The model is based on cycle 27 of the ECMWF forecast model but with various modifications as documented by Slingo et al. (1993). The most significant of these, for the research described here, was the incorporation of the convective adjustment scheme developed by Betts and Miller (1993), in which both shallow and deep convection are parametrized in terms of a relaxation towards observed thermodynamic structures. As in the AMIP integration, the model has been integrated with 19 levels in the vertical, represented by a hybrid sigma–pressure coordinate. In the horizontal, the prognostic variables are represented by a truncated series of spherical

* Advanced very high resolution radiometer.
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Harmonics with a resolution of T42; that is isotropic or 'triangular' truncation at total wavenumber 42. Contributions to tendencies from parametrization schemes are calculated on a Gaussian grid of 128 (longitude) by 64 (latitude) points, where the mesh size is approximately 2.8 degrees. The basic features of the UGCM are summarized in Table 1.

The sensitivity experiments described in this paper are based on a '10-year' integration of the UGCM, with the observed SSTs for 1979–88, performed as part of the AMIP. As described by JS, the UGCM has successfully captured the large-scale features of the Asian summer monsoon and the rapid transition of the flow during northern spring and summer. In individual years the onset of the monsoon over the Indian peninsula is realistic, although the simulated monsoon has a tendency to retreat prematurely by the end of August. Whilst the large-scale monsoon circulation is well simulated by the model, the precipitation distribution is less satisfactory. Following the onset, the model rapidly develops a monsoon break, with the main areas of precipitation occurring over the equatorial Indian Ocean and up against the Tibetan plateau. As a result, the precipitation over India is substantially underestimated, in common with a number of GCMs which participated in the AMIP (Sperber and Palmer 1996). Thus, as discussed by JS, the use of the seasonal mean rainfall over India as a measure of monsoon interannual variability may be problematic. However, as is evident in Fig. 1, the UGCM has been successful in reproducing the interannual variability of the large-scale monsoon circulation and it is this measure of the monsoon activity which has been used in this paper.
3. Experimental design

Figure 1 shows that the simulated monsoon circulation was at its weakest during the El Niño year of 1983, and at its strongest during the La Niña year of 1984. Also, as shown in Fig. 7(a) of JS, the simulated monsoon onset was very late in 1983 and very early in 1984. These extreme years therefore were chosen to investigate the relative roles of warm and cold SST anomalies, in the tropical Pacific Ocean, related to the El Niño pattern.

For both 1983 and 1984, a series of integrations were performed to separate the effects of the principal SST anomalies in the central and east Pacific associated with El Niño, versus those in the western and subtropical Pacific which develop as a complementary pattern during the mature phase of El Niño. Integrations were performed with ‘warm-only’ or ‘cold-only’ tropical Pacific SST anomalies, where the SST anomalies were modified over the tropical Pacific Ocean only between latitudes 30°N and 30°S. In the ‘warm-only’ case, all the negative SST anomalies in the tropical Pacific Ocean were set to zero. Conversely in the ‘cold-only’ case, all the warm anomalies were set to zero. To allow a smooth transition between the original SST anomaly field outside the tropical Pacific and the modified field within the tropical Pacific, a linear interpolation between the original and modified fields was carried out, covering three model grid points at the edge of the tropical Pacific domain.
The SST anomalies over the tropical Pacific were changed for each month of 1983 and 1984, and Figs. 3 and 4 give examples of the distribution of SST anomalies for May for the El Niño (1983) and La Niña (1984) cases, respectively. In each case the anomalies have been computed against the 10-year mean for the AMIP period.

The sensitivity experiments to study the effect of either ‘cold-only’ or ‘warm-only’ SST anomalies used, as initial data, the history data for 1 January for the relevant year (1983 or 1984) from the AMIP integration. Each integration continued through to the end of September with seasonally evolving SSTs, the SSTs being updated daily by interpolation between adjacent months. A control integration was also performed using the 10-year mean monthly SSTs, thus providing a measure of the strength of the monsoon under mean conditions. This experiment was initiated on 1 October and integrated for 12 months. (Note that the control integration had a different initial time because it had been completed for another purpose prior to these experiments.) The integration with ‘all anomalies’ is the AMIP integration already described by JS.

4. RESULTS

In the following discussion, the sensitivity of the simulated monsoon to different SST anomaly distributions is assessed in terms of the strength of the monsoon flow rather
than changes in the regional precipitation over India. As noted in the Introduction, the UGCM possesses reasonable skill in reproducing the response of the large-scale monsoon circulation to interannually varying SSTs. In the following subsections the results have been divided up to emphasize the distinct roles of the SST anomalies in determining the strength of the monsoon, the timing of the onset, and the perturbation to the upper tropospheric zonal winds, particularly during the preceding winter and spring.

The onset of the Asian summer monsoon is marked by a rapid strengthening of the lower tropospheric westerlies and upper tropospheric easterlies, with an associated increase in moisture content of the atmosphere and rainfall. The onset and subsequent strength of the monsoon can be described by the evolution of the kinetic energy (KE) of the 850 hPa winds, averaged over a region covering India and the Bay of Bengal (5°N–20°N, 60°E–100°E). In this study the evolution of the daily KE will be used to describe the response of the simulated monsoon to the various SST anomaly distributions. Whilst KE is a good descriptor of the evolution of the large-scale monsoon it can also be related to the onset date of the monsoon over India (Soman and Kumar 1993), and thus is an appropriate parameter with which to relate the model’s behaviour to the observed variation in onset date.

(a) Influence of SST anomalies on the strength of the monsoon

The daily time series of the KE at 850 hPa from April to September for both the 1983 and 1984 experiments are presented in Fig. 5. Note that the data have been smoothed with a 15-day running mean to eliminate short timescale variations. For comparison, the KE from climatological mean SST integration has been included in each set of curves. Owing to the much stronger simulated monsoon in 1984, different ordinate scales have been used in Figs. 5(a) and 5(b).

Associated with the onset of the monsoon, all the integrations display a rapid increase in KE in late May or early June, from a typical pre-monsoon value of less than 10 m² s⁻². A peak value of KE is usually reached in late June or early July; thereafter, the KE shows a gradual decrease. In all cases the withdrawal of the monsoon occurs about a month earlier in the UGCM than observed. In this study the emphasis will be placed on the sensitivity of the onset and peak strength of the monsoon to the imposed SST anomalies, following the findings of WY and JS which showed that the influence of ENSO is seen primarily during the early part of the monsoon and not during the retreat phase.

The results from the suite of experiments for 1983 (the El Niño case) show that the low-level flow in the integration with observed SSTs for that year (‘all anomalies’) is substantially weaker than that simulated by the UGCM with climatological mean SSTs. This is consistent with the monsoon indices shown in Fig. 1. As Fig. 5(a) demonstrates, in the ‘warm-only’ integration the evolution of the monsoon is hardly affected when the cold anomalies in the west and subtropical Pacific are removed and only the warm anomalies in the east and central Pacific are imposed (Fig. 3(a)). In contrast, when the warm anomalies (i.e. the main El Niño signal) are removed in the ‘cold-only’ integration (Fig. 3(b)), the strength of the monsoon increases dramatically and is restored to that simulated by the control integration with climatological mean SSTs.

The implication from the results shown in Fig. 5(a) is that the weakening of the monsoon in 1983 is associated with the warm SST anomalies in the central/east Pacific rather than with the cold SST anomalies in the west Pacific. This suggests that in 1983, and possibly in other El Niño years, the strength of the monsoon is determined by a modulation of the Walker circulation in which there is implied additional subsidence over the west Pacific and south-east Asia (see Fig. 20 of JS). This modulation of the Walker circulation arises from the systematic eastwards shift of the convection (Fig. 2(a)) in response to the extension of the warm pool into the central Pacific associated with the principal warm
anomalies in the El Niño pattern. It is generally accepted that this change in the Walker circulation is the main reason for the tendency for monsoons to be weaker during El Niño years (e.g. Palmer et al. (1992)).

The results from the suite of experiments for the La Niña case (1984) give a very different picture. The year 1984 is characterized by a stronger than normal monsoon circulation (Fig. 1), and the KE from the integration with ‘all anomalies’ is much larger than the KE simulated by the control integration with climatological SSTs (Fig. 5(b)). When the main La Niña pattern of cold anomalies in the central and eastern equatorial Pacific is removed (i.e. the ‘warm-only’ integration; Fig. 4(a)), there is very little impact on the peak KE achieved in the simulated monsoon. This is in stark contrast to the results for
1983 (Fig. 5(a)), where removal of the main El Niño signature in the SSTs had a substantial impact on the strength of the monsoon flow. On the other hand, in the La Niña case, the warm anomalies in the west Pacific, which are part of the remote pattern associated with La Niña, appear to be considerably more important. As Fig. 5(b) demonstrates, the ‘cold-only’ integration gives a monsoon flow whose strength is much weaker and near to that simulated by the control integration. These results suggest that the tendency for stronger monsoons in La Niña years may be associated with the warm SST anomalies in the west Pacific and not through a modification of the Walker circulation, as in the El Niño case described above. This result supports the hypothesis of JS that warm anomalies in the west Pacific, even when small, can have a direct effect on the monsoon. The reasons for this are discussed in section 5.

(b) Influence of SSTs on the timing of the onset of the monsoon

Joseph et al. (1994) made a comprehensive study of the role of SST anomalies in determining the timing of the onset of the monsoon. They found that the onset tended to be significantly delayed in El Niño years, particularly when the mature phase of El Niño was reached during the preceding winter and spring (e.g. as in 1983). As discussed in the Introduction, the UGCM shows a similar sensitivity of the onset date to the phase of ENSO. Joseph et al. (1994) used the rainfall over Kerala, southern India, as an indicator of the onset of the monsoon. For the purposes of the present study, a sharp and sustained increase in the KE at 850 hPa to above 10 m$^2$ s$^{-2}$ is considered, in the model, as the onset of the large-scale monsoon. This value of the KE is close to the average observed value in the layer 1000–700 hPa over peninsular India during the onset of the monsoon over Kerala (Soman and Kumar 1993). This criterion gives the onset date for the control integration, with climatological mean SSTs, as 29 May, which is nearly the same as the mean observed onset date of 30 May over the southern tip of India (Ananthakrishnan and Soman 1988), derived on the basis of precipitation alone.

For the El Niño case (Fig. 5(a)), the simulated onset is late (9 June) and remains the same irrespective of whether the Pacific SSTs include ‘all anomalies’ or only parts of the anomalies (i.e. ‘cold-only’ or ‘warm-only’). This is an intriguing result which suggests that both the modulation of the Walker circulation by the east and central Pacific warm anomalies, and the direct effect on the convection of the cold anomalies in the west Pacific, can influence the timing of the onset. Joseph et al. (1994) assumed that the main reason for the delayed onset in El Niño years was the predominance of warm water in the central equatorial Pacific, whereas these results suggest that the cold anomalies in the west Pacific, which are also part of the El Niño pattern, may be equally responsible.

The results for the strong monsoon year (1984) are quite different (Fig. 5(b)). The date of onset is far more variable, with the earliest onset associated with the ‘all anomaly’ integration. The results from the ‘warm-only’ integration show that the onset date is still advanced by about 10 days, although the main acceleration of the flow occurs somewhat later. When the warm anomalies in the west Pacific are removed (i.e. ‘cold-only’) then the onset is delayed substantially. These results again suggest the important role of the warm, but small amplitude, anomalies in the west Pacific. They are consistent with those of Joseph et al. (1994) who found that the strongest correlations between SST and onset date occurred over the west Pacific, north of the equator.

The intransigence of the onset date in the El Niño suite of experiments (Fig. 5(a)) is an interesting result which merits further study. In the following sections a series of sensitivity experiments are described which are aimed at elucidating the role of the cold anomalies in the west Pacific in determining the onset date during El Niño conditions.
(i) Sensitivity to initial conditions. As noted in section 3, the initial conditions for all the sensitivity experiments for the El Niño case were taken from the state of the model on 1 January 1983 from the AMIP integration with ‘all anomalies’. Since El Niño had already been in place for several months, the model’s atmosphere and surface will almost certainly contain the signature of El Niño. For example, JS showed that the UGCM simulates a systematic equatorwards shift of the upper tropospheric subtropical jet during the preceding winter and spring, in association with El Niño. It may be possible therefore that in the ‘cold-only’ integration, in particular, a residual effect of El Niño may persist through into the spring and may thus influence the timing of the onset.

To test whether this might be the case, the ‘cold-only’ integration was repeated with initial conditions for 1 January 1984 from the AMIP integration. This represents a fairly extreme change in the initial conditions because January 1984 was already under the influence of the developing La Niña. Figure 6 shows the sensitivity of the evolution of the KE at 850 hPa to this change in initial conditions. Not only is the strength of the monsoon similar but the onset date remains delayed, when compared with the integration with climatological SSTs. This result confirms that the initial conditions have little effect on the onset of the monsoon and suggests that the SST anomalies are the controlling factor. Furthermore it supports Joseph et al.’s contention that SST anomalies in the west Pacific may be an important factor in determining the date of onset.

Figure 6 also includes the KE evolution from the ‘warm-only’ experiment, which shows that the two integrations with ‘cold-only’ anomalies, but different initial conditions, are much nearer to each other than to the ‘warm-only’ integration, at least in the early stages of the monsoon. This is a reassuring result, suggesting that the peak strength of the simulated monsoon may also be relatively insensitive to the initial conditions, and implying that the inferred influence of the SST anomalies on the strength of the monsoon, discussed in section 4(a), may be fairly robust. An ensemble of integrations with differing initial conditions would be needed to show this conclusively, but to do this was beyond the computing resources available at the time of this study.
Figure 7. Example of the SST anomaly pattern for May 1983 showing typical distributions for the sensitivity experiment with no anomalies in the tropical Indian Ocean and 'cold-only' anomalies in the tropical Pacific Ocean. Contours are drawn at ±0.2, 0.5, 1, 1.5, 2, 2.5, 3 and 4 K. Positive values in excess of 0.2 K are shaded. The zero contour is dotted.

Figure 8. Temporal evolution of the kinetic energy at 850 hPa (m² s⁻²) for the region 5°N–20°N, 60°E–100°E, from the 'cold only' integrations for 1983 with and without tropical Indian Ocean SST anomalies. For comparison the results from the integration with climatological SSTs are also shown.

(ii) **Sensitivity to Indian Ocean SST anomalies.** The previous experiments have all been concerned with the influence of the Pacific SST anomalies on the evolution of the monsoon. However, it is well known that the Indian Ocean SSTs display specific anomaly patterns associated with El Niño. Although these anomalies are usually less than those in the Pacific, Joseph *et al.* (1994) suggested that the pattern of positive SST anomalies in the equatorial Indian Ocean and negative SST anomalies in the Arabian Sea, associated with El Niño (see Fig. 3(a), for example), may also cause a delay in the onset of the monsoon. To test this hypothesis, a parallel integration to the 1983 'cold-only' case was performed in which the SST anomalies in the Indian Ocean and Arabian Sea were set to zero (Fig. 7).

Figure 8 shows the evolution of the KE at 850 hPa from the 'cold-only' integrations with, and without, Indian Ocean SST anomalies. The results from the integration with climatological SSTs are also included for comparison. Again the onset date is delayed, the
intrinsigence of the onset behaviour supporting the hypothesis that the west Pacific SST anomalies could be a controlling factor in the timing of the monsoon. The result further confirms the observation by Joseph et al. (1994) that the only area where a significant correlation between SST and onset date was found, was in the west Pacific, north of the equator. The response of the monsoon strength to the removal of the Indian Ocean SST anomalies, implied by Fig. 8, is a potentially interesting result which needs to be studied further with additional realizations.

(iii) **Sensitivity to Pacific Ocean SST anomalies.** The results of the additional sensitivity experiments described above have suggested that the cold anomalies in the west Pacific, associated with El Niño, may be important in delaying the onset of the monsoon over India. To test this hypothesis further, an experiment was performed using 1983 conditions.
in which all the anomalies in the tropical Pacific Ocean were removed (Fig. 9). In all the remaining oceans, the SST anomalies used were from the year 1983.

The time series of the KE at 850 hPa simulated by this integration is presented in Fig. 10 along with the KE from the climatological SST and ‘cold-only’ integrations. When all the Pacific SST anomalies are removed, the onset date returns to normal, coinciding almost exactly with the onset in the integration with climatological SSTs. This provides yet further evidence that cold anomalies in the west Pacific may delay the onset of the monsoon. Similarly, the results shown in Fig. 5(b) for the La Niña case suggest that warm anomalies in the west Pacific may advance the onset of the monsoon. Both scenarios support the hypothesis that SST anomalies in the west Pacific may have a dominant role in the evolution of the Asian summer monsoon. This point will be returned to later in section 5.

(c) Influence of SSTs on the upper tropospheric zonal winds

Seasonal rainfall prediction for the Indian portion of the Asian summer monsoon has traditionally been based on statistical methods using a variety of precursory signatures during the preceding winter and spring (e.g. Krishna Kumar et al. 1995). It is important that the reasons why these precursory signatures exist are understood. Both WY and JS noted that weak monsoons were characterized by increased upper-level westerlies in the early months of the year (JS; their Fig. 12). These were associated with an equatorwards shift of the subtropical westerly jet (JS; their Fig. 13). It has been suggested that these changes in the mid-latitude circulation may be caused by, or have an influence on, the Eurasian snow cover. However, the UGCM displayed very little interannual variability in Eurasian snow cover, implying that the coherent signal seen in the model’s upper-level winds was primarily a remote response to the equatorial Pacific SST anomalies. JS noted that weak monsoons tended to be associated with El Niño conditions in the preceding spring, and posed the question whether this precursory signature in the upper tropospheric flow might be a remote response to El Niño.

Figure 11 shows the time series of the zonal winds at 200 hPa from January to June over the area covering latitudes 5°N to 20°N and longitudes 60°E to 100°E, for the suite of experiments for the El Niño (1983) and La Niña (1984) cases. The winds from the integration with climatological SSTs have been included in both sets of curves. The winds have been smoothed with a 30-day running mean to remove short timescale fluctuations. The results for the El Niño case (Fig. 11(a)) show the enhanced westerlies during the northern spring when the observed SSTs for 1983 were used (‘all anomalies’), as reported by JS. In the ‘warm-only’ integration, (when the cold anomalies in the west and subtropical Pacific were removed and only the main El Niño signal in the east and central Pacific was retained, Fig. 3(a)), the enhanced upper-tropospheric westerly flow in the preceding spring is maintained. In contrast, when the warm anomalies (i.e. the main El Niño signal) were removed in the ‘cold-only’ integration (Fig. 3(b)), the precursory signature of enhanced westerlies is lost. The strength of the westerlies is then similar to that simulated by the control integration with climatological mean SSTs. This suggests that the increased upper tropospheric westerlies in weak monsoon/El Niño years may be associated with the presence of warm anomalies in the central and east Pacific, and thus may be a remote response by the circulation to those anomalies, as hypothesized by JS.

In contrast, the results from the suite of experiments for 1984 (Fig. 11(b)) suggest that the Pacific SST anomalies associated with La Niña have very little influence on the upper-tropospheric zonal winds in the preceding winter and spring. The westerlies are very similar to those simulated by the control integration with climatological SSTs. This suggests that the enhanced heating over the west Pacific in La Niña years, implied by
the OLR anomalies shown in Fig. 2(b), does not perturb the upper-tropospheric flow in such a way as to produce a remote effect on the position of the subtropical jet over south-east Asia. The results shown in Fig. 11(b) imply that there is unlikely to be a robust precursory signature in the upper-tropospheric zonal wind for strong monsoon years. This is in agreement with the results from ECMWF analyses, which suggest that the upper-tropospheric zonal wind in strong monsoon/La Niña years is very close to that for normal monsoon years, at least during the AMIP decade. In El Niño years, however, the transition of the heating from the west to central Pacific Ocean, implied by the OLR anomalies shown in Fig. 2(a), produces a modulation of the upper-tropospheric flow and a robust precursory signature in the upper-tropospheric zonal winds over south-east Asia. An extended study
of the interannual variability of the upper tropospheric flow over Asia is planned with a ECMWF reanalyses.

5. Discussion

The results described in the previous section have suggested that the strength of the monsoon and the timing of its onset, in dynamical terms, are sensitive to the SST anomalies in both the east/central Pacific (El Niño), and the west Pacific, associated with the complementary anomaly pattern that develops during the mature phase of El Niño. This result is important because, previously, studies of the relationship between monsoon activity and ENSO have focused traditionally on correlations between all-India rainfall and the Niño 3 SST anomaly, where the Niño 3 anomaly is computed over the domain bounded by latitudes 5°N to 5°S and longitudes 150°W to 90°W. The possibility that SST anomalies in other parts of the tropical Pacific Ocean might have an important influence on the monsoon has largely been overlooked.

In the following sections the possible mechanisms whereby the pattern of SST anomalies in the Pacific Ocean might influence the Asian summer monsoon are investigated on the basis of the sensitivity experiments described above. The results are then applied to the case of the Asian summer monsoon of 1994 which was unexpectedly strong considering that an El Niño was in progress and that the Southern Oscillation Index (SOI) was negative.

(a) Role of the west Pacific tropical convective maximum

The recent study by Joseph et al. (1994) drew attention to the potential role of the latitudinal position and strength of the TCM over the west Pacific in determining the timing of the monsoon onset. JS (their Fig. 19) noted a similar relationship between the dynamical strength of the monsoon during its onset phase (May and June) and the anomalous convection over Indonesia and the west Pacific in the preceding spring (March–May). The observed latitudinal evolution of the TCM over the west Pacific for 1983 and 1984, as depicted by the OLR measured by the AVHRR, is shown in Fig. 12. The OLR over Indonesian and west Pacific longitudes (100°E to 150°E) has been computed and plotted as a time–latitude diagram. During the spring of 1983 (Fig. 12(a)), the main areas of convection are confined south of the equator, with much higher values of OLR (i.e. very suppressed conditions) north of the equator, compared with 1984 (Fig. 12(b)). The northwards movement of the TCM in May is markedly delayed in 1983 and the TCM is not as intense.

JS introduced the idea that modulation of the TCM may arise, firstly, from the remote influence of warm SST anomalies in the east/central Pacific which may modulate the Walker circulation, and/or, secondly, from the in situ response by the convection to the local SST anomalies in the west Pacific. The results from the sensitivity experiments described in the previous sections suggest that both scenarios are possible. During an El Niño, the remote effects of the warm east/central Pacific anomalies control the strength of the monsoon, whilst in La Niña years it is the response to the west Pacific warm SST anomalies that is dominant. With regard to the timing of the onset, however, the experiments suggested a sensitivity to both cold and warm anomalies in the west Pacific, indicative of the role of the latitudinal position, as well as the strength, of the TCM in determining the onset of the monsoon.

To investigate the role of the west Pacific TCM in determining the response of the monsoon to the various SST anomaly distributions described above, time–latitude diagrams of the OLR over Indonesia and the west Pacific (longitudes 100°E to 150°E) are presented in Figs. 13 and 14 from the various sensitivity experiments for the 1983 and 1984 cases,
Figure 12. Time–latitude diagrams of the outgoing longwave radiation (W m$^{-2}$) observed by the NOAA AVHRR satellite, and averaged between longitudes 100°E and 150°E for (a) 1983; (b) 1984. The period from March to June is plotted. The contour interval is 10 W m$^{-2}$ and values less than 240 W m$^{-2}$ are shaded.

respectively. As in Fig. 12, the evolution of the TCM through the spring preceding the monsoon and also its progression northwards associated with the onset of the monsoon in early summer is described.

In the El Niño case (Fig. 13), the TCM in the model lies persistently south of the equator and its transition northwards occurs very late in the season. This is essentially in agreement with the observed behaviour (Fig. 12), although the model has a tendency to produce a rather intense and latitudinally confined TCM, as noted by JS. Compared with the results for 1984 (Fig. 14), more suppressed conditions are simulated north of the equator, which agrees with the satellite observations (Fig. 12). Figure 13 shows that the impact of removing elements of the Pacific SST anomaly pattern has little effect on the
Figure 13. Time–latitude diagrams of the outgoing longwave radiation (W m$^{-2}$) averaged between 100°E and 150°E from the UGCM 1983 experiments. The period from March to June is plotted. (a) 'All anomalies', (b) 'warm-only', and (c) 'cold-only'. The contour interval is 20 W m$^{-2}$ and values less than 260 W m$^{-2}$ are shaded.
Figure 14. As Fig. 13, but for the UGCM 1984 experiments.
position of the TCM in the spring. This supports the result that the delayed onset in El Niño years is associated with a southwards displacement of the TCM in spring, and weaker northwards progression in the early summer.

The role of the TCM in determining the strength of the monsoon in the El Niño case is less evident. In terms of the strength of the TCM, the difference between the ‘warm-only’ and ‘cold-only’ experiments (Figs. 13(b) and (c)) is minimal compared with the difference between the ‘cold-only’ case for 1983 (Fig. 13(c)) and the ‘cold-only’ case for 1984 (Fig. 14(c)), both of which gave monsoons of similar intensity (Fig. 5). This suggests that in the El Niño case the behaviour of the TCM is not the dominant mechanism in determining the strength of the monsoon.

Weaker monsoons in El Niño years have usually been associated with the modification of the Walker circulation by the warm SSTs in the east and central Pacific, and hence implied additional subsidence over India and south-east Asia. Figure 15 shows the differences in the velocity potential, χ, at 200 hPa for May, June and July between the ‘warm-only’ and ‘cold-only’ experiments for 1983 and 1984. Associated with warm SST anomalies in the east and central Pacific in 1983, the model has simulated a strong positive χ-anomaly.
(i.e. implied additional subsidence) over India and south-east Asia during the onset phase of the monsoon (Fig. 17(a)), and a substantial modulation of the planetary-scale east–west (Walker) circulation. Thus it can be concluded that this change in the divergent circulation over India and south-east Asia, rather than a modulation of the TCM over Indonesia and the west Pacific, may be the dominant factor in determining the strength of the monsoon in 1983. However, it can also be concluded that the late onset in 1983 is directly related to the latitudinal position of the TCM as shown in Fig. 13, so that the TCM does have a part to play in El Niño years. In summary, for 1983, the warm anomalies in the east Pacific appear to influence the TCM through modulation of the Walker circulation, giving rise to a late onset. In the ‘cold-only’ case, the northwards movement of the TCM is affected directly by the in situ cold anomalies leading to a late onset, but not affecting the strength of the monsoon.

As already discussed, the TCM in the La Niña experiments (Fig. 14) is consistently slightly further north in the spring preceding the monsoon, and conditions north of the equator are less suppressed (OLR values are not as high). This agrees with the observed behaviour shown in Fig. 12. The transition of the TCM northwards occurs much earlier in the La Niña experiments than in those for El Niño. As noted earlier in section 4(b)(ii), the influence on the timing of the onset of the various elements of the La Niña SST anomaly pattern is less clear. However, there is a suggestion that the transition of the TCM northwards is more coherent and earlier in the ‘all anomalies’ (Fig. 14(a)) and ‘warm-only’ cases (Fig. 14(b)) than in the ‘cold-only’ case (Fig. 14(c)). In the ‘cold-only’ case, the main area of convection near the equator persists throughout June.

With regard to the role of the TCM in determining the strength of the monsoon in 1984, the same arguments, as used above for the onset of the monsoon, can be applied. Convection is consistently slightly further north than in 1983 (Fig. 13), and the transition of the TCM northwards occurs earlier. For example, in the ‘warm-only’ case, where a strong monsoon was simulated, the pre-monsoon TCM was strong and the transition of the convection north of latitude 10°N was very pronounced in June. On the other hand, the tendency for the convection to persist near the equator in the ‘cold-only’ case (Fig. 14(c)) after the onset of the monsoon, may account for the strength of the monsoon being near that simulated with climatological SSTs.

JS already noted that the modulation of the Walker circulation in La Niña years was not very strong and therefore may not be a dominant mechanism for determining the strength of the monsoon. The difference in the velocity potential between the ‘warm-only’ and ‘cold-only’ experiments for 1984 is shown in Fig. 15(b). Although these results suggest that the Walker circulation is indeed modified, there is very little local modulation of the divergent circulation over India and south-east Asia compared with the equivalent results for 1983 (Fig. 15(a)). However, a substantial change in the Hadley circulation over Indonesia and the west Pacific is implied, with additional ascent north of the equator. Thus it can be concluded that the modulation of the local Hadley circulation over the warm-pool region may be a dominant factor in determining the behaviour of the TCM, and hence the strength and timing of the monsoon in the La Niña case.

(b) Interpretation of the response by the TCM to the SST anomaly patterns

The results of the sensitivity experiments described above suggest that the modulation of the monsoon strength by tropical Pacific SST anomalies occurs predominantly in association with warm anomalies in either the east/central Pacific or the west Pacific. There is no evidence from the model results that cold anomalies affect the strength of the monsoon. In other words there is not an equal and opposite response to the warm and cold
SST anomalies. In very simple terms this can be understood by considering the nonlinearity of the Clausius–Clapeyron equation which, essentially, implies that the equivalent heat content of air over higher SSTs is correspondingly greater than that for air over lower SSTs. The propensity for convection to occur in association with SSTs in excess of 28 °C has been widely observed (e.g. Slingo et al. 1994; their Fig. 8(a)) and has been explained by Webster (1990) through the Clausius–Clapeyron equation as quite simply a manifestation of the nonlinearity of moist processes.

Figure 16 shows the longitudinal variation of the SSTs across the Pacific Ocean, averaged between the equator and latitude 14°N for March, April and May for the 1983 and 1984 experiments. Both sets of curves include the SST profile for the climatological
mean state. The substantial warming of the east Pacific in the ‘all anomalies’ and ‘warm-only’ cases for 1983 is evident, with SSTs in excess of 28 °C (Fig. 16(a)). In fact the SSTs are close to 28 °C across the entire Pacific Ocean in those cases. Thus, in simple terms, conditions favourable for convection occur at all longitudes and can explain the pronounced eastwards shift of the TCM with the concomitant changes in the Walker circulation noted in Fig. 15(a). In the ‘cold-only’ case, however, the longitudinal extent of SSTs in excess of 28 °C is much less than in the other cases, and this would again support the convectively suppressed conditions north of the equator, noted in Fig. 13(c).

The corresponding curves for the La Niña case (Fig. 16(b)) show a much less extensive region of SSTs near to, or in excess of, 28 °C. The warm anomalies in the west Pacific serve to enhance the local SSTs to temperatures in excess of 29 °C, near the maximum observed. The enhanced TCM in the ‘all anomalies’ and ‘warm-only’ cases can be explained simply in terms of the higher SSTs, as discussed above. However, in the ‘cold-only’ case the extent of the high SSTs is much reduced, even compared with the ‘climatological’ case and a more suppressed TCM, and hence, potentially, a less active monsoon might have been expected. As Figs. 14(c) and 5(b) show, this was not the case, and the simulated monsoon strength is close to that with climatological SSTs. The reasons for this may be clear if the pattern of the SST distribution is considered in more detail.

As noted above, cumulus convection occurs predominantly in association with SSTs in excess of 28 °C, typical of those occurring in the warm pool; but there are many notable exceptions. Over the central and eastern tropical Pacific Ocean, for example, the convection in the ITCZ is frequently associated with lower SSTs. Similarly there is not a simple relationship between SST anomalies and changes in the distribution of convection. Lindzen and Nigam (1987) introduced the concept that the gradients in the SST field, as well as the SST itself, may be influential in determining the distribution of tropical precipitation. They emphasized the importance of the low-level convergence anomalies produced solely in response to the gradients in the SST (and thereby in the surface pressure field), as distinct from those that arise explicitly through the thermodynamic effects of the SST itself. As noted by Lindzen and Nigam, the meridional gradients in SST in the east and central Pacific play a dominant part in the development of convection along the ITCZ. They also note that both the zonal and meridional gradients are relatively weak in the west Pacific where the low-level convergence may be generated primarily by the cumulus heating which arises as a direct response to the high SSTs in that region.

Lindzen and Nigam concluded that the pattern of SSTs may be of prime importance in determining the low-level flow and hence the distribution of precipitation over the tropical oceans. In Fig. 17 the SSTs for March, April and May of 1983 are compared with those of 1984. Also shown are the absolute values of the meridional and zonal gradients in SST. During the El Niño of 1983 the SST pattern was dominated by meridional SST gradients with negligible zonal gradients. Lindzen and Nigam noted that variations in the lower tropospheric branch of the Walker circulation are forced as much by zonal variations in the meridional SST gradient field as by changes in the zonal gradients themselves. Hence the change in the meridional SST gradient, as well as the magnitude of the SSTs, supports the transition of the convection from the warm pool to the central Pacific during El Niño.

In 1984, the meridional gradients are markedly reduced but the zonal gradients in the west Pacific are much stronger. Lindzen and Nigam found that the convergence forced by the zonal SST gradients in the west Pacific could be considerable and represented a major contribution to the total convergence. This occurs despite the zonal gradients being relatively much weaker than the meridional gradients. Thus it could be postulated, from the change in the zonal gradients shown in Fig. 17, that the enhancement of the TCM over the west Pacific during La Niña may arise from the increased zonal SST gradients.
Figure 17. Distribution of the mean SST and its zonal and meridional gradients for March, April, May of 1983 ((a) to (c)), and 1984 ((d) to (f)). (a) and (d) show the mean SST with contours drawn at 20, 22, 24, 26, 27, 28, 28.5, 29, 29.5 and 30 °C; values in excess of 27 °C are shaded. (b) and (e) show the zonal gradients in SST; the contour interval is 0.025 K/deg and values in excess of 0.05 K/deg are shaded. (c) and (f) show the meridional gradients in SST. Contours are drawn at 0.05, 0.1, 0.15, 0.2 K/deg and every 0.1 K/deg thereafter; values in excess of 0.2 K/deg are shaded.
Figure 18. Distribution of the mean SST anomalies for March, April, May, for (a) 1991; (b) 1994. Contours are drawn at ±0.2, 0.5 and 1 K. Positive values in excess of 0.2 K are shaded. The zero contour is dotted.

(as a result of the cold SST anomalies in the central and east Pacific), as well as from the in situ warm SST anomalies. Thus, in the ‘cold-only’ case for 1984, the maintenance of the TCM may well be due to the enhanced zonal gradients in SST, which would increase the moisture convergence into the Indonesian/west Pacific region.

(c) Application of the results to the Asian summer monsoon of 1994

The results of the sensitivity experiments described in this paper have suggested that all aspects of the SST field in the tropical Pacific need to be considered if the teleconnections between El Niño and the Asian summer monsoon are to be understood, and progress made in the seasonal prediction of the monsoon. A good example of this is evident when the monsoons of 1991 and 1994 are compared. Both were El Niño years with almost identical, negative Southern Oscillation Indices (SOI) in the preceding spring. Nevertheless, the Asian summer monsoon in each year displayed a totally different character. The monsoon in 1991 was weaker than usual, with the June–September all-India rainfall (AIR) being 7.6% below normal. This was correctly forecast by statistical methods, using a number of antecedent atmospheric and oceanic parameters. In contrast, the monsoon in 1994 was very good with about 10% excess AIR, although, as in 1991, the statistical forecast had indicated a weaker than normal monsoon. In terms of the large-scale flow, the monsoon was also stronger in 1994 than in 1991.

A possible explanation for the very different monsoons of 1991 and 1994 can be found in the SST anomaly patterns in the preceding spring (Fig. 18), and the associated OLR anomalies (Fig. 19). Examination of Pacific SST anomalies shows that, although the SOI for both years was the same (−1.3), the spatial distribution of SSTs was quite different. Both years show warm anomalies in the central and east Pacific and have comparable Niño 3 anomalies, consistent with the negative SOIs. However, whilst 1991 displayed cold
SST anomalies in the west Pacific, typical of an El Niño distribution (cf. Fig. 3(a)), 1994 did not; instead 1994 had substantial warm anomalies in the west Pacific, particularly north of the equator. In 1991, the OLR anomalies observed during the spring preceding the monsoon (Fig. 19(a)) show the expected eastwards extension of convection over the Pacific typical of that associated with El Niño. In contrast, the OLR anomaly pattern for 1994 (Fig. 19(b)) is very reminiscent of the composite pattern for strong monsoon years shown in Fig. 2(b), and is more typical of that observed in La Niña years. The implication from these observations, and from the sensitivity experiments described in this paper, is that the warm SST anomalies in the west Pacific in 1994 may well have been instrumental in enhancing the strength of the TCM in that region. On the basis of the results of JS and of the sensitivity experiments described in this paper, an enhanced TCM in the preceding spring should lead to a stronger than normal monsoon, as was observed. This case-study emphasizes the important influence that west Pacific warm SST anomalies can have on the Asian summer monsoon.

6. Conclusions

This paper has described a range of sensitivity experiments with a global circulation model of the atmosphere, designed to investigate the relative impacts of various aspects of the El Niño SST anomaly pattern in the tropical Pacific Ocean on the Asian summer monsoon. The research has concentrated on two particular cases with the aim of highlighting possible mechanisms that could explain the response of the monsoon to variations in the tropical SSTs, and as such it must be seen as a pilot study. Owing to the restrictions imposed by computing time, only a single realization for each case has been possible, and
thus no statistical significance can be attached to the results. Also, the limitations imposed by the quality of the model’s basic simulation of the monsoon, particularly its regional manifestations, are recognized. Nevertheless, the consistent response by the model in the range of experiments performed, and in the test with differing initial conditions, allow the following tentative conclusions to be drawn:

(i) During El Niño, modulation of the Walker circulation by the warm SST anomalies in the central and east Pacific appears to be the dominant mechanism controlling the strength of the monsoon; it also delays the onset of the monsoon through its remote effect on the TCM. However, the cold SST anomalies in the west Pacific, which develop during the mature phase of El Niño, also influence the TCM and delay the onset of the monsoon.

(ii) During La Niña, the local warm SSTs in the west Pacific modulate the strength and position of the TCM and provide the dominant mechanism for controlling the strength of the monsoon. They also influence the latitudinal position of the TCM and hence tend to advance the time of the monsoon onset.

(iii) The stronger upper-tropospheric subtropical westerlies over India and south-east Asia, in the winter and spring, preceding weak monsoons, are a remote response to El Niño.

(iv) The interannual behaviour of the Asian summer monsoon is sensitive to the pattern of SST anomalies in the Pacific Ocean. The results from a case-study of the monsoon of 1994 suggest that warm anomalies in the west Pacific can be as important as those associated with El Niño. Thus statistical methods for seasonal monsoon prediction should include a measure of the SST anomalies in the west Pacific and the ensuing response by the convection.

This paper has focused on the behaviour of the monsoon in two particular years and, clearly, the study needs to be extended to include a range of cases. The use of ensembles with differing initial conditions is also seen as the necessary extension of this research. A 3-year programme of research, funded by the European Community and commenced in 1996, is aimed specifically at understanding the mechanisms involved in the intraseasonal and interannual variability of the Asian summer monsoon. This program, entitled ‘Studies of the hydrology, influence and variability of the Asian summer monsoon (SHIVA)’, involves all the major modelling groups within Europe, and will, among other activities, consolidate the preliminary results presented in this paper.

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