Seasonal effects of Indian Ocean freshwater forcing in a regional coupled model

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Abstract

Effects of freshwater forcing from river discharge in the Indian Ocean on oceanic vertical structure and the Indian monsoons are investigated using a fully coupled, high-resolution, regional climate model. The effect of river discharge is included in the model by restoring sea surface salinity (SSS) towards observations. The simulations with and without this effect in the coupled model reveal a highly seasonal influence of salinity and barrier layer (BL) on oceanic vertical density stratification, which is in turn linked to the changes in sea surface temperature (SST), surface winds and precipitation.

During both the boreal summer and winter, SSS relaxation leads to a more realistic spatial distribution of salinity and BL in the model. In summer, BL in BoB enhances the upper ocean stratification and increases the SSTs near the river mouths where the freshwater forcing is largest. However the warming is limited to the coastal ocean and the amplitude is not large enough to give a significant impact on the monsoon rainfall.

The strengthened BL during the boreal winter, on the other hand, leads to the shallow mixed layer with SSS relaxation. Atmospheric heat flux forcing acting on a thin mixed results in an extensive reduction of SST over the northern Indian Ocean. Relatively suppressed mixing below the mixed layer warms the subsurface layer, leading to a temperature inversion. This oceanic condition, through the cooling of the sea surface, induces a large-scale adjustment in the winter atmosphere with amplified northeasterly winds. This impedes atmospheric convection north of the equator while facilitating it in the austral summer inter-tropical convergence zone, resulting in a hemispheric-asymmetric response pattern. Overall, our results suggest that freshwater forcing from the river charges play an important role during the boreal winter by affecting SST and the coupled ocean-atmosphere interaction, with potential impacts on the broad-scale regional climate.
The Bay of Bengal (BoB) receives substantial freshwater flux both from local precipitation and river discharge during the summer Southwest (SW) monsoon. Precipitation well exceeds evaporation (Harenduprakash and Mitra 1988, Prasad 1997). In addition, major rivers from the bordering countries, the Irrawaddy, the Krishna, the Godavari, the Mahanadi, the Ganges, and the Brahmaputra, discharge an annual freshwater volume estimated from $1.5 \times 10^{12}$ m$^3$ (Thadathil et al. 2002) to $1.83 \times 10^{13}$ m$^3$ (Varkey et al. 1996) into the BoB, making the BoB the freshest region in the Indian Ocean. Seasonally, three fourths of all riverine influx occurs during the SW monsoon period from May until September, indicating that this substantial summertime freshwater influx will affect the hydrography of the Bay during the Northeast (NE) monsoon (Shetye et al. 1996).

This unique summertime and wintertime oceanic conditions associated with the presence of a low-salinity layer lead to marked variations in density stratification and vertical mixing in BoB both in summer (Vinayachandran et al. 2002) and winter (Shetye et al. 1996). The strong halocline associated with the surface freshwater cap shoals the density mixed layer (ML), while the isothermal layer depth (ILD) is usually deeper. Thus the salinity has a crucial control on the depth at which mixing effect is confined. The ILD is defined in this study as the depth at which temperature drops from SST by $\Delta T = 1^\circ C$. The ML depth (MLD) is defined as the depth at which the salinity effect on the density variation is equivalent to the variation of temperature by $\Delta T = 1^\circ C$ from the sea surface. Following the criteria of Sprintall and Tomczak (1992), the positive difference in the MLD and ILD is termed the barrier layers (BLs, Lukas and Lindstrom 1991), since it acts as a barrier to the mixing of cold thermocline water into the surface ML, essentially decoupling the dynamics and the thermodynamics. That is, the stratified BL retains atmospheric heat and momentum inputs predominantly within the ML while reducing the entrainment of the cooler thermocline water from the bottom (Vialard and Delecluse 1998). A survey of BL thickness (BLT) by Sprintall and Tomczak (1992) using the Levitus climatology (1982) indicates that BLs exist throughout the year in BoB, with the thickness of 25 m during August-October in the northwestern part of the Bay.
(Vinayachandran et al. 2002). Rao and Sivakuma (2003) also discuss the formation mechanism and seasonal cycle of BL in the northern Indian Ocean. They showed that the build-up of BL during summertime becomes most prominent by February in the following year, reaching maximum thickness of 50 m. The boreal winter is the period when the hydrological forcing generates its largest freshening effects through the river discharge and the local rainfall. BLT subsequently diminishes from February to the minima in May before the onset of the summer monsoon. Based on more recent datasets with better spatial and temporal coverage, various investigators confirm the seasonal cycle and the observed thickness of the BL in the BoB and the northern Indian Ocean (e.g., Thadathil et al. 2007, Mignot et al. 2007 and de Boyer Montégut et al. 2007 (DMLC07, hereafter)).

High SST in BoB is conducive to deep convection in the atmosphere and hence precipitation during the SW monsoon season (Gadgil et al. 1984, Shenoi et al. 2002). Altered atmospheric convection results in variations in air-sea heat flux, which can further influence the convection itself via SST changes (Krishnamurti et al. 1988). Furthermore, through the aforementioned impact of BL on the ocean mixing, surface fresh water affects SST and ocean-atmosphere coupling (Vinayachandran et al. 2002). From the measurements of surface meteorological fields and near-surface salinity structure at the head of the Bay during the MONTBLEX-90 field experiment during August-September 1990, Sanilkumar et al. (1994) showed that the low-salinity layer leads to the accumulation of heat in the upper 30 m of the ocean, which triggers the genesis and the propagation of atmospheric moist convection and monsoon depressions. This is due to the nonlinearity and coupling of moisture, convection and SST, enabling small variations in SST to affect atmospheric circulations and convection (Soman and Slingo 1997, Shankar et al. 2007, Zhou and Murtugudde 2008), with an important implication for the SW monsoon winds and rainfall over the northern Indian Ocean (Ueda et al. 2009).

The mechanism by which BL influences the summertime stratification and ML temperature of BoB is essentially the same in winter when the BL attains its maximum thickness (Thadathil et al. 2007). The effect of wintertime heat flux and the continental
wind (Hastenrath and Lamb 1979) is retained within this shallow ML, while subsurface layer is insulated from the surface cooling. The vertical temperature distribution of cold surface and warmer subsurface layer leads to a temperature inversion (Shetye et al. 1996, Thadathil et al. 2002, DMLC07).

The wintertime low salinity water in BoB is then advected by the winter monsoon current towards the southeast Arabian Sea (SEAS). This arrival of cold less saline water over the warm saline Arabian Sea water is considered to contribute to the formation of BL and inversion layer (Lukas and Lindstrom 1991, Thadathil and Gosh 1992, Shenoi et al. 1999) and the so-called mini-warm pool through the reduction of vertical mixing (Vialard and Delecluse 1998), in the Lakshadweep Sea during the pre-summer monsoon season (Joseph 1990, Rao and Sivakumar 1999, Vinayachandran et al. 2007). Note Kurian and Vinayachandran (2007) suggest from a series of experiments that low-wind condition due to the orographic effects of Western Ghats and the resultant reduced latent heat loss over the SEAS greatly contribute to the formation of mini-warm pool. Masson et al. (2005) showed that this BL and subsurface inversion in SEAS substantially contribute to the warming of SST during the pre-monsoon season and rainfall, facilitating the formation of the monsoon onset vortex and the early onset of the SW monsoon (Rao and Sivakumar 1999).

BLs and the temperature inversion in SEAS are good examples that the oceanic process can impact the characteristics of the Indian summer monsoon. However it is not clear in the literature if, and how, the activity of the winter monsoon is affected by the BL. Recalling that BLs reach the maximum thickness during the boreal winter, it is natural to ask whether this BL-SST-monsoon connection is at work in winter.

Answering such a question requires the use of the numerical models that capture the BL effects. However, a conclusive identification of the influence of freshwater cap on SST and the Indian monsoons is difficult with ocean general circulation models (OGCMs), where a simple atmospheric boundary model has to be employed to compute surface heat and momentum flux (Vialard and Delecluse 1998, Han et al. 2001, Han and McCreary 2001, Masson et al. 2002, Durand et al. 2004). This simple approach limits models’ ability to evaluate salinity effect on the SST due to the lack of fully coupled heat and
wind forcing. Murtugudde and Busalacchi (1998) and Howden and Murtugudde (2001, hereafter HM01) used an advective atmospheric mixed layer model (Seager et al. 1995) that computes its own heat flux, but the wind is still imposed. Furthermore, the advective atmospheric mixed layer could be problematic over BoB because the land temperatures are very different from the SSTs.

Masson et al.’s (2005) study is probably the only one that used the fully coupled GCM (CGCM) to examine the effect of salinity stratification and BL on the Indian monsoon. In their sensitivity experiment, the potential density was computed as a function of temperature only in order to suppress the turbulent processes associated salinity stratification (Vialard and Delecluse 1998). In their model, however, there was no significant change in monsoon over continental India due to the BL dynamics in SEAS, while the effect of BL is rather local to the region where the perturbation is added. This was attributed partly to the coarse-resolution atmospheric model, which fails to capture the mature phase of the Indian monsoon and its extent over the continent.

The present study uses a higher resolution, regional coupled climate model with a goal of assessing climate sensitivity in the Indian Ocean sector to freshwater forcings and BLs. By allowing ocean-atmosphere interactions on relatively fine ocean and atmospheric grids, the regional coupled model could provide an assessment of the salinity effect on the SST and the monsoons by resolving more detailed structure of the coupled system. To this end, we perform two coupled sensitivity tests, with and without the effect that mimics the freshwater forcing from the river discharges and will demonstrate the sensitivity of the ocean, atmosphere and the coupled system to freshwater forcing.

Briefly, our modeling results show that it is in boreal winter that the dynamics of BL is most influential on the Indian monsoon. It is because BLs are thicker in winter when the surface heat flux more efficiently cools the thin ML and the sea surface. Significant cooling of the ocean in turn induces a large-scale response in the boreal winter atmosphere by strengthening winter monsoon wind and displacing the inter-tropical convergence zone (ITCZ) southward.
The paper is organized as follows. In Section 2, a description of the model the observational data is presented. Section 3 introduces the experimental design. Various aspects of the model biases in seasonal mean climate are discussed in Section 4. In Section 5, we discuss the impact of low salinity on the oceanic vertical structure in summer and winter. Section 6 examines a link between SW and NE monsoons to the altered SSTs by SSS. Conclusions and discussion follow in Section 7.

2. Model and observational data

a. Scripps Coupled Ocean-Atmosphere Regional (SCOAR) Model

The coupled model used for the present study is the Scripps Coupled Ocean-Atmospheric Regional (SCOAR) model. Two regional models, the Experimental Climate Prediction Center (ECPC) Regional Spectral Model (RSM, Juang and Kanamitsu 1994) for the atmosphere and the Regional Ocean Modeling System (ROMS, Haidvogel et al. 2000, Shchepetkin and McWilliams 2005) for the ocean, are coupled through the flux-SST coupler by the method discussed in detail by Seo et al. (2007).

For this study, we have set up the SCOAR model over the entire Indian Ocean from 29°E-112°E and 30°S-38°N, extending from eastern Africa to the eastern Indian Ocean (see Figure 1 for the model domain). The horizontal resolution for the ocean and atmospheric component of the coupled model is identical at 0.26°. The ocean (atmospheric) model uses 20 (28) sigma layers in the vertical. There are roughly 12 layers in the upper 100 m in the BoB. The vertical mixing in the model is parameterized using a K-profile parameterization scheme (Large et al. 1994) including penetrative shortwave heating effects (Paulson and Simpson, 1977) with a background vertical viscosity (diffusivity) of 5x10^{-5} (5x10^{-6}) m^2 s^{-1}.

To obtain the oceanic initial condition for the coupled model, the ROMS was first spun up for 8 years forced with monthly climatological atmospheric forcings interpolated from the Comprehensive Ocean-Atmosphere Data Set (da Silva et al. 1994), and the monthly climatological oceanic boundary conditions derived from the World Ocean Atlas 2005.
(WOA05, Locarnini et al. 2006, Antonov et al. 2006). The end state from this spun-up forced ocean run is used for the initial condition for ROMS in the coupled mode. The RSM in the coupled run is initialized on January 1st of 1993 using National Centers for Environmental Prediction (NCEP) / Department Of Energy (DOE) Reanalysis 2 (RA2, Kanamitsu et al. 2002), which is also used to constrain the low-wave number atmospheric flows over the domain. The coupled run is then performed for 12 years from 1993 to 2004 with a daily exchange of surface fluxes (total heat flux, net freshwater flux, downward shortwave radiation, and momentum fluxes) and SSTs. The model monthly climatology is constructed based on the 12-year model outputs, and we will mainly display the difference fields of the monthly climatology of the two experiments, emphasizing the seasonal response during summer (June-July-August, JJA) and winter (December-January-February, DJF).

The RSM employs a scale-selective spectral nudging technique documented in Kanamaru and Kanamitsu (2007) and updated Yoshimura and Kanamitsu (2008). This technique is designed to minimize systematic large-scale errors in the downscaling procedure while retaining the regional scale processes such as river plumes from the mouth of rivers in BoB. As will be shown in Section 4, however, despite this spectral nudging technique, the simulated large-scale mean feature deviates from its base field (RA2), resulting in noticeable mean bias errors in the model.

b. Observational Data

For the purpose of verification of the model simulation (1993-2004) against the observations, the following observational products are obtained and interpolated to the model’s horizontal and vertical grids. The ocean temperature and salinity (T/S) from the WOA05 are used to compare the surface and subsurface climatology. We also estimate BLT, MLD, ILD and the depth of 20°C isotherm (Z20), as a proxy for the thermocline depth, from the T/S climatology of WOA05. The observed monthly SST data are obtained from the NOAA Optimum Interpolation SST Analysis on 1° grid (Reynolds et al. 2002). Three precipitation products are used in this study: 1) the monthly Global
Precipitation Climatology Project (GPCP) version 2 (Adler et al. 2003) on a 2.5° grid, (2) the CPC Merged Analysis of Precipitation (CMAP) on a global 2.5° grid (Xie and Arkin 1997) and finally (3) the monthly Tropical Rainfall Measuring Mission (TRMM) precipitation product 3B43 Version 6 (Huffman et al. 2007) that combines the TRMM data with the estimate from the global gridded rain gauge data on 0.25° grid. The TRMM data are used for comparison only during the shorter period of 1998-2004 but at the better matching resolution to the model. Finally the 10-m wind fields are obtained from the monthly averaged NCEP Reanalysis 2 on a global T62 Gaussian grid.

3. Experiment design

In order to capture the effect of freshwater forcing introduced by river discharge and to connect E-P errors in the open ocean, two coupled simulations are performed. In the first experiment, termed SR (Salinity Restoring), SSS is relaxed to the WOA05 monthly climatological SSS over the entire domain. Thus in SR, SSS is computed based on the climatological evaporation, precipitation and the river runoff. In a complementary coupled experiment, no SSS restoring (NoSR) is applied in the ocean. SSS in the NoSR run hence is determined by the simulated evaporation minus precipitation (E-P) only, without the freshwater forcing from the river charges.

Figure 1 illustrates the difference in E-P, comparing two experiments for the boreal summer and the boreal winter. The spatial patterns of E-P are dominated by the seasonal mean precipitation patterns associated with the Indian monsoon (Prasad 1977, Harenduprakash and Mitra 1988) in both runs. Magnitude of the E-P during JJA exceeds 1 cm day\(^{-1}\) in the Arabian Sea, BoB, and the western Indian Ocean. Note the difference in freshwater flux in JJA exhibits a spatial pattern that resembles that of SST and precipitation near the equator where extrema in E-P fluxes are located (Figure 12 in the later section). This indicates that difference in E-P due to SSS restoring may cause some responses in SST, wind and rainfall. Note in BoB, however, that there is no difference in E-P between the two runs. This strongly suggests that inconsistency in E-P plays a minor
role in BoB, and the dominant factor in determining SSS is the freshwater flux from the SSS restoring.

Figures 2 and 3 illustrate how the simulated salinity fields in the model with and without the SSS relaxation compared with the observations. In the observations (Figures 2a,b), low salinity waters fill the entire Bay throughout the year, with the freshest water mass less than 28 practical salinity unit (psu) along the coast adjacent to the major river mouths. This results in a very large lateral haline gradient. From a time-series of hydrographic observations conducted in the northern BoB during the summer of 1984 on board ORV Sagar Kanya, Murty et al. (1992) reported the large SSS gradient from the open Bay (35 psu) toward the northern BoB (16 psu). The WOA05 well captures this spatial distribution with, for example, low salinity in the region from the river runoff (e.g., 20.5 psu in December at 88°E, 21°N) that increases sharply towards the center of the Bay (32.8 psu at 88°E, 15°N in December). Figures 2c,d show the differences of simulated SSS from SR and the WOA05 while Figures 2e,f show the difference between SR and the NoSR. Figures 2c,d illustrate that NoSR run produces too high a salinity both in the Bay and in the northern Indian Ocean in summer and winter, which is due to the absence of river discharge. In the Arabian Sea and the rest of the domain, NoSR produces generally lower salinity than that is observed. This seems to be related to the model errors in wind and rainfall.

When the SSS relaxation is imposed in the SR run, these errors in SSS are substantially reduced throughout the domain. The strong lateral salinity gradient is much better captured in SR run in both seasons and the difference in SR and NoSR clearly demonstrates the improvement in the salinity fields in the SR relative to NoSR. Note, however, there are patches of lower SSS (-2 psu) and higher SSS (+0.5 psu). These patches are seen as errors arising from restoring of salinity in the regions with large salinity gradient to the smoothed salinity products such as WOA05 (Yu and McCreary 2004). In such regions, SSS restoring is particularly sensitive to the chosen relaxation timescale. We have chosen a strong relaxation (τ=10 days), which, as is revealed in Figure 2, reproduces the spatial patterns of the observed salinity quite well, but can
sometimes overshoot the SSS differences and switch its sign in the regions of large salinity gradient.

The subsurface salinity distribution in BoB also improves in SR. Figure 3 shows the depth-latitude transects of the salinity along 90°E in BoB. The simulated salinity field in SR exhibits a strong halocline at the observed depth during the both seasons, which is missing in the NoSR case. Although SR run underestimates the isohaline sloping down toward the coast as reported in the observations (Shetye et al. 1996), the simulated salinity structure, both in the surface and the subsurface, produces far more realistic salinity distribution in SR than in NoSR case.

HM01 included the river discharges locally as surface freshwater flux at the mouth of rivers in their ocean model. This methodology is deemed more appropriate to incorporate the river discharge in the numerical model than our approach. Their simulated SSS pattern nevertheless does not completely remove the SSS errors and is somewhat similar to what is shown in Figures 2e-f, lending some justification on the use of a simple SSS relaxation. Note a study by Kurian and Vinayachandran (2007) well reproduced the climatological SSS patterns by prescribing the river discharge in their high-resolution OCGM. Their success appears to be due to the relatively higher horizontal and vertical resolution used in their model along with the realistic surface forcings. While our horizontal resolution is comparable to what is used in their study, the number of vertical layers used in our ocean model is only half of theirs and in fact closer to 19 vertical layers in HM01. Since the BL effects on SST and the definitions of the oceanic variables such as MLD and BL critically depend on the vertical processes and the resolutions of the model (Kara et al. 2000), we expect the results to be more similar with those from HM01, if the river discharge were imposed locally.

In summary, despite the limitation of the experimental setup, the comparison of E-P fluxes and SSS indicate that SSS relaxation method captures important aspects of Indian Ocean salinity variations reasonably well (e.g., seasonal cycle) and hence capable of offering useful insights on the role of the freshwater forcing in the large-scale atmospheric circulation.
4. Model Validation

Before we investigate the sensitivity of the system to surface freshening, this section discusses the seasonal mean biases displayed by the current model. Figure 4 show the plots of SST, 10-m wind, and precipitation from the observations (OBS), SR run and the difference or error (SR-OBS).

The model summer climatology in rainfall, wind and SST broadly resembles the observed patterns, with the Somali Jet directed from the southwest to northeast along the coast of Somalia, resulting in coastal filament and the Great Whirl (Schott and McCreary 2001, Seo et al. 2008). The southwesterly monsoon wind impinges on India and Indochina, with the windward rainfall maxima. From satellite observations, Xie et al. (2006) discussed the orographic lifting acting on the SW monsoon wind over coastal mountains. The model appears to mimic the observed pattern, but overestimates the magnitudes of the orographic rainfall. Excessive rainfall is found over the Indochina and the western Arabian Sea (see Table 1) and it requires further investigations to understand the origin of these errors in simulated rainfall.

Summertime warm bias errors are found in SST over the western Indian Ocean, the Arabian Sea, and the broad regions of the south Indian Ocean. As discussed in Seo et al. (2008), the summertime warm bias in the Arabian Sea is because the Somali Jet is significantly underestimated in the model, which appears to be linked to model errors in large-scale atmospheric wind (See their Figure 2). This results in weak coastal upwelling and the higher SST in the Arabian Sea and the Omani coast, which produce excessive rainfall during the SW monsoon. This pattern of summertime mean errors is common, to a varying degree, among the existing CGCMs (Annamalai et al. 2007, Bollasina and Nigam 2008).

The warm bias persists in the Arabian Sea throughout the year, which is associated with the model’s tendency to produce too deep thermocline depth all years (Figure 5). More importantly, in the southwestern Indian Ocean, the model only vaguely captures the salient feature of the observed oceanic thermocline dome. Note that the amplitude and
location of this thermocline dome is reasonably well simulated in many CGCMs, while some models still produce similar patterns as shown in Figure 6b (Saji et al. 2006). Vecchi and Harrison (2004) and Izumo et al. (2008) noted that anomalously warm SST over the thermocline dome and the western Arabian Sea can result in excess precipitation in the Western Ghats, indicating the importance of capturing the thermocline structure for precipitation simulation during the SW monsoon. The current model places its shallow thermocline too far west at 45°E, 5°S, which is forced by the cyclonic wind forcing. In the western Indian Ocean, excessively deep thermocline roughly coincides with the region of warm SST bias at 50°W, 5°S (Figure 4i) and the corresponding excess precipitation (Figure 4k). Thus the errors in mean SST and precipitation are dynamically consistent with those of subsurface thermocline structure.

Furthermore, the basin-scale pattern of the warm bias in the west and the cold bias in the eastern equatorial ocean and the BoB resemble that associated with the Indian Ocean Zonal Dipole Mode (IOZDM, Webster et al. 1999, Saji et al. 1999) which generates the Indian summer monsoon rainfall anomalies that are qualitatively similar to Figure 4k (Behera et al. 1999, Annamalai et al. 2003). This indicates that portion of the summertime model bias is linked to the mean state errors that bear some resemblance to the spatial patterns of IOZDM. All this suggests that the mean bias errors described so far can be viewed as a result of model’s errors in simulating realistic large-scale winds and the amplification by air-sea interaction. The overall amplitude of mean bias errors that is more pronounced during the boreal summer.

Over BoB, SST displays cold biases especially during the summer, possibly due to errors in surface wind and thermocline depth. Despite the cold bias, the model simulates too strong precipitation during summer over BoB except on the east coast of India, where the spurious upwelling occurs. Thermohaline forcing is strong in BoB and Sections 5 and 6 investigate salinity stratification on SST and the atmosphere.

Table 1 compares the simulated precipitation in two key regions of the largest bias with the observations for JJA and DJF: one along the Western Ghats and the other over the BoB and the Indochina. Seasonally, the summer rainfall bias increases slightly in the SR run compared to NoSR, while the winter rainfall bias displays a moderate improvement.
The difference in rainfall between SR and NoSR is smaller compared to their mean values in summer (~3-6%) than in winter (~11%). Weaker sensitivity of the model precipitation to the freshwater forcing in summer implies that other factors such as wind anomalies (Shukla 1987, Murtugudde and Busalacchi 1999, Murtugudde et al. 2007) and their feedback on to the SST (Wang et al. 2005) can be more important for the Indian summer monsoon. In the wintertime, the model error is smaller and the model exhibits greater sensitivity to the salinity. For the rest of the paper, we will discuss sensitivity of the coupled system in the Indian Ocean due to the freshwater forcings for both seasons, but the wintertime response will be more emphasized for such reasons.

The large rainfall biases are common in the state-of-the-art climate models (Annamalai et al. 2007). Because of the complicated interaction of ocean, atmosphere and land, such mean state errors limit the ability of models as a tool to study climate dynamics. The salinity difference between SR and NoSR experiments amounts to 3-4 psu over BoB (Figure 2), This is sufficiently large that our results about its climate effects are likely to hold despite mean biases. Further study is necessary as coupled models improve in the future.

5. Oceanic Changes

Influence of low salinity input on the oceanic vertical structure can be seen in Figures 6 and 7. The top (bottom) panel of Figure 6 shows the mean seasonal cycle of salinity and temperature as a function of depth in BoB from SR (SR-NoSR). Figure 7 shows the mean profiles of temperature ($T$), salinity ($S$) and the density ($\sigma_\theta$) averaged over the same region for summer and winter. The largest variation in salinity between SR and NoSR can be found in the upper 80 m of the ocean with the greatest reduction of salinity in SR greater than 4 psu at the surface from June to February (Figure 6c). Accordingly, a strong halocline develops in salinity profiles in SR at ~40 m, while the salinity profile is more or less uniform in NoSR. The enhanced upper ocean salinity gradient leads to a density stratification in SR, indicating that salinity plays a dominant role in determining upper ocean stratification and the MLD.
Difference plot in temperature (Figure 6d, Figures 7c,f) shows two interesting features. First, during the summer monsoon period, the vertical temperature distribution over the upper 60 m is by and large uniform with weak subsurface inversion, and the overall temperature change is small. Second, during the wintertime, the upper 20 m is cooled by up to 1.1°C while the subsurface ocean at depths between 20 m to 180 m become warmer, with maximum warming of 0.8°C occurring at the depth of 40 m. The implied subsurface inversion is much stronger in the winter than in summer.

Figures 6 and 7 clearly indicate that the salinity signals remain strong in winter. The difference in temperature profiles implies a strong inversion in winter at the depth of around 40 m. The summertime temperature profile difference (Figure 7c) show a subsurface inversion-like vertical distribution of temperature with the amplitude of 0.3°C. During the Bay of Bengal Monsoon Experiment (BOBMEX), Vinayachandran et al. (2002) observed such an inversion layer in the northern Bay in August with the maximum amplitude of 0.5°C at 30 m depth. This was attributed to the net surface cooling during cloudy days (Bhat et al. 2001) and some contribution from the Ekman transport of low saline water to the observation site. During JJA in the model, the net heat flux is slightly positive (warming the ocean) in both runs but total heat flux in SR is smaller than in NoSR (Figure 8a). The reduction in net heat flux in SR is partly attributable to the decreased downward shortwave radiation, in support of the observational study of Vinayachandran et al. (2002).

Figure 9 shows the seasonal cycle of the BL averaged over the BoB derived from the WOA05 and from the models. Also shown from the most recent estimate of the global BLT by DMLC07 derived from the NODC, WOCE, and ARGO databases of measured temperature and salinity profiles. In Figure 10 we also shows the spatial maps of the BLT from the WOA05 and the model simulation for the two seasons. In WOA05, BLs start to develop in the beginning of the summer monsoon season with the maximum thickness during the boreal winter. DMLC07 and Mignot et al. (2007) showed that BoB BL is a quasi-permanent feature, persisting for 10 months per year and reaching its maximum thickness in the boreal winter. The seasonal cycle of BL from the WOA05 is qualitatively consistent with their findings, yet the accurate timing of the maximum of the
BLT is in January from DMLC07, not in February from WOA05. BLT is also generally thinner throughout the year in DMLC07 than in WOA05. Note the estimate of BLs from DMLC07 is based on $\Delta T=0.2^\circ C$ criterion that is applied to the individual profiles (with higher vertical resolutions) before spatially mapped, while the BLs from WOA05 and from the model are based on spatially and temporally averaged data (Kara et al. 2003). So the difference in seasonal cycle and accuracy exists between two estimates. In this study we use the estimate from the WOA05 because the rather coarse vertical resolution in the model (20 vertical layers) favors comparison with the already smoothed fields in WOA05. Furthermore, the monthly WOA05 climatology is used for the lateral boundary conditions of the ocean model and the SSS is restored to the WOA05 fields.

In JJA, WOA05 shows the BL of 40-50 m thicknesses is spread along the coast of the Bay, with the maximum thickness of 50-60 m thicknesses along the eastern coast (Figure 10a). In DJF, BLT reaches 70 m in the northern Bay and the SEAS (Figure 10b). The SR run qualitatively captures key aspects of spatial patterns. Specifically, 20-40 m thickness of BL in BoB and Arabian Sea in JJA and thick BL centered in BoB and SEAS in DJF. However, the BLT in SR is generally underestimated throughout the domain compared to the WOA05 and some features of BL in the eastern equatorial ocean are not reproduced. In particular, the BLT in the SEAS during DJF is thicker and spatially more extensive towards the Arabian Sea in SR than in WOA05. The NoSR run essentially fails to reproduce the important features of the BL in the Indian Ocean, including the spatial distribution and its seasonal cycle.

b. Impact on summer SST

Modulation of vertical stratification by salinity and BL has an important implication on the ML temperature and SST. Figure 12a illustrates the spatial patterns of change in SST during the JJA season. The local increase in SST can be found in the northwestern coast of the Bay adjacent to the mouths of the major rivers, which warm up to 0.4$^\circ$C with the inclusion of river runoff. However the area extents of this increase in SST are limited to the coastal boundaries and hence when averaged over the entire Bay, the increase in SST
is rather small, 0.1°C, in this model. This is clearly not sufficient to offset the cold bias shown in Figure 5i and implies that the effect of river input on SST is weak during the summer and the cold bias in the model is more closely related to wind anomalies as discussed in Section 4.

Since the isothermal layer is deep and the net surface heat flux is weak over the summer BoB, the warming effect due to the weakened mixing is not pronounced. Furthermore, the surface heat flux acts as a damping on slight increase in SST. Figure 8a compares the seasonal mean net heat flux (positive warming ocean) between SR and NoSR, and Figure 8b shows the difference in net and each component of the net heat flux. The net heat flux in JJA in SR (NoSR) is 9 Wm$^{-2}$ (17 Wm$^{-2}$), indicating that heat flux warming is reduced in SR over the warm water. This is explained largely by greater evaporative cooling and increase in cloudiness that reduces the shortwave radiation reaching at the surface.

In the forced ocean model studies, Han and McCreary (2001) and Han et al. (2001) suggested a significant local increase in SST at the northwest corner of the Bay during the summer monsoon with the inclusion of river discharges. However our results suggest that the basin-average increase in SST is small. HM01 also showed that BoB SST indeed would warm with the inclusion of river influx into the BoB, but this increase is generally small and occurs only without the effect of the penetrative solar heat flux loss due to the thinning of the ML (Godfrey and Lindstrom 1989, Murtugudde et al. 2002). With the inclusion of penetrative shortwave heating effect, which is parameterized in the ROMS model (Large et al. 1994), SST generally decreases in their experiment. From the analysis of the buoy data in BoB, Sengupta and Ravichandran (2001) also showed that low salinity offsets the heat flux warming by allowing more sunlight to penetrate below the ML. Thus there is substantial modeling and observational evidences that low salinity due to the riverine input could increase the SST in BoB but the overall change is small due to the offset.

Regarding the SST changes in summer, the equatorial Indian Ocean deserves some attention. A large increase in SST is found at 55°E and the moderate cooling is located at 70°E, both at the equator. These changes in SST are similar in the spatial patterns of difference in SSS (Figure 2g) with the higher salinity at 50°E and the lower salinity at
70°E, indicating that the SSS restoring is partly responsible. These anomalies are also consistent with the changes in the thermocline depth (Figure 5c), which deepens at 50°E where it warms but shoals at 70°E where it cools. The change in equatorial SST bears little direct influence from the river discharges, and likely associated with the model mean errors in locating thermocline dome and the basin-wide SSS relaxation. The atmospheric response to the summertime SST pattern will be discussed in Section 6.

c. Impact on winter SST

The influence of salinity on upper ocean temperature is more striking in the wintertime when SST is cooled and the subsurface layer warms, hence forming a subsurface inversion. For these differences we consider two effects: the shallower ML and the wintertime surface cooling. Advection also can be an important element (Thadathil et al. 2002), but since we are averaging over the entire Bay, advection within the Bay could be a less decisive factor and the process reduces to a nearly 1-dimensional balance between the surface heat flux and oceanic vertical mixing (Rao and Sivakumar 2000).

Figure 12b illustrates the reduction in SST in our model in response to surface freshening. Large cooling occurs over the entire Bay and the SEAS, while the northwestern Arabian Sea becomes warmer. The maximum SST decrease in SR, up to 3°C, is found at the northernmost Bay. The area-averaged DJF cooling is 1.1°C in BoB and 0.3°C in SEAS. A comparison between the difference in BLT in Figure 10 and SST in Figures 12a,b indicates that the spatial patterns of changes in two quantities are much more coherent in winter than in summer, which again lends some support to the argument that salinity plays a more important role in regulating SSTs in winter in the current model.

Figures 11b,d show that the wintertime mean MLD over BoB is about 15-25 m. ML is shallower by 25 m in SR than in NoSR (Han et al. 2001, HM01). A shallow ML with the river flux limits the surface cooling effects within thin layer during the winter, resulting in SST decrease. Figure 8 showed that DJF climatological net heat flux in SR (NoSR) is -25 (-52) W m$^{-2}$, both providing a net surface cooling. The reduction of SST in SR occurs
despite the reduced net cooling (by reduced evaporative cooling over the cold water),
demonstrating that it is the oceanic process that causes the SST cooling.

6. Atmospheric response to SST changes

The previous section showed that lowering salinity via SSS relaxation induces small
change in local SST (where added freshwater forcing is large) during the summer, but in
winter it leads to a striking basin-wide cooling that is spatially coherent with the change
in SSS and BL. The simulated changes in surface winds and rainfall also exhibit a spatial
structure more coherent with the underlying SST in winter than in summer. This section
discusses the atmospheric response to the change in SST. Figures 12c,d show the changes
in rainfall and 10-m wind fields in JJA and DJF.

During the summer monsoon period, the westerly and the southwesterly winds tend to
become stronger towards the Western Ghats (WG) and in the northern BoB and the
Indochina in SR. The increased wind leads to the enhanced precipitation in these regions.
The increase in rain could be understood in part due to the weak increase in SST in the
northern Indian Ocean including BoB. Generally the amplitude and the area of significant
increase in SST are small and limited near the coast. There is a tendency for summertime
convection to be intensified over the BoB, but the change is weak in magnitudes. Thus
the direct connection between SST by BL and the summer monsoon rainfall in WG and
BoB is only vaguely suggested. Change in the summer monsoon precipitation and rainfall
is maybe related to significant changes in SST at the equator, rather than to the local BL
effects in the northern Indian Ocean. The warming (cooling) at 50°E (70°E) at the
equator is associated with the significant increase (decrease) in local precipitation and
changes in near-surface convergent (divergent) winds. In particular the surface divergent
flow from 70°E appears to contribute to the strengthened southwesterly, and the
increased orographic precipitation west of India.

In contrast, the atmospheric response in winter is more coherent with the significant and
spatially extensive reduction in SST over the northern Indian Ocean. The cooling of SST
shown in SR (Figure 12b) leads to the formation of near-surface divergent flow
originating from the BoB and the SEAS and amplifies the simulated NE monsoon wind (Figure 12d). This response in simulated wind is consistent with a linear baroclinic response of the troposphere (Gill 1980) to the diabatic cooling associated with the reduced SST. Note that local response (reduction) in precipitation over the northern Indian Ocean to the SST cooling is not remarkable because of lack of moisture during the winter monsoon period (Table 1). The intensified northeasterly wind however tends to reach far southward beyond the northern Indian Ocean, rendering a nonlocal response in the atmosphere. The intensified northeasterly wind leads to a surface divergence over the broad region at 5°N, causing a significant reduction in convection and rainfall of 1-2 mm day\(^{-1}\). Further to the south, it intensifies the ITCZ convection. Thus the large-scale atmospheric response to the changes in SST due to BL is to displace the ITCZ southward.

Note that CGCMs typically have large BL biases in the BoB and the northern Indian Ocean. In particular in the latest Community Climate System Model (version 3.5) the BL is rather weak (J. Mignot, pers. comm. 2008) and in DJF there is too much precipitation north and near the equator and too little south of the equator at 10-15°S (Jochum and Potemra, 2008). This suggests that in current GCMs, the weak BL physics, among other factors, could be partially responsible for the tropical precipitation biases. By analyzing the BLs in the several CGCMs, Breugem et al. (2008) hypothesized a positive BL-SST-ITCZ feedback that maintains models’ warm and fresh biases in the Southeastern Equatorial Atlantic Ocean. Within their feedback process, BL is an active and dynamic impetus for generating climate anomalies. The results from previous and present studies suggest, it is critical to capture the realistic BLs in the coupled climate models for the improvement in the simulation and prediction of regional climate variability including Indian monsoons.

7. Summary and discussion

We have used a high-resolution, regional coupled ocean-atmosphere model to investigate the effect of river water on the seasonal cycle in Indian Ocean climate. The model mean state exhibits deviations from observations, with biases in thermocline depth, SST, wind
and rainfall. These biases are somewhat similar in magnitude and spatial pattern to those in current CGCMs over the same area (see reviews of the monsoon biases in IPCC coupled GCMs in Annamalai et al. 2007 and Bollasina and Nigam 2008).

Our initial hypothesis that motivated this study is that the introduction of freshwater cap in the BoB would alleviate the summertime cold bias present in the model through the influence of the BL on the mixing SST. To our surprise, there is no significant change in summertime SST in the northern Indian Ocean despite more realistic oceanic vertical structure due to the freshwater forcing. There is no significant direct response in wind and rainfall during the southwest (SW) monsoon period to the local change in SST over BoB. However, we find a significant influence of the BL on local SST during boreal winter with a large atmospheric response in the northeast (NE) monsoon.

The main results are summarized as follows. During the SW monsoon period, a large freshwater influx from rivers via SSS relaxation generates a strong halocline in the BoB, leading to the formation of barrier layers (BLs), and suppressing the mixing between surface and subsurface water. With a reduced warming by surface heat flux and the persistent subsurface warming from the previous winter, difference in summertime temperature profile exhibits a weak inversion-like distribution. Because of the deep isothermal layer, the weakened mixing effect is not sufficient to further raise the SSTs. Overall, the increase in summer SST averaged over the Bay is 0.1°C (not significant at 90%), except for the coastal regions adjacent to the mouths of major rivers where SST warms up to 0.4°C. This is qualitatively consistent with the finding from Han et al. (2001) and HM01 who reported greater warming of 0.5-1.0°C. The SST warming in SR is despite the enhanced damping by the evaporative cooling effect and the reduced shortwave radiation flux, indicating that greater warming in forced ocean models underestimates negative atmospheric damping.

The SW monsoon wind somewhat strengthens in the Western Ghats and the northern BoB, causing rainfall increase in the northern Indian Ocean. Insignificant increase in SST in the northern Indian Ocean does not appear to be sufficient to result in these increases in wind and rainfall. Instead the divergent flows in the equatorial Indian Ocean in
response changes in SST there appears to exert a greater atmospheric influence over the northern Indian Ocean.

During the winter monsoon season, BLT attains its maximum in BoB, and the mixed layer (ML) consequently become shallower. The vertical trapping of the wintertime cooling efficiently cools the SST by an average of 1.1°C. SR run exhibits a strong subsurface inversion layer in BoB (and southeast Arabian Sea as discussed by Thadathil and Gosh 1992) with amplitudes of 0.8°C (Figures 6 and 7), consistent with observations.

The wintertime decrease in SST in the BoB results in nearly no changes in local rainfall and convection, because of the lack of moisture convection and cold SST. Instead, the reduced SST in the northern Indian Ocean amplifies the NE monsoon winds through a linear baroclinic response in the atmosphere. This results in a significant reduction of convection and rainfall around at 5°N, while enhancing it south of the equator. This indicates a possible connection of ocean stratification to the broader scale, nonlocal atmospheric circulation during the NE monsoon period.

Masson et al. (2005) suggested from CGCM experiment a link between the spatial structure of the salinity in the SEAS and the onset of the summer monsoon. This led them to hypothesize that the spatial extent of BL during the pre-monsoon season could be a useful predictor of the onset of the summer monsoon. Our model results differ in that BL had no significant impact on summer SST and the SW monsoon. Instead we find that the BL in BoB leaves its clear signature in the wintertime SST and subsequently induces large-scale adjustments in convection and rainfall across the equator during the winter monsoon. The importance of understanding the BL dynamics in BoB in winter is further supported by the fact that simulated monsoon response is not limited to the BoB, but extends far into the Southern Hemisphere.

Further studies are necessary to determine the detailed structure of the remote atmospheric response and assess the potential impacts on climate. There is a clear need for efforts in understanding the performances of coupled GCMs in reproducing the observed BLs (Kara et al. 2003, de Boyer Montégut et al. 2007) in the Indian Ocean. The results of Breugem et al. (2008) in the Atlantic suggest that errors in BL can cause the climate biases through multiple feedback processes in the coupled climate models (Lin et
al. 2007). A better assessment of impact of BLs on climate requires an inter-comparison exercise using the results from those climate models with and without the salinity and BL effects. This kind of studies will help determining the extent to which common biases in the climate models can be attributed to BL errors. It will also help detecting common and robust features on the sensitivity of the regional climate system to the BLs. Such research will greatly improve our understanding on the potential role of the salinity and BL to the large-scale climate variability including the Indian monsoon.

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regions, *Nature*, 312, 141-143.


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Table 1: Precipitation (mm day⁻¹) in Western Ghats (60-80°E, 5-20°N) and BoB and Indochina (80-105°E, 13-23°N)
Figure 1. Evaporation minus precipitation (E-P, cm day$^{-1}$) simulated from (top) SR, and (bottom) difference SR-NoSR averaged in (left) June-July-August (JJA) and (right) December-January-February (DJF) for 1993-2004.
Figure 2. Sea surface salinity (SSS, psu) averaged in (left) JJA and (right) DJF for 1993-2004. (top) the World Ocean Atlas 2005 (WOA05), (middle) difference in SSS between NoSR and WOA05, (bottom) SSS difference between SR and NoSR. The original domain covers 29°E-112°E, 30°S-38°N, but here the domain is shown excluding the lateral boundary sponge layer in east and south and the land points in west and north.
Figure 3. Depth-latitude diagrams of salinity along $90^\circ$E for (top) WOA05, (middle) NoSR, and (bottom) SR averaged in (left) JJA and (right) DJF from 1993-2004.
Figure 4. Thirteen-year mean (top) observations (OBS), (middle) model (SR run), and (bottom) SR-OBS. Left (right) two panels show SST [°C] (precipitation [mm day⁻¹] and 10-m wind [m s⁻¹]), with JJA (DJF) averages for first and third (second and fourth) columns. Observations are from the NOAA OI SST, the GPCP precipitation and the NCEP/DOE RA2 10- winds, all of which are averaged for the same time period as in the model. Reference vectors are shown in Figure (h) and (l). Wind vectors with wind speed less than 3 m s⁻¹ are not omitted in (k)-(l) for clarity.
Figure 5. Annual mean (1993-2004) thermocline depths from (a) WOA05 (b) SR, and (c) SR-NoSR. Thermocline depths are derived from the depth of the 20°C isotherms.
Figure 6. Time-depth diagrams of mean seasonal cycle of (a) SR salinity [psu] over the Bay of Bengal (80°-100°E, 10°N-25°N), (c) salinity difference (SR-NoSR), (b) SR temperature [°C] as in (a), and (d) temperature difference (SR-NoSR). The contour intervals in (a) and (c) are 1 psu and 0.5 psu, respectively. Values less than -0.5 psu are shaded in (c). The contour intervals for (b) and (d) are 1°C in and 0.2°C, respectively. The positive values are shaded in (d).
Figure 7. (top) JJA mean profiles of salinity ($S$, salinity-10 psu), temperature ($T$, °C) and density ($\sigma_\theta$) averaged over Bay of Bengal (80°-100°E, 10°N-25°N) from (a) SR, (b) NoSR, and (c) SR-NoSR. (bottom) Same as (top) except for DJF values. The quantities are averaged for 12 years (1993-2004). There are approximately 14 vertical layers in the upper 200 m (1.9, 5.7, 9.7, 14.0, 18.9, 24.7, 31.6, 40.1, 50.9, 64.9, 83.3, 107.7, 140.6, and 185.1 m).
Figure 8. (top) Surface net heat flux climatology [Wm$^{-2}$] from SR and NoSR as a function of seasons, and (bottom) difference (SR-NoSR) of the net heat flux and each component of net heat flux. A positive heat flux is into the ocean. Averaged over the BoB (80°-100°E, 10°N-25°N).
Figure 9. Seasonal cycle of BLT in BoB (80°-100°E, 10°N-25°N) estimated from the observations and the models. DMLC07 denotes de Boyer Montégut et al. (2007). BLT is calculated by the positive difference between the depth at which temperature becomes cooler than SST by $\Delta T = 1^\circ C$ and the depth at which salinity effects on density variations are equivalent to $\Delta T = 1^\circ C$ decrease of temperature.
Figure 10. Twelve-yr (1993-2004) averaged barrier layer thickness (BLT) estimated from (top) the WOA05 in (a) JJA and (b) DJF, (middle) from SR and (bottom) from NoSR. The contour intervals are 20 m and values greater than 20 m are shaded.
Figure 11. (top) Twelve-yr (1993-2004) averaged mixed layer depth (MLD) estimated from NoSR averaged for (a) JJA, and (b) DJF. (bottom) MLD differences (SR-NoSR). MLD is defined as the depth at which salinity effects on density variations are equivalent to $\Delta T=1^\circ$C decrease of temperature. The contour intervals are 10 m and the negative values (shoaling of ML) is shaded in gray in (c) and (d).
Figure 12. Changes (SR-NoSR) in (top) SST [°C] and (bottom) rainfall [mm day⁻¹] and 10-m wind speed [m s⁻¹] and vectors, averaged for (left) JJA and (right) DJF from 1993 to 2004. The overlaid black contours denote that the region where change is statistically significant at 90% confidence level. The reference vector is shown in the lower-right corner of the figure and applies to both the thin and thick vectors in (c) and (d).