Air-Sea Processes in the Indian Ocean and the Intraseasonal Oscillation

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ABSTRACT

Some issues related to the role of the air-sea interaction in the triggering and the evolution of tropical intraseasonal convective events are briefly reviewed. We discuss in particular (i) the link between large-scale organized convection and the intraseasonal variability; (ii) the potential role of the formation of ocean Diurnal Warm Layers in the triggering of large-scale organized convective perturbations; (iii) the ability of the current GCMs to reproduce the main characteristics of the intraseasonal variability of the convection.

1. The intraseasonal variability of the SST

In situ (Sengupta and Ravichandran 2001) and satellite (Harrison and Vecchi 2001; Duvel et al 2004; Duvel and Vialard 2007) observations reveal that the SST can be strongly modulated at intraseasonal timescales (20-90 days). The strong intraseasonal perturbations of the tropical convective activity could thus be primarily controlled by air-sea interaction processes. Modelling studies indeed suggest that air-sea interactions could play an important role in the physics of this intraseasonal variability (e.g. Waliser et al. 1999; Inness and Slingo 2003; Woolnough et al. 2007).

This interaction between the SST and the convection at the intraseasonal time scale is based mainly on the perturbation of the surface fluxes by the atmospheric convective activity organized at large scales (1000-5000km). In convectively suppressed conditions, the ocean surface warms up because of the stronger solar surface flux. This effect is larger if these suppressed conditions are associated to low wind conditions (reduced surface turbulent heat fluxes and formation of ocean warm layers). During a convective event, enhanced evaporation and reduced solar flux related to the convective activity cool the surface and thus tend to reduce the convective instability. This effect is local but has also remote effect due to the large-scale dynamical Matsuno-Gill-type response to the convective heating of the mid-troposphere. The amplitude of the resulting ISV of the SST will also depend on the local ocean Mixed Layer Depth (MLD). A relatively small MLD is necessary to have a large (i.e. a few K) warming and cooling at the intraseasonal time-scale due to perturbation of the surface fluxes. However, the MLD has also to be thick enough to contain sufficient energy to maintain the convective instability for around 10-15 days, and this despite an enhanced surface latent heat flux and a reduced solar input. Using an atmospheric GCM coupled to a slab ocean with a fixed depth, Maloney and Sobel (2004) showed that the ISV of the convection during NH winter is indeed maximal for a MLD around 20-30 meters.

Results reported in Duvel and Vialard (2007) and in Vialard et al (2008) tend to show that the surface fluxes and the MLD are important factors governing the intraseasonal perturbations of the SST. However, other factors such as surface ocean diurnal warm layers (DWL) have also to be taken into account to fully reproduce the perturbation of the SST fields at intraseasonal time scales. Woolnough et al (2007) showed that the prevision of the equatorial ISV (Madden-Julian) in winter is improved by using a simple ocean mixed layer model with a high vertical resolution (enabling the simulation of ocean diurnal warm layers) instead of a full OGCM with a first layer with a depth of 10m.

For some regions such as the thermocline ridge south of the equator in the Indian Ocean, or the northwest Australian shelf, other processes have also to be investigated to fully understand the physical origin of the strong SST variability at intraseasonal time-scales. In particular, the role of this thermocline ridge in increasing the efficiency of exchanges with deeper water and thus the cooling of ocean mixed layer during strong wind episodes has to be investigated.

As noted in Bellenger and Duvel (2007, 2008), all these processes are not necessarily independent. For example, region with low surface wind speed (that is generally the case for ITCZ regions) will favour at the same time a relatively small MLD and the formation of DWL. If a large-scale convective event is generated by the high SST following a suppressed episode, the MLD will deepen and cool and DWL will dissipate under the action of the wind generated by large-scale organized convection. If the surface wind is generated only by this local convection, low-wind conditions, small MLD and DWL will reappear as soon as the convective instability decreases. However, for the northern Indian Ocean (Arabian Sea and Bay of Bengal), there is a large ISV in May and June due to large intraseasonal convective events before the onset or associated to the monsoon onset. In this last case, the deepening of the ocean mixed layer and the suppression of the DWL are more persistent since they result from the setting of the monsoon jet following the convective event. According to the mechanism described above, this could explain the decrease of the ISV amplitude in the heart of the monsoon season (July-September) above the Northern Indian Ocean.

The following sections report results of recent studies undertaken at LMD. Some of these results are already reported in submitted or accepted papers.

2. Diurnal warm layer characteristics

A long (1979-2002) time series of the diurnal amplitude of the SST is constructed to study, among other processes, the potential impact of DWL formation on the ISV of the SST. We use a single approach (Bellenger and Duvel 2008) that can be used to: (i) diagnose the main characteristics of the DWL over the whole tropical zone and; (ii) compute the perturbation of the surface fluxes related to the presence of DWL in AGCM. The approach is based on simple DWL models forced by large-scale surface meteorological field. The approach is evaluated by constructing a long time series with two DWL models forced by hourly-interpolated surface parameters given by the ECMWF Re-Analysis (ERA40) product. One advantage of this approach is the homogeneity of the results given by the relative homogeneity of ERA40 fields. This study uses both the DWL models of Fairall et al. (1996) and of Zeng and Beljaars (2005) in order to test the sensitivity of the results to the scheme considered. The DSA are computed for each region and each day by forcing these DWL models by hourly fields constructed from the 6-hourly fields of ERA40. This is more precise than to evaluate the DSA from empirical relation based on daily statistics and this gives an approach usable in GCMs. Temperature measurements of the Surface Velocity Program (SVP) drifters of the Marine Environmental Data Service (MEDS) and previous empirical models obtained from satellite measurements are used to validate the approach at the global scale. Results show that the Fairall algorithm tends to underestimate the diurnal SST amplitude (DSA) for very strong DWLs but gives correct estimate for low and medium (the more common) values. The Zeng and Beljaars algorithm tends to overestimate quite strongly the DSA for these low and medium DWLs. Despite these discrepancies, these evaluations show that such simplified models (especially the Fairall et al. model) can be used to parameterise the effect of the DWL for an improve computation of the surface fluxes in GCMs (see next section).

The largest average DSA are obtained over the northern Indian Ocean before the monsoon onset (Fig.1), over Timor and Arafura seas (north of Australia) around October, over eastern equatorial Pacific from February to April, and near the Gulf of California almost throughout the year (with average values up to 1.8K in July). Strongest seasonal variations of the DSA are observed in the Indian Ocean in relation with the seasonal march of the monsoon. In particular, the strong DSA over the whole northern Indian Ocean decays in June south of the Bay of Bengal in association with low-level wind burst related to pre-monsoon intraseasonal events and monsoon onset events (Bellenger and Duvel 2007). Between June and August, the strong low-level monsoon jet prevents the formation of DWL in the whole Northern Indian Ocean, but a relative DSA maximum appears over the northwestern Pacific Ocean where the low-level wind is still weak (Bellenger and Duvel 2007). During boreal winter, the DSA is maximal south of the Equator in the Indian Ocean and north of Australia. Some DWL activity remains over the equatorial Indian Ocean during boreal summer (June to September) with calm conditions between the trade winds to the south and the monsoon jet to the north. There are also calm conditions giving large DSA in the Mozambique Channel between October and January.

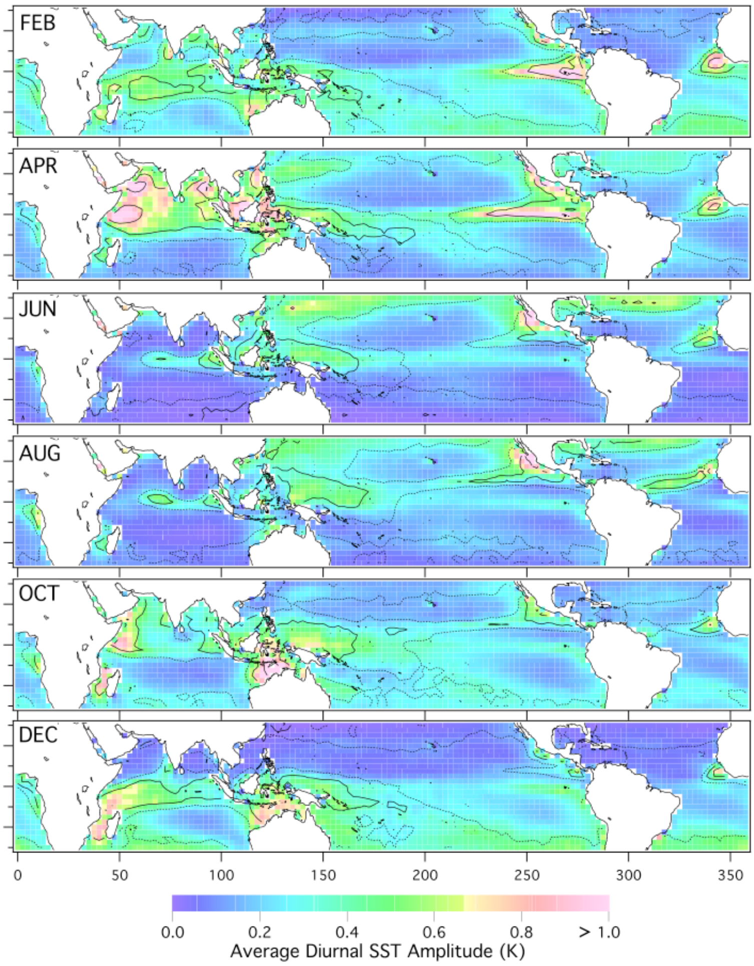

Figure 1:Monthly average DSA (colors, in K) and DWL depth (5m and 7.5m thin solid line, 10m dotted line) for the period 1979-2002. The average DSA is computed using all days (even those without DWL, but the average DWL depth is computed only for days with a DWL.

Monthly mean DWL depth (Fig. 1;computed from the Fairall algorithm) is clearly related to the monthly average DSA with a depth generally smaller than 5m for DSA larger than 0.8-1.0 K. The effect of the solar flux is also clearly evident by comparing the northern and the southern hemisphere in June and in December. For identical average DWL depths, the DSA is obviously larger in the summer hemisphere. The relation between the DWL depth and the DSA is however not simple and the relation between monthly average values should be considered with care. For daily values (not shown), the dispersion of the DSA values increases for smaller DWL depth with DSA values between 0 and 3.5K for a depth of 1m, between 0 and 1K for a depth of 5m, and between 0 and 0.5K for a depth of 10m.

The whole DWL time series (1979-2002) constructed using the Fairall et al. (1996) algorithm and ERA40 is used to analyze the potential role of the DWLs in the variability of the tropical climate. The perturbation of the surface fluxes by the DWL can give a cooling of the ocean mixed layer as large as 2.5 K/year in some tropical regions (Fig. 2). On a daily basis, this flux perturbation is often above $10Wm^{-2}$ and sometimes exceeds $50Wm^{-2}$, representing a large fraction of the surface turbulent flux, especially for these low surface wind speed episodes. DWLs can be organized on regions up to a few thousand kilometers and can persist for more than 5 days (Fig.3). To give an idea of the temporal occurrence of these large DWLs, equivalent radius larger than 1000 km appears between 2 and 3 times by year for a threshold of 1.42K and between 2 and 3 times by week for a threshold of 0.68 K.

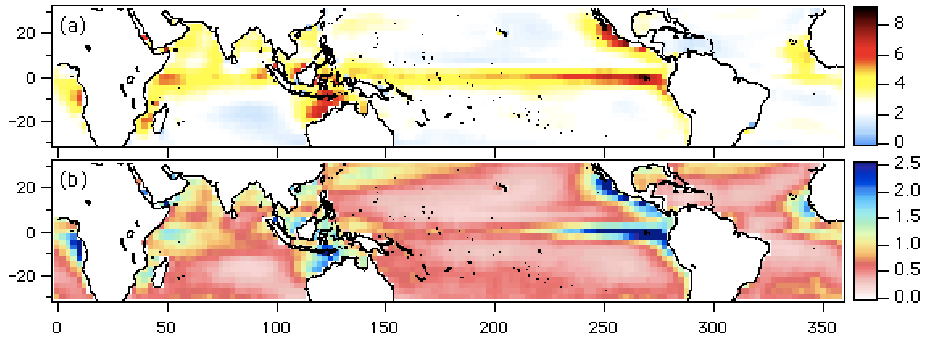

Figure 2: (a) Annual mean surface flux perturbation (surface cooling in Wm^{-2}) due to DWL and (b) corresponding annual mean cooling of the ocean mixed layer (K/year). (Bellenger and Duvel 2008)

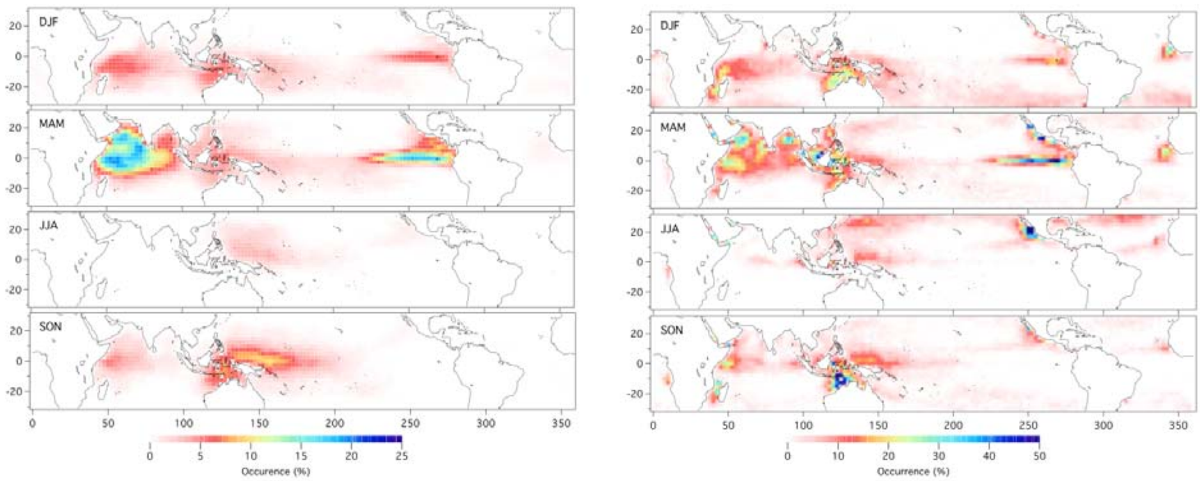

Figure 3: Occurrence (% of day) of (left) DWL with an equivalent radius larger than 1000 km and (right) DWL with duration larger than 5 days, for a DSA > 0.68 K for 1979 to 2002. DWLs reaching 30° of latitude are not considered.

Regions and seasons with large average DSA are thus also regions and seasons with persistent and extended DWLs. It is striking that extended and persistent DWLs occur mostly over monsoon regions (northern Indian Ocean and Timor Sea), just prior to the monsoon onsets (respectively April and November). These seasons also precede the large intraseasonal events associated to these onsets (Bellenger and Duvel 2007). During the summer monsoon season, the DWLs nearly disappear over the northern Indian Ocean but still persist over the Northwestern Pacific, in agreement with the seasonal migration of the intraseasonal signal. During DJF, the season with the largest intraseasonal signal south of the equator, persistent and extended DWLs are present over the western Indian Ocean and over the NW Australian shelf.

As already mentioned above, these DWLs can be related to suppressed wind conditions preceding the triggering of the convection, and give an enhanced variability of the SST at the intraseasonal time scales. The intraseasonal activity is detected simply on the OLR time series for the equatorial region at 90°E. This OLR time series is filtered between 20 and 90 days using the method described in Bellenger and Duvel (2007). Intraseasonal events with a filtered signal larger than 80Wm^{-2} are considered. The intraseasonal perturbation of the DSA is especially strong in the Indian Ocean with a clear increase of the DSA during the convectively suppressed phase of the ISO. The DSA decreases progressively as the convection develops first to the west (Fig.4b) and then over the entire basin (Fig.4c). The DSA begin to increase over the western Indian Ocean the following phase (Fig.4d). The DSA is a maximum northwest of Australia in a phase corresponding to low surface wind conditions when the convection is maximal over the eastern Indian Ocean (Fig.4c). This strong DSA to the East of the convective perturbation will favour its eastward propagation since a strong DSA is associated to a higher SST that increases the convective instability.

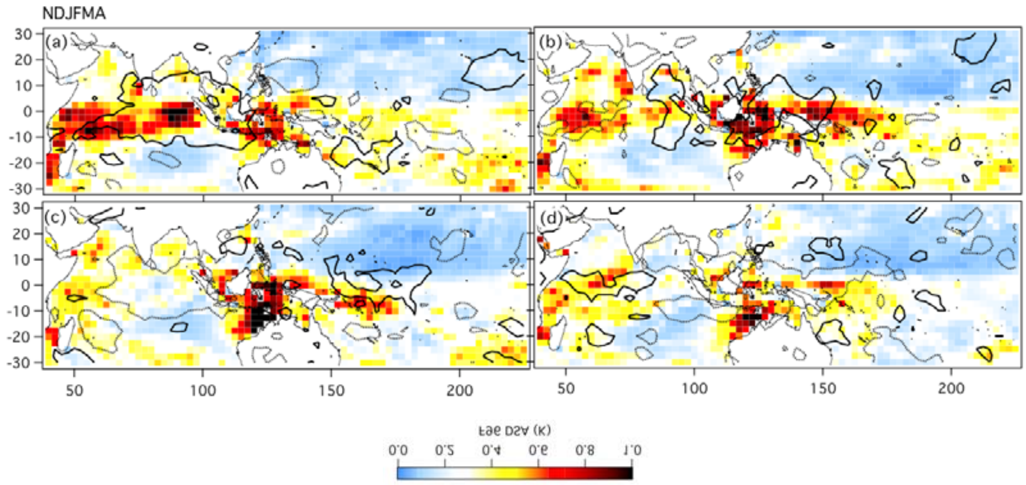

Figure 4: Modulation of the DSA (colors) by the intraseasonal oscillation (ISO) for the November-April season (18 strong ISO events between 1979-2002). The four ISO phases are detected based on the filtered OLR signal around (0°N, 90°E). Panel (a) corresponds to maximum OLR around (0°N, 90°E) and panel (c) to minimum OLR. The corresponding OLR anomaly is also reported (contours, dashed: -10Wm^{-2}; solid: +10Wm^{-2}).

3. DWL in the LMD GCM and convective response

The LMD atmospheric GCM is used to study the impact of these DWL on the intraseasonal variability of precipitation and SST. To this end, a simple test was designed using the zoom capability of the model. A zoom region was defined in the Indian Ocean. The model is forced with Reynolds SST and guided by ERA40 reanalysis with a relaxation time of 48 days (i.e. nearly no effect) in the zoomed region (roughly 50°E-90°E; 5°N-15°S) and with a relaxation time of 2 hours (i.e. nearly ERA40) farther than around 10° outside of this region. Intermediate relaxation times are applied in the transition “ring” of around 10°. Two such guided and zoomed simulations were performed for 3 months in JFM 1999, one with the standard version of the LMD GCM and the other one with a version including cool-skin and the DWL parameterisations based on the Fairall at al. (1996) algorithms.

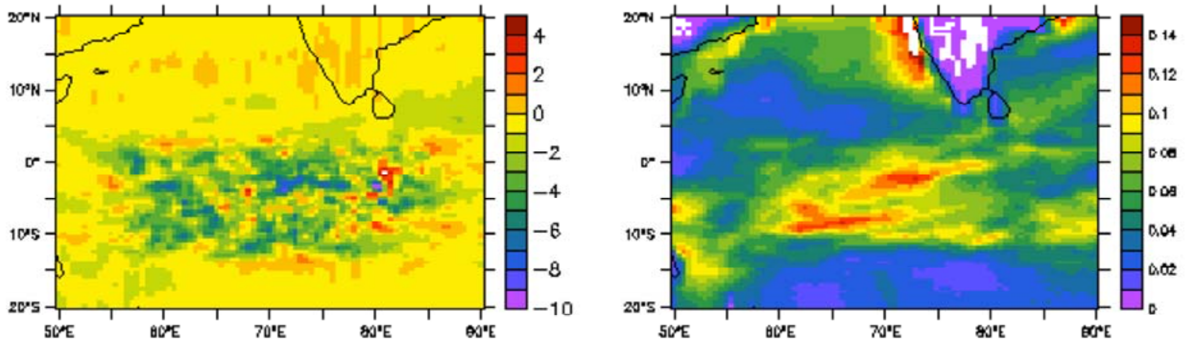

Figure 5: Difference between the control simulation and the simulation with DWL and cool skin parameterisations for the intraseasonal standard deviation of (right) the SST [K] and (left) the OLR [Wm^{-2}].

The intraseasonal standard deviation of the SST is increased for all regions with maximum values south of the equator where relatively strong westerly wind bursts occur during JFM 1999 after periods of relatively calm conditions. For the convection however, the impact is less clear with even a generally weaker intraseasonal standard deviation of the OLR south of the equator. This is due to the diurnal response of the convection during DWL episodes that tends to increase the deep convective activity during suppressed conditions. This decreases the OLR contrast between suppressed and active intraseasonal phases and thus the intraseasonal standard deviation of the OLR. This shows that the parameterisation of the deep convection is too responsive to the SST over tropical oceans, independently of the large-scale dynamical conditions. The increase of the convection as soon as the SST increases (at either the diurnal or the intraseasonal time-scales) is a known problem of GCMs. In a first approach, this could be simply interpreted as a too weak triggering criterion for the deep convection in the model. Therefore, a more stringent triggering criterion for the deep convection could be important to improve the response of the convection to DWL over the tropical ocean, and in particular to generate convective perturbation organized at large scale. Note however that shallow and mid-level convection are associated to the formation of DWL with a maximum in the afternoon, as observed during the MISMO experiment (maximum convection in early morning for perturbed conditions with deep convection). Tentatively, a good convective scheme should thus be able to trigger shallow and mid-level convection during the suppressed phase when DWLs develop, but also to conserve some conditional convective instability at large scale (including a relatively high SST), in order to be able to generate large-scale organised deep convection at the beginning of the perturbed phase.

A more stringent triggering criterion for the deep convection is indeed already known to improve the representation of the intraseasonal variability in GCM. A better representation of the ISV can be obtained by modifying selected parameters of the convective and/or cloud parameterisations (e.g. Wang and Schlesinger 1999; Vitart et al 2007). Generally, criterion leading to a better large-scale organization of the convection improves the simulation of the intraseasonal variability. This can be achieved by modifying for example the triggering criterion for the deep convection or cloud radiative processes that leads to reinforce the low level convergence in the convective area. For example, putting a more stringent triggering criterion that enables the accumulation of moist static energy in the lower troposphere weakens the variability of the convection at short space and time scales in the GCM and increases the large-scale organisation at the intraseasonal time-scales. Also, inhibiting the convection under certain threshold increases the variance in the tropical winds by enabling a greater proportion of deep convective heating to be represented by grid-scale motions (Vitart et al. 2003).

It can thus be useful to test the ability of GCMs to simulate large-scale organized convection at intraseasonal time-scales. It is also useful to test if a model is able to give the observed reproducible intraseasonal perturbations patterns. This is the aim of the next section.

4. Evaluation of the intraseasonal variability in GCMs

The intraseasonal oscillation of the deep convection is an intermittent coupled phenomenon with large variations in amplitude and occurrence and various perturbation patterns depending mostly on season (the position of the ITCZ). During boreal winter, the large-scale perturbation pattern is relatively well reproducible and the average pattern represents the canonical Madden-Julian Oscillation characterised by an eastward propagation from the Indian Ocean to the Central Pacific with maximum convective and wind perturbations south of the equator. On the opposite, during boreal summer, the large-scale Indo-Pacific pattern of the intraseasonal oscillation is less reproducible and the average mode is poorly representative of the intraseasonal events succeeding one another in time (Goulet and Duvel 2000). Considering each basin (Indian Ocean, Indonesia, Western Pacific) separately instead of the whole Indo-Pacific area, the local intraseasonal perturbation pattern for a given season is well reproducible from one event to another (Duvel and Vialard 2007). This suggests that local, basin scale processes are important to understand the physical origin of the intraseasonal oscillation and that air-sea interactions could be a key process for the triggering of an intraseasonal convective event on a particular basin.

Due to the intermittency of the intraseasonal perturbations, the evaluation of its representation in GCM is not straightforward. In particular, as discussed above, average large-scale Indo-Pacific patterns computed from a composite or from principal component analyses may be not representative of actual events succeeding one another in time (Goulet and Duvel 2000). In addition, due to the strong seasonal variation of the ISV patterns, the evaluation should ideally be done month-by-month, and not season-by-season, especially during boreal spring and summer. At the local basin scale, the ISV patterns show for example large differences between May and June over the northern Indian Ocean. At large-scale, the maximum ISV amplitude shift from the Indian in May-June to the Pacific Ocean in July-August.

To take into account the intermittency of the ISV and the strong variability of the intraseasonal perturbation patterns from one event to another, Goulet and Duvel (2000) developed the Local Mode Analysis (LMA) approach. This technique detects large-scale organised perturbations and characterise the perturbation pattern of each event using a simple mathematical formulation. The LMA is based on the computation of the first complex EOF for short time windows (around 3 times the average period of the phenomenon; 120 days for the intraseasonal variability). This window is running along the full time series and only windows corresponding to a “local” maximum of variance percentage of the first complex EOF are retained. This approach gives an ensemble of events that shows the diversity of the perturbation patterns and is thus well adapted to characterise an intermittent phenomenon. The LMA approach was recently used to assess the representation of the ISV in the DEMETER hindcasts. These hindcasts are done using coupled simulations for 7 models from the 1st of May of 22 years (Xavier et al 2008). Two applications of the LMA are used for this evaluation.

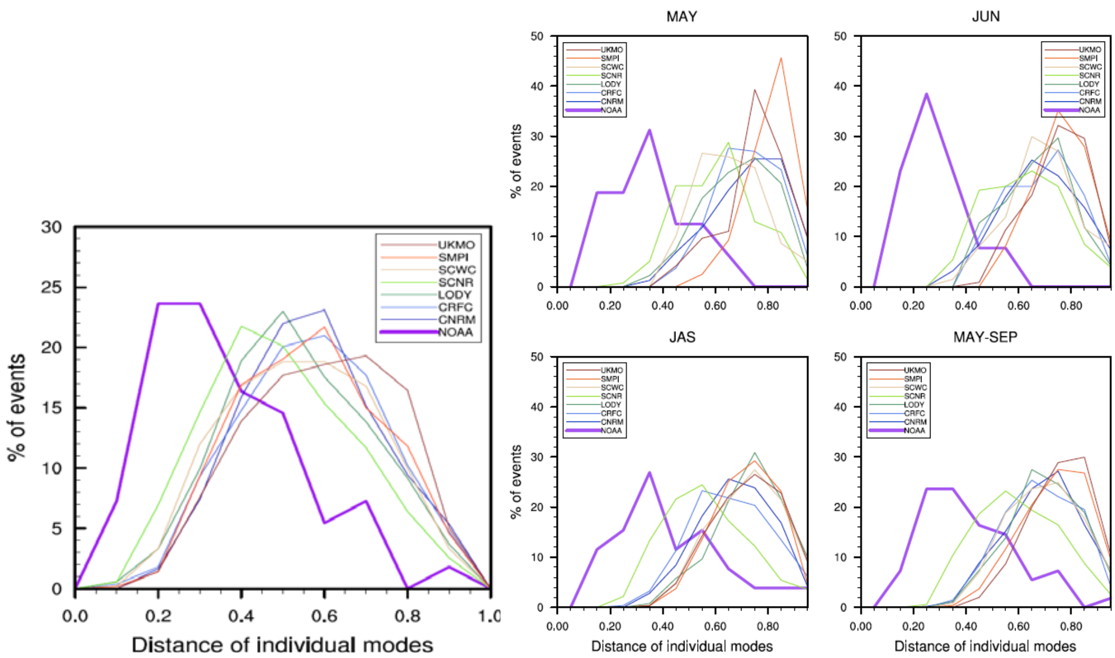

Figure 6: Distribution of the normalised distances between individual local modes and an average mode for the OLR time series. (left) distance between each mode and the corresponding average for the DEMETER hindcasts and for observation for May-September. (right) distance between each OLR mode and the corresponding average OBSERVED pattern for different months. Region [25°N-20°S;40°E-110°E].

First, the LMA gives a metrics of the similarity between two events. This metrics is a normalised distance between two perturbation patterns taking into account both the amplitude of the signal and its relative lag between regions. The distribution of the normalized distances between the pattern of each event and an average pattern give the ability of this average pattern to represent the ensemble of event. The average pattern used for the comparison can be the actual average pattern for each coupled model (Fig.6 left) or the observed average pattern (Fig.6 right). Using this metrics, the observed OLR intraseasonal perturbations patterns are more reproducible than the simulated one (Fig.6). This lack of reproducibility of the perturbation patterns in the coupled GCMs hindcasts tends to show that the simulated intraseasonal perturbations of the convection miss some important process.

Second, the LMA results are used to test the impact of the large-scale organised perturbation on the intraseasonal variability. For a given region, the part of the OLR variance explained by a given Local Mode represents the weight of the large-scale organised perturbation of the convection in regard to local, unorganized convection. The ratio between the regional variance of the mode and the raw intraseasonal variance thus gives a metrics of the impact of the organised convection on the variability of this region. The model generally underestimates this ratio (Fig.7), showing that the large-scale organised perturbations have a too weak impact on the variability at intraseasonal time-scales. In addition, comparing the results for different models, there is a tendency to have at the same time a better reproducibility of the perturbation pattern, more realistic patterns and a larger weight of the large-scale organised perturbation. This tends to show that an important process for a better representation of the ISV in GCM could be the ability to generate large-scale convective perturbations. This is confirmed by an ongoing study using long-term simulations done by 19 coupled GCMs in the IPCC framework.

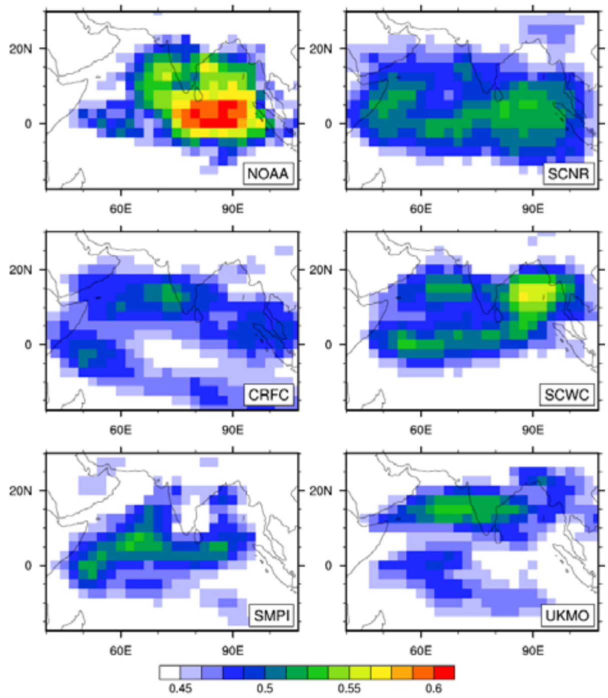

Figure 7: Average ratio between the variance of the modes and the raw intraseasonal variance.

5. Questions

The control of the representation of the ISV in GCMs is still a challenge. This extended abstract presents some results related to the role of air-sea interaction in the triggering of intraseasonal convective events. The MLD is a factor that may control both the amplitude and the time-scale of intraseasonal convective events above the tropical ocean. However, other factor like the formation of DWL could also play an important role. An emphasis of this abstract is put on the potential role of the formation and the dissipation of DWL on the amplitude of the ISV of the SST. In addition, during the convectively suppressed phase of the intraseasonal perturbation, the DWL formation could favour the triggering of large-scale organised convective perurbation. More analyses based on observations and GCM simulations are necessary to validate this hypothesis. Concerning GCM simulations, the present studied shows that a simple parameterisation of the DWL, incorporated in the Atmospheric GCM, is able to estimate the perturbation of the surface fluxes due to the presence of DWL. The two simple DWL models used here have however some bias in comparison to satellite estimates and could be improved. Also, the incorporation of DWL in the atmospheric GCM can give a diurnal response of the deep convection in the suppressed phase that tends to reduce rather than increase the amplitude of the ISV of the convection. The incorporation of a DWL parameterisation gives thus also an opportunity to test the response of the convective schemes to SST variations at different time scales. In particular, it is certainly important to test the ability of the convective scheme to trigger shallow and mid-level convection during suppressed phase, when DWLs develop, but also to conserve some conditional convective instability at large scale (including a relatively high SST) to trigger large-scale organised deep convective events at the intraseasonal time-scale.

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