29. WHY CLIMATE MODELERS SHOULD WORRY ABOUT ATMOSPHERIC AND OCEANIC WEATHER

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The local air-sea feedback diagnostic presented here shows that in many regions of the tropical ocean and the atmosphere primarily drives interannual sea surface temperature variability. This diagnostic is applied to both uncoupled AGCM simulations and coupled simulations. The results support the claim that uncoupled AGCM simulations fail to capture the co-variability between the atmosphere and ocean particularly in warm regions of the Indo-Pacific. This has implications in terms of how well the model is able to reproduce the observed tropical teleconnections. In addition, the diagnostic reveals that the coupled models typically fail to capture the observed local air-sea feedbacks in the western Pacific. Based on simple theoretical calculations the authors argue that: (i) this error leads to ENSO events that extend too far to the west and (ii) that to reduce this error addition stochastic forcing at the air-sea interface needs to be added to the coupled system. This second point is supported by CGCM experiments.

1. Introduction

Changing oceanic conditions, as manifest through sea surface temperature (SST), can influence atmospheric circulation through a variety of processes, largely by changing enthalpy fluxes across the surface. Thus, the SST anomalies play an important role in atmospheric variability and predictability (Charney and Shukla 1981; Shukla 1998; Trenberth *et al.* 1998; Kang *et al.* 2002; Wang and Zhang 2002; Wang *et al.* 2004). However, atmospheric variability exists that is independent of SST forcing. Both forced and internal aspects of atmospheric climate variability impact oceanic conditions, both through local momentum, freshwater and enthalpy fluxes, and through the remote response of oceanic circulation via wave modes. Some of these oceanic changes further impact atmospheric conditions, including those in the monsoon regions (e.g., Vecchi and Harrison 2000; Vecchi *et al.* 2006). Describing, understanding and representing the coupled interactions between these two fluid systems is a major focus of the scientific community, both as a source of predictability of climate conditions around the world, on a variety of time- and space-scales, and as a basic scientific research problem. For example, Hendon (2003) showed that

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seasonally varying air-sea interactions, particularly associated with latent heat flux were critical to interannual Indian Ocean SST anomalies and Indonesian rainfall. The importance of air-sea heat exchanges in the Indo-Pacific region in terms of capturing the monsoon variability was also noted by Krishna Kumar *et al.* (2005), Wu and Kirtman (2007), Wu and Kirtman (2005), Wu *et al.* (2006), Wang *et al.* (2003, 2005), Kucharshki *et al.* (2007), Bracco *et al.* 2007, Vecchi and Harrison 2004, and Lau and Nath (2000, 2003, 2004). Observed and simulated Indian Ocean SST variability and its relationship with the monsoon has also received considerable attention in recent years (e.g., Krishnamurty and Kirtman 2003; Xie *et al.* 2002; Annamalai *et al.* 2003; Behera *et al.* 2000; Huang and Kinter 2002; Song *et al.* 2007; Iizuka *et al.* 2000; Jin and An 1999; Li *et al.* 2002; Murtugudde and Busalacchi 1999; Murtugudde *et al.* 2000; Saji *et al.* 1999; Webster *et al.* 1999). There have also been simulations and observational studies demonstrating the role of Atlantic and Pacific Ocean variability on the Indian Ocean monsoon system (e.g., Kucharski *et al.* 2008; Shinoda *et al.* 2004). A detailed discussion of Indo-Pacific variability and its relationship with the monsoon can be found in the review article by Webster *et al.* (1998).

Simulations of atmospheric general circulation models (AGCMs) forced by prescribed SSTs (either observed or idealized in order to isolate particular mechanisms) allow us to assess aspects of the SST control on climate variability, and exploit this atmospheric response to SST anomalies for predictive purposes and to increase our understanding of the climate system. However, AGCM experiments forced by observed SST show both consistencies with and discrepancies from observations (e.g., Sperber and Palmer 1996; Kumar and Hoerling 1998; Kang *et al.* 2002; Wang *et al.* 2004). In addition to fundamental predictability limitations arising directly from internal atmospheric dynamics, two major reasons for the model-observation discrepancies are: (1) the biases in the model physics and (2) the lack of air–sea coupling in the forced simulations. The discrepancies due to (1) are model dependent and can be reduced with the improvement in the representation of atmospheric physical processes in the model. The discrepancies due to (2) are fundamental and common to all of the forced simulations, and arise because some SST features are actually the result of atmospheric conditions that arise due to either remote SST forcing or internal atmospheric dynamics independent of SST changes: the SST used as a forcing in some conditions is actually a response. In turn, using these SST anomalies that are forced by the observed atmosphere as prescribed SST in uncoupled atmospheric model simulation can result in improper local air–sea relationships in some regions resulting in unrealistic atmospheric variability (Saravanan 1998; Saravanan and McWilliams 1998; Bretherton and Battisti 2000; Wang *et al.* 2004, 2005; Krishna Kumar *et al.* 2005; Trenberth and Shea 2005; Wu *et al.* 2006). Some discrepancies due to the lack of air–sea coupling have been demonstrated in previous studies (Roebber *et al.* 1997; Barsugli and Battisti 1998; Wittenberg and Anderson 1998; Wu and Kirtman 2005).

To illustrate an aspect of discrepancy (2), Fig. 1 shows that even the interannual correlation of seasonal SST and evaporation anomalies can differ in various regions around the tropics, when comparing SST-forced and coupled climate integrations using the same atmospheric component. There are substantial regions of negative correlation in the coupled climate model, indicating regions where enhanced evaporation is associated with cool conditions, while the correlation tends to be more positive in the AGCM, as warm conditions tend to favor evaporation in an SST-forced framework. The reader is also referred to Fig. 3 (top left panel) where we show the same correlation based on observational estimates. The observational estimates indicate significant regions of negative correlation that are either complete absent on the AGCM forced simulation or are weaker than observed in the CGCM simulations.

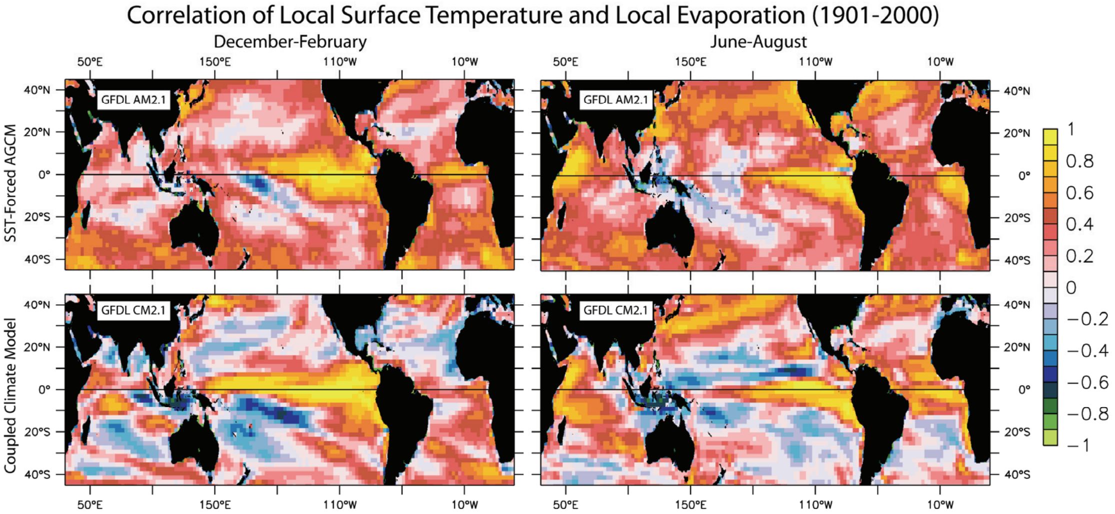


Figure 1. Correlation of seasonal-mean SST anomalies and evaporation anomalies from SST-forced AGCM (upper panels) and coupled climate model (lower panels) integrations over a 100-year period. The two model systems share the same atmospheric component. The atmospheric component is the finite volume version of the GFDL atmospheric model (AM2.1; GAMDT 2004; Lin *et al.* 2006), and the coupled model is a version of the GFDL coupled climate model (CM2.1; Delworth *et al.* 2006; Gnanadesikan *et al.* 2006; Stouffer *et al.* 2006; Wittenberg *et al.* 2006; Song *et al.* 2007).

When and where the discrepancies due to the lack of air–sea coupling occur depends on what causes the SST anomalies. In the case that the local SST anomalies are primarily due to internal oceanic processes, it is likely that the forced simulations can capture the observed atmospheric variability. This is the case in the tropical central and eastern Pacific where the observed SST anomalies are mainly due to oceanic processes with surface heat fluxes mainly acting as a damping effect (e.g., Jin and An 1999; Kang *et al.* 2001) and SST forced simulations perform well (e.g., Kumar and Hoerling 1998; Kang *et al.* 2002; Wang *et al.* 2004). In the case that the observed SST anomalies are largely due to atmospheric forcing, erroneous atmospheric response can result in the specified SST simulations. This occurs in the extratropics and the tropical Indo–western Pacific Ocean regions where the atmospheric forcing plays an important role in inducing SST anomalies (e.g., Lau and Nath 1996; Alexander *et al.* 2002; Lau and Nath 2000, 2003; Wang *et al.* 2003; Krishnamurthy and Kirtman 2003). In these regions, the forced simulations deviate from observations (e.g., Sperber and Palmer 1996; Wang *et al.* 2004; Wu *et al.* 2006) and coupled model simulations (e.g., Kitoh and Arakawa 1999; Wu and Kirtman 2005; Wu *et al.* 2006). We focus on the western tropical Pacific in more detail below and show how the air-sea feedbacks in the western Pacific ultimately impact the remote ENSO variability in coupled models.

2. Diagnosing Air-Sea Feedbacks

The nature of local air–sea interaction can be understood from the evolution of lag–lead correlation between the atmospheric variables and SST (von Storch 2000; Wu *et al.* 2006). Using a simple stochastic model, Barsugli and Battisti (1998) identified distinct lagged linear regression between sea and air temperature for coupled and uncoupled cases. von Storch (2000) provided a conceptual interpretation of how the different shapes of lag cross-correlations relate to different forcing-response relationships. The author identified very different evolution of the lag correlation between surface heat flux and SST in the mid-latitude North Pacific and the equatorial central Pacific. Wu and Kirtman (2005) demonstrated that the local lag–lead correlation between SST, rainfall, and surface evaporation can indicate an atmospheric negative feedback in the coupled model. The analysis of lag–lead correlations has been used to understand the atmosphere–ocean relationship in observations and models (Frankignoul *et al.* 1998; von Storch 2000; Frankignoul and Kestenare 2002; Frankignoul *et al.* 2002; Frankignoul *et al.* 2004; Kitoh and Arakawa 1999; Wang *et al.* 2005; Wu and Kirtman 2005; Wu *et al.* 2006). However, because atmosphere-ocean interactions are seasonally dependent (e.g., Hendon 2003; Wang *et al.* 2003), it is not so simple to analyze the lag–lead correlations.

Another way to reveal the air–sea relationship is to combine simultaneous atmosphere–SST and atmosphere–SST tendency correlations (Wu *et al.* 2006). SST anomalies can induce anomalous convection through surface evaporation and low-level moisture convergence. Because the atmospheric response to SST forcing is relatively fast, a large positive simultaneous correlation, for example, between rainfall and SST may indicate that the SST is forcing the atmosphere. On the other hand, anomalous atmospheric convection can change the SST through cloud-radiation and wind-evaporation effects and wind-induced oceanic mixing and upwelling. These atmospheric feedbacks can be detected in the SST tendency. Thus, the magnitude of simultaneous rainfall–SST and rainfall–SST tendency correlations can indicate the relative importance of SST forcing and atmospheric forcing. Wu and Kirtman (2005) showed that in regions where the atmosphere has a strong negative feedback on SST (e.g., tropical western North Pacific in boreal summer and tropical southwestern Indian Ocean in austral summer), the negative rainfall–SST tendency correlation is larger than the rainfall–SST correlation. This differs from the equatorial central-eastern Pacific where the positive rainfall–SST correlation is much larger than the rainfall–SST tendency correlation. Using simple model simulations, Wu *et al.* (2006) demonstrated that the surface turbulent heat flux–SST/SST tendency correlation displays marked differences for the case when atmospheric forcing dominates versus when SST forcing dominates. An analysis of heat flux–SST tendency correlation has been performed to identify the atmospheric forcing of SST in the North Pacific (e.g., Cayan 1992) and in the tropical Indo-western Pacific Ocean regions (e.g., Wu and Kirtman 2005; Wu *et al.* 2006).

The importance of accurately capturing the western Pacific air–sea feedbacks correctly (or the implications of failing to capture these feedbacks) is exemplified in Fig. 2, which shows an example from several models participating in the Intergovernmental Panel on Climate Change (IPCC) Assessment Report Four (AR4). In this figure we have plotted the spatial pattern of the first empirical orthogonal function (EOF1) of the SST anomaly in the equator in the Pacific from five different state-of-the-art coupled models and observational estimates. All of the coupled models shown here have dominant ENSO modes that extend too far to the west. Often, but not always, the models have ENSO periodicities that are too fast compared to the observations. The conventional wisdom is that the westward extension of the ENSO events and the fast periodicity is due to the cold tongue mean state errors. Simply, errors in the mean state are the cause for the errors in the anomalies. Here we suggest that the errors in the simulated ENSO are due to errors in the statistics of the tropical atmospheric weather and the associated air–sea feedback in the western Pacific. In other words, if there are large errors in the simulation of the weather statistics in the western Pacific and the associated air–sea feedbacks, then the climate simulation is seriously degraded.

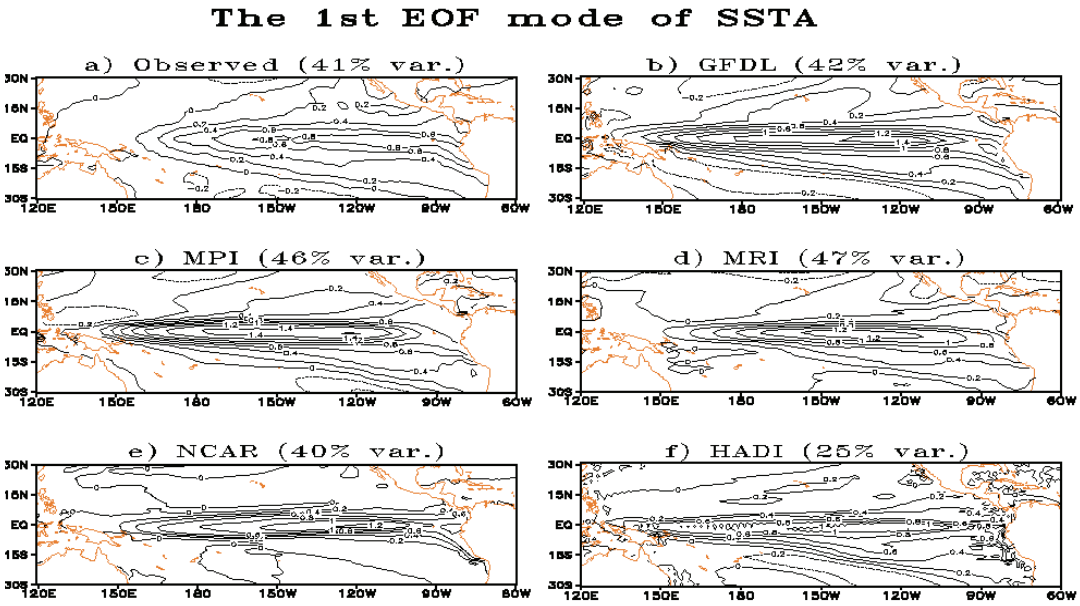
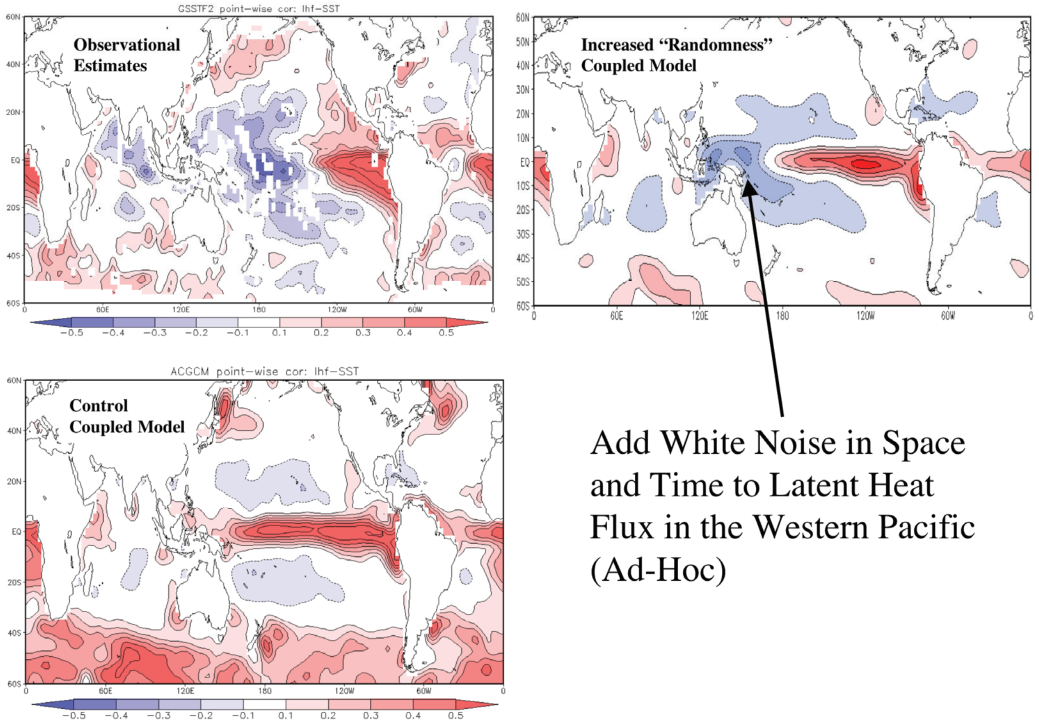


Figure 2. SSTA EOFs calculated from various coupled model intercomparison project simulations (CMIP3). The domain plotted corresponds to the domain of the EOF calculation. In each figure 100-years of data was used from simulation with fixed climate forcing at 1990 levels.

3. Linking Theory with Simulation

The theoretical coupled model presented in Wu *et al.* (2006) suggests that the source of the western Pacific problem is due to incorrect latent heat flux – SST feedbacks, and the theory suggest a potential solution. Wu *et al.* (2006) show that when the correlation between the latent heat flux (our convention here is that latent heat flux is positive upward) and SST anomalies is strongly negative, the SST variability can be viewed as strongly forced by atmospheric variability (noise). Conversely, when the ocean forcing dominates the correlation is positive. Figure 3 (in part taken from Wu *et al.* 2006) shows this correlation from satellite based observational estimates (top left) and the Center of Ocean-Land-Atmosphere Studies (COLA) anomaly coupled model (bottom left; Kirtman *et al.* 2002). Clearly, near the equator in the western Pacific the coupled model fails to capture the observed relationship. This is also true in significant regions of the tropical Indian and Atlantic Oceans. Similar errors have been identified with Climate Forecast System (CFS) (e.g., Wu *et al.* 2007) and with Community Climate System Model 3.0 (CCSM3) (not shown). The theoretical model suggests two possible interpretations of this result: (a) the ocean is too strongly forcing the atmosphere or (b) the atmosphere is not forcing the ocean enough. Wu *et al.* (2006) describes the theoretical basis for these possible interpretations.

Contemporaneous Latent Heat Flux - SST Correlation



Add White Noise in Space and Time to Latent Heat Flux in the Western Pacific (Ad-Hoc)

Figure 3. Contemporaneous correlation between latent heat flux anomalies (positive upward) and SST anomalies based on observational (top left) estimates from version 2 of the Goddard Satellite-Based Turbulence Fluxes (GFSST2) data, the COLA coupled model simulation (bottom left) and the COLA model forced with Gaussian white noise in the latent heat flux in the western Pacific (top right).

In the case of the atmosphere forcing the ocean, the theoretical model of Wu *et al.* (2006) adopted from Barsugli and Battisti (1998) is as follows:

$$\frac{dT_a}{dt} = \alpha(T_o - T_a) + N_a \qquad (1)$$

$$\frac{dT_o}{dt} = \beta(T_a - T_o) - \gamma_o T_o \qquad (2)$$

In the above, T_a and T_o refer to air and sea temperature, respectively. Air–sea heat flux (latent and sensible) is represented by the air–sea temperature difference, N_a represents atmospheric white noise forcing and α and β are exchange coefficients. This theoretical model implies a negative contemporaneous correlation between the atmosphere and the ocean. In contrast, Wu *et al.* (2006) also present a simple theoretical model for the ocean forcing the atmosphere, i.e.,

$$\frac{dT_a}{dt} = \alpha(T_o - T_a) \qquad (3)$$

$$\frac{dT_o}{dt} = \beta(T_a - T_o) - \gamma_o T_o + N_o \qquad (4)$$

In this case, N_o represent oceanic forcing of the atmosphere and the air–sea correlation in positive.

The theoretical model described above also suggests a possible solution to this air–sea feedback problem, namely we need to change the relative strength of the atmosphere forcing of the ocean or the ocean forcing of the atmosphere. In other words, we can simply modify N_a or N_o to change the air-sea correlation. We present here an ad-hoc preliminary attempt at modifying the relative forcing strength. Here we modify the effective N_a in the CGCM by simply add Gaussian white noise (in both space and time) to the latent heat flux that is used to force the ocean. In this test, the noise amplitude is arbitrarily chosen to be 15% of the day-to-day variance produced by a control run of the model and is only applied in the far western Pacific (5°N-5°S, 120°E-160°E). This Gaussian white noise forcing was applied to a 100-yr simulation of the COLA anomaly coupled model. The resulting correlation is also shown in Fig. 3 (top right). As predicted by the theoretical model, the correlation has changed sign in the western Pacific. We emphasize that this is more than simply reducing the amplitude of the correlation – it has actually changed sign. The entire ENSO system in this simulation has shifted further to the east with a consistent increase in the periodicity. This suggests the air–sea physics in the western Pacific can have a profound impact on the ENSO simulation. This impact is more than merely making the ENSO more irregular; it is shifting the system eastward modifying the oceanic time scales (via wave dynamics) and even modifying the global teleconnections by shifting the region of maximum rainfall anomalies to the east. The changes in the periodicity and the eastward shift of the variability can easily be detected in Fig. 4, which shows the lag-lead regression of Nino3.4 SSTA onto equatorial Pacific SSTA. In essence, adding noise in the western Pacific heat flux has modified the coupled signal without explicit changes to either the atmospheric or oceanic component model.

Another possible solution to the problem is to restrict the uncoupled SST forcing of the atmospheric model to a region where SST can be considered a local forcing (e.g., the eastern and central equatorial Pacific Ocean), and allowing the atmospheric model to couple to a thermodynamically or dynamically active oceanic model elsewhere. Model configurations of this type have been used to explore the response of the monsoon and midlatitude climate systems to forcing from various tropical basins (e.g., Alexander *et al.* 2002; Lau and Nath 1996, 2000, 2003, 2004; Bracco *et al.* 2007), and to explore the impact of decadal oceanic variations in the Atlantic on global climate conditions (e.g., Zhang *et al.* 2007).

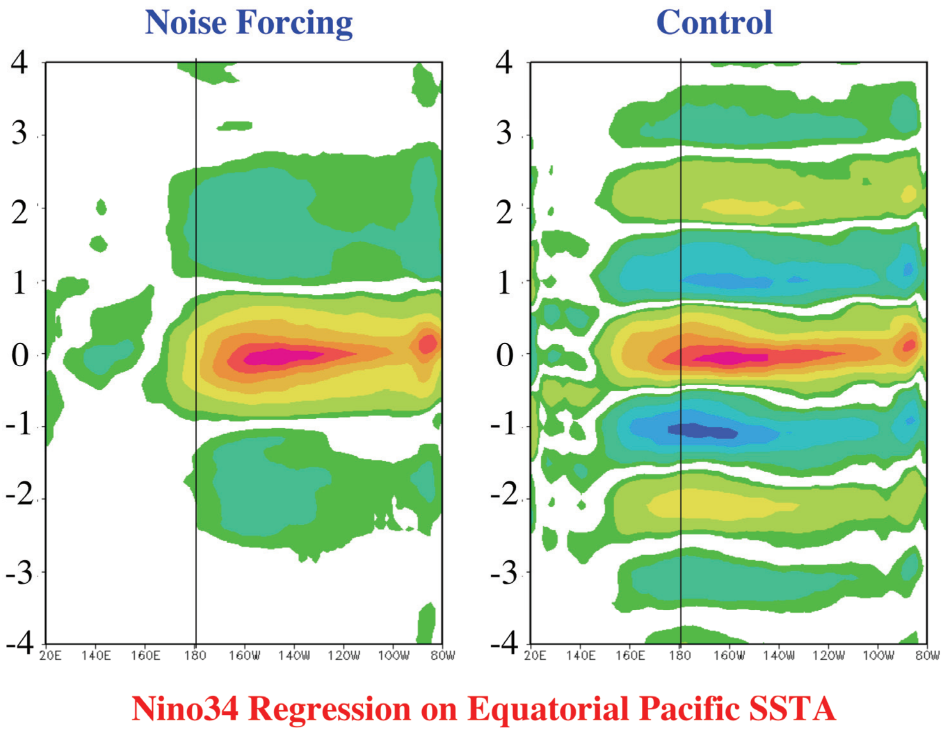


Figure 4. Lag-lead regression between Nino3.4 SSTA and equatorial Pacific SSTA. The left panel corresponds to the noise forcing experiment and the control is shown in the right panel. The contour interval is the same for both panels and starts at ±0.2. The lags and leads noted on the left of each panel are in years.

While the previous discussion has largely focused on large-scale errors arising from inadequate coupling – largely arising due to internal atmospheric variability, it is interesting to briefly consider the possibility of analogous issues on the oceanic mesoscale, arising from internal oceanic variability. There is now considerable evidence that the sharp SST gradients induced by oceanic mesoscale features (e.g., upwelling filaments, eddies, tropical instability waves, sharp fronts, warm western boundary currents, etc.) can drive changes in the atmosphere, through local air-sea interaction (e.g., Chelton *et al.* 2001, 2004, 2005, 2007; Hashizume *et al.* 2001; Xie 2004; Vecchi *et al.* 2004; Seo *et al.* 2007; Minobe *et al.* 2008; Small *et al.* 2008). These atmospheric changes on the oceanic mesoscale result in variations to the enthalpy and momentum fluxes of sufficient magnitude to impact the oceanic structures that drove them (e.g., Chelton *et al.* 2005; Vecchi *et al.* 2004; Seo *et al.* 2007). Thus, in order to correctly represent the physical processes behind these oceanic mesoscale features, one may be required to correctly represent the impact of this air–sea coupling. However, high resolution ocean models are generally forced by winds from either global analysis products (like ECMWF and NCEP) or by winds derived from satellite scatterometry (such as NSCAT or QuickSCAT). Wind from the global analyses do not include features on the oceanic mesoscale, so the effects of this coupling will be absent from a forced experiment, while scatterometer winds include the impacts of coupling that correspond to the internal oceanic structures present in the real world, which need not correspond to those in the model. Thus, as eddy-permitting and eddy-resolving models continue to be developed and implemented in climate-scale integrations, solutions to, perhaps analogous to those discussed above, the problem of inadequately representing air–sea interactions on the oceanic mesoscale must be explored and developed.

4. Final Remarks

Dynamical numerical modeling systems are an essential tool in describing, understanding, representing and predicting the atmospheric and oceanic conditions of the global climate system, including those in the monsoonal regions of the world. Boundary forced configurations of these models can represent many aspects of the variations of the ocean and atmosphere climate system, but discrepancies can arise from incorrectly specifying the boundary values as a forcing, when they are actually largely a response to variations in the system one is modeling. Solutions to this problem have been and should continue to be developed, and modelers should be keenly aware of these potential problems.

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