Simulation of the Madden–Julian Oscillation in a Coupled General Circulation Model. Part I: Comparison with Observations and an Atmosphere-Only GCM

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ABSTRACT

The simulation of the Madden–Julian oscillation (MJO) has become something of a benchmark test for the performance of GCMs in the Tropics over recent years. Many atmospheric GCMs have been shown to reproduce some aspects of the MJO but have had problems representing its amplitude, propagation speed, and seasonality. Recent observational and modeling studies have suggested that the MJO is, at least to some extent, a coupled phenomenon. Thus with the complex interactions between convection and large-scale dynamics, together with the interactions between the sea surface and boundary layer, the MJO provides a rigorous test for many aspects of a GCM formulation.

In this study, the ability of an atmosphere–ocean coupled global climate model to represent various aspects of the MJO will be examined, and compared with the performance of the atmosphere-only component of the same model forced with slowly varying sea surface temperatures. One impact of coupling this GCM to an interactive ocean is to improve the eastward propagation of convection across the Indian Ocean. Surface flux anomalies associated with the MJO are in reasonable agreement with observations, although the resulting SST variability is found to be slightly weaker than observed. There is no propagation of convection into the west Pacific in the coupled model. It is proposed that this deficiency is due to errors in the basic state of the coupled model, in particular the lack of low-level westerly winds over the west Pacific.

1. Introduction

The Madden–Julian oscillation (MJO) represents the major mode of variability in tropical convection on timescales of less than a season. The MJO organizes convection on a regional scale, and by modulation of the strength and position of the main tropical heat source it can generate teleconnection patterns that affect the weather in the sub-and extratropics. Ferranti et al. (1990) have shown that during periods of MJO activity in the Tropics, the skill of a medium-range numerical weather prediction (NWP) model in the extratropics is significantly improved if the state of the tropical convection associated with the MJO is correctly prescribed. Hence a realistic simulation of the MJO in numerical models used for medium-range and seasonal prediction is desirable to improve forecast skill not only in the Tropics but also globally. Despite this, recent studies with the current generation of operational medium-range prediction models such as Jones et al. (2000) have shown that the representation of the MJO in such models is still rather poor.

Studies have also shown that many atmospheric GCMs do not simulate the MJO particularly realistically (e.g., Slingo et al. 1996, Sperber et al. 1997). Although some models are capable of producing strong variability in the Tropics on intraseasonal timescales, closer inspection often reveals errors in period, amplitude, seasonality, and geographical dependence of the MJO. The eastward propagation of convection from the Indian Ocean into the west Pacific in particular is often poorly represented in GCMs that instead tend to produce, at best, standing oscillations of convective activity over the Indian Ocean and west Pacific. Indeed, a study by Inness et al. (2001) showed that when the vertical resolution of their atmosphere-only GCM was increased, the model was able to capture many aspects of the MJO but the eastward propagation of the convective envelope from the Indian Ocean to west Pacific was still absent. However, Maloney (2002) showed that one atmosphere-only GCM was able to reproduce the eastward propagation of enhanced convection although this was displaced somewhat to the east of the observed signal.

Experiments such at that by Waliser et al. (1999) have shown that the representation of the MJO in an atmospheric GCM can be improved by using a simplified slab ocean parametrization to allow the organized convection to interact with ocean surface temperatures through the modulation of surface fluxes. Wang and Xie
(1998) developed a theoretical coupled model for the warm-pool climate system and showed that, unless the atmospheric component was coupled to the ocean mixed layer, there were no atmospheric unstable modes in the model. The thermodynamic coupling to the ocean resulted in a coupled high-frequency Kelvin mode which resembled the MJO in many respects. They suggested that the coupling not only provided an energy source to destabilize weakly stable atmospheric modes but also provided a mechanism for selecting the time and space scales of the eastward propagating modes which closely matched those of the MJO.

Studies such as these, together with a mounting body of observational evidence showing coherent fluctuations in surface fluxes and SST correlated with MJO events (Zhang 1997; Hendon and Glick 1997; Woolnough et al. 2000) have prompted explanations for the MJO in terms of coupled atmosphere–ocean processes. This suggests that most atmospheric GCMs should be coupled to some form of ocean model to achieve a realistic MJO simulation. Indeed Gualdi et al. (1999) have shown that a fully coupled GCM can reproduce many observed aspects of the MJO. However, coupling a GCM to an interactive ocean model may not be a universal solution to the problems of simulating the MJO. There may still be deficiencies in the physics of the atmospheric component, the response of the upper ocean to variability of atmospheric fluxes or errors in the mean climate of the coupled model which may prevent a realistic MJO simulation. Hendon (2000) showed that in the case of one particular GCM, coupling to a slab ocean had minimal impact on the simulation of the MJO. This was shown to be due to a combination of errors in the basic state climatology of the model together with rather weak latent heat flux anomalies at the ocean surface in response to intraseasonal surface wind stress variations.

All of these points suggest that modeling studies of the MJO may need to be conducted with an atmospheric GCM coupled to some form of ocean model in which SSTs can vary in response to changes in surface fluxes brought about by intraseasonal variability of tropical convection. In this study we will investigate the ability of one coupled atmosphere–ocean GCM to simulate the MJO, making comparisons with the atmosphere-only component of the same model forced by seasonally evolving observed SSTs with no intraseasonal variations. As well as introducing physical processes that may be important to the maintenance of the MJO, coupling an atmospheric GCM to an interactive ocean model will almost certainly introduce changes to the basic state of the atmosphere, which may well affect the simulation of the MJO. The impact on the simulation of the MJO, of basic-state errors in the climatology of low-level winds and sea surface temperatures (SSTs), will be examined in more detail in Part II (Inness et al. 2003) of this paper. Throughout this study we will concentrate on the period October to April when the MJO is at its strongest and most coherent (Hendon and Salby 1994).

2. Model description and mean state

The coupled GCM used in this study is the third Hadley Centre Coupled Ocean–Atmosphere General Circulation Model (HadCM3). The model has a stable mean climate and so is able to run without the need for heat flux corrections. The atmospheric component of this model has a horizontal resolution of 2.5° latitude × 3.75° longitude, with 30 levels in the vertical. The choice of 30 vertical levels is based on a study by Inness et al. (2001), which showed that the atmospheric component of this GCM forced with observed SSTs produced a more realistic simulation of some aspects of the MJO when the vertical resolution was increased from 19 to 30 levels, halving the layer thickness in midtroposphere to 50 hPa. Convection is parameterized using the mass-flux scheme of Gregory and Rowntree (1990), with the addition of convective momentum transports (Gregory et al. 1997). The ocean component of the model has a horizontal resolution of 1.25° × 1.25° with 20 levels in the vertical. The top three ocean model levels are each 10 m thick. Horizontal eddy mixing is parameterized using a version of the Gent and McWilliams (1990) adiabatic thickness diffusion scheme. Near-surface vertical mixing is modeled using a combination of a Kraus and Turner (1967) mixed-layer submodel and a K-theory scheme. Coupling between ocean and atmosphere occurs once per day. The study of Shinoda and Hendon (1998) has shown that the diurnal variations of SST, particularly during the suppressed phase of the MJO, may be important for fully capturing the ocean mixed layer interaction with the MJO. Due to the once-per-day coupling, this variability is not present in HadCM3.

Comparisons will be made with the atmosphere-only version of this model run for 17 yr forced with the observed SSTs between December 1978 and December 1995. The SST fields are updated every 5 days by smooth interpolation between monthly mean values. This version also has 30 levels in the vertical, and will be referred to as HadAM3. A fuller description of the standard versions of these models and their climatologies (albeit with 19 vertical levels) can be found in Pope et al. (2000) for the atmosphere-only version and Gordon et al. (2000) for the coupled version. Features of the GCM basic state that are pertinent to this study will be described briefly here.

a. Coupled GCM basic state

While the coupled model climate is stable and so the model runs without flux adjustment, there are still some large errors in the SST pattern of the model. In particular much of the ocean surface is too cold, with the exception of the upwelling regions on the eastern side of the subtropical ocean basins. The largest errors occur in the North Pacific. There are also errors in the tropical Pacific of greater than −3°C. Figure 1 shows the October to
April (ONDJFMA) Global Sea Ice and Sea Surface Temperature (GISST)-3.0 climatology (Rayner et al. 1996) for the period 1979 to 1995, together with the ONDJFMA climatology of HadCM3 averaged over the 20 yr of the integration. The GISST-3.0 dataset also provides the SST forcing for the atmosphere-only model HadAM3 integration and so Fig. 1a also shows the ONDJFMA SST climatology for HadAM3. The cold SSTs in the equatorial Pacific in HadCM3 are very clear, with values in excess of 29°C confined to the waters immediately surrounding the Maritime Continent, the complex group of islands between Australia and Southeast Asia.

Figure 2 shows the ONDJFMA precipitation climatology of Xie and Arkin (1996) together with the equivalent fields from HadAM3 and HadCM3. The tropical precipitation distribution in HadAM3 is quite realistic apart from a deficit around the Maritime Continent and excessive precipitation in the western Indian Ocean and in the West Pacific warm pool region. These errors are related and are discussed in detail by Neale and Slingo (2003). The precipitation climatology of HadCM3 is strongly affected by the SST errors in the model, with excessive precipitation over the Maritime Continent and a deficit over the very cold SSTs of the equatorial Pacific.

The ONDJFMA 850-hPa zonal wind climatology of the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis (ERA; Gibson et al. 1997), together with those from HadAM3 and HadCM3, is shown in Fig. 3. Excessively strong easterly trade winds in the central Pacific are a feature of both GCMs, with the easterlies being strongest in HadCM3. A lack of westerly winds on and just south of the equator in the west Pacific is seen in HadCM3 whereas the HadAM3 climatology shows the extent of the westerlies in this region to be well captured, although a little stronger than ERA. In the eastern Indian Ocean, HadCM3 represents the extent of the westerlies quite well but their magnitude is too large. It is HadAM3 which lacks westerly flow in this region.

Some part of these tropical errors in SST, precipitation and low level winds in HadCM3 can be understood in terms of a coupled mechanism. Strong convection over Indonesia drives a strong Walker circulation with strong low level easterly trade winds over the equatorial Pacific. In the center of the Pacific these anomalously strong easterlies drive Ekman divergence at the ocean...
surface along the equator resulting in the upwelling of cold water. The subsequent cold SST anomaly along the equator acts to keep the convection confined to Indonesia and also sets up an anomalous surface temperature gradient in the west Pacific. Both factors contribute to the systematic weakening of the low level westerly wind in this region.

b. Variance of convective activity

The seasonal mean climatology of the Tropics is in fact made up of weather systems on many different time and space scales. In particular, the mean precipitation field is a result of the compositing of precipitating systems on scales from individual cumulus clouds through mesoscale convective complexes to the large scale regions of enhanced convection associated with the MJO. This degree of variability modulates the strength and position of the mean convective region over Indonesia and the west Pacific warm pool leading to teleconnection patterns with an almost global influence. Hence it is not enough just to examine the tropical mean climate of a GCM without also taking into account the variability about that mean.

Figure 4a shows the variance of daily observed outgoing longwave radiation (OLR) from the National Oceanic and Atmospheric Administration (NOAA) Advanced Very High Resolution Radiometer (AVHRR) (Liebmann and Smith 1996) for the October to April period. A 20–100-day filter has been applied to the data to highlight variability on intraseasonal timescales. On the whole, the largest variance of OLR is coincident with the regions of heaviest mean precipitation shown in Fig. 2a. There is a local minimum of variance around the Indonesian islands. This minimum is probably due to the fact that, over the islands themselves, much of...
the variability of convection occurs on shorter timescales than the 20–100-day period selected by the filter.

The equivalent plots from the HadAM3 (Fig. 4b) and HadCM3 (Fig. 4c) show a broadly similar pattern but with some notable exceptions. The most obvious is the minimum in variance centered on the equator in both GCM simulations, particularly in the eastern Indian Ocean and through the Indonesian region. In the Indian Ocean sector, HadCM3 shows higher values of intraseasonal variance of OLR than HadAM3, but averaged over the whole tropical belt between 20°N and 20°S, the variance in HadAM3 is slightly higher than in HadCM3. However, the intraseasonal variance of OLR in both HadAM3 and HadCM3 explains about 25% of the total variance between 20°N and 20°S.

Part of the reason for the low variance of OLR near the equator in the GCM is persistent cirrus cloud which means that there is not as strong a relationship between low OLR and active convection as is seen in observations. For this reason, we will use convective precipitation from the GCMs directly rather than OLR. It will not be possible to make quantitative comparisons of convective precipitation from the GCM and OLR from observations, but some qualitative assessment will be possible. Convective precipitation will tend to have a shorter decorrelation length scale than OLR and so we would not expect precipitation anomalies associated with the MJO to be as extensive as OLR anomalies. However, as in the study of Hendon (2000), the maxima of convective precipitation in the GCM are collocated with minima in OLR on intraseasonal timescales, so the fields can be used interchangeably to determine phase relationships between convection and other variables.

3. MJO simulation by the coupled GCM

As shown in the previous section, there are clearly problems with both the basic climatology and variability of convection in the equatorial Indo–Pacific region in the coupled GCM. Modeled intraseasonal variance of OLR is weak throughout this region and since much of this variability is associated with the MJO, this suggests that the MJO simulation in this GCM will also be somewhat weak. This will be investigated in detail in this section.

In order to assess the MJO signal in the upper-level winds throughout the Tropics we use an index of MJO activity defined in Slingo et al. (1999). This index is formed by calculating the variance of the zonal mean 200-hPa zonal wind component averaged between 10°N and 10°S for each daily mean value of the winds. The resulting time series is then filtered using a 20–100-day bandpass filter and finally the time series is plotted in a 100-day moving window. As discussed extensively by Slingo et al. (1999), this index provides a useful measure of the envelope of intraseasonal variability of the tropical upper-level winds which is well-correlated to periods when the MJO is active. This index has been calculated for the 20 yr of the HadCM3 integration, the 17 yr of the HadAM3 Atmospheric Model Intercomparison Project-II (AMIPII) integration and also for 20 yr of ERA and post-ERA data. All three indices are plotted in Fig. 5. Note that the coupled model index is displayed on a different time axis than the ERA index and that from HadAM3. This is because the coupled GCM has a 360-day calendar whereas the atmosphere-only model has the true Gregorian calendar from 1979 to 1995.

This index shows that the MJO-related variance of the upper winds in HadCM3 is rather weak compared to observed values and to the atmosphere-only GCM. Although in some years, for instance 1984, there is very little MJO activity in the observed time series, HadCM3 produces activity which is consistently too weak. However, the seasonality of the model index is correct, with peaks in activity generally occurring during Northern Hemisphere winter and spring.

The fact that this index is weaker in HadCM3 than HadAM3 suggests that the MJO simulation in the coupled model will be worse than that in HadAM3. However, the distinction between the MJO simulation of the two GCMs cannot be made purely on the basis of looking at the intraseasonal variance of the tropical upper-tropospheric winds. The index presented here is a measure of the global intraseasonal variability of upper level winds but gives no information on the structure and propagation characteristics of the MJO. Sperber et al. (1997) showed that, while some GCMs can produce strong intraseasonal variability of tropical upper-tropospheric winds and convection, the signal in convective activity can still look rather unrealistic when ex-
Fig. 6. Lag-correlation plots of OLR or convective precipitation averaged between 10°N and 10°S with 200-hPa velocity potential at 90°E, also averaged between 10°N and 10°S. (a) NOAA AVHRR OLR correlated with ECMWF reanalysis velocity potential, (b) HadAM3 precipitation and velocity potential, and (c) HadCM3 precipitation and velocity potential. Shading indicates negative correlations between precipitation and VP, and positive correlations between OLR and VP. Contour interval is 0.1. All data are 20–100-day bandpass filtered.

amined in more detail, particularly in terms of the eastward propagation of the enhanced convective region. This emphasizes the importance of not relying on a single index or diagnostic to assess the quality of the MJO simulation in a GCM. The eastward propagation of convection will now be examined in more detail.

Figure 6a shows a lag correlation of daily values of OLR averaged between 5°N and 5°S with 200-hPa velocity potential (VP) at 90°E, also averaged between 5°N and 5°S. Both sets of data have been filtered to retain variability at periods between 20 and 100 days. Fifteen years of October–April data are used. The eastward propagation of convective activity is clearly seen with positive correlations extending from the western Indian Ocean at a lag of −15 days to the date line at a lag of around +20 days.

The equivalent plot for the 17 yr of the HadAM3 integration is shown in Fig. 6b although in this case convective precipitation is used directly instead of OLR, resulting in correlations of the opposite sign to those in Fig. 6a. The main signal is a standing oscillation centered between 90° and 120°E. Although there are negative correlations in the center of the Indian Ocean at a lag of −30 days which appear to propagate slowly eastwards, these correlations are very weak and are insignificant at 95%. So the atmosphere-only GCM HadAM3, which is itself a considerably updated version of one of the GCMs in the Sperber et al. (1997) study, shows no eastward propagation of the MJO convective envelope despite showing realistic intraseasonal variability of the tropical upper tropospheric zonal wind as depicted by the MJO index shown in Fig. 5.

Figure 6c shows the same lag-correlation analysis using data from the 20 yr of the HadCM3 integration. Although there is still a standing component to the east of 90°E there is also a statistically significant eastward propagating signal across the Indian Ocean, extending to about 120°E with a phase speed of about 5 m s⁻¹. However, to the east of 120°E no farther eastward propagation is evident. The correlations across the rest of the Pacific are rather weak and show, if anything, a rapid westward moving signal or an excitation of convection right along the equator at the same time as the maximum at 90°E.

a. Lag-regression analysis

More detail of the propagation of convection can be obtained by reconstructing the spatial patterns of the MJO cycle using lag regression. In this case we take the same base time series as in the lag correlations of Fig. 6, the 20–100 day bandpass-filtered 200-hPa velocity potential on the equator at 90°E. This base time series is an indicator of the large-scale upper-level divergence and implied ascent associated with the MJO. As for all the other plots in the paper we use just the data from the October–April period. Data are only plotted where the regressions are statistically significant at 95% and are calculated for one standard deviation from the mean at each grid point.

Figure 7 shows the OLR and ERA 850-hPa wind vectors regressed onto the 200-hPa VP base time series, also from ERA. This regression is plotted at 6-day intervals from 18 days before the VP minimum at 90°E through to 24 days after this minimum. The progression of the MJO is well captured in this figure. Initially there is suppressed convection over Indonesia with a small area of enhanced convection in the western Indian Ocean. This enhanced convection develops and moves eastward through the sequence, reaching the west Pacific by day +12. The westerly wind anomaly region to the west of the convective maximum becomes well stab-
Fig. 7. Lag regressions of NOAA AVHRR OLR onto a base time series of ECMWF reanalysis 20–100-day-filtered 200-hPa velocity potential at 90°E, averaged between 10°N and 10°S. The regressions of ECMWF reanalysis 850-hPa wind vectors onto the same base time series are superimposed. Data are only plotted at grid points where the local regression is significant at the 95% level. Fields are plotted at time intervals of 6 days, from (a) −18 to (h) +24 days, with (d) day 0 indicating the day of minimum 200-hPa VP at 90°E.
lished by −6 days, with the easterly anomalies to the east also strengthening. As the convection emerges into the west Pacific, the westerly wind anomaly extends right through the region of enhanced convection. Such a systematic change in phasing between the convection and the low level zonal wind anomalies was also noted in observational data by Woolnough et al. (2000) although in their case they were looking at the surface zonal wind stress anomalies. This will be discussed further in subsequent sections.

The equivalent sequence of lag-regression plots for HadCM3 is shown in Fig. 8, although in this case convective precipitation is used instead of OLR. The spatial scale of the regions of positive and negative precipitation anomalies is somewhat smaller than for the OLR anomalies shown in Fig. 7 because for a large scale region of convective precipitation, the high-level cirrus cloud shield extends beyond the actual region of precipitation.

In the Indian Ocean between −18 and 0 days, the patterns of convective and wind anomalies are similar to the observed data, with enhanced convection developing in the western Indian Ocean and progressing eastward, accompanied by a change from easterly wind anomalies at day −18 to westerly anomalies by day 0. Over Indonesia and the Pacific, however, the model picture is rather different from the observations. The region of suppressed convection initially over the Maritime Continent splits into two centers by day −6, one in the South China Sea and one on the north coast of Australia. At the same time, convection is enhanced to the north of the equator along the Pacific ITCZ, a pattern which persists until day +6. As the region of enhanced convection in the Indian Ocean approaches the Maritime Continent, it too undergoes splitting into two centers, one to the north and one to the south of the equator. Throughout the period, the only convective anomalies near the equator in the Pacific are the enhancement and then suppression of convection right along the Pacific ITCZ. The progression of 850-hPa wind anomalies looks quite similar to the observed pattern with an extension of westerly anomalies into the Pacific at +18 and +24 days, although the anomalies themselves are about half as strong as observed.

The enhancement of convection right along the ITCZ centered around day zero in HadCM3 is possibly related to a mechanism discussed by Matthews and Kiladis (1999), whereby extratropical Rossby wave energy emerging from the Asian subtropical jet exit propagates into the Tropics and excites convection. The fact that convection flares up right along the ITCZ in HadCM3 instead of being confined east of the date line suggests that rapid, westward moving disturbances may occur in the ITCZ, triggered by the initial flaring of convection in the east Pacific. Slingo (1998) discussed just such a paradigm. To determine the exact mechanism operating in HadCM3 requires further analysis of the higher frequency transients in HadCM3 and how they interact with the intraseasonal variability. This is beyond the scope of the current study.

The lag regression analysis for HadAM3 is shown in Fig. 9. The patterns are similar to HadCM3 but there are some significant differences. The precipitation anomalies are weaker in the atmosphere-only version, with the maximum positive precipitation anomaly in HadAM3 being 1.5 mm day$^{-1}$ compared to 3 mm day$^{-1}$ in HadCM3. HadAM3 does produce positive precipitation anomalies and just to the south of the equator in the west Pacific between days +6 and +24, which are not seen in HadCM3. However, when the HadAM3 precipitation anomalies are viewed on a daily basis it becomes apparent that the convection in this region develops in situ rather than as a result of propagation from the west. This is best seen when the lag-regression time series is viewed as an animation. The 850-hPa wind anomalies are also weaker and very much less coherent in HadAM3 throughout the period.

The lag-regression analysis confirms that, although there is a reasonable MJO signal over the Indian Ocean in HadCM3, neither the enhanced nor suppressed phase of the MJO extends into the equatorial west Pacific. Instead the centers seem to split into two over the Maritime Continent with the northern branch moving into the South China Sea and the southern branch moving across northern Australia. This leads to a possible conclusion that, while the model appears to be able to represent the underlying physical mechanisms that generate the eastward propagation of the convective envelope, the lack of MJO-related activity in the west Pacific may be related to the basic-state errors in the GCM in this region which were highlighted in section 2a. While the atmosphere-only GCM does produce intraseasonal variability of convection, the magnitude of the anomalies is about half those seen in the coupled GCM. The variability in HadAM3 is also dominated by standing oscillations, which also suggests that the eastward propagation of convection may be due to a mechanism which is only present in the coupled GCM.

b. The relationship between convection and surface fluxes

The theory for the eastward movement of the enhanced convective envelope put forward by Flatau et al. (1997) relies on a basic state low-level westerly wind in the region through which the convection propagates. The zonal wind anomalies associated with the inflow into the convective region, when superimposed onto the background wind field will enhance or reduce the surface wind stress and will thus reduce or enhance the latent heat flux (LHF) at the ocean surface. There will also be variations in solar shortwave flux (SWF) at the surface associated with changes in cloudiness associated with the MJO. The combination of these flux variations can induce SST anomalies to the east and west of the
FIG. 8. Same as Fig. 7 except for HadCM3, and using convective precipitation instead of OLR.
Fig. 9. Same as Fig. 8 except for HadAM3.
convective region, which may induce the convection to move toward the east.

Various observational studies (e.g., Zhang 1997; Hendon and Glick 1997; Woolnough et al. 2000) have shown that there are indeed coherent relationships between MJO convection, surface fluxes and SST. Here we use the lag-correlation method of Woolnough et al. (2000, hereafter WSH00), to examine these relationships in HadCM3. Daily model anomalies of convective precipitation, SST, zonal wind stress and surface latent and short-wave fluxes are filtered with a 20–100-day bandpass filter and averaged between 5°N and 5°S. Lag correlations between pairs of fields are then computed for lags of between −50 and 50 days. In all cases we will take surface flux anomalies as being positive into the ocean surface. So a positive LHF anomaly corresponds to a reduction in evaporation from the sea surface.

Figures 10a to 10c reproduce some of the lag-correlation plots from WSH00. Figure 10a shows the relationship between convection and SST, with warm (cold) SST anomalies leading (lagging) convection by about 10 days across the Indian Ocean, Maritime Continent, and west Pacific. Figure 10b shows that enhanced (suppressed) convection is correlated at zero lag with reduced (enhanced) shortwave flux reaching the ocean surface. The relationship between convection and surface latent heat flux is shown in Fig. 10c. Across the Indian Ocean and into the west Pacific as far east as the date line, enhanced LHF into the ocean surface (i.e., reduced evaporation) leads convection by about 10 days and reduced LHF into the surface lags convection by about 5 days. There is a slight shift in this relationship between the Indian Ocean and west Pacific with both the positive and negative LHF anomalies in the west Pacific occurring 2–3 days earlier relative to the convective maximum than in the Indian Ocean.

Of course, the sign and magnitude of the LHF anomalies depends on the magnitude of the total wind. Easterly wind stress anomalies will lead to reduced evaporation if the background state is westerly and to enhanced evaporation if the background state is easterly, as long as the anomalies are smaller than the magnitude
Fig. 11. Same as Fig. 10 except for the HadCM3 simulation. OLR is replaced by convective precipitation in (a), (b), and (c). Positive correlations in (a), (b), and (c) indicate that enhanced convection (increased precipitation) is correlated with a positive SST or flux anomaly. Negative correlations are shaded.

of the basic state wind. Figure 10d shows a lag correlation between LHF and zonal wind stress (UST) for the ERA data used by WSH00. This shows that between 60°E and the date line the correlation at lag zero is always strongly negative which shows that positive (i.e., westerly) UST anomalies lead to reduced LHF into the ocean surface (enhanced evaporation) while negative (i.e., easterly) UST anomalies increase the LHF into the ocean surface by reduced evaporation. This is consistent with the westerly basic state surface wind shown in Fig. 3a.

Figure 11 shows the same lag correlations from HadCM3. These can be directly compared with Fig. 10 but note that we have used convective precipitation directly from HadCM3 whereas for the observational figures we have used OLR as a proxy for deep convection. Thus the signs of all the correlations involving precipitation are reversed. However, to aid comparison we have shaded the regions of positive correlation in the plots using OLR, and the regions of negative correlation in the plots using precipitation. Figure 11a shows the lag correlations between convective precipitation and SST from HadCM3. Across the Indian Ocean and as far west as 140°E the correlation pattern shows that SST maxima (minima) lead (lag) convective precipitation maxima by 10 days. To the east of 140°E the correlations become weak and insignificant. Although there is a strong negative correlation at lag zero between convective precipitation and SWF right across the Pacific, the relationship between convective precipitation and LHF becomes very weak to the east of 140°E. This abrupt weakening in the west Pacific can be understood by considering the sign of the LHF anomaly together with the sign of the mean zonal wind. Figure 11d shows that the correlation of LHF and UST changes sign east of 140°E. Here, easterly wind anomalies superimposed onto the easterly mean wind shown in Fig. 3c will lead to an increase in evaporation and hence a negative LHF anomaly. This results in a breakdown in the relationship between SST and convection over the west Pacific.

A phasing difference is also apparent when the relationship between surface LHF and convective precipitation in HadCM3 is compared with that seen in the WSH00 study. They found that, from about 30 days before a convective maximum, reduced evaporation to the east of enhanced convection acts to warm the SST
in this region, with the greatest positive LHF anomaly occurring about 12 days before the peak in convective activity. In HadCM3, the peak LHF anomaly occurs 15–20 days before the maximum in convection, and from about 10 days prior to the convective peak onward, the sign of the LHF anomaly changes such that it is already acting to cool the SST prior the convective maximum. The maximum negative LHF anomaly occurs close to lag zero and is thus coincident with the maximum negative SWF anomaly. However, using data from the Tropical Atmosphere Ocean (TAO) moored buoys in the west Pacific, Zhang and McPhaden (2000) found that the peaks in negative LHF and SWF during MJO events were indeed often coincident.

It should be noted that both the current study and WSH00 use fluxes taken from NWP model reanalyses rather than in situ observations. Zhang and McPhaden (2000) point out that fields from NWP analyses in the Tropics will be heavily dependent on the physical parameterizations of the model and so should be treated with a degree of caution. However, a lack of long term observations in the Indian Ocean make it impossible to say whether the actual phasing of LHF and SWF anomalies in this basin is different from that deduced from NWP reanalyses and whether it is closer to that observed during Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE). The phasing of the LHF and SWF anomalies associated with MJO convection will be discussed further in section 4.

c. MJO events and the background zonal wind

The issue of the sign of the basic state wind being crucial to the eastward propagation of convection will now be investigated further. Although the observed seasonal mean low level wind field is westerly during the boreal autumn, winter, and spring, there is considerable variability in the sign of the low level zonal wind, particularly across the Maritime Continent and the west Pacific. Figure 12a shows a time–longitude plot of the 1000-hPa zonal wind component from ECMWF reanalysis monthly mean fields averaged between 4°N and 4°S for the period January 1979 to December 1998. While in the Indian Ocean the zonal wind component is generally westerly, there is a high degree of variability.
to the east of 115°E on various timescales from intra-
seasonal to interannual. The boreal winters of 1983–84
and 1995–96 are dominated by easterly flow in the west
Pacific, contrasting with the winters such as 1986–87
and 1992–93 when westerlies dominate.

Figure 12b shows the equivalent plot from HadCM3.
The most striking difference from ERA is in the com-
plete lack of westerly flow east of 130°E except in year
2 and year 5, which both correspond to strong El Niño
events in the model. The model winds are also too easterly
in the western part of the Indian Ocean.

We use the selection criteria of WSH00, with a slight
modification, to identify individual eastward propagat-
ing convective events at various longitudes in the band-
pass-filtered convective precipitation or OLR fields.
This technique is based on two criteria to select large
magnitude, eastward propagating events. First, at each
longitude, all maxima in convective precipitation (or
minima in OLR for observations) with magnitudes
greater than one standard deviation from the mean at
that longitude are selected. Second, as a test for eastward
propagation, each of the selected events must satisfy at
most one of these criteria:

1) the convective precipitation (OLR) 45° to the west
    must be positive (negative) for the entire period be-
    tween 10 and 19 days earlier, or
2) the convective precipitation (OLR) 45° to the east
    must be positive (negative) for the entire period be-
    tween 10 and 19 days later, or
3) the convective precipitation (OLR) 25° to the west
    3 to 14 days earlier and the convective precipita-
    tion (OLR) 25° to the east 3 to 14 days later must be
    positive (negative).

WSH00 describe in detail the periods and phase speeds
of the waves which these selection criteria permit. In
the current study a further visual check is applied to
eliminate cases where the technique identifies two stand-
ning oscillations separated by 45° of latitude which are
not connected by eastward propagation. This final check
does not greatly affect the number of observed events
in the OLR data, but does eliminate up to 50% of the
precipitation events in the coupled model data at some
longitudes, and over 50% of the events in the atmos-
phere-only model data.

The tick marks on Fig. 12a show the occurrence of
eastward propagating events at 60°, 90°, 120°, and
150°E. At 60°E, most but not all the MJO events occur
in westerly mean winds. At 90°E, all the events occur
in basic-state westerlies, but at this longitude the wind
is westerly on the equator virtually all the time except
during strong ENSO warm events. At 120°E, the picture
is less clear with 10 of the total 34 MJO events occurring
in basic-state easterlies. At 150°E, 4 of the total 32
events occur when the monthly mean wind is easterly.
Hence it is not simply the case that the eastward prop-
gating convective events on the MJO time scale will
only occur if the basic-state low level wind is westerly.

The same analysis is carried out for the coupled GCM
and the eastward propagating events are shown as tick
marks on Fig. 12b. The number of events identified is
also summarized in Table 1. The model produces fewer
eastward propagating convective events than are found
in the OLR data at all the longitudes, but particularly
over the Maritime Continent. Part of the reason for there
being fewer events in the model data may be due to the
use of convective precipitation rather than OLR. Con-
vective precipitation is more noisy and has a shorter
decorrelation length scale than OLR and so it is possible
that fewer events meet the above criteria. At 60° and
90°E most of the model eastward moving events occur
in a westerly basic state but at 120°E as many events
occur in easterly winds as in westerlies. Of the three
eastward propagating events identified in the model data
at 150°E, all occur in basic-state easterlies.

For completeness, the number of eastward propagat-
ing events identified in HadAM3 has also been included
in Table 1. Since there are only 17 yr in this model
integration, data are only available from 16 boreal win-
ters compared to 19 winters in the coupled GCM. If
there had been no eastward propagating events in the
atmosphere-only model this would have been quite com-
pelling evidence for the theory that the MJO requires
coupling to an ocean surface to bring about eastward
propagation. This is not the case and there are some
occasions when the atmosphere-only GCM produces
eastward propagating convective events on the MJO
timescale. However, the numbers of events identified in
HadAM3 are much less than HadCM3 at 60° and 90°E
while being similar at 120° and 150°E. This suggests
that the coupled mechanism is important since it is in
the Indian Ocean that the surface winds are westerly
in HadCM3, whereas in the west Pacific the surface winds
are easterly and here the coupled GCM shows no im-
provement over the atmosphere-only GCM.

In the reanalysis data and particularly in HadCM3
there are a significant number of eastward propagating
events at 120°E which occur in basic-state easterly low-
level winds. This would also seem to be evidence against
the coupled theory for the MJO that requires background
westerly flow to induce eastward propagation of the
convection. However, at this longitude the presence of
large island landmasses to the east and west must com-
plicate the issue. This is perhaps a particularly important

<table>
<thead>
<tr>
<th>Method</th>
<th>No. of eastward propagating events at each base longitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>AVHRR OLR (16 yr)</td>
<td>30 37 34 32</td>
</tr>
<tr>
<td>HadCM3 (19 yr)</td>
<td>17 21 11 3</td>
</tr>
<tr>
<td>HadAM3 (16 yr)</td>
<td>10 9 9 5</td>
</tr>
</tbody>
</table>
factor in the GCM, where the islands are only coarsely resolved. The question of the impact of the Maritime Continent on the propagation of the MJO will be discussed further in section 4.

d. Composite MJO events

Having identified individual eastward propagating convective events, we can now produce composite events at each longitude by meaning together the individual events that pass through that longitude. This has been done, not only for convective precipitation, but also for the SST and surface flux anomalies which accompany these events. Composites are only displayed for events passing through 60°E, 90°E, and 120°E since both versions of the GCM only produce a few rather weak events at 150°E. The nature of the compositing technique means that the signal will be strongest at the base longitude and will decay away from this longitude due to the slightly different character of each individual event. Therefore care must be taken when interpreting the composites in terms of where MJO events initiate and decay.

Figure 13 shows composites from observations, HadAM3, and HadCM3. The observed composites are formed from OLR which is inherently a smoother field than precipitation. These show eastward propagating events which extend from the Indian Ocean to the date line. Embedded standing oscillations can be seen in the observed composites, centered on the eastern Indian Ocean and the west Pacific and the strength of these components varies depending on the base longitude.

The HadCM3 precipitation composites at all three longitudes show an eastward propagating signal across the Indian Ocean which goes no farther east than the Maritime Continent, even when the base point is 120°E. There are also standing oscillations at around 90°E and 120°E, which appear more dominant than those seen in the OLR. The composites from HadAM3 are based on fewer events and so are likely to be more noisy. Even allowing for this the composites show much less coherence than those from the coupled model. Only the composite with the base longitude at 60°E shows any convincing eastward propagation, and this does not extend as far east as the equivalent composite from the coupled GCM. The other composites are all completely dominated by standing components.

SST composites calculated for the events included in the precipitation composites are shown in Fig. 14. The observational SST composites show that while the SST anomalies tend to propagate eastward with time in the Indian Ocean, in the west Pacific the SST variability has more of a standing oscillation pattern. Across the
Indian Ocean, the negative SST anomaly appears larger in magnitude than the positive anomaly—a 0.2°C cold anomaly behind the convective maximum compared to a 0.1°C warm anomaly ahead of the convective maximum. The compositing tends to reduce the magnitude of the anomalies; observed SST anomalies associated with individual MJO events often exceed 0.5°C.

The HadCM3 SST composites show a similar pattern to the precipitation composites, with anomalies of the order of ±0.1°C propagating across the Indian Ocean but not extending into the west Pacific. In this case the positive and negative anomalies are of the same size. There appears to be a slight phasing difference between the SST anomalies in HadCM3 and the observed anomalies. In the observations, the zero SST anomaly line passes through the base longitude at lag zero. However, in the model, there are warm SST anomalies at the base longitude at lag zero and the zero anomaly occurs between 2 and 5 days after the convective maximum depending on longitude. Woolnough et al. (2001) performed an experiment using an aqua-planet version of HadAM3 forced with an eastward-moving SST anomaly dipole to investigate how the atmosphere responds to such SST anomalies. They found that the positive convective anomaly remained over the maximum of the SST gradient (coincident with the zero SST anomaly) in agreement with observations. This implies that the shift in phasing in the coupled model may be due to a rather slow response of the upper ocean to the surface flux anomalies. Instability waves near the date line are also apparent in the HadCM3 SST composites. Such features also occur in reality but are more marked in the model as the thermocline is closer to the surface than observed.

In order to shed light on the patterns of the SST anomaly composites, Fig. 15 shows the composites of the sum of the SWF and LHF anomalies. As noted in WSH00, these two flux components make up the largest part of the intraseasonal variance of the surface flux in the equatorial Indo–Pacific and so the other components of the net surface energy flux (sensible heat and longwave radiation) can be neglected with very little loss of precision. These composites, which use ERA fluxes, are probably the least reliable of the “observational” fields, particularly around the complex island structures of the Maritime Continent. Lin and Johnson (1996) compared ERA surface latent heat flux with observed values from the Intensive Flux Array during the TOGA

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**Fig. 14.** Same as Fig. 13 except for SST anomalies: (a)–(c) Reynolds SST, and (d)–(f) HadCM3. Contour interval is 0.025°C. Negative anomalies are shaded.
COARE experiment in the west Pacific. They found that, although the reanalysis fields generally agreed quite well with observations, there could be large discrepancies (up to 40 W m$^{-2}$) during strong westerly wind events. The ERA composites appear rather noisy, particularly around the Maritime Continent. In general, the negative flux anomalies are larger than the positive anomalies, consistent with the fact that the negative SST anomalies are also larger than the positive ones in the observations. The flux anomalies in HadCM3 are generally of comparable magnitude to those from ERA and show a tendency to propagate quite smoothly across the Indian Ocean.

The intraseasonal SST variability in the warm pool is largely a thermodynamic response to variations in surface flux, with advection of heat playing a minor role due to the small horizontal temperature gradients and weak surface currents (Shinoda et al. 1998; Shinoda and Hendon 2001). If this is the case then it would be expected that the net surface flux and SST variations should be in quadrature, with the flux variations leading the SST variations by one-quarter cycle. This is the case for the Reynolds SST and ERA flux composites. However, in the HadCM3 composites the SST cooling seems to occur more slowly, with the zero SST anomaly line being displaced 2 to 5 days later than the peak in the convective precipitation as shown in Figs. 13e–g. One possible reason for such a delayed response is that the shoaling of the mixed layer during the clear sky/light wind phase of the MJO has not been fully captured by the ocean model. Thus when the cooling phase begins, the SST response is delayed as the cooling is taking place through a greater depth of the upper ocean. A detailed diagnosis of the ocean mixed layer processes in the model will be necessary to fully resolve this difference between the model and observations. This will form the focus of future work.

4. Discussion of results

The lag-correlation, lag-regression, and compositing techniques used in this study have shown that a coupled GCM is capable of reproducing many of the observed aspects of the MJO, at least in the Indian Ocean basin where the basic state climatology of low-level zonal winds from the model is in reasonable agreement with
observations. The eastward propagation of large-scale convective maxima across the basin and the relationship between convection, SST and surface fluxes are all well captured by the coupled model, although the number of MJO occurrences is too few. This shows improvement over the MJO simulation by the atmospheric component of the coupled GCM forced with observed, slowly varying SSTs, in which the intraseasonal variability of convection is dominated by standing oscillations.

In the west Pacific, the climatology of the coupled GCM is such that it has low level mean easterly flow on and immediately south of the equator during the boreal autumn, winter, and spring instead of the low level mean westerly flow, which is actually observed during this period. The MJO simulated by the coupled model does not extend into this basin but tends to break down over the Maritime Continent. This suggests that the MJO is unable to propagate through the erroneous low level easterly winds in the west Pacific. The lag-regression analysis of Fig. 7 shows that the MJO convective anomaly tends to split into two centers as it approaches the Maritime Continent, with one branch moving northwards into the South China Sea and one branch moving across northern Australia and into the South Pacific convergence zone (SPCZ). The fact that this split starts to occur in the eastern Indian Ocean where the winds in the model are still westerly suggests that there is some other factor involved. Possibly the rather coarsely resolved islands of Indonesia act as some sort of barrier to eastward propagation. This issue will form the basis of future work.

In both the coupled model and observations, most, but not all, eastward moving convective events at MJO time and space scales occur when the background state wind is westerly. The atmosphere-only GCM also produces some eastward moving MJO-like events, although these are fewer in number, weaker and less coherent than those seen in the coupled GCM. These two strands of evidence suggest that the coupled mechanism of Flatau et al. (1997) is not the only way of producing eastward propagating MJO events. Rather, it would appear that this mechanism perhaps acts as an amplifier of existing propagating convective anomalies. The coupling mechanism may also be acting as a timescaler selector for such events as proposed by Woolnough et al. (2001). Of course, in a region with low-level easterly flow, the coupled mechanism will act to destroy eastward moving events, which explains the lack of extension of the MJO into the west Pacific in the coupled model. The results of Waliser et al. (1999) also show that the impact of coupling the Goddard Laboratory of Atmospheres (GLA) atmospheric GCM to an interactive ocean was to amplify and slow the rather weak and rapidly propagating disturbances simulated by the GCM forced with a slowly evolving seasonal cycle of SST. They debate whether the MJO is a truly coupled mode in the sense that it would not exist if it were not for interaction between the convective signal and the SST field. On the strength of their results they conclude that the MJO is inherently an atmospheric mode, which is modified and amplified through the interaction with the ocean surface. The evidence from this study would seem to support that argument.

Of the four longitudes chosen as base points in this study, the longitude at which MJO events occur most often during periods of mean easterly wind in both model and observations is 120°E. Through the Maritime Continent, any ocean–atmosphere coupling will be complicated and disrupted by the complex arrangements of islands. So in this region, it is possible that coupling between the large-scale region of enhanced convection and the SSTs is less important to the maintenance of the MJO than in the ocean basins to either side. It could be that when the large scale envelope of the disturbed part of the MJO reaches the Maritime Continent, the island regions are already predisposed to convect regardless of SST anomalies in the surrounding oceans. Thus the MJO is really acting to modulate the in situ convection over the Maritime Continent, resulting in a standing component embedded within the eastward propagation.

Hendon (2000) investigated why the MJO simulation by one particular GCM did not improve when the model was coupled to an interactive mixed layer ocean model. One reason was that the latent heat flux anomalies produced by MJO-related surface wind stress anomalies were too weak or even of the wrong sign. In the current study, the flux anomalies produced by the coupled model—that is, the sum of the latent heat and shortwave radiative fluxes—are of similar magnitude to those seen in the ERA data. The SST anomalies which result from this intraseasonal flux variability are also similar to those in the observations, although the cold anomalies to the west of the convective core are slightly weaker in the model than in observations. There is also a shift in the phasing of the precipitation maximum relative to the SST dipole in HadCM3. The zero SST anomaly in HadCM3 occurs 2–3 days after the maximum convective anomaly instead of being coincident with it. This suggests that the upper ocean in HadCM3 responds rather too weakly and slowly to intraseasonal surface flux variability, a problem which may be addressed by increasing the vertical resolution in the upper layers of the ocean.

The relative phasing of the latent heat and shortwave flux anomalies diagnosed in HadCM3 differs from that diagnosed from ERA, although it is in reasonable agreement with the study of Zhang and McPhaden (2000) who used data from moored buoys in the West Pacific. In HadCM3, the minima of the two flux components are almost coincident, both with each other and with the maximum in convective precipitation associated with the MJO. This implies that the low-level westerly wind anomalies in HadCM3 extend right through the region of convection with the zero wind anomaly displaced to the east. In the ERA data, the minimum in
LHF into the ocean (i.e., the maximum in evaporation) tends to lag the minimum in SWF by several days. A lack of in situ observations in the Indian Ocean makes it currently impossible to say whether the true phase relationships between convective precipitation and the various surface flux anomalies really are as diagnosed from NWP reanalyses or whether they are closer to the patterns observed by in situ moorings in the west Pacific. The MJO structure in HadCM3 corresponds most closely with model II of the MJO presented by Zhang and McPhaden (2000) whereas the structure of the MJO diagnosed from ERA corresponds more to model I of the MJO in that study.

We have postulated that the reason the MJO does not extend into the west Pacific in the coupled GCM simulations is that the lack of westerly surface winds on and just to the south of the equator prevents the coupled mechanism of Flatau et al. (1997) operating in this basin. The fact that the MJO does look reasonably realistic in the Indian Ocean where the low level winds are westerly suggests that if the model’s basic state were correct in the west Pacific then the MJO would be able to propagate out to the date line. It is also possible that the lack of MJO in the west Pacific is purely a result of the SSTs being too cold to support the convective activity. In part II (Inness et al. 2003) of this paper we will investigate this question further by performing an experiment in which we modify the climatology of the GCM in the tropical Pacific in order to produce a basic state with low-level westerly winds in the west Pacific. We will also examine the simulation of the MJO by a coupled GCM that has the same atmospheric component as HadCM3, but a different ocean component. This combination of experiments will allow us to draw some firmer conclusions on the importance of the basic state to the maintenance of the MJO.

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