The MJO in Global Climate Models

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Abstract

The impacts of various changes to the configuration of the Hadley Centre GCM on the simulation of the Madden-Julian oscillation (MJO) are reviewed in the light of theoretical and observational studies. It is shown that the representation of tropical cumulus congestus clouds, the inclusion of an interactive ocean model and the quality of the basic state climatology of a GCM are all important factors in simulating the MJO. However, fundamental problems still remain in the simulation of tropical variability on a wide range of time and space scales which prevent a truly realistic representation of the MJO and other tropical wave phenomena.

1. Introduction

Since its discovery in the 1970's, the Madden-Julian Oscillation (MJO) has proved a fascinating but elusive challenge to the numerical climate and weather prediction modelling community. The desire to simulate the MJO in models has been driven both by the potential for predictability at medium range to seasonal lead times, and also by the desire to show that models can reproduce the many complex processes and interactions which go together to determine the nature of the MJO. In some senses, simulating the MJO has become something of a benchmark test for a climate model. There have been many studies published which show how changes to model formulation have affected the simulation of the MJO. These have included examples of how apparent improvements to the physical realism of a GCM have had a negative impact on the MJO as well as many cases when model changes have improved the MJO simulation. Such studies have tended to focus on the impact of changes to the convective parametrization of numerical models, but changes to other aspects of model formulation could also potentially impact upon the MJO simulation. In this review, the impact of 3 different changes to a GCM will be discussed in the light of current theories and observational studies. Firstly, the impact of increased vertical resolution and the associated change in the tropical cloud population will be examined. Secondly, the impact of coupling an atmospheric GCM to an interactive ocean will be described, and finally the impact of an artificially forced change to the tropical basic state of the coupled GCM will be considered. The model used throughout is the Hadley Centre climate model in its atmosphere-only (HadAM3) and atmosphere-ocean coupled (HadCM3) versions – see Pope *et al*. (1999) for a description of the atmospheric component of this GCM.

2. Increased vertical resolution

Inness *et al.* (2001) studied the impact on the MJO simulation of a change to the vertical resolution of HadAM3. Whilst the vertical resolution remained the same in the boundary layer, resolution was doubled through the depth of the troposphere, giving 30 levels (L30) compared to the standard 19 (L19). The levels in the boundary layer remained unchanged and the resolution in the mid-troposphere increased from about 100 hPa to 50 hPa. It was found that the MJO index, described by Slingo *et al.* (1999) was quite different in the 2 versions. Whilst the standard version of the GCM had a rather weak signal, the MJO index in the 30 level version was comparable to the index derived from re-analyses (see fig. 1).

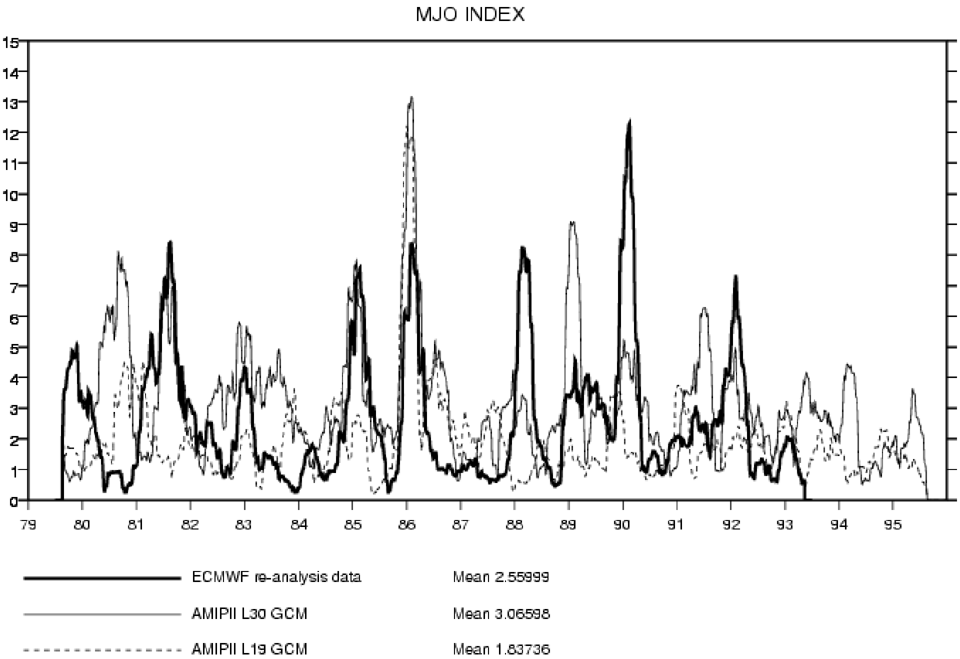

Figure 1: MJO index based on intraseasonal variability of 200 hPa zonal mean zonal wind averaged between 10°N and 10°S.

An aquaplanet configuration of the GCM was used to investigate differences in the behaviour of tropical convection in the L19 and L30 versions. It was found that, over warm tropical SSTs, the L19 version produced deep, precipitating convection virtually all the time. In the L30 version, convection varied between troposphere-deep, precipitating cumulonimbus and shallower cumulus congestus with tops close to the 0°C level (around 600 hPa). These clouds only precipitate weakly and detrain moisture into the mid-troposphere. Figure 2 shows the contribution of the convection scheme to specific humidity for the 2 versions. Convection in the L19 version was always acting to dry the troposphere, whereas in the L30 version there were significant periods when convection was actually moistening the lower and mid-troposphere. Observational studies from the TOGA-COARE (e.g. Lin and Johnson, 1996) show that, during the suppressed phase of the MJO in the West Pacific, convection does indeed significantly moisten the mid-troposphere with a capping inversion near the 0°C level. The formation of this inversion is associated with melting precipitation during the active phase of the MJO. Such an inversion was also seen in the GCM results, and was more clearly resolved in the L30 version.

Inness *et al*. (2001) proposed that this moistening by cumulus congestus was a key process in representing the intraseasonal variability of convection in the GCM, acting to precondition the free troposphere during the suppressed phase of the MJO for the next burst of deep convection. This is consistent with the “recharge-discharge” mechanism of Bladé and Hartmann (1993), and also with more recent work on the feedback between tropospheric moisture and convection (e.g. Tompkins, 2001). However, it was noted that even the L30 version of the GCM was barely resolving the 0°C inversion and the congestus phase, and so it may be better to try to account for these effects more explicitly within the convective parametrization. It was also noted in this study that in both versions of the GCM, the MJO convective region did not propagate eastwards as observed. Instead it tended to oscillate in situ, with centres of action in the eastern Indian Ocean and West Pacific.

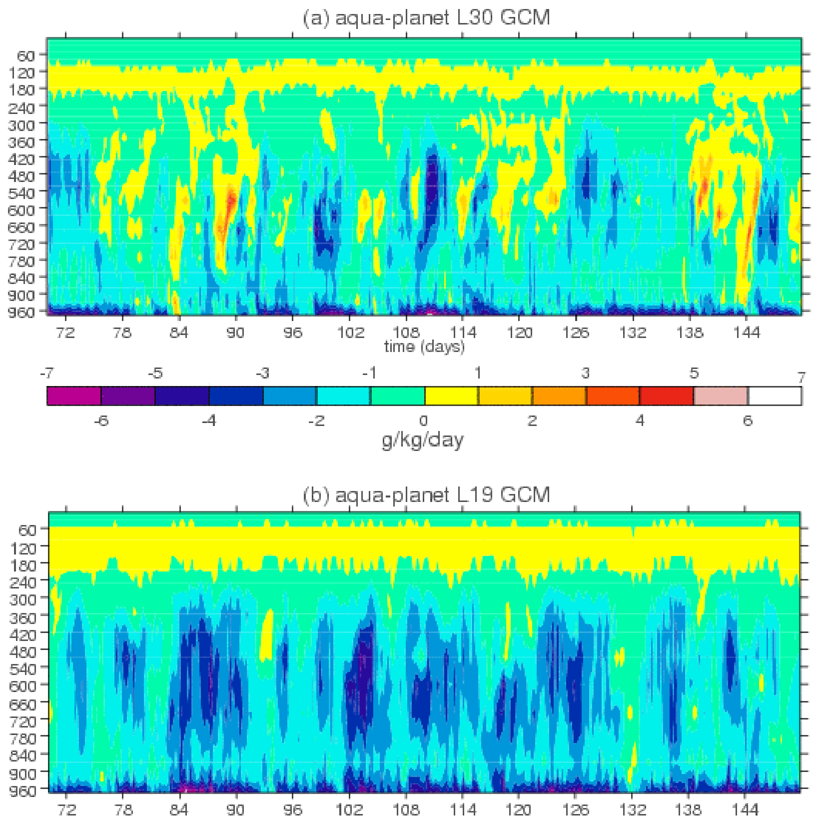

Figure 2: Time-height sections of convective increment to specific humidity averaged over 9 equatorial grid-boxes, from; (a) L30 and (b) L19 versions of HadAM3

3. The impact of air-sea interaction

Over recent years there have been several observational, theoretical and modelling studies which have suggested that air-sea interaction may be important in the maintenance of the MJO (e.g. Zhang, 1996; Flatau *et al*., 1997; Waliser *et al*., 1999). In particular it has been proposed that the characteristic phase speed of the eastward propagating convective maximum may be regulated by intraseasonal SST variability. This SST variability is itself forced by surface flux anomalies associated with the convection. It is certainly the case that atmospheric GCMs forced by prescribed SSTs tend to produce an MJO signal which propagates too quickly (e.g. Slingo *et al*., 1996) and closer study of the intraseasonal variability in such models (e.g. Sperber *et al*., 1997) has shown that the eastward propagation of convection can be rather poorly captured.

Inness and Slingo (2003) examined the impact of coupling an atmospheric GCM to an interactive ocean model. They used the Hadley Centre coupled GCM HadCM3 with 30 vertical levels in the atmosphere – the same atmospheric component used in Inness *et al*. (2001). The model was run for 20 years and the MJO behaviour examined. Figure 3 summarizes the eastward propagation of intraseasonal convection for observations (AVHRR outgoing longwave radiation), and the atmosphere-only and coupled versions of the GCM. The observations show eastward propagation from the Indian Ocean, across the Maritime Continent and into the West Pacific. The signal dies out at around the date-line as the convection moves over the cooler waters of the equatorial Pacific cold tongue. The convection in the atmosphere-only GCM shows no eastward propagation, with the intraseasonal variability being characterised by a standing oscillation as described above. By contrast, the convection in the coupled model does propagate eastwards across the Indian Ocean and Maritime Continent, but the signal dies out at around 120°E rather than extending to the date-line.

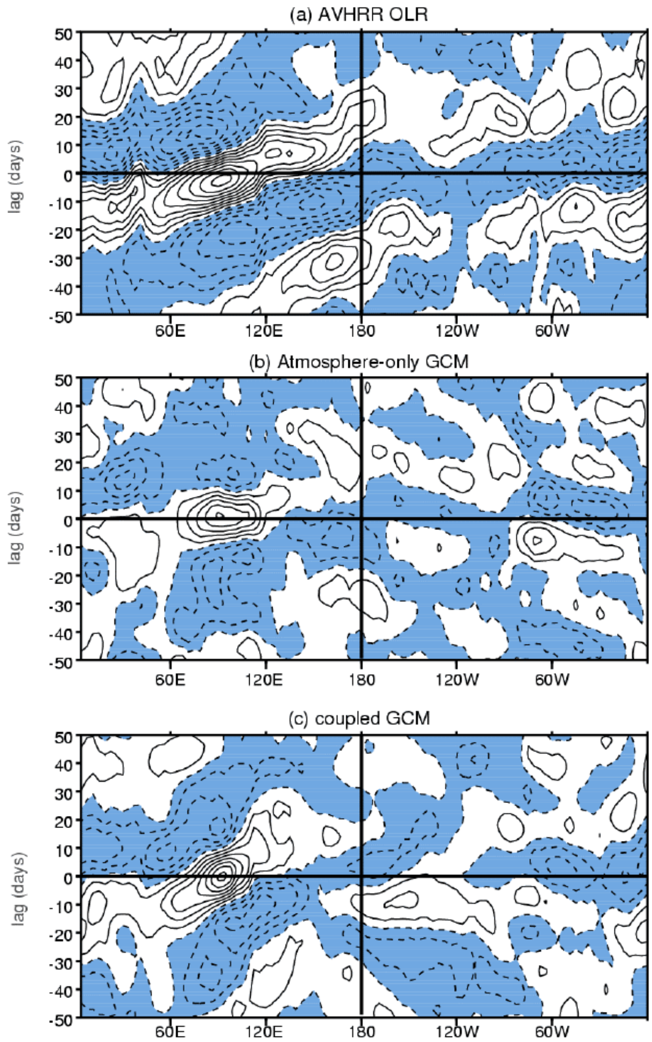

Figure 3: Lag correlations of OLR with 20-100 day filtered 200 hPa velocity potential (VP) at 90°E from; (a) AVHRR OLR and ECMWF re-analysis 200 hPa VP, (b) HadAM3 atmosphere-only GCM, (c) HadCM3 coupled GCM. Contour interval is 0.1 and negative contours are dashed. All fields averaged between 10°N and 10°S.

The SST and surface fluxes in the coupled model were examined and compared with observations. The correlations of various surface fields with the intraseasonal convective signal are shown in fig. 4. In the Indian Ocean the SST pattern (fig. 4a) shows a warming which peaks about 10 days prior to the convective maximum, followed by a cooling which peaks about 10 days after the convective maximum. These SST variations are due mainly to variations in the surface latent heat and shortwave fluxes shown in fig. 4(b) and (c). SST warming prior to the onset of convection is due to shortwave heating under clear skies, together with reduced latent heat flux from the ocean surface in light winds. The onset of convection reduces the shortwave flux to the surface and westerly wind anomalies coincident with, and to the west of, the convection lead to increased evaporation from the surface, giving SST cooling. These patterns are fairly consistent with the observed patterns described in Woolnough *et al*. (2000), but the relationship between convection and latent heat flux anomalies only holds out to about 140°E instead of extending to the date-line. Figure 4(d) shows that, to the east of this longitude, the correlation between latent heat flux anomalies and surface zonal wind-stress actually changes sign, implying that westerly wind stress anomalies would lead to *reduced* evaporation and thus SST *warming* east of 140°E. The reasons for this will be examined in the next section, but the result is that the relationship between surface fluxes and intraseasonal SST variations breaks down at this point, close to where the eastward propagation of convection also stops in the model.

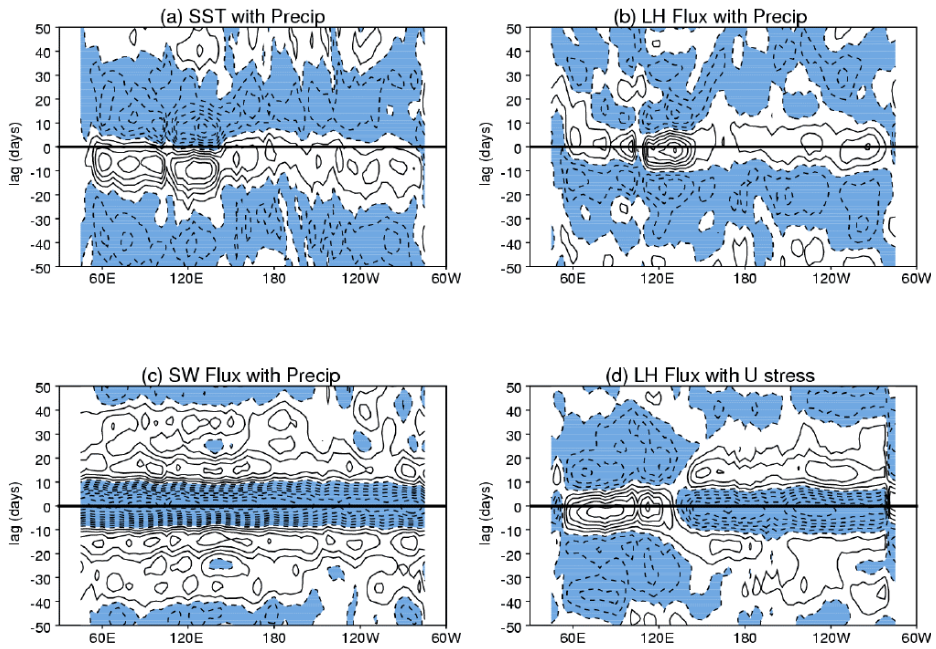

Figure 4: Lag correlations of surface quantities with convective precipitation or zonal wind-stress from HadCM3 coupled GCM. (a) SST with precipitation, (b) latent heat flux with precipitation, (c) SW flux with precipitation, (d) latent heat flux with zonal wind-stress. All fields averaged between 10°N and 10°S and filtered with a 20-100 day band-pass filter. Contour interval is 0.1 and negative values are dashed. The sign convention for fluxes is that a positive flux is into the ocean surface.

The results of this study suggested that air-sea interaction is important for the eastward propagation of the MJO convective signal. However, it was also shown that the intraseasonal surface flux and SST variations simulated by the model were weaker than those in various observational studies. The MJO convective signal was also weaker than observations. The weaker than observed SST variability may in part be due to the inability of the model to reproduce the extreme shoaling of the ocean mixed layer during the suppressed phase, and the strong diurnal cycle of SST that occurs at this time. These aspects are discussed by Steve Woolnough elsewhere in these proceedings.

Whilst the coupling of the atmospheric GCM to an ocean model improved the simulation of the MJO in this study, there are examples where coupling has had a negligible impact on the MJO simulation of a GCM. Hendon (2000) showed that coupling did not materially change the nature of the MJO simulation in one particular GCM. This was shown to be due to a combination of weaker than observed surface latent heat flux variations, together with a change in the phasing of the fluxes relative to the convection. This second problem was related in turn to problems with the basic state climate of the coupled GCM. This issue of errors in the basic state forms the third part of this review.

4. The role of the basic state

It is very difficult to study the impact of air-sea interaction on the MJO simulation of a GCM in isolation. Coupling the model to an interactive ocean will not only introduce physical processes which may be important to the maintenance of the MJO, but will also result in changes to the basic state climate of the model. The nature of the MJO is itself affected by the basic state through which it is propagating and so it is possible that, in a coupled GCM, errors in the basic state will offset any positive impacts on the MJO of coupling to an interactive ocean.

Inness *et al*. (2003) investigated the lack of propagation of the MJO in the West Pacific in HadCM3 which was described in the previous section. They hypothesised that this was due to errors in the basic state of the coupled model, and in particular that the low level westerly winds observed on and to the south of the equator in the West Pacific were not present in the coupled GCM. Instead the flow in this region was easterly. This led to the opposite sign of latent heat flux anomalies associated with the MJO to that observed so that, in the model, there was enhanced latent heat flux out of the ocean surface to the east of convection, with reduced flux to the west. This explains the change in the sign of the correlation between surface zonal wind stress and latent heat flux at about 130^{o}E shown in fig. 4(d).

In order to test this hypothesis, flux adjustments were applied to the ocean to reduce the cold SST bias along the equator in the Pacific. This in turn weakened the easterly trade winds in the Pacific and introduced surface westerly winds in the West Pacific. In this version of the coupled model, the MJO convection was found to propagate through the West Pacific out to the date-line. The possibility remained that it was the change in SST that was affecting the MJO directly, with the increased SST in the West Pacific supporting deep convection where previously the SST was too cold. A version of the coupled model in which the same atmospheric component was coupled to a different ocean model was also examined. In this configuration the SSTs were actually slightly too warm in the West Pacific but the westerly winds were still absent. Analysis of the MJO in this version showed that the convection still failed to propagate into the West Pacific. A lack of westerly basic state surface winds in the West Pacific was also cited by Hendon (2000) as a reason why coupling had a negligible impact on the MJO simulation of a different GCM.

5. Outstanding issues

It should be emphasised that, despite the various improvements described in this review, the MJO simulation of the HadCM3 GCM is still far from perfect. In particular, the overall variability of convection near the equator is very weak on synoptic to intraseasonal timescales. Hence the MJO convective signal is itself much weaker than observed. The convective activity in this GCM is also not well coupled to the various equatorially trapped wave modes (G-Y. Yang, personal communication). This points to more fundamental deficiencies of the model, related to the interaction of various aspects of the model physics and dynamics, and in particular the behaviour of the convective parametrization. This is an issue with no clear solution and is certainly not unique to this particular GCM. Over recent years there has been an increasing interest in analysing the sub-annual climate variability simulated by GCMs as well as the simulated mean state and interannual variability. Much attention has been focused on tropical intraseasonal variability in GCMs (e.g. this workshop), but variability on other timescales may impact both the mean climate and the entire spectrum of climate variability through the kind of scale interactions discussed by Slingo *et al*. (2003). The diurnal cycle of tropical convection both over land and over the oceans is one example of how high frequency variability can affect both the mean state (e.g. Yang and Slingo, 2001) and variability at lower frequencies (e.g. Johnson *et al*., 1999). So in order to make real breakthroughs in the simulation of the MJO and tropical variability in general, we need to focus considerable effort on understanding the behaviour of the physics in our models at the most fundamental levels.

There are other issues that may be easier to address. The SST and surface flux variability associated with the MJO is too weak in HadCM3, and this may be due to the coarse vertical resolution of the upper ocean and the fact that the ocean and atmosphere exchange information only once per day. Modifications to the upper layers of the ocean model, and to the ocean-atmosphere coupling frequency could bring about improvements here. Another issue is that of tropical cumulus congestus cloud. Even in the L30 version of the GCM this type of cloud still occurs less frequently than observed over the warm-pool region. Careful consideration of how to model the physical processes important to the formation of this type of cloud should lead to a more realistic representation of its occurrence in GCMs.

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