The organization of tropical convection by intraseasonal sea surface temperature anomalies

By S. J. WOOLNOUGH*, J. M. SLINGO and B. J. HOSKINS

University of Reading, UK

(Received 1 July 2000; revised 4 December 2000)

SUMMARY

The organization of convection by intraseasonal sea surface temperature (SST) anomalies is investigated to examine the role that atmosphere–ocean coupling may play in the maintenance of the Madden–Julian Oscillation. An atmospheric general-circulation model with an ocean-covered surface is used to investigate the response of tropical convection to idealized imposed intraseasonal SST anomalies, and the sensitivity of this response to their propagation speed. The convection is found to be organized on the spatial and temporal scales of the imposed SST anomalies and the location of the maximum in precipitation relative to the SST anomaly is in good agreement with observations, and different from that for a stationary SST anomaly. The magnitude of the precipitation anomalies increases with decreasing propagation speed of the SST anomalies. The role of the free-tropospheric humidity is crucial for determining the location and magnitude of the precipitation response.

KEYWORDS: Convection  Madden–Julian Oscillation  Sea surface temperature anomalies

1. INTRODUCTION

Observations from the Tropical Ocean Global Atmosphere–Coupled Ocean Atmosphere Response Experiment (TOGA–COARE) have shown that sea surface temperature (SST) in the warm pool is modulated by the passage of the Madden–Julian Oscillation (MJO) (e.g. Weller and Anderson 1996; Hendon and Glick 1997; Lau and Sui 1997) and the variations in surface flux associated with it. Results from studies covering a larger domain and longer time records (e.g. Hendon and Glick 1997; Shinoda et al. 1998; Woolnough et al. 2000) have confirmed that the relationships between SST and convection hold over the tropical Indian Ocean and west Pacific to near the dateline. This coherent relationship between the convection and SST in the tropics supports the hypothesis that the MJO may be a coupled atmosphere–ocean phenomena, as suggested by Platini et al. (1997) and Sperber et al. (1997). In this coupled mechanism the surface heat flux anomalies generated by the convection act to warm the ocean to the east of the convection and cool the ocean in situ and to the west of the convection, thereby leading to an eastward propagating SST anomaly. It is hypothesized that this eastward propagating SST anomaly acts to generate, or strongly modify, the eastward propagation of the convection. Waliser et al. (1998) have demonstrated that coupling may be important by conducting sensitivity experiments with an atmospheric general-circulation model (GCM) coupled to a simple slab ocean.

The observational studies have clearly demonstrated that the atmosphere can force the ocean on intraseasonal time-scales and this has been supported by studies with an ocean mixed-layer model forced with observed fluxes (e.g. Shinoda and Hendon 1998). However, for a coupled atmosphere–ocean phenomenon, the ocean must, in turn, be able to force the atmosphere. These observational studies cannot easily determine whether the SST variations have an impact on the strength, period or propagation characteristics of the convection associated with the MJO. To fully determine this, modelling studies are needed.

Convection in the tropics is seen to be organized by structures that can be identified with theoretically determined equatorially trapped waves, modified to take account of

* Corresponding author, present address: Met Office, London Road, Bracknell, Berkshire RG12 2SZ, UK.
the effects of moisture (Wheeler and Kiladis 1999). The WAVE–CISK (Conditional Instability of the Second Kind) mechanism of Lau and Peng (1987) and the evaporation–wind feedback (WISHE) mechanisms of Neelin et al. (1987) and Emanuel (1987) for the MJO are also based around the ideas of the modification of these equatorially trapped modes. It is possible that the organization of convection by intraseasonal SST anomalies may also involve these equatorially trapped modes. If the SST anomalies provide a way of organizing the convection on the spatial and temporal scales of equatorially trapped waves, then a resonant response may be produced, corresponding to the natural modes of the system. However, it is possible that the SST anomalies may organize the convection independently of the equatorially trapped waves, which develop purely as a dynamical response to the heating associated with organized convection without feeding back on the organization of the convection. In these cases one might expect to see a large response in dynamical fields around the natural modes of the system, but not in the precipitation itself.

One approach to investigate the two-way interaction between the SST anomalies and the convection is to use a fully coupled atmosphere–ocean model, but it is difficult to control these models to perform conclusive sensitivity studies. This approach also relies on the correct representation of the basic state of the atmosphere–ocean system in the coupled model to enable the atmosphere to force the ocean in the correct sense. In particular, the sign of the latent-heat flux anomalies is dependent on the sign of the ambient surface winds, with westerly basic-state winds producing the required sign of the latent-heat flux anomalies and easterly basic-state winds generating latent-heat flux anomalies which would oppose the easterly propagation of the SST anomalies (Woolnough et al. 2000). In light of these difficulties, this paper addresses the specific question of how the ocean may organize convection on intraseasonal time-scales by examining the effect that imposed intraseasonal SST anomalies have on the intraseasonal variability of convection in an aqua-planet configuration of an atmospheric GCM. Using imposed SST anomalies allows fully controlled sensitivity experiments to be performed. The aqua-planet configuration gives the simplest context for the investigation and increases the proportion of the tropical domain which is ocean covered and where deep convection is active, thus increasing the sample size. Once the ability of the ocean to force the atmosphere on intraseasonal time-scales, and the mechanisms by which it does so, have been established, then fully coupled modelling studies can be designed and diagnosed with confidence.

In section 2 the model configuration is described and the experimental design explained. The results of the sensitivity experiments are discussed in section 3, and the implications of these results for the coupled mechanism of the MJO, and suggestions for extensions to this work are discussed in section 4.

2. Model Configuration and Experimental Design

The numerical model used in this study is the UK Met Office Unified Model version HadAM3. A detailed description of the components of the model and its performance in Atmospheric Model Intercomparison Project (AMIP)-type integrations can be found in Pope et al. (2000) and references therein. Inness et al. (2001) have shown that HadAM3 is able to simulate an intraseasonal oscillation whose periodicity is shorter than observed and whose propagation is less coherent, particularly in terms of the convective signal. Nevertheless, the Unified Model is known to be one of the more successful GCMs for capturing some aspects of the intraseasonal oscillation (Slingo et al. 1996).
The model is run with a horizontal resolution of 3.75° longitude \times 2.5° latitude, with 19 levels in the vertical, corresponding to a layer thickness of about 100 hPa in the mid troposphere with higher resolutions in the boundary layer and around the tropopause. Because of the nature of this study it is worth summarizing the convection scheme here. The convection scheme is the mass-flux scheme of Gregory and Rowntree (1990) including a representation of convective downdraughts (Gregory and Allen 1991) and the vertical transport of momentum by convection (Gregory et al. 1997). The scheme uses a bulk cloud model, based on parcel theory and taking into account the effects of mixing entrainment and detrainment between the cloud parcel and the environment, to represent an ensemble of convecting plumes within the grid box. Forced detrainment is included to represent the termination of less buoyant plumes within the ensemble. Convection is initiated as a parcel with a prescribed small buoyancy excess (equivalent to 0.2 K) and the initial mass flux is determined from the buoyancy excess of the parcel after lifting from the initiation level to the next level. The environmental profiles of moisture, temperature and momentum are modified by the effects of the detrainment and the compensating subsidence within the cloud-free area.

In this study the model is used in an aqua-planet configuration (Neale and Hoskins 2001a): the surface is water covered, although two latitude rows are left as land at each pole for computational reasons. The incoming solar radiation is fixed at an equinoctial value. Other than these two modifications the aqua-planet configuration retains all the physical parametrizations of the GCM.

A series of experiments were performed using prescribed SSTs as boundary conditions. The control experiment has a fixed zonally symmetric SST with an analytical profile of the form,

$$\text{SST} = 29 \, ^\circ C - 29 \, ^\circ C \left( \frac{\sin^2(\phi/60^\circ) + \sin^4(\phi/60^\circ)}{2} \right),$$

between the equator and 60° of latitude; poleward of 60° the SST is fixed at 0 °C. This profile is chosen to mimic the magnitude and observed latitudinal gradients of SST in the tropical Indian Ocean and west Pacific warm pool (Fig. 1(a)) and to produce a reasonable control climatology. The control integration is 540 days long and is initialized with a model dump from an existing aqua-planet integration, the first 180 days are disregarded to allow the model to adjust to changes in the SST profile. The zonal-mean zonal wind from the last 360 days of the control integration is shown in Fig. 1(b). The flow in the tropics is easterly throughout the depth of the troposphere, with subtropical jets centred at about 30° of latitude, with maximum wind speeds of about 60 m s\(^{-1}\).

The easterly flow at the surface in the tropics is opposite to that required for the coupled mechanisms proposed by Flatau et al. (1997) and Sperber et al. (1997). This is particularly important for the impact that the convection associated with the MJO has on the ocean and may not be so important for the atmospheric response to the SST anomalies; this problem will be addressed further in section 4. The zonal-mean precipitation (Fig. 1(c)) at the equator is about 12 mm day\(^{-1}\), with the maximum precipitation occurring just off the equator, there are secondary maxima in the precipitation at about 40° of latitude associated with the mid-latitude synoptic weather systems. Due to slight asymmetries in the initial atmospheric conditions, the solar forcing, and the natural variability of the model, there are slight hemispheric asymmetries in the climatology of the control integration.
Figure 1. Zonal-mean climatology of the control integration. (a) Model sea surface temperature (SST) (solid line) and observed (1982–97) December–March SST for the Indian Ocean and west Pacific (60°E–180°E) (dotted line) for comparison. (b) Zonal-mean zonal wind from 360 days of the control integration (contour interval is 5 m s⁻¹, negative contours dashed, zero contour dotted). (c) Zonal-mean precipitation from 360 days of the control integration.

Figure 2(a) shows a Hovmöller diagram of the precipitation averaged from 5°N to 5°S in the control integration. The dominant structures in the precipitation field are eastward moving signals which propagate round the globe in about 25 days and westward moving signals travelling at about one third of this speed. There are occasions when groups of westward moving precipitation maxima show a tendency to propagate eastwards (e.g. days 75–90 between 60°W and 30°W), although these signals tend to persist for only 10 days or so. Thus, in common with other GCM studies (e.g. Slingo et al. 1996), the aqua-planet model does not display an intraseasonal oscillation with the observed periodicity, although it clearly has a sub-seasonal organization of convection.

Figure 2(b) shows a wave-number–frequency decomposition of the daily rainfall, following the technique of Wheeler and Kiladis (1999). The wave-number–frequency spectra are calculated at each latitude between 5°N and 5°S and the total power spectrum is the mean of the individual spectra. The total power spectrum is smoothed using 10
Figure 2. (a) Hovmöller diagram of the precipitation averaged from 5°N to 5°S, for one year of the control integration. (b) Log-power spectra of the convective precipitation averaged from 5°N to 5°S normalized by a background spectrum calculated from successive smoothings of the total power spectrum (see text for details). Negative wave numbers indicate westward moving waves, the frequency is in days⁻¹. The solid lines indicate eastward propagating waves with periods of 25 days and westward propagating waves with periods of 75 days.
successive 1-2-1 filter passes in both the wave-number and frequency directions to produce a background spectrum. The raw power spectra are then divided by the background spectrum to emphasize peaks in the wave-number–frequency decomposition. In the control integration the two dominant regions in the wave-number–frequency spectrum correspond to the eastward moving equatorial Kelvin waves and the westward moving equatorially trapped modes including the equatorial Rossby waves. The Hovmöller and wave-number–frequency spectrum shown in Fig. 2 indicate that in the control integration the dominant organization of the convection is by waves with similar characteristics to theoretical equatorially trapped waves, as shown by Wheeler and Kiladis (1999) for the atmosphere.

The ability of intraseasonally propagating SST anomalies, compared with those associated with the MJO, to influence the organization of the convection is now investigated by performing a series of sensitivity experiments with SST anomalies given by,

\[
SST' = A_0 \sin[2(\lambda - ct)] \exp \left\{ - \left( \frac{\lambda - ct}{45^\circ} \right)^2 - \left( \frac{\phi}{7.5^\circ} \right)^2 \right\}
\]

\forall -90^\circ < \lambda - ct < 90^\circ,

where \(A_0\) is an amplitude factor (here \(A_0\) is chosen such that the amplitude of the SST anomaly is 1 K) and \(c\) is the prescribed propagation speed. This SST anomaly corresponds to an SST dipole, centred at the equator, propagating eastwards around the globe with a fixed magnitude and speed (Figs. 3(a) and (b)).

The spatial scale of the SST anomaly is chosen to mimic the composite SST anomalies associated with the MJO found in Woolnough et al. (2000). The amplitude of the SST anomalies used here is greater than that of the composite SST anomalies,
but is comparable to the magnitude of individual SST anomalies associated with the passage of the MJO. A series of experiments were carried out with different values of $c$: $c = 0^\circ$ day$^{-1}$ (i.e. stationary) and $c = 4, 6, \text{ and } 12^\circ$ day$^{-1}$ (corresponding to periods of 90, 60 and 30 days, respectively) this permits investigation of the sensitivity of the response in the convection to the speed of the SST anomaly. The anomaly experiments were initialized with the final dump from the control integration and were run for 1260 days. Once again, the first 180 days were excluded from the analysis as a spin-up period.

3. Results

(a) Precipitation response

Figure 4(a) shows a Hovmöller diagram of the precipitation averaged between $5^\circ$N and $5^\circ$S for the integration with a propagating SST anomaly with a period of 60 days. Whilst the eastward moving equatorial Kelvin waves and the westward moving equatorial Rossby waves are still apparent there is a strong organization of the convection on the time-scale of the imposed SST anomaly. Figure 4(b) shows the wave-number-frequency spectrum for this anomaly integration. In addition to the peaks associated with the equatorially trapped waves, there is a dominant peak in the power spectrum at wave numbers 1–4 with a phase speed corresponding to that of the SST anomaly. This suggests that the SST anomaly organizes the convection, not only on the time-scale of the imposed SST anomaly, but also on the large space scales of the anomaly.

In addition to the longitudinal variations in the precipitation generated by the introduction of the SST anomalies there are changes in the zonal-mean fields relative to the control integration; these changes will be considered later. However, they are small enough that the organization of the convection by the SST anomalies can be discussed in terms of zonal variations within each integration. All the fields are composited relative to the centre of the SST anomaly by averaging in a coordinate frame moving with the SST anomaly. The compositing technique used here means that, for the propagating SST anomalies, longitude, when scaled by the propagation speed, can also be interpreted as time.

Figure 5 shows the composite equatorial ($2.5^\circ$N–$2.5^\circ$S) precipitation anomaly, from the zonal mean averaged over the last 1080 days of the integrations with SST anomalies with propagation speeds, $c = 0, 4, 6, \text{ and } 12^\circ$ day$^{-1}$. The longitudinal variations in the composite represent the mean effect of the organization of the precipitation by the SST anomaly. With a stationary SST anomaly imposed there is a maximum in equatorial precipitation collocated with the maximum in SST. This agrees well with observations of the tropical atmosphere where the climatological-mean maxima in precipitation are well correlated with the maxima in SST. Over the cold SST anomaly there is a minimum in equatorial precipitation. This response is at least partly forced by the local SST anomaly, but experiments by Neale and Hoskins (2001b) show that a localized positive SST anomaly will lead to suppression of the equatorial precipitation to the west, even in the absence of a cold SST anomaly, as a result of the dynamical response to the heating associated with the enhanced convection.

When an eastward moving SST anomaly is applied there is a westward relative shift of the maximum in precipitation away from the maximum in SST. The maximum in precipitation is reduced in magnitude, as the speed of the SST anomaly is increased. The minimum in precipitation is also shifted to the west of the cold SST anomaly. There is a much larger reduction in the amplitude of the negative precipitation anomaly, as the propagation speed is increased, than for the positive precipitation anomaly. The position
Figure 4. As for Fig. 2 except for the case of a moving sea surface temperature anomaly with a phase speed of 6° day\(^{-1}\), corresponding to a period of 60 days. The dashed line indicates eastward propagating waves with a period of 60 days.

of the precipitation maximum relative to the moving SST anomalies is also in good agreement with the observations of the MJO (e.g. Woolnough et al. 2000).

Figure 6(a) shows the undiluted convective available potential energy (CAPE) from each integration calculated for a parcel ascent from 975 hPa, averaged over the grid points at 1.25°N and 1.25°S. CAPE is a measure of the amount of energy which can be released in convective motions, and thus one may expect it to indicate regions where
Figure 5. Precipitation anomaly from the zonal mean averaged between 2.5°N and 2.5°S composited relative to the stationary sea surface temperature (SST) anomaly (solid line), and to the SST anomalies moving with speeds of 4° day⁻¹ (dot-dashed line), 6° day⁻¹ (dashed line) and 12° day⁻¹ (dotted line). The thick solid line shows the SST anomaly on the equator.

Figure 6. (a) Undiluted convective available potential energy (CAPE) and (b) convective inhibition (CIN), averaged over the grid points at 1.25°N and 1.25°S, for a parcel lifted from 975 hPa. Composited relative to the stationary sea surface temperature (SST) anomaly (solid line), and to the SST anomalies moving with speeds of 4° day⁻¹ (dot-dashed line), 6° day⁻¹ (dashed line) and 12° day⁻¹ (dotted line).

There will be deep convection. The longitudinal variations in CAPE do not match the variations in the precipitation, that is to say, the maximum and minimum in precipitation do not occur at the same location as the maximum and minimum in CAPE. However, CAPE measures the total energy which can be released in convective motions and it may be that the energy required to raise the parcel to the level of free convection is too great to allow the CAPE to be realized. The convective inhibition (CIN) is a measure of the amount of energy required to lift a parcel to the level of free convection and so minima in CIN may also indicate regions of deep convection. Figure 6(b) shows the CIN for a parcel ascent from the same level. Once again the maximum in precipitation is not collocated with the minimum CIN and also the variations in CIN between the
integrations with moving SST are very small. These features in the measures of low-level instability indicate that the character of the convective precipitation anomalies is not just a simple response to the low-level instability.

Both the CAPE and CIN are based on undiluted parcel ascents, whereas convection in the real atmosphere and in the model involves mixing with the environmental air. This mixing with the environmental air can affect the convection and precipitation in two ways. Changes in the humidity and temperature resulting from the mixing will affect the parcel buoyancy and hence the strength of the convection. In addition, changes in the humidity associated with mixing can affect the rate of generation of precipitation by the convection, even if the parcel buoyancy remains unaffected. The impact of the mixing on the convection means that the convective precipitation is dependent on the properties of the air in the free troposphere as well as those of the boundary layer.

Figure 7 shows the composite specific-humidity anomalies from the zonal mean averaged over 1.25°N and 1.25°S for each of the SST anomaly integrations. In each integration the positive boundary-layer specific-humidity anomaly is approximately collocated with the positive SST anomaly, indicating that the boundary layer responds very quickly to the surface forcing, in agreement with the CAPE and CIN. However, in the moving SST anomaly integrations the free-tropospheric humidity anomalies are shifted to the west of the surface anomaly. Because the anomaly plots are composited relative to the moving SST anomaly, a westward shift in space is equivalent to a lag in time. This westward shift of the specific-humidity anomaly indicates that there is a time lag between the moistening of the boundary layer and the moistening of the free troposphere. As this moistening occurs, the enhanced convection triggered by the low-level instability will take place in an increasingly moister environment. This means that the parcels will dry out less through entrainment. The moister parcels will retain their buoyancy excess for longer and will also generate more precipitation. This time lag for the moistening of the lower troposphere (900–600 hPa) explains the westward shift of the precipitation anomaly away from the maximum SST anomaly. For the stationary SST anomaly, there is no westward phase shift between the surface anomaly and the free-tropospheric anomaly, the moistening of the free troposphere being collocated with the SST anomaly and an equilibrium response to the forcing.

As the warm SST anomaly moves away, the source of the low-level instability is removed and the convection and hence the precipitation will begin to decrease. The removal of the source of the low-level instability as the cold SST passes depends on the spatial structure of the SST anomaly rather than its speed. As the spatial structure of the SST anomaly remains unchanged between each of the experiments, the location of the maximum precipitation anomaly is essentially fixed relative to the SST anomaly. However, the magnitude of the positive precipitation anomaly does depend on the speed of the anomaly, since the slower moving SST anomaly can influence the atmosphere for longer, the moistening of the free troposphere is greater, leading to larger humidity anomalies and greater precipitation. There is also a steady increase in the upper-tropospheric (600–300 hPa) specific-humidity anomaly as the speed of the SST anomaly decreases. This upper anomaly is an indication of the transport of moisture into the upper troposphere as the deep convection becomes stronger.

As the cold SST anomaly passes, the boundary-layer negative specific-humidity anomaly develops as a response to the SST anomaly, and is collocated with it. However, the reduction in boundary-layer specific humidity and temperature (not shown) results in a reduction in the low-level instability as can be seen from the CAPE and CIN (Fig. 6). The reduction in low-level instability leads to a reduction in the convection. But such a reduction in the convection means that low-level dry humidity anomalies cannot be
carried into the free troposphere in the same way as they are for moist anomalies. In fact, the negative specific-humidity anomalies in the free troposphere arise as a response to the anomalous descent which is associated with the suppression of the convection. Once again the moisture anomalies in the free troposphere act to modify the convection and so the slower moving anomalies have a larger negative precipitation anomaly. The lag between the surface anomaly and the free-troposphere anomalies is greater for the negative anomaly than the positive anomaly and this is indicative of the different
Figure 8. As for Fig. 7 except for vertical pressure velocity. (a) Stationary sea surface temperature anomaly, and 
SST anomalies moving with speeds of: (b) 4° day⁻¹, (c) 6° day⁻¹ and (d) 12° day⁻¹. The contour interval is 
0.5 mb hr⁻¹, negative contours (ascent) are solid, positive contours (descent) are dashed and the zero contour is 
dotted. Heavy contours are multiples of 1 mb hr⁻¹.

time-scales involved for the different processes determining the humidity anomalies. The stationary SST anomaly has a large negative specific-humidity anomaly associated with it and this is again an indication of the equilibrium response to the cold SST rather than the transient response which develops to the moving SST anomalies. The downward motion which develops in the region of the cold SST anomaly is enhanced by the vertical motion associated with the dynamical response to the heating. Hence the negative precipitation anomaly is modified by this dynamical response.
Figure 9. As for Fig. 7 except for zonal wind. (a) Stationary sea surface temperature SST anomaly, and SST anomalies moving with speeds of: (b) 4° day⁻¹, (c) 6° day⁻¹ and (d) 12° day⁻¹. The contour interval is 1 m s⁻¹, positive contours are solid, negative contours are dashed and the zero contour is dotted. Heavy contours are multiples of 2 m s⁻¹.

(b) The dynamical response

Figure 8 shows the vertical-velocity anomalies from the zonal mean for the four SST anomaly experiments. The maximum upward velocity is collocated with the maximum in precipitation, increases with the magnitude of the precipitation anomaly, and is associated with the heating anomaly generated by the enhanced precipitation. For the stationary SST anomaly and the 4° day⁻¹ and 6° day⁻¹ SST anomalies, the descent is concentrated in a region to the west of the ascent. However, for the SST anomaly moving at 12° day⁻¹ there are regions of anomalous descent to the west and east of
the ascent anomaly. The phase speed of $12^\circ$ day$^{-1}$ for the fastest moving SST anomaly is very close to that of the moist equatorial Kelvin waves in the control integration, and hence the heating associated with the enhanced convection is efficient at generating and maintaining Kelvin waves. The slower moving anomalies, however, do not have this phase locking and are less efficient at generating Kelvin waves. The ratio of the maximum upward anomaly to the maximum downward anomaly for the $6^\circ$ day$^{-1}$ and the $4^\circ$ day$^{-1}$ cases are similar at 0.48 and 0.46, respectively, but for the $12^\circ$ day$^{-1}$ case the ratios of the maximum descent (to the west) to the maximum ascent is 0.31. The strong Kelvin-wave response to the east for the $12^\circ$ day$^{-1}$ anomaly is consistent with this comparatively weak descent anomaly to the west of the convection. The weakness of the descent to the west of the heating anomaly for the $12^\circ$ day$^{-1}$ case results in a weaker free-tropospheric humidity anomaly in the region of the cold SST anomaly (Fig. 7), and this in turn contributes only weakly to the suppression of the precipitation there, giving rise to the relatively small negative precipitation anomaly.
Figure 11. As for Fig. 10 except at 850 mb.

Figure 9 shows the zonal-wind anomaly from the zonal mean for the four anomaly experiments. The zonal-wind anomalies confirm the increased strength of the Kelvin-wave response for the \(12^\circ\) day\(^{-1}\) SST anomaly case, with a low-level easterly anomaly to the east of the ascent region, characteristic of the Kelvin-wave structure. The slower moving SST anomalies have weaker zonal-wind anomalies to the east of the enhanced precipitation and stronger anomalies to the west. The upper-level structure in the outflow region of the anomalous ascent varies in the same way, with a strong westerly anomaly to the east of the ascent in the \(12^\circ\) day\(^{-1}\) case. For the slower moving SST anomalies the westerly anomaly is reduced and the upper-level flow is dominated by an easterly anomaly to the west of the ascent region.

Figures 10 and 11 show the anomalous wind fields for the four experiments at 200 mb and 850 mb, respectively, over the tropical region 30°N to 30°S. Again, the Kelvin-wave signal can be seen at both 850 mb and 200 mb to the east of the anomalous ascent for the \(12^\circ\) day\(^{-1}\) case (Figs. 10(d) and 11(d)). An upper-level quadrupole is seen with anticyclones to the west of the precipitation anomaly and cyclones to the east of the precipitation anomaly, consistent with Sverdrup balance and, for example, the Gill
(1980) linear model of the response to tropical heating. As the speed of the SST anomaly decreases the upper-level cyclones to the east of the precipitation anomaly weaken, and the anticyclones shift eastwards relative to the precipitation. In the stationary SST anomaly case (Fig. 10(a)) the upper-level anticyclones strengthen, with their centres even slightly east of the maximum ascent. The strong poleward flow associated with these anticyclones, makes a large contribution to the upper-level divergence in the region of the ascent. At 850 mb, for the slower moving SST anomalies in particular, the inflow into the ascent region is dominated by westerly anomalies to the west of the enhanced convection, with relatively weak flow to the east. The low-level cyclones are shifted to the east relative to their position in the Gill (1980) model.

(c) The zonal-mean response

The changes in the zonal mean induced by the SST anomalies are small. However, it is of interest to discuss them briefly. The region of enhanced convection acts as a source of Rossby waves which propagate out of the tropics. Associated with the Rossby waves is a flux of westerly momentum into the tropics which leads to an acceleration of the zonal flow and a westerly zonal-mean zonal-wind anomaly in the upper troposphere in the tropics. Figure 12 shows the difference in the zonal-mean wind from the control integration for the four SST anomaly experiments. For the moving anomalies the difference in the zonal-mean wind between the anomaly experiment and the control
experiment decreases as the anomaly moves faster because, as the precipitation anomaly reduces, the strength of the Rossby-wave source reduces. Associated with the westerly acceleration of the zonal flow in the tropics is a deceleration of the flow in the subtropical jets. The asymmetries in the zonal-mean anomalies may simply be sampling errors, either in the control integration or in the anomaly experiments. However, if there are slight asymmetries in the solar forcing which lead to slight asymmetries in the control integration, then the differences in the wave propagation characteristics associated with these asymmetries in the basic state could lead to large asymmetries in the zonal-mean anomaly fields.

In addition to the changes in the zonal-mean zonal-wind there are changes in the zonal-mean precipitation because of the nonlinearity of the processes involved. Figure 13 shows the zonal-mean precipitation anomalies for each of the four experiments. The largest changes in the zonal-mean precipitation occur for the stationary anomaly. For the stationary anomaly the equilibrium response to the convection leads to a small reduction in the equatorial precipitation and an enhancement of a twin intertropical convergence-zone structure in the precipitation which is a characteristic of the control integration. For the moving SST anomalies, the structure of the zonal-mean precipitation anomalies is similar to that for the stationary case, but the magnitude of the anomalies is reduced. Associated with the changes in the zonal-mean precipitation is a small reduction in the strength of the Hadley circulation (not shown).

4. DISCUSSION AND CONCLUSIONS

(a) Conclusions

The SST anomaly experiments have demonstrated that tropical convection can be organized on intraseasonal time-scales by SST anomalies with structures and magnitudes comparable to those observed in the Indian Ocean and west Pacific associated with the passage of the active phase of the Madden–Julian Oscillation. There are aspects of the
model’s response to intraseasonally varying SSTs, such as the location of the maximum in precipitation, which are in agreement with the observed structure of the MJO (e.g. Woolnough et al. 2000). This suggests that the processes described in this paper by which the atmosphere responds to varying SSTs may be relevant to the real system.

The strength of the response in the convection to the SST anomalies is dependent on the phase speed of the anomalies. Faster moving anomalies generate a weaker response than slower moving anomalies. This sensitivity to the phase speed arises because of the way the increase in the precipitation, due to the enhancement of the low-level convective instability generated by the passage of the positive SST anomaly, is affected by the moisture anomalies in the free troposphere. The relatively dry air in the free troposphere is entrained into the convecting plumes, thus reducing the parcel buoyancy and/or the moisture available for precipitation. However, the enhanced convection acts to moisten the free troposphere, through vertical advection associated with the latent heating and detrainment from the convecting plumes, and so reduces the difference in moisture between the parcels and environment. For the slower moving anomalies the low-level instability remains for longer and so a greater moistening of the free troposphere can occur and larger precipitation anomalies can develop. Once the low-level SST anomaly moves away, the surface instability is removed and the precipitation begins to decrease again. Hence the position of the maximum in precipitation relative to the SST anomaly is determined by the spatial structure of the anomaly rather than its time-scale, and for the moving SST anomalies the maximum in precipitation occurs at approximately the centre of the SST anomaly dipole. The location of the precipitation near the centre of the SST dipole anomaly is in agreement with observations of the convection associated with the MJO (e.g. Woolnough et al. 2000). For the stationary SST anomaly the moistening of the free troposphere is collocated with the SST anomaly. Hence the precipitation maximum is collocated with the SST anomaly, as observed for stationary SST anomalies, e.g. the west Pacific warm pool.

The dynamical structure which develops in response to the convection is also sensitive to the propagation speed of the SST anomalies. For the SST anomaly moving at 12° day\(^{-1}\), the phase speed is close to that of the intrinsic moist Kelvin waves of the control integration and, as such, the anomalies are efficient at forcing Kelvin-wave-like responses to the heating. The slower moving anomalies are less efficient at generating Kelvin waves because of the mismatch in period and the response is dominated by westerly inflow at low levels and easterly outflow at upper levels. This strong westerly inflow is similar to the observed surface wind-stress anomalies in Woolnough et al. (2000) associated with the MJO.

\(b\) Implications for the Madden–Julian Oscillation

The organization of tropical convection by intraseasonal SST anomalies consistent with the observations of the MJO, demonstrated here, together with the forcing of SST by the tropical convection shown in Woolnough et al. (2000) indicates that there is a clear potential for the MJO to be a coupled atmosphere–ocean phenomenon. It is not possible at this stage to categorically state that its existence depends on the coupling. However, it is clear that some of the characteristic properties of the MJO must be determined by coupled processes.

The mechanism by which the convection responds to the SST anomalies involves a time lag due to the moistening of the free troposphere. This time lag favours longer time-scales than the convectively coupled equatorially trapped waves, which are invoked in the WAVE–CISK mechanism of Lau and Peng (1987) or the WISHE mechanisms of Emanuel (1987) and Neelin et al. (1987). Although it is not possible to estimate
exactly what this time lag might be from these experiments, since the SST is constantly changing, the results suggest that it is of the order of a few days. This time-scale is consistent with that found for convection to respond to a reversal in the underlying SST gradient in the cloud-resolving model experiments of Tompkins (2001), where the time delay in the response of the convection to the SST changes was again related to the moistening of the free troposphere.

Furthermore, for the coupled mechanism proposed, the eastward propagation of the SST anomalies is driven by the surface fluxes associated with the convection. For slow moving SST anomalies, the associated enhancement of the convection will be stronger, generating larger surface flux anomalies and the SST anomalies will tend to grow and decay faster. For fast moving SST anomalies the surface fluxes will be weaker, hence the SST tendency will be weaker and the SST anomalies will change more slowly. This feedback will act to mediate the speed of the SST anomalies and provide a characteristic speed for such a coupled phenomenon.

In this study, the response to intraseasonally varying SSTs appears to involve an adjustment of the humidity of the free troposphere by the buoyancy-driven convection associated with the warm SSTs. In contrast, Waliser et al. (1998) and Wang and Xie (1998) suggest that the enhanced convection in response to the varying SSTs arises from dynamically forced convergence, although the component of the wind field that gives rise to this convergence varies between the two studies.

As mentioned in section 2, the control integration of the aqua-planet model has surface easterlies at the equator, in contrast to the westerlies required for the proposed coupled mechanism to work. The surface westerlies are only required for the coupled mechanism in that they determine the sign of the surface latent-heat flux anomalies, which are partly responsible for the eastward propagation of the SST anomalies. However, the evaporation also determines the boundary-layer humidity anomalies which are partly responsible for the increase in low-level instability as measured by the CAPE. Thus the sign of the basic-state winds may also have an effect on the boundary-layer humidity. It can be seen from Fig. 11 that, with the exception of the anomaly with a propagation speed of 12° day⁻¹, the easterly wind anomalies to the east of the enhanced convection are weak and so any modification of the latent-heat flux by these winds will be small. However, Fig. 7 shows that the boundary layer responds very quickly to the surface forcing, indicating that the low-level humidity response is forced largely by the surface temperature variations. Thus it is considered that the easterly low-level winds in the experimental set-up used here do not invalidate the current study.

(c) Future work

It has been demonstrated here and in Woolnough et al. (2000) that the two components of the coupled system, the atmosphere and the ocean, can act on intraseasonal time-scales to force the other component of the system, indicating the potential for truly coupled processes to exist. However, it has not been demonstrated that the phase relationships and strength of the forcings by the atmosphere on the ocean, and ocean on the atmosphere, are arranged in such a way as to generate or maintain an eastward propagating mode of convective modulation. To fully determine this, experiments must be carried out in a model which allows coupling between the atmosphere and ocean. Once again there are two approaches which can be adopted for such work. A fully coupled GCM should contain all the processes required to exhibit such coupled modes of behaviour should they exist, but the weakness in the representation of the basic climate which current coupled models exhibit means that such a coupled process can not be ruled out purely because of its absence in a coupled GCM.
Such a coupled modelling approach is already underway, but more idealized experiments would serve to investigate the sensitivity of the coupled processes to the parameters which may determine the presence and properties of the coupled modes, and which may be misrepresented or absent in current coupled GCMs. These idealized experiments should include a continuation of the aqua-planet experiments, with a fully interactive coupled ocean model, of varying complexities, including slab oceans or simple mixed-layer models. However, investigation of the coupled modes in an aqua-planet GCM would require modifications of the control integration of the atmospheric component to generate the surface westerlies at the equator which are crucial to the coupled mechanism.

A further series of sensitivity experiments could be carried out, within the aqua-planet framework used here, to investigate the sensitivity of the response to the SST anomalies to the type of convection scheme, or parameter choices within a given convection scheme. Such experiments have already been suggested by Neale and Hoskins (2001a) for the response to stationary SST anomalies in the context of testing physical parametrizations and their interaction with the dynamics. Experiments by Wang and Schlesinger (1999) have indicated that in a full atmospheric GCM the representation of the MJO is dependent on the parameters within a given convection scheme. The mechanism proposed for the response of the convection to propagating SST anomalies will clearly result in a dependence of the response on the choice of convective parametrizations and parameters within the convection scheme. In the convection scheme used here for example, one would expect the response to be sensitive to the entrainment rate specified within the convection scheme.

ACKNOWLEDGEMENTS

Dr S. J. Woolnough was supported by the Natural Environmental Research Council (Grant No. GR3/10798). The authors would like to thank Richard Neale and Pete Inness for useful discussions. Jeff Cole and Lois Steenman-Clark are thanked for their help with the GCM integrations.

REFERENCES

Gregory, D. and Rowntree, P. R. 1990 A mass flux convection scheme with the representation of cloud ensemble characteristics and stability dependent closure. Mon. Weather Rev., 118, 1483–1506


Wang, W. and Schlesinger, M. E. 1999 The dependence on convection parameterization of the tropical intraseasonal oscillation simulated by the UIUC 11-layer atmospheric GCM. *J. Climate*, 12, 1423–1457

