Can a Regional Ocean–Atmosphere Coupled Model Improve the Simulation of the Interannual Variability of the Western North Pacific Summer Monsoon?

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(Manuscript received 9 December 2011, in final form 2 October 2012)

ABSTRACT
A flexible regional ocean–atmosphere–land system coupled model [Flexible Regional Ocean Atmosphere Land System (FROALS)] was developed through the Ocean Atmosphere Sea Ice Soil, version 3 (OASIS3), coupler to improve the simulation of the interannual variability of the western North Pacific summer monsoon (WNPSM). The regionally coupled model consists of a regional atmospheric model, the Regional Climate Model, version 3 (RegCM3), and a global climate ocean model, the National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG)/Institute of Atmospheric Physics (IAP) Climate Ocean Model (LICOM). The impacts of local air–sea interaction on the simulation of the interannual variability of the WNPSM are investigated through regionally ocean–atmosphere coupled and uncoupled simulations, with a focus on El Niño's decaying summer. Compared with the uncoupled simulation, the regionally coupled simulation exhibits improvements in both the climatology and the interannual variability of rainfall over the WNP. In El Niño's decaying summer, the WNP is dominated by an anomalous anticyclone, less rainfall, and enhanced subsidence, which lead to increases in the downward shortwave radiation flux, thereby warming sea surface temperature (SST) anomalies. Thus, the ocean appears as a slave to atmospheric forcing. In the uncoupled simulation, however, the atmosphere is a slave to oceanic SST forcing, with the warm SST anomalies located east of the Philippines unrealistically producing excessive rainfall. In the regionally coupled run, the unrealistic positive rainfall anomalies and the associated atmospheric circulations east of the Philippines are significantly improved, highlighting the importance of air–sea coupling in the simulation of the interannual variability of the WNPSM. One limitation of the model is that the anomalous anticyclone over the WNP is weaker than the observations in both the regionally coupled and the uncoupled simulations. This results from the weaker simulated climatological summer rainfall intensity over the monsoon trough.

1. Introduction
The western North Pacific (WNP) summer monsoon (WNPSM), which is an important component of the broad Asian summer monsoon, is an oceanic monsoon driven primarily by meridional gradients of sea surface temperature (SST) (Murakami and Matsumoto 1994; Tanaka 1997; Wang and Lin 2002; also see review papers by Li and Wang (2005); Zhou et al. (2011)). The WNPSM exhibits a pronounced interannual variability (e.g., Chang et al. 2000a,b; Wu and Wang 2000; Chou et al. 2003; Sui et al. 2007; Chung et al. 2011) and has considerable impacts on the summer climate over East Asia through the well-known Pacific–Japan (Nitta 1987) or western Pacific–East Asian teleconnection pattern (Huang and Sun 1992). The activity of the WNPSM also dominates the interannual variation of low-level atmospheric circulation over the Philippine Sea (Zhang et al. 1999; Chang et al. 2000a; Wang et al. 2000; Wang and Zhang 2002).

The interannual variability of the WNPSM is closely related to El Niño–Southern Oscillation (ENSO) (Chou et al. 2003). In El Niño’s decaying summer, a huge anomalous anticyclone is witnessed over the WNP (see review papers by Li and Wang 2005; Lau and Wang...
The anticyclone (WNPAC) is formed over the Philippine Sea in the previous autumn of El Niño’s developing year as a Rossby wave response to suppressed convection over the western Pacific Ocean (Wang et al. 2000; Wang and Zhang 2002) or as a result of the eastward propagation of a low-level anticyclone anomaly over the northern Indian Ocean (Chen et al. 2007). The maintenance of the WNPAC from El Niño’s mature winter to the subsequent early summer is attributed to a positive thermodynamic air–sea feedback (Wang et al. 2003; Wu et al. 2010).

The strengthening WNPAC from the spring to the summer after El Niño’s peak is attributed to the remote forcing from the tropical Indian Ocean basinwide warming (Yang et al. 2007; Wu et al. 2009; Xie et al. 2009). The Indian Ocean basinwide warming is established after El Niño’s mature winter and persists into the subsequent summer (Yang et al. 2007; Du et al. 2009; Hong et al. 2010). Linear model solutions (Watanabe and Jin 2003) and nonlinear atmospheric general circulation model (AGCM) experiments (Annamalai et al. 2005) suggest that, from December through May after the mature phase of El Niño, the Indian Ocean basinwide warming may contribute to maintain the WNPAC. Later studies argue that the Indian Ocean basinwide warming during El Niño’s mature winter and the subsequent spring does not have a significant impact on the WNPAC, because of the suppressed convection over the tropical Indian Ocean (Wu et al. 2009). In the subsequent summer, the basinwide warming enhances local deep convection, which emanates a baroclinic Kelvin wave into the equatorial Pacific. The Kelvin-type easterlies have the maximum amplitude on the equator and decrease with latitude. The induced anticyclonic shear results in a divergence in the atmospheric planetary boundary layer over the WNP. The divergence suppresses the local convection, which induces an anomalous anticyclone (Wu et al. 2009). The interaction between equatorial waves and moist physics is important in determining the WNPAC (Annamalai 2010).

Climate models are useful tools in predicting the variation of the WNPSM and in understanding the formation of the WNPAC during El Niño’s decaying phase. Zhou et al. (2009a) analyzed the output of AGCMs that participated in the Climate Variability and Predictability (CLIVAR) International Climate of the Twentieth Century (C20C) project and found that the interannual variability of WNPSM circulation during 1950–99 was forced and well modeled. Examinations of 11 AGCMs that participated in the Atmospheric Model Intercomparison Project II (AMIP II) found that the seasonal evolution of the WNPAC in the AMIP simulation strictly matches that of El Niño remote forcing, while in the observations it is not absolutely in phase with ENSO forcing, indicating the importance of air–sea interaction (Zhou et al. 2009b). Based on the result of a global ocean–atmosphere coupled model, Yang et al. (2007) state that most climate anomalies over the WNP in the summer subsequent to an El Niño event are caused by Indian Ocean warming. Sensitivity experiments employing an AGCM indicate that the contribution of Indian Ocean basinwide warming to the maintenance of the WNPAC is gradually enhanced from June to August (Wu et al. 2010). Chowdary et al. (2010) evaluated the predictability of summer WNPSM climate in 11 coupled model hindcasts and found that the remote forcing from tropical Indian Ocean SST warming is important for WNPSM summer rainfall anomalies following the mature phase of El Niño in 4 of the 11 models. Both local forcing and remote forcing are important in the other seven models. The WNPAC weakens considerably and reduces its westward extension in a seasonal forecast system without an interactive tropical Indian Ocean (Chowdary et al. 2011).

Previous modeling studies indicate that many climate models show limitations in the simulation of WNPSM variability, particularly in SST-driven AGCM simulations. The anomalous WNPSM activities during El Niño decaying year summers are largely forced by the tropical Indian Ocean. The influence of Indian Ocean basinwide warming on climate anomalies over the WNP may be separated into two stages: the responses of Kelvin-type atmospheric circulation anomalies to basinwide warming and the local responses of WNPSM climate to the Kelvin-type circulation anomalies. The bias in global model simulation may result from either stage or from both. In this study, we developed a regional ocean–atmosphere–land coupled model [Flexible Regional Ocean Atmosphere Land System (FROALS)] to improve WNPSM simulation. In the regional climate model (RCM) simulation, since the lateral boundary condition is forced by the circulation fields derived from reanalysis data, the large-scale Kelvin-type atmospheric circulation response to tropical Indian Ocean warming is included in the lateral boundary condition of the RCM. We thus hope that the RCM would show a better performance in WNPSM simulation. The following question will be addressed: Can the regional ocean–atmosphere coupled model improve the simulation of the interannual variability of the WNPSM, especially the WNPAC during El Niño decaying year summers?

The rest of the paper is organized as follows: Section 2 gives a brief introduction to the regional ocean–atmosphere–land coupled model FROALS and the experimental design. Section 3 evaluates the performance of FROALS in simulating SST over the WNP
region. Section 4 provides comparisons on rainfall simulation over the WNP between the uncoupled atmosphere-only model and FROALS, with emphasis on interannual variability during 1983–2007. Section 5 further discusses the performance of regionally coupled and uncoupled simulations during El Niño’s decaying summer. Section 6 summarizes the major conclusions.

2. Regional coupled model and experiments

a. Regional atmospheric model

We name the regional ocean–atmosphere coupled model used in this study as FROALS. The atmospheric component of FROALS is the Regional Climate Model, version 3 (RegCM3), which was developed at the Abdus Salam International Centre for Theoretical Physics (ICTP) (Pal et al. 2007). RegCM3 is the advanced version of RegCM2 (Giorgi et al. 1993a,b). It is a hydrostatic, compressible model with terrain following a sigma vertical coordinate system. The following physics schemes are employed in our study: the cumulus parameterization scheme of Grell (1993); the Subgrid Explicit Moisture Scheme (Pal et al. 2000); the radiation package of the Community Climate Model, version 3, of the National Center for Atmospheric Research (Kiehl et al. 1996); the nonlocal planetary boundary layer (Holtslag et al. 1990); the Biosphere–Atmosphere Transfer Scheme (BATS) of Dickinson et al. (1993); and the ocean–atmosphere flux algorithm proposed by Zeng et al. (1998). RegCM3 exhibits a reasonable performance in the simulation of the East Asian summer monsoon (Lee et al. 2004; Gao et al. 2006; Li and Zhou 2010) and the South Asian summer monsoon (Dash et al. 2006; Ratnam et al. 2009).

To prevent deep convection from occurring during dry conditions, convection is activated when the relative humidity averaged from the cloud top to the cloud base is larger than a critical value (in our application: 0.70). This convection suppression criterion has also been employed in regional climate models over East Asia (Emori et al. 2001; Chow et al. 2006) and a regional ocean–atmosphere coupled model over the WNP (Zou and Zhou 2011).

In the present experiment, the model domain covers 0º–40ºN, 105º–160ºE, with a horizontal resolution of 45 km. In the following discussion, this domain is termed the “target domain.” There are 113 grid points in the meridional direction, 136 grid points in the zonal direction, and 18 sigma levels in the vertical. The initial and lateral boundary conditions of the atmosphere are derived from the National Centers for Environmental Prediction/Department of Energy (NCEP/DOE) Global Reanalysis 2 (NCEP-2) (Kanamitsu et al. 2002) and are updated every 6 h. The buffer zone of RegCM3 is 12 grid points.

b. Ocean model

The oceanic component of FROALS is the National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG)/Institute of Atmospheric Physics (IAP) Climate System Model, version 2.0 (LICOM2.0; Liu et al. 2012). LICOM2.0 is an update of the previous version of the LASG/IAP climate ocean model (Jin et al. 1999; Zhang et al. 2003; Liu et al. 2004). LICOM has been employed as the oceanic component of the LASG/IAP Flexible Global Ocean Atmosphere Land System Model (FGOALS; Zhou et al. 2007, 2008; Yu et al. 2008; Bao et al. 2010).

The model domain covers 75ºS–90ºN. South of 75ºS is covered by land, and the boundary conditions at the 75ºS boundary are set to be rigid. The latitudinal grid spacing is 1º, while the meridional grid spacing is 0.5º between 10ºS and 10ºN and is gradually increased to 1º between 20ºS and 20ºN. There are 30 levels in the vertical, with 15 equally spaced levels in the upper 150 m. Vertical mixing is based on a second-order closure turbulence model (Canuto et al. 2001, 2002).

LICOM2.0 was spun up for 3 yr, from 1979 through 1981, from an initial condition with World Ocean Atlas 2005 (WOA05) (Locarnini et al. 2006; Antonov et al. 2006) temperature and salinity. Surface salinity is restored to the WOA05 monthly climatology with a time constant of 50 days. The daily mean 2-m air temperature, 2-m specific humidity, surface pressure, and 10-m wind derived from the NCEP-2 are used to calculate the driven fields of wind stress and surface turbulent heat flux along with the simulated SST (Large and Yeager 2004). The solar radiation and the longwave radiation derived from NCEP-2 are also employed.

c. Ocean–atmosphere coupling

The regionally coupled simulation begins from 1 January 1982. During the regionally coupled simulation, the ocean is forced by the NCEP-2 surface variables outside of the target domain, while inside the target domain shown in Fig. 1 the ocean receives fluxes from RegCM3. RegCM3 provides sea surface heat flux and wind stress to LICOM, while LICOM supplies the SST field to RegCM3. The coupling is performed once per day. This regionally coupled configuration has also been employed in Aldrian et al. (2005) for the Maritime Continent and in Xie et al. (2007) for the eastern Pacific region. The coupling here is accomplished through the Ocean Atmospheric Sea Ice Soil, version 3 (OASIS3), coupler (Valcke 2006). Because of different resolutions of atmospheric and oceanic models, the coupling fields
from source grid to target grid are interpolated using the “mosaic” method inside the coupler.

We carried out three sets of simulation. The regionally coupled model is integrated over 26 yr from 1982 through 2007. This regionally coupled simulation is termed the “regionally coupled run” in the following discussion. To investigate the role of air–sea coupling within the target domain, the stand-alone RegCM3 simulation forced by weekly Optimum Interpolation Sea Surface Temperature V2 (OISST2) (Reynolds et al. 2002) is performed. The stand-alone RegCM3 simulation is termed the “control run” in the following discussion. The stand-alone RegCM3 simulation forced by the SST from the regionally coupled run is also conducted and termed the “CPLSST run.” The first year, 1982, is regarded as the “spinup” time of simulations, and the results are excluded in the following analysis. Our analysis is based mainly on the outputs of the regionally coupled run and the control run.

d. Validation data

The following datasets are used to validate the model results: 1) the observational precipitation field derived from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) for the period 1983–2007 (Xie and Arkin 1997); 2) SST derived from OISST2 for 1983–2007 (Reynolds et al. 2002); and 3) the circulation fields, sea surface latent, and solar radiation fluxes derived from the NCEP-2 for the period 1983–2007 (Kanamitsu et al. 2002). The uncertainties of sea surface heat fluxes from the NCEP-2 should also be noted owing to the paucity of observed products of this nature.

3. Evaluation of FROALS on the simulation of SST

Figure 2 shows the performance of the regionally coupled model in the simulation of climatological June–August (JJA) mean SST averaged during 1983–2007. The north edge of the warm pool, where SST exceeds 28°C, extends to 25°N in the observations (Fig. 2a). The spatial pattern of SST is reasonably simulated in the regionally coupled simulation (Fig. 2b), as evidenced by a spatial pattern correlation coefficient (PCC) of 0.87, which is statistically significant at the 5% level. However, the regionally coupled run underestimates the SST over the coastal region of the East Asian continent and the Kuroshio region. The largest bias is about 2°C. The cold biases of SST have been also found in previous regional ocean–atmosphere coupled simulations over East Asia–western Pacific (Li and Zhou 2010; Fang and Zhang 2011; Zou and Zhou 2012). The poleward extension of the warm pool over the western Pacific is also limited in many of phase 3 of the Coupled Model Intercomparison Project (CMIP3) models (Annamalai et al. 2007).

Figure 3 shows the standard deviation (std) of JJA mean SST. A strong variability is found over the Kuroshio region and north of 30°N, whereas weak variability is evident over the WNP (Fig. 3a). The std of simulated SST has a spatial correlation coefficient of 0.71 with the observations (Fig. 3b), but the strength is weaker than in the observations.

Figure 4 shows the time series of JJA mean SST anomalies (SSTA) averaged within (5°–35°N, 110°–155°E). The regional average of climatological SST is 28.48°C in the observations versus 28.32°C in the simulation. The interannual variation of observed SST is reasonably reproduced by the regionally coupled run. The temporal correlation coefficient (TCC) between the simulation and the observations is 0.65, which is statistically significant at the 5% level. The major bias of the regionally coupled run is that the amplitude of interannual variability is weaker than the observations,
with an std value of 0.13°C in the simulation versus 0.25°C in the observations. In addition, the regionally coupled model fails to capture the major extreme events: that is, 1985, 1998, and 2001. How to improve the SST simulation deserves further study.

**Fig. 2.** Spatial patterns of June–August mean SST (°C) averaged during 1983–2007 from (a) observations and (b) the regionally coupled run. (c) Spatial map for the difference of SST between the regionally coupled run and observations.

**Fig. 3.** Interannual std of the JJA mean SST (°C) from (a) OISST and (b) the regionally coupled run.

**Fig. 4.** Time series of JJA mean SST anomalies (°C) averaged over (5°–35°N, 110°–155°E). The numbers in parentheses show the regional average of climatological SST.
4. Rainfall simulation in the control run and the regionally coupled experiment

a. Climatology

Figure 5 shows the spatial distributions of JJA mean rainfall averaged from 1983 to 2007. The observed major rainbands are located over the South China Sea region, east of the Philippines, and south of Japan. The control run fails in reproducing the rainbands over the South China Sea region and east of the Philippines. The intensity of rainfall south of Japan is overestimated in the control run. The spatial PCC (root-mean-square error) between the control run and the observations is $-0.17 \times (5.27 \text{ mm day}^{-1})$.

The regionally coupled run, which includes regional air–sea coupling processes, partly improves the simulation of climatological rainfall distribution. Compared with the control run, the intensity of the rainfall center south of Japan is reduced, while that over the South China Sea region and east of the Philippines is increased. The spatial PCC (root-mean-square error) between the regionally coupled run and the observations is 0.33 \times (4.77 \text{ mm day}^{-1}). Both are significantly improved in comparison with the control run, indicating that the air–sea coupling improves the simulation of climatological summer rainfall.

b. Interannual variability

To examine the performance of the model in simulating the interannual variability of summer rainfall, we divide the WNP into three subregions, as shown in Fig. 5c. Given the relatively small coverage of the simulated domain, an analysis of regionally averaged rainfall time series over these three subregions is enough. Figure 6 shows the regionally averaged percentage variations of rainfall anomalies relative to rainfall climatology. A strong interannual variation is evident in the central region of the WNP in the observations (Fig. 6a). However, the control run exhibits low skill in producing the observed variation, as evidenced by the insignificant correlation coefficient of 0.14. An encouraging result is seen in the regionally coupled run; the correlation coefficient between the simulated and the observed time series of rainfall anomalies is 0.50, which is statistically significant at the 5% level. The improvements in the simulation of the interannual variability of rainfall anomalies are partly related to the improvements in the simulation of climatological rainfall in the regionally coupled run (Fig. 5).

The amplitude of the interannual variability of summer rainfall is weak over the South China Sea in the observations (Fig. 6b). Both the control run and the
regionally coupled run overestimate the amplitude of the year-by-year variation of rainfall, with the result of the regionally coupled run being slightly better than that of the control run. The temporal correlation is 0.55 (0.37) between the regionally coupled run (control run) and the observations. This result indicates that the air–sea coupling improves the simulation of rainfall variability over the South China Sea, but the difference between the control run and the regionally coupled run are statistically insignificant at the 10% level.

The analysis above indicates that the process of air–sea coupling is crucial to the simulation of the interannual variability of summer rainfall over the WNP, especially over the central domain of 10°–25°N, 120°–150°E.

To further examine the interannual variability of WNP summer rainfall, empirical orthogonal function (EOF) analysis is done to extract leading modes of rainfall variability. Figure 7a shows the first leading mode derived from the observation. The first observed EOF mode, which explains 23% of the total variance, exhibits a meridional dipole pattern, with an elongated band of positive rainfall anomalies extending from the lower reaches of the Yangtze River valley (110°–120°N, 26°–32°E) to the northern Pacific (26°–35°N, 125°–155°E) and negative rainfall anomalies extending from the South China Sea to east of the Philippines (Fig. 7a). This mode is similar to that extracted with multivariate EOF analysis in Wang et al. (2008) and that extracted with season-reliant EOF analysis in Wu et al. (2009), despite the different lengths of samples and the different scopes of the analyzed domain.

In the control run, neither the simulated EOF1 nor the simulated EOF2 exhibits the meridional dipole pattern of the observed EOF1. The simulated EOF1 is featured with an elongated band of positive rainfall anomalies extending northeastwardly from the north of the South China Sea to south of Japan, while the simulated EOF2 is featured with negative rainfall anomalies from the north of the South China Sea to the central region of the WNP (figures not shown).

The third EOF mode derived from the control run (Fig. 7b) corresponds to the observed EOF1. The PCC is 0.72 between the regressed pattern of simulated rainfall anomalies against the observed principal component of EOF1 and the simulated third EOF mode, while it is 0.40 (0.28) for the simulated first (second) EOF mode. The simulated EOF3 accounts for only 9.13% of the total variance in the control run. This mode is statistically indistinguishable from EOF1 and EOF2 based on North et al. (1982). The observed positive rainfall anomalies north of 22°N are partly reproduced in the control run but with slightly southward replacement and stronger magnitude. The negative rainfall anomalies are mainly located in the South China Sea, and the positive rainfall anomalies are found east of the Philippines in the control run (Fig. 7b). The PCC of EOF3 derived from the control run with the observed EOF1 is only 0.20.

The first EOF mode derived from the regionally coupled run (Fig. 7c) corresponds to the observed
Compared with the control run, an improvement is seen in the regionally coupled run (Fig. 7c). The observed deficient rainfall east of the Philippines is now reproduced in the regionally coupled run, although the magnitude of the negative rainfall anomalies is still weaker than the observational counterpart. This mode accounts for 16.8% of the total variance and has a PCC of 0.50 with that derived from the observation. Thus the air–sea coupling significantly improves the simulation of rainfall variability mode, with PCC = 0.50 versus PCC = 0.20 in the control run.

Figure 8 shows the corresponding normalized principle components (PC) (PC1 for observations and the regionally coupled run and PC3 for the control run). The leading mode in the observations shows robust year-by-year variations during 1983–2007. In many of the large amplitude swings (e.g., 1984–85, 1988, 1990, and 1998), the control run is close to both the regionally coupled run and the observations. The TCC of PC between the control run (regionally coupled run) and the observations is 0.58 (0.74). Both the control run and the regionally coupled run reasonably reproduce the interannual variability of the leading mode, because of the specified lateral boundary condition derived from the NCEP-2. The major deficiency of the control run is the poor representation of the spatial pattern of the mode. The weak amplitude of the PC1 derived from the regionally coupled run may be from the weak std of simulated SST.

c. Atmospheric circulation and SST patterns associated with the leading mode of WNP rainfall variability

Previous studies have shown that the observed first rainfall mode over the WNP generally occurs following the peak of the ENSO event (Wang et al. 2003), and the Indian Ocean basin mode (IOBM) is a crucial remote forcing factor during the ENSO decaying summer (Xie et al. 2009; Wu et al. 2009). To examine this relationship...
in our regional modeling, Fig. 9 shows the lead–lag correlations of observed and simulated PC (PC1 for observations and the regionally coupled run and PC3 for the control run) and (a) the Niño-3.4 index and (b) the Indian Ocean basinwide warming index. The Indian Ocean basinwide warming index is defined as area-averaged SSTAs in the tropical Indian Ocean (10°S–10°N, 40°–110°E).

Figure 10 shows the atmospheric circulation and SST patterns associated with the EOF mode (EOF1 for observations and the regionally coupled run and EOF3 for the control run). Here, the JJA mean 850-hPa wind, 500-hPa vertical pressure p velocity, and SST anomalies are regressed against the PC time series. Figure 11 shows the regression coefficients of net solar radiation flux anomalies against the PC time series. In the observations, associated with the suppressed rainfall (Fig. 7a), the regions extending from the South China Sea to east of the Philippines are dominated by anomalous descending motions and a prominent anticyclone (Fig. 10a). The suppressed convection results in a decrease in clouds and an increase in the downward shortwave radiative flux (Fig. 11a). The enhanced downward shortwave radiative fluxes lead to warm SSTAs (Fig. 10b). As proposed by Wu et al. (2009), the initial divergence at a low level over the WNP region is induced by the anticyclonic shear of the easterly anomalies that are forced by the basinwide convective heating over the Indian Ocean. The initial divergence at a low level is amplified because of the positive local circulation–convection feedback (Wu et al. 2009; Xie et al. 2009).

In the control run, since the SST is specified, the SSTA pattern associated with the corresponding rainfall mode resembles that of the observations (Fig. 10d). Following a weak anomalous anticyclone extending northeastwardly from the South China Sea to the south of Japan (Fig. 10c), weak descending motions at 500 hPa (Fig. 10c) and weak increases of the downward shortwave radiative flux (Fig. 11b) are seen. The warm SSTAs located east of the Philippines and north of 22°N drive ascending motions at 500 hPa (Fig. 10c) and induce the increases of clouds (not shown), which decrease the downward shortwave radiative flux reaching sea surface (Fig. 11b). These results indicate that in the control run the air–sea interactions appear as the ocean forcing the atmosphere, unlike the condition in the observations, which appears as the atmosphere forcing the ocean.

Compared with that in the control run, the low-level anomalous anticyclone over the WNP region is more prominent in the regionally coupled run (Fig. 10c), although it is less zonally extensive and still weaker than the observations. The associated descending motion anomalies at 500 hPa (Fig. 10c) and the increases of the downward solar radiation flux (Fig. 11c) are enhanced, which can be found over the South China Sea. The increased downward solar radiation flux is also followed by warm SSTAs (Fig. 10f) but with slightly smaller spatial coverage.

In particular, the anomalous ascending motions at 500 hPa east of the Philippines in the control run are
now replaced by weak anomalous descending motions in the regionally coupled run. The weak descending motions are followed by weak increases in the downward solar radiation flux (Fig. 11c). However, weak cold SSTAs are found (Fig. 10f). In the regionally coupled run, the SSTA averaged over \((8^\circ-14^\circ N, 125^\circ-145^\circ E; \text{the box in Fig. 10f})\) is \(-0.04^\circ C\). The associated value is 0.45 W m\(^{-2}\) in net solar radiation flux (downward is positive, hereafter), 1.75 W m\(^{-2}\) in latent heat flux, and 0.35 W m\(^{-2}\) in sensible heat flux. These results imply that the simulated cold SSTAs may be caused by the ocean dynamics processes in the regionally coupled run, since the surface heat flux anomalies favor warm SSTAs.

**Fig. 10.** (left) Spatial patterns of 850-hPa wind (m s\(^{-1}\); vector) and 500-hPa vertical \(p\)-velocity (shaded; \(10^{-2}\) hPa s\(^{-1}\)) anomalies regressed onto the principal component of the EOF mode (EOF1 for observations and the regionally coupled run and EOF3 for the control run) from (a) NCEP-2, (c) the control run, and (e) the regionally coupled run. (right) The corresponding JJA mean SSTA (°C) from (b) OISST, (d) the control run, and (f) the regionally coupled run. The black box in (f) indicates \(8^\circ-14^\circ N, 125^\circ-145^\circ E\). Note that the interval of the scale is irregular. In the left (right) panels, negative shading indicates ascending motion anomalies (cold SST anomalies).
5. Discussion

a. Why is the interannual variability of the WNPSM during El Niño’s decaying summer not simulated well by RegCM3 forced by observed SST?

The analyses above show that the low-level anomalous anticyclone over the WNP generally occurs during El Niño’s decaying summer. The summer climatology over the WNP plays an important role in determining the preferred location of the anomalous anticyclone. The summer maximum convective heating located along the monsoon trough favors a much greater response to the Kelvin wave anomalies induced by Indian Ocean basinwide warming (Wu et al. 2010).

Since the Kelvin-type circulation anomalies induced by the Indian Ocean basinwide warming have been prescribed as the lateral boundary condition of RegCM3, the ability of RegCM3 forced by observed SST (control run) in simulating the climatological summer monsoon trough over the WNP determines the reproducibility of WNP climate anomalies during El Niño’s decaying summer. Because the JJA mean rainfall intensity over the monsoon trough in the control run is much weaker than the observations (Fig. 5), the response of RegCM3 to Kelvin wave forcing is weaker than the observation. Thus, a weaker anticyclone and weaker descending motions are observed (Fig. 10c).

Since the ocean is primarily forced by the overlying atmosphere over the WNP in observations during El Niño’s decaying summer, the weak responses of the atmosphere to the lateral boundary forcing expect weak SSTA responses. This is indeed the case seen in the regionally coupled run. In the stand-alone RegCM3 run, the specified SST exerts an unrealistic forcing to the atmosphere (Figs. 7b, 10c). The “too strong” SSTA forcing and the “realistic but weak” response of the model to boundary forcing lead to the poor performance of the control run in rainfall simulation.

b. Why does RegCM3 with the inclusion of local air–sea interaction improve the simulation of the interannual variability of the WNPSM during El Niño’s decaying summer?

Since the climatological JJA mean rainfall intensity over the monsoon trough in the regionally coupled run is still weaker than the observations (Fig. 5c), the response of the atmosphere to Kelvin wave forcing induced by Indian Ocean basinwide warming in the regionally coupled run is still weak, as evidenced by the weaker-than-observations low-level anticyclone and vertical velocity (Figs. 10e, 11c).

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**Fig. 11.** JJA mean shortwave radiative flux anomalies (W m$^{-2}$) regressed onto the principal component of the EOF mode (EOF1 for observations and the regionally coupled run and EOF3 for the control run) from (a) NCEP-2, (b) the control run, and (c) the regionally coupled run. Positive shading indicates increased downward shortwave radiation flux.
In the regionally coupled run, the weaker low-level anticyclone results in a weak warm SSTA, as expected (Fig. 10f). Thus, the feedbacks from the underlying SSTAs to the overlying atmosphere are far weaker than those in the control run. In addition, the cold SSTAs located east of the Philippines in the regionally coupled run (Fig. 10f) also enhance the intensity of the low-level anticyclone. The air–sea coupling improves the rainfall variability simulation by suppressing the unrealistic SST forcing seen in the control run.

To further confirm this argument, we carried out a set of sensitivity experiments (CPLSST run), in which RegCM3 is forced by the daily SST provided by the regionally coupled run. The performance of the CPLSST run is generally similar to the regionally coupled run (figures not shown here). The PCC of climatological JJA mean rainfall (first EOF mode of JJA mean rainfall anomalies) is 0.99 (0.90) between the CPLSST run and the regionally coupled run. The TCC of PC1 is 0.93 between the CPLSST run and the regionally coupled run. The resemblance between the CPLSST run and the regionally coupled run confirms our argument that the weak forcing from the underlying SST is the major factor that affects the rainfall simulation in the regionally coupled run.

6. Summary and concluding remarks
   a. Summary

In this study, we developed a flexible regional ocean–atmosphere–land system model (FROALS) based on the OASIS3 coupler and applied it to the simulation of the WNPSM. RegCM3 served as the regional atmospheric model, while the global ocean model LICOM served as the oceanic model. The land surface process BATS was treated as part of the atmospheric component. The performance of FROALS in simulating the interannual variability of the WNPSM during 1983–2007 was compared with the stand-alone RegCM3 simulation. The impacts of regional air–sea coupling in simulating the interannual variability of the WNPSM during the ENSO decaying summer were also discussed. The major conclusions are summarized as follows:

1) FROALS reasonably reproduces the broad characteristics of climatology (PCC = 0.87) and interannual variability (TCC = 0.65) of JJA mean SST over the WNP region during 1983–2007. Compared with the observation, the climatology of the simulated JJA mean SST is 2°C colder over the coastal region of East Asia and the Kuroshio region and the amplitude of interannual variability of JJA SST is weaker.

2) Compared with the stand-alone RegCM3 simulation, FROALS improves the simulation of both the climatology and the interannual variability of rainfall during 1983–2007 over the WNP. The first dominant interannual variability mode (EOF1) of observed rainfall anomalies, appearing as a meridional dipole pattern, generally occurs during the ENSO decaying summer. This mode and its associated temporal variation simulated by FROALS (the first simulated EOF mode) are better than those derived from RegCM3 stand-alone simulation (the third simulated EOF mode), with PCC = 0.50 versus 0.20 and TCC = 0.74 versus 0.58.

3) Data diagnosis indicates that, in El Niño’s decaying summer, the ocean over the WNP appears as a slave to atmospheric forcing. An anomalous anticyclone leads to decreases in clouds, increases in the downward shortwave radiation flux, and thereby warm SSTAs. In the uncoupled simulation, however, the atmosphere is a slave to oceanic SST forcing, and the warm SSTAs located east of the Philippines unrealistically produce excessive rainfall. In the regionally coupled run, the unrealistic SST forcing was suppressed because of the inclusion of air–sea interaction.

4) The strength of the WNP anticyclone in the model is affected by the climatological monsoon trough. Because the climatological JJA mean rainfall intensity over the monsoon trough simulated by stand-alone RegCM3 is much weaker than the observations, the low-level anticyclone simulated by stand-alone RegCM3 is weaker than the reanalysis. The weak atmospheric response expects a weak SSTA response, but in the stand-alone RegCM3 simulation the observational SSTA is specified as boundary forcing. The local warm SSTA exerts a strong forcing on the above atmosphere, as evidenced by the excessive rainfall and the associated atmospheric responses east of the Philippines.

5) In our model world, regional air–sea coupling effectively improves the simulation of WNPSM variability. The final response of FROALS may be regarded as a balance between the specified lateral boundary condition forcing and the underlying SST forcing. During El Niño’s decaying summer, in the FROALS simulation, the low-level anomalous anticyclone is still weaker than the reanalysis because of the weaker climatological rainfall intensity over the monsoon trough. The unrealistic and too strong SST forcing east of the Philippines in the offline RegCM3 run is greatly suppressed in the FROALS simulation through the reduction of feedback from underlying SST anomalies, thus leading to a significant...
improvement of the simulation of the WNPSM rainfall variability. The realistic response of the regional model to lateral boundary condition forcing is not offset by the unrealistic SST forcing as seen in the stand-alone RegCM3 run, so significant improvements are achieved in the simulation of rainfall anomalies and associated atmospheric circulation fields.

b. Concluding remarks

The limitations of the current study should be acknowledged. In the study, both the stand-alone RegCM3 and the regionally coupled simulation experience large systematic biases. The first systematic bias is the underestimation of climatological summer rainfall over the monsoon trough, which is evident in both the stand-alone RegCM3 and the regionally coupled simulation. Another systematic bias is the cold SST bias in the regionally coupled simulation. Our additional sensitivity experiment (CPLSST run) confirmed the importance of simulated SST in improving the simulation of WNPSM variability. How to improve these systematic biases deserves further study, which will be addressed through a series of sensitivity experiments in the future.

Previous AGCM sensitivity experiments suggest that the observed local cold SST anomalies over (10°–25°N, 160°E–180°) may also play a role in maintaining the WNP anticyclone during El Niño’s decaying summer, as evidenced by the weaker-than-observations low-level anticyclone in both the stand-alone RegCM3 and the regionally coupled simulation. Another systematic bias is the cold SST bias in the regionally coupled simulation. Our additional sensitivity experiment (CPLSST run) confirmed the importance of simulated SST in improving the simulation of WNPSM variability. How to improve these systematic biases deserves further study, which will be addressed through a series of sensitivity experiments in the future.

Acknowledgments. The helpful comments from the three anonymous reviewers and the editor John Chiang are highly appreciated. This work was supported by the Strategic Priority Research Program of the Chinese Academy of Sciences (XDA05110301), National Program on Key Basic Research Project of China (2010CB951904, 2013CB956204), National Natural Science Foundation of China (41205080, 40890054, and 41023002), and Public Science and Technology Research Funds (Projects of Ocean 201105019-3).

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