**Climate impacts of** **stochastic atmospheric perturbations on the ocean**

Short Title: **Impacts of atmospheric noise on the ocean**

Jie ZHANG1, Wei XUE2, Minghua ZHANG3, Huimin LI2, Tao ZHANG2, Lijuan LI2,4, Xiaoge XIN1

1 National Climate Center, China Meteorological Administration, Beijing, China

2 Center of Earth System Science, Tsinghua University, Beijing, China

3 Institute for Terrestrial and Planetary Atmospheres, [School of Marine and Atmospheric Sciences](http://www.msrc.sunysb.edu/), [Stony Brook University](http://www.sunysb.edu/)

3 LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

(1st round)

Submitted to ***International Journal of Climatology***

**(29 Jul. , 2013)**

**Corresponding author:**

Xue WEI

Center of Earth System Science, Tsinghua University,

Beijing, China

E-mail: xuewei@tsinghua.edu.cn

Phone: 86-10-62783505-7

Fax : 86-10-62771138

**Abstract**

An Interactive Ensemble (IE) Platform was established based on a standard coupled climate model (SC) with seven atmosphere-land model realizations coupled to a single ocean model and a single sea-ice model. The IE strategy reduces stochastic noise generated by atmospheric dynamics and therefore can be used to estimate the impact of atmospheric perturbations on the ocean. The atmospheric noise reduction shallows the mixed layer, modifies the oceanic upwelling/downwelling, and cools the surface. It is assumed that the deterministic part of the surface cooling is due to the reduced heat capacity of the mixed layer in high latitudes and the changed surface wind curl that causes upwelling/downwelling in the middle latitudes and tropics. The positive albedo feedback is responsible for the polar amplified changes. The weakening of MOC further suppresses ocean heat transport to the Northern Hemisphere. The model results also suggest that the irregular ENSO cycle partly arises from the stochastic perturbation in the atmosphere. The annual cycle of SST in the tropical Pacific is more reasonably simulated in the IE platform.

**Keywords:** Atmospheric perturbations, Interactive Ensemble, Sea surface temperature, the oceanic meridional overturning circulation, the ENSO variability

1. **Introduction**

The atmosphere-ocean interaction at the air-sea interface is one of the important processes that affect both atmosphere and ocean. By exchanging air-sea fluxes of heat, freshwater, and momentum at ocean surface, the atmosphere can control sea surface temperature (SST), surface mixing and drive ocean currents. Despite their high frequency and small spatial scale, the atmospheric perturbations (the high-frequency weather fluctuations) can drive low-frequency changes in the ocean. By explicitly adding water/heat components of the buoyancy flux, noise-induced mean climate drift and noise-enhanced variability were found in Williams (2012). The decadal variability of the tripole mode of North Atlantic SST is forced primarily by the local weather noise surface heat flux (e.g., Fan and Schneider, 2011; Seager *et al.*, 2000). Using a so-called “interactive coupled ensemble” technique, which allows for reductions of the stochastic part of the atmospheric circulation at the air-sea interface, Wu *et al.* (2004) demonstrated that low frequency SST variability in the subtropical North Atlantic was mainly induced by stable coupled feedbacks in which the weather noise played a central role.

The “interactive coupled ensemble” was introduced by Kirtman and Shukla (2002) and used to study the influence of stochastic atmospheric perturbations. It is a new strategy in which the ensemble averaging of fluxes produced by multiple identical atmospheric members with different initial states continuously interact with the single realizations of the other component models. The “interactive coupled ensemble” diminishes the impact of stochastic atmospheric noise and highlights the signal generated by internal dynamics of the climate system. Previous studies found that the “interactive coupled ensemble” generated colder sea surface temperature (SST) in sub-tropics and tropics (Kirtman *et al.*, 2011), improved the development of ENSO events and the associated global impact (Kirtman and Shukla, 2002), and suggested the importance of coupled ocean-atmosphere feedbacks to SST especially in the tropics (Kirtman *et al.*, 2009).

Studying the roles of atmospheric noise at air-sea interface can help to understand the chaotic climate variability, enable new applications of climate theory in air-sea coupling processes. Following Kirtman’s work, based on a Standard Coupled (SC) climate model, an Interactive Ensemble (IE) platform is newly established at the Center of Earth System Science, Tsinghua University. Different from the strategy in Kirtman and Shukla (2002), each atmospheric member in the IE platform is firstly coupled to a land surface model and then coupled to a single ocean model and a single sea-ice model with the ensemble fluxes. That is, the IE strategy here is active at the air-sea and air-ice interfaces but not at the air-land interface. In Kirtman *et al.* (2011), they also used the same IE strategy over land in which only one land model simulation is coupled to an esemble of atmospheric simulations. They found that surface temperature changes are largest over land and ice. In our IE platform, all the significant surface temperature changes are in the ocean grids. Using the IE platform and the SC model, this paper is intended to estimate the influence of stochastic noise generated by atmospheric dynamics on the ocean, including mixed layer depth, sea surface heat budget, , wind curls, the oceanic meridional overturning circulation (MOC), and the ENSO variability. We present results from the SC model and the IE platform. The differences can illustrate the impact of atmospheric stochastic forcings. The remainder of the paper is organized as follows. The SC model, the IE platform and the experiments designs are introduced in section 2. Section 3 examines the responses of surface temperature and precipitation to surface noise reduction. The changes in the ocean are examined in section 4. Major findings are concluded in section 5.

1. **The SC model and the IE platform**

**2.1 The SC model**

The SC model consists of four components of the earth system, connected by coupler CPL6 (Craig *et al.*, 2005): the atmospheric component GAMIL 2.0 with a hybrid horizontal grid, Gaussian grid of 2.8º between 65.58ºS and 65.58ºN and weighted equal-area grid poleward of 65.58º, and 26 levels in a sigma vertical coordinate system with the model top at 2.194 hPa (Li *et al.*, 2013; Li *et al.*, 2007); the oceanic component POP1.4.3 with a horizontal resolution of roughly 1º lon \* 0.5º lat, refined meridional spacing in the equatorial region, and 40 unevenly spaced cells in the vertical (Smith and Gent, 2002); the land surface component CLM3 (Dickinson *et al.*, 2006); and the sea-ice component CSIM5 (Briegleb *et al.*, 2004).

An 850-year pre-industrial equilibrium run with the SC model was conducted. As shown in Figure 1, after some initial adjustment, the simulated climate experiences minimal drift. Over the last 450 years, the globally averaged SST is stable with only a very small globally averaged trend of -0.0186oC per century. The SST change over a century is less than the standard deviations of annual mean SST (0.055oC), implying that the small drift may not affect the internal variability (Lin *et al.*, 2013). The global sea-ice concentration (SIC) also shows very small trend of 0.036% per century. The maximum value of the oceanic stream function on the depth-latitude plane in the Atlantic, i.e. the Atlantic Meridional Overturning Circulation (AMOC) index, reached a quasi-equilibrium state with the value about 16 Sverdrup (Sv, 1Sv = 106m3s-1) after the spin-up integration (Figure 1c). The strength of AMOC in the SC model is slightly weaker than the observational estimate at 26.5oN (about 18.5 ± 4.9 Sv) (Hirschi *et al.*, 2003).

The mean climate in the SC model from year 421 to 450 is assessed in simulating the SST and precipitation (Figure 2). We also examined the SC results from year 541 to 570 and the spatial structures were almost the same. Evaluated against contemporary observations (Figure 2a and 2b), the SC fairly simulated the large scale patterns such as the western Pacific warm pool (Figure 2c), the wet/dry areas over the western/eastern subtropical ocean basins, and a local minimum in precipitation in the tropical North Africa (Figure 2d). However, the meridional extension of warm pool (Figure 2c) is too confined in the tropics and the cold bias south of 10oS make its orientation and the South Pacific Convergence Zone (Figure 2d) too zonally distributed. The warm bias in the eastern tropical Pacific favors the development of so-called “double ITCZ”. Zhang *et al.* (2007) suggested the important roles of ocean dynamics in the development of the “double ITCZ”. In a case study of the CCSM, Liu *et al.* (2011) demonstrated that the initial warm bias is caused by excessive surface shortwave radiation and further amplified by biases in both surface latent heat flux and horizontal heat transport in the upper ocean. The positive feedback between the oceanic advection with SST is presumed to be applicable for models other than the CCSM.

Surface temperature largely depends on the net radiative flux at the top of the atmosphere (TOA). The annual mean local downward radiative flux is largest in the tropics and decreases poleward (Figure 3a and 3c). The SC model underestimates the net incoming radiation in tropical area except for the eastern Pacific, and overestimates it at around 30oN and 30oS. At the global scale, the net shortwave radiation is nearly compensated by the outgoing longwave radiation at the TOA with the net radiation of 4.87 W m-2 in ERBE and 3.41 W m-2 in the SC model. Clouds are responsible for about half the outgoing shortwave radiation. Errors in cloud radiative forcing will strongly modify local radiation balance. As shown in Figure 3e and 3f, the biases of radiation balance bear a great resemblance to the cloud radiative forcing biases. Despite the general biases, general performances of the SC model are comparable to those of other current-generation coupled models.

**2.2 The IE platform**

The IE platform was established at the Center of Earth System Science, Tsinghua University based on the SC model. Because of the high computational cost, the interactive-ensemble in this study is performed with seven sets of the atmosphere-land surface components. Yeh and Kirtman (2004) demonstrated that the amplitude of internal atmospheric variability at air-sea interface decreases proportionally to the increasing of number of AGCM realizations. Seven sets of initial conditions selected from the SC equilibrium run 5 years apart from year 420 to 450 are used to initialize the atmosphere and land-surface models in the IE platform. The initial states for the single ocean model and the sea ice model are from the same initial set as in the SC model. The single ocean model and the sea ice model are coupled with the ensemble mean flux of the multiple atmospheric members at every coupling step, one day for the ocean model and 60 minutes for the sea ice model.

The IE platform was integrated for 128 years from year 450 to 577. As shown in Figure 4a, the surface temperature dropped rapidly by -0.31°C at the end of the first year, slightly rebounded to approximately -0.18°C in the following 50 years, and then dropped back to a quasi-equilibrium state. Surface temperature in the IE platform is about 0.35°C colder in comparison with the SC model. After the spin-up integration, the global averaged surface temperature is about 13.3°C. In this study, climate states averaged from year 421 to 450 in the SC model and from year 541 to 570 in the IE platform are used to examine the changes.

**3. Surface temperature and precipitation changes**

For surface temperature (Figure 4b), the IE platform produces a moderate surface cooling in tropics and subtropics in relative to the SC model. Significant cooling is evident at the polar and sub-polar regions in the Northern and Southern Hemispheres, the eastern tropical Pacific, the tropical Atlantic and some spots in the North Indian Ocean. The cooling changes at high latitudes in the IE platform share the same centers with the standard deviation of surface temperature of the SC ensemble runs (Figure not shown), which arises largely from the model’s internal variability. Atmosphere strongly forces the ocean at high latitudes. Therefore, the place where the atmospheric stochastic noise is more active is also the region with stronger surface temperature variability. Warming anomalies are also visible, significant in the sub-polar South Atlantic and South Indian Ocean. Surface heat budget and ocean circulation may play roles and will be discussed in section 4. The results indicate that the large-scale atmospheric turbulence at the air-sea interface can affect the coupled model performances in simulating surface temperature, especially at middle and high latitudes.

For precipitation (Figure 4c), although the ‘double ITCZ’ problem is still evident in the IE platform, the too eastward extended rain belts are somehow suppressed. Surface cooling, which results in a reduction in the upward fluxes of heat and moisture from the ocean, leads to slightly decrease in global mean precipitation by about 0.015 mm/day. In accompany with the significant cooling, there is a stronger atmospheric subsidence in the eastern tropical Pacific and the North Indian Ocean. Atmospheric convection is enhanced in the tropical central and western Pacific, creating a local excessive rainfall. Besides the local thermal effect, precipitation changes may also be attributed to remote teleconnection. The sea level pressure changes between the IE platform and the SC model characterize a significant positive AAO-like changes, negative anomalies centered in the Antarctic and positive anomalies centers about 40 ~ 50°N (Figure not shown). Gao *et al.* (2003) suggested that strong Antarctic oscillation (AAO) is accompanied with increased East Asian summer monsoon rainfall. The AAO anomalies can change the position and intensity of the Mascarene High and the Australian High in May, the western Pacific subtropical high and the South Asian high in July, which are believed to be crucial to summer rainfall in eastern China.

As shown in Figure 4a, the cooling is slightly smaller in the air than at the surface. The significant changes are all located in the ocean grids (Figure 4b). That is, the impacts of interactive ensemble are more pronounced in the ocean. We therefore focus on oceanic changes, including the sea surface heat budget, the changes in oceanic meridional overturning circulation, as well as the ENSO variability hereafter.

**4 Changes in the ocean**

The stochastic surface stress perturbation at the air-sea interface can enhance oceanic mixing, intensifying downwelling momentum and heat transport. In comparison with the mean climate in SC (Figure 5a), pronounced surface stress weakening in IE (Figure 5b) appears at the peripheries of the storm tracks, especially on the poleward side, where the atmospheric stochastic variability is large. The surface stress changes at high latitudes show good coincident with surface temperature anomalies, decreasing in sub-polar North Atlantic, North and South Pacific, increasing in the sub-polar South Atlantic and South Indian Ocean. The spatial correlation coefficient between the surface stress and surface temperature anomalies in the ocean grids is 0.25. The cancellations among the ensemble members are responsible for the surface stress weakening. The surface wind stress weakening may be partly explained, in addition to the ensemble averaging, by the positive feedback of surface stress to surface temperature changes. Convective instability over warmer water will lead to a deeper boundary layer in the atmosphere and result in intensified surface stress over the warmer water, and vise versa (Samelson *et al.*, 2006).

The mix-layer depth (MLD) of the ocean is greatly influenced by surface stress at air-sea interface and can to a certain extent reflect the strength of air-sea coupling. The MLD is defined here as the depth at which the temperature difference with the top level exceeds 0.5°C. Changes in the MLD were shown in Figure 5c. The spatial correlation coefficient between the surface stress and MLD anomalies in the ocean grids is 0.24. The power of oceanic mixing in the IE platform decreases significantly in sub-polar regions around 60°S and 60°N. The mixed layer depth is more than 10 meters thinner than that in SC in association with the greatly suppressed surface stress (Figure 5b). As a direct response to the decreased surface stress, the ocean surface should get warm due to weakened entrainment of upper-thermocline cool waters into the mixed layer. It is somewhat surprising that our results show the opposite (Figure 4b). Therefore, in order to gain some insight into the heat budget in the upper ocean, the sea surface heat budget and the roles of ocean currents are examined.

**4.1 The sea surface heat flux budget**

The sea surface heat budget is examined in terms of the net radiation at sea surface (shortwave radiation and longwave radiation), surface latent and sensible heat fluxes (Figure 6). In the IE platform, the shallower mixed layer corresponds to smaller heat capacity. As a result, in high latitudes where the there is net surface cooling, the mixed-layer becomes colder than in SC. the net surface radiation income is also decreased by about -0.18W m-2, likely due to surface albedo feedback. Evolution of the net surface radiation is highly correlated with the surface temperature with correlation coefficient of 0.52, significant at the 1% level using student’s ***t*** test. The recovering of net downward surface radiation around year 480 to 510 is associated with the rebounding surface temperature during the adjustment period. As climate response to the surface cooling, the upward surface latent heat flux (Figure 6b) is reduced and is significantly correlated with the downward net surface radiation with correlation coefficient of 0.55.. The increased upward surface sensible heat flux of 0.17 W m-2 (Figure 6c) is likely a response to the above changes that contributes to the cooling.

The changes of the global distribution of the climatological annual mean surface radiation between the IE platform and the SC model are shown in Figure 7. Positive values represent anomalous radiation gain by the ocean. The shortwave radiation contributes to the surface temperature changes in the sub-polar North/South Pacific, the North Atlantic and the warming in the South Indian Ocean. The surface cooling suppresses the upward released longwave radiation, which is responsible for the increase in net downward surface radiation in the Arctic and South Pacific (Figure 7c).

Surface temperature changes (Figure 4b) exhibited a polar amplification. The polar amplification is recognized as an inherent characteristic of the Earth’s climate system, with multiple intertwined causes (Serreze and Barry, 2011). Ice-Albedo feedback may be a key driver. A decrease in atmospheric heat flux can cool the surface, leading to growing sea ice concentration (SIC). The consequent increasing surface albedo enhances the shortwave radiation reflection and then amplifies the surface cooling, which favors sea ice grows. We examine the changes of SIC in Figure 8. The SIC increase by up to 20% in the Bering Sea, Irminger Sea, Greenland Seas and the sub-polar South Pacific. The net shortwave radiation reaching the surface decreases by more than 10W m-2 over sea ice growing centers. Besides the albedo feedback, increase in SIC can also block the air-sea interaction and the heat flux exchange at the air-sea interface.

As feedbacks, surface cooling will dampen the surface heat flux and lead to decreases in atmospheric water vapor and cloud cover that reduce the downward longwave returns but increase the incoming shortwave radiation. The cloud cover changes in the IE platform are mainly due to the anomalous low-level cloud amount, decreasing/increasing over cold/warm sea surface (Figure not shown). Thicker low-level clouds over warmer water than over cooler water are also observed in satellite observation (O'Neill *et al.*, 2003). As explored in Figure 7d, the cloud radiative forcing tends to amplify the role of net surface radiation at middle and low latitudes. In accompany with the significant surface cooling, cloud cover decreases in the tropical eastern Pacific and leads to the increase in cloud shortwave radiation. The cloud radiative forcing partly offsets the warming effect of net incoming radiation increase in the Arctic and coastal region along 70°S.

The net surface radiation cannot fully explain the spatial changes of surface temperature in detail, i.e. the significant cooling anomalies in the eastern tropical Pacific (Figure 4b). We presume that the regional ocean current adjustments to surface wind stress curl may be responsible for the changes at regional scales. The anomalous sub-surface current velocity potential at 5 meters depth is shown in Figure 9a. Significant convergence can be found in most parts of the Pacific and the latitude belts along 30°S and 30°N. The associated downwelling warms the mixed layer, which is partly responsible for the relatively weak surface cooling at low and middle latitudes. Upwelling associated with the divergence anomalies can well explain the significant cooling in the tropical eastern Pacific, tropical Atlantic, as well as the strong cooling in the polar regions. The upwelling also accounts for the thinner mixed layer depth at high latitudes in the IE platform (Figure 5c).

The oceanic vertical transports at the bottom of the mixed layer from wind stress curl (Figure 9b) follow the equation as in Smith (1968):

  (2)

where  is the Ekman pumping velocity,  is the curl of the wind stress vector,  is the density of seawater, and  is the Coriolis parameter.  is positive in the Northern Hemisphere (NH), negative in the Southern Hemisphere (SH). Positive surface wind stress curl induces an upwelling motion and negative curl stimulates a downwelling motion in the NH, and vice versa for SH. Positive wind stress curl in Figure 9b favors the strong upwelling and surface cooling in the Arctic, as well as the subsidence at approximately 40°S and along the coast of the Antarctic which contribute to the slight regional warming. The pronounced off-coast warming south of Africa and cooling in the subpolar South Pacific are associated with the positive and negative surface wind stress curl, respectively. In the Eastern tropical Pacific, the easterly anomaly associated northward/southward Ekman transport north/south of the equator leads to an upwelling motion that corresponds to the regional cooling.

**4.2 Meridional overturning circulation and heat transport**

To establish a stable climate state, the meridionally distributed net surface heating changes should be balanced by the ocean meridional advection, otherwise there will be a net drift in the depth averaged zonal mean ocean potential temperature. The upper ocean heat budget will further lead to changes in the ocean dynamic and thermodynamic processes in the deep ocean.

There is a strong MLD shallowing in the subpolar North Atlantic and in the Southern Ocean near the Antarctic continent (Figure 5c), where deep water occurs and drives the so-called thermohaline circulation. The MLD shallowing is strongest during the boreal winter (figure not shown). Although the annual mean results are weaker than that during the wintertime, they can well describe the MLD changes in both the Northern and Southern Hemispheres. The close linkage between the changes in MLD and the thermohaline circulation has been pointed out by many previous studies (e.g., Bentsen *et al.*, 2004; Delworth and Greatbatch, 2000; Dickson *et al.*, 1996; Dong and Sutton, 2005; Jungclaus *et al.*, 2005). The thermohaline circulation is also referred to as meridional overturning circulation (MOC). In NH, it transports warm and salty seawater northward above 1 km, sinks at higher latitudes and return to lower latitudes at around 1 km to 3 km depth (Cunningham and Marsh, 2010). The SC model can well reproduce the basic MOC structure (Figure 10b). The northward-flowing at the upper ocean and the returning southward-flowing comprise the upper cell of the MOC and it is mainly contributed by its Atlantic branch(Figure 10c). The Atlantic MOC transports heat from the South Atlantic to the North Atlantic, playing a central role in the meridional deep circulation of the global ocean. The lower cell of the MOC comprises the northward Antarctic Bottom Water and the southward North Atlantic Deep water flowing above.

As shown in Figure 10a, the Atlantic MOC strength in the IE simulation is about 1.2 Sv weaker than that in the SC run in the first 40 years, drops almost linearly in the following 20 years by about 0.8 Sv, and then reaches a quasi-equilibrium state. The upper cell of the global MOC in NH is dampened by about 0.8 Sverdrup (Figure 10d). The amplitude of the MOC weakening is about three times of the MOC inter-annual variability in the standard SC simulation (0.3 Sverdrup). The positive anomalies of the upper MOC limb in SH is associated with the negative surface wind stress curl and the associated upwelling around 50~60oS. The intensified downwelling south of 60oS is related to the strong surface wind stress trough anomalies (positive wind stress curl along the Antarctic coast). It should be noticed that the thermohaline circulation extends from the surface to the abyssal ocean but the integration length with the IE platform is only 150 years. The integration length is not long enough for the deep ocean to come into equilibrium. It is thus plausible that the MOC changes in the IE platform could be simply multi-decadal variability.

Marotzke (2000) reported that if the gross density structure of the ocean was sustained, a reduction of the MOC in NH would accompany with an increase of the MOC in SH. This leads to cooling in NH and the high-latitude warming in SH. The IE result is consistent with this theory. As shown by the global mean ocean heat transport changes in Figure 11a (yellow dotted line), the IE platform suppresses the northward heat transport in the Northern Hemisphere from 0o to 70oN but favors the southward heat transport in the Southern Hemisphere from 0o to 50oS. The blocked anomalous southward heat transport south of 50oS is at least partly due to the Antarctic Circumpolar Current as an impediment to strong ocean heat transport into high southern latitudes. Li and Conil (2003) demonstrated that the SST changes south of 30oS are closely related to the strong natural variability of the deep convection in the Southern Ocean. The poleward heat transport changes can be mostly attributed to transport in the Atlantic basin (Figure 11b). Generally, the poleward heat transport is suppressed by about 0.04 PW (1015 watt) in the Northern Hemisphere and intensified by about 0.03 PW in the Southern Hemisphere.

**4.3 The ENSO variability**

Besides the climate mean states, the stochastic perturbation at the air-sea interface can also affect the climate variability. The NINO3.4 auto-correlation is examined. In the observation (Figure 12a), ENSO has a wide span of interannual variability extending from 1~4 year, strongest at the 4 year period. In our SC model, the irregular ENSO variability with wide ranges of period was reasonably reproduced with weaker spectrum power as shown by the color bar than in observation. Different from the observation and the SC model, the ENSO interannual variability in the IE platform is more concentrated at biennial period with weaker spectral power. Therefore, we suggest that the atmospheric noise is at least partly responsible for the irregular ENSO variability and can explain a large amount of its spectral power.

Spatially, the SST changes during the positive ENSO phase, i.e. the El Niño period, exhibit a boomerang type cooling changes extending from the marine-continent to the northeast/southeast Pacific at middle latitudes that surround the positive anomalies in the eastern Pacific (Figure 13a). The surface warming extends from eastern Pacific westward to around 160oE. Surface warming is also evident in the western part of North Indian Ocean. The major bias in the SC simulation is the extended surface warming about 30o too far to the west in the tropical Pacific, the underestimated warming in the Indian Ocean, and the weak/missing cooling the subtropical South/North Pacific (Figure 13b). Although the general bias in the SC model is still evident in the IE platform, there are some improvements, such as the intensified warming signal in the Indian Ocean and the cooling developed in west Pacific and the sea surface south of Australia (Figure 13c).

Improvement is also evident in simulating the equatorial annual cycle of SST in the Pacific. In both the HadISST data and simulations (Figure 14), a clear westward propagation of the SST pattern can be seen. There is an annual SST cycle in the eastern Pacific, a semi-annual cycle in the western Pacific and their transition phases in the central Pacific. The semi-annual cycle in the western Pacific is attributable to the variation of solar forcings, whereas the tropical eastern and central Pacific is also controlled by air-sea interactions (Xie, 1994). The strength of the simulated eastern Pacific SST annual cycle in the SC simulation is about 60% of that in the HadISST. The strength grows up to 71% in the IE simulation. The strength of semi-annual cycle in the western tropical Pacific is also more reasonably simulated in IE, although it is still weaker than that in the observation. Therefore, the dampened atmospheric perturbations in IE increase the signal-to-noise ratio in the tropical Pacific and benefit the seasonal regional climate study.

**5. Conclusions**

The impacts of atmospheric stochastic perturbations on the ocean are investigated based on an equilibrium simulation with a newly developed Interactive Ensemble (IE) platform. The IE platform diminishes the atmospheric stochastic noise at the air-sea interface. In comparison with the Standard Coupled (SC) climate model, the IE platform reduces the time-mean sea surface temperature..Major sea surface temperature changes in the IE platform are evident at high latitudes in both Northern and Southern Hemispheres.

Surface wind stress shows positive correlation with the mixed layer thickness. The dampened surface stress in IE suppresses oceanic mixing, which results in thinner mixed layer. The thinner mixed layer and the smaller heat capacity are suggested to be responsible for the sea surface cooling. The upwelling and downwelling associated with the anomalous surface wind stress curl contribute to the local variations of sea surface temperature. The sea ice is enhanced and causes increased shortwave radiation reflection at the surface. This amplifies the cooling at high latitudes. In order to adjust surface energy budget, the ocean meridional overturning circulation is weakened especially in the Atlantic, which further sustains the surface cooling.

The wavelet spectrum of the Niño 3.4 index suggests that with reduced atmospheric noise the ENSO variability is preferentially on the quasi-biennial time scale. However, improvements in IE are evident in the simulation of ENSO related SST changes and the intensity of annual cycle of the tropical Pacific.

Our study shows that the model’s ability to capture the atmospheric perturbations at the air-sea interface is important to faithfully reproduce the climate mean state. This is consistent with Williams (2012). It should be noticed that the changes in our IE platform are mainly wind-driven. The dampened stochastic perturbations shallows the mixed layer depth. The smaller heat capacity of the mixed layer together with the decrease in the net surface radiation and wind curl changes caused thesea surface cooling In Williams (2012), the water and heat components were perturbed separately in two experimental runs, emphasizing the effects of the surface buoyancy fluxes. The stochastic modifications in Williams (2012) are more pronounced in the tropics. The approach of Williams (2012) requires only single model integration, so it is cheaper in terms of computational costs and is complementary to our study. The atmospheric noise reduction may also lead to changes in weather patterns and regimes. It is worthwhile in the future to investigate the impact of atmospheric noise on the climate in IE when the variations of external forcing are considered.

**Acknowledgments:** This work was jointly supported by the National Basic Research Program of China (973 Program) under Grant No. 2010CB951903 and Grant No. 2010CB951803, the National Natural Science Foundation of China under grant No. 41205043 and No. 41105054.

**References**

Adler, R., G. Huffman, A. Chang, R. Ferraro, P. Xie, J. Janowiak, B. Rudolf, U. Schneider, S. Curtis, D. Bolvin, A. Gruber, J. Susskind, P. Arkin, and E. Nelkin. 2003. The version-2 global precipitation climatology project (GPCP) monthly precipitation analysis (1979-present). *Journal of Hydrometeorology* **4** (6):1147-1167. DOI: http://dx.doi.org/10.1175/1525-7541(2003)004%3C1147:TVGPCP%3E2.0.CO;2.

Barkstrom, B., E. Harrison, G. Smith, R. Green, J. Kibler, and R. Cess. 1989. Earth radiation budget experiment (ERBE) archival and April 1985 results. *Bulletin of the American Meteorological Society* **70** (10):1254-1262. DOI: http://dx.doi.org/10.1175/1520-0477(1989)070%3C1254:ERBEAA%3E2.0.CO;2.

Bentsen, M., Drange, H., Furevik, T. and Zhou, T., 2004. Simulated variability of the Atlantic meridional overturning circulation. *Climate Dynamics*, **22**(6-7): 701-720.DOI:10.1007/s00382-004-0397-x.

Briegleb, B. et al., 2004. Scientific description of the sea ice component in the Community Climate System Model, version three. Technical Note TN-463STR, NTIS# PB2004-106574, National Center for Atmospheric Research, Boulder, CO.

Craig, A.P. et al., 2005. CPL6: The new extensible, high performance parallel coupler for the Community Climate System Model. *International Journal of High Performance Computing Applications*, **19**(3): 309-327.DOI:10.1177/1094342005056117.

Cunningham, S.A. and Marsh, R., 2010. Observing and modeling changes in the Atlantic MOC. *Wiley Interdisciplinary Reviews: Climate Change*, **1**(2): 180-191.DOI: 10.1002/wcc.22.

Delworth, T.L. and Greatbatch, R.J., 2000. Multidecadal Thermohaline Circulation Variability Driven by Atmospheric Surface Flux Forcing. *Journal of Climate*, **13**(9): 1481-1495.DOI:10.1175/1520-0442(2000)013<1481:MTCVDB>2.0.CO;2.

Dickinson, R.E. et al., 2006. The Community Land Model and Its Climate Statistics as a Component of the Community Climate System Model. *Journal of Climate*, **19**(11): 2302-2324.DOI:10.1175/JCLI3742.1.

Dickson, R., Lazier, J., Meincke, J., Rhines, P. and Swift, J., 1996. Long-term coordinated changes in the convective activity of the North Atlantic. *Progress in Oceanography*, **38**(3): 241-295.DOI:10.1016/S0079-6611(97)00002-5.

Dong, B. and Sutton, R.T., 2005. Mechanism of Interdecadal Thermohaline Circulation Variability in a Coupled Ocean–Atmosphere GCM. *Journal of Climate*, **18**(8): 1117-1135.DOI:10.1175/JCLI3328.1.

Fan, M. and Schneider, E.K., 2011. Observed Decadal North Atlantic Tripole SST Variability. Part I: Weather Noise Forcing and Coupled Response. *Journal of the Atmospheric Sciences*, **69**(1): 35-50.DOI:10.1175/JAS-D-11-018.1.

Gao, H., Xue, F. and Wang, H.J., 2003. Influence of interannual variability of Antarctic oscillation on mei-yu along the Yangtze and Huaihe River valley and its importance to prediction. *Chinese Science Bulletin*, **48**(s2): 61-67.

Hirschi, J. et al., 2003. A monitoring design for the Atlantic meridional overturning circulation. *Geophysical Research Letters*, **30**(7): 1413.DOI:10.1029/2002GL016776.

Jungclaus, J.H., Haak, H., Latif, M. and Mikolajewicz, U., 2005. Arctic–North Atlantic Interactions and Multidecadal Variability of the Meridional Overturning Circulation. *Journal of Climate*, **18**(19): 4013-4031.DOI:10.1175/JCLI3462.1.

Kirtman, B., Schneider, E., Straus, D., Min, D. and Burgman, R., 2011. How weather impacts the forced climate response. *Climate Dynamics*, **37**(11-12): 2389-2416.DOI:10.1007/s00382-011-1084-3.

Kirtman, B.P. and Shukla, J., 2002. Interactive coupled ensemble: A new coupling strategy for CGCMs. *Geophysical Research Letters*, **29**(10): 5-1-5-4.DOI:10.1029/2002GL014834.

Kirtman, B.P., Straus, D.M., Min, D., Schneider, E.K. and Siqueira, L., 2009. Toward linking weather and climate in the interactive ensemble NCAR climate model. *Geophysical Research Letters*, **36**(13): L13705.DOI:10.1029/2009GL038389.

Li, L. et al., 2013. Evaluation of grid-point atmospheric model of IAP LASG version 2 (GAMIL2). *Adv. Atmos. Sci.*, **30**(3): 855-867.DOI:10.1007/s00376-013-2157-5.

Li, L., Wang, B., Yuqing, W. and Hui, W., 2007. Improvements in climate simulation with modifications to the Tiedtke convective parameterization in the grid-point atmospheric model of IAP LASG (GAMIL). *Adv. Atmos. Sci.*, **24**(2): 323-335.DOI:10.1007/s00376-007-0323-3.

Li, Z.X. and Conil, S., 2003. A 1000-year simulation with the IPSL ocean-atmosphere coupled model. *Annals of Geophysics*, **46**(1): 39-46.

Lin, P., Liu, H., Yu, Y. and Zhou, T., 2013. Long-term behaviors of two versions of FGOALS2 in preindustrial control simulations with implications for 20th century simulations. *Adv. Atmos. Sci.*, **30**(3): 577-592.DOI:10.1007/s00376-013-2186-0.

Liu, H., Zhang, M. and Lin, W., 2011. An Investigation of the Initial Development of the Double-ITCZ Warm SST Biases in the CCSM. *Journal of Climate*, **25**(1): 140-155.DOI:10.1175/2011JCLI4001.1.

Marotzke, J., 2000. Abrupt climate change and thermohaline circulation: Mechanisms and predictability. *Proceedings of the National Academy of Sciences*, **97**(4): 1347-1350.

O'Neill, L.W., Chelton, D.B. and Esbensen, S.K., 2003. Observations of SST-Induced Perturbations of the Wind Stress Field over the Southern Ocean on Seasonal Timescales. *Journal of Climate*, **16**(14): 2340-2354.DOI:10.1175/2780.1.

Rayner, N., D. Parker, E. Horton, C. Folland, L. Alexander, D. Rowell, E. Kent, and A. Kaplan. 2003. Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *Journal of Geophysical Research: Atmospheres (1984–2012)* **108** (D14). DOI: 10.1029/2002JD002670.

Samelson, R.M. et al., 2006. On the Coupling of Wind Stress and Sea Surface Temperature. *Journal of Climate*, **19**(8): 1557-1566.DOI:10.1175/JCLI3682.1.

Seager, R. et al., 2000. Causes of Atlantic Ocean Climate Variability between 1958 and 1998. *Journal of Climate*, **13**(16): 2845-2862. DOI:10.1175/1520-0442(2000)013<2845:COAOCV>2.0.CO;2.

Serreze, M.C. and Barry, R.G., 2011. Processes and impacts of Arctic amplification: A research synthesis. *Global and Planetary Change*, **77**(1): 85-96.DOI: 10.1016/j.gloplacha.2011.03.004.

Smith, R. and Gent, P., 2002. Reference manual for the Parallel Ocean Program (POP), ocean component of the Community Climate System Model (CCSM2. 0 and 3.0), Technical Report LA-UR-02-2484, Los Alamos National Laboratory, Los Alamos, NM, http://www. ccsm. ucar. edu/models/ccsm3. 0/pop.

Smith, R.L., 1968. Upwelling. *Oceanogr. Marine Biol. Ann. Rev.*, **6**: 11-46.

Williams, P.D., 2012. Climatic impacts of stochastic fluctuations in air–sea fluxes. *Geophysical Research Letters*, **39**(10): L10705.DOI:10.1029/2012GL051813.

Wu, Z., Schneider, E.K. and Kirtman, B.P., 2004. Causes of low frequency North Atlantic SST variability in a coupled GCM. *Geophysical Research Letters*, **31**(9): L09210.DOI:10.1029/2004GL019548.

Xie, S.-P., 1994. On the Genesis of the Equatorial Annual Cycle. *Journal of Climate*, **7**(12): 2008-2013.DOI:10.1175/1520-0442(1994)007<2008:OTGOTE>2.0.CO;2.

Yeh, S.-W. and Kirtman, B.P., 2004. Tropical Pacific decadal variability and ENSO amplitude modulation in a CGCM. *Journal of Geophysical Research: Oceans*, **109**(C11): C11009.DOI:10.1029/2004JC002442.

Zhang, X., Lin, W. and Zhang, M., 2007. Toward understanding the double Intertropical Convergence Zone pathology in coupled ocean-atmosphere general circulation models. *Journal of Geophysical Research: Atmospheres*, **112**(D12): D12102.DOI:10.1029/2006JD007878.