Impact of Changjiang River Discharge on Sea Surface Temperature in the East China Sea

SHIN'ICHIRO KAKO AND TOMOFUMI NAKAGAWA
Graduate School of Science and Technology, Department of Ocean Civil Engineering, Kagoshima University, Kagoshima, Japan

KATSUMI TAKAYAMA, NAOKI HIROSE, AND ATSUHIKO ISOBE
Research Institute for Applied Mechanics, Kyushu University, Kasuga, Japan

(Manuscript received 26 August 2015, in final form 9 March 2016)

ABSTRACT

This study investigated how the Changjiang River discharge (CRD) emptying into the East China Sea (ECS) affects the upper-ocean stratification [hence, sea surface temperature (SST) changes], based on ocean general circulation modeling with and without CRD. A new finding in this study is that CRD contributes significantly to a reduction in summer SST in the ECS. Comparison between the two model runs revealed that vertical one-dimensional processes contribute considerably to SST warming in the ECS, while horizontal advection plays an important role in lowering SST in summer. The results of a particle-tracking experiment suggested that the cold water mass formed along the Chinese coast during the previous winter contributes to the SST reduction in the following summer. From the end of the summer monsoon season, the less saline CRD advected toward the Chinese coast generates a shallow mixed layer (ML), which inhibits heat exchange between the ML and thermocline. In winter, heat loss of the ML through the sea surface results in a reduction in SST over a broad region. Water exchange through the bottom of the ML is relatively suppressed by robust stratification, which prevents cooling of the thermocline and leads to a temperature inversion. The north-eastward ocean current associated with the summer monsoon carries the cold water mass in the ML across the ECS; therefore, SST decreases during the following season. These results suggest that CRD has a critical role on both the ocean circulation system and the coupled air–sea interactions in the ECS.

1. Introduction

It is well known that the Changjiang River (or Yangtze River) is the major source of freshwater entering the Yellow and East China Seas (YECS; e.g., Chang and Isobe 2003). Although many rivers discharge into the YECS, the Changjiang River discharge (CRD) accounts for 90% of the total (e.g., Shen et al. 1998; Chang and Isobe 2003; see Fig. 1). Figure 2 shows the annual variation of the CRD, as reported by Chang and Isobe (2003), which demonstrates that the CRD has a clear seasonal variation, that is, it is largest in July and smallest in January.

In addition to river discharge, precipitation and the Taiwan Warm Current are considered significant sources of freshwater for the YECS (Chang and Isobe 2003). Nevertheless, in summer, the freshening effect due to the Taiwan Warm Current is not greater than that of the CRD, and the freshwater influx from the CRD is nearly 4 times greater in magnitude than the average summer (May–August) influx from precipitation (Chang and Isobe 2003; see their Fig. 3). Thus, in summer, the Changjiang River is the dominant freshwater source for the YECS. In fact, results derived from both numerical experiments (e.g., Chang and Isobe 2003; Moon et al. 2009) and analyses using in situ observational data (e.g., Siswanto et al. 2008; Kim et al. 2009) have revealed that the summertime distribution of sea surface salinity (SSS) in the East China Sea (ECS) is affected significantly by the intrusion of the Changjiang diluted water.

The intrusion of the Changjiang diluted water near the surface induces a thin mixed layer (ML) above the thermocline, while a deeper isothermal layer (IL) that includes the ML and the pycnocline is also formed. The
difference between the IL and ML depths is defined as the depth of the barrier layer (BL). This layer isolates the warm ML water from the underlying cool water (e.g., Sprintall and Tomczak 1992). Thus, the less saline intrusion water near the surface plays an important role in determining the depth over which mixing effects are confined. Higher sea surface temperature (SST) is maintained by the enhancement of both the ML and the BL, which reduces the heat exchange between the warm ML water and the cool water deep below the ML, regardless of the season. Essentially, the freshwater-induced ML and the BL predominantly retain the effects of atmospheric forcing within the ML (Vialard and Delecluse 1998; Seo et al. 2009).

There have been no satisfactory explanations of how CRD affects the distribution and variation of SST in the ECS because of the lack of high-resolution spatiotemporal measurements of temperature and salinity. Delcroix and Murtugudde (2002) showed that the CRD induces SST warming in August, although this is contrary to the findings of other studies. They demonstrated that the northward advection of warmer water is the main contributor to this surface warming and that there is no SST warming related to the BL, because BL enhancement is prevented by the intrusion of a water mass with different salinity and temperature. Chen et al. (2006) suggested that the subsurface warming observed off the Chinese coasts in autumn could be associated with BL formation, resulting from the strong halocline associated with CRD that reduces heat exchange between the ML and the layer below the ML. However, few studies have addressed the ML and BL processes in summer when the distribution of Changjiang diluted water is broadest. In a recent study based on ocean general circulation model experiments, Park et al. (2011) demonstrated that CRD induces a BL in the ECS and that the CRD-induced BL contributes to the increase of SST in the ECS in August.

Despite the importance of CRD on SST and the circulation system of the ECS, the role of CRD in the determination of upper-ocean stratification (i.e., SST) has been investigated in only a few studies; consequently, agreement concerning the relationship has yet to be reached. Therefore, based on high-resolution ocean general circulation model experiments, this article discusses the effects of CRD on the SST distribution and the
upper-ocean thermal structure of the ECS from the viewpoint of the ML and BL processes induced by CRD.

### 2. Model and data

#### a. Observational data

The long-term averaged monthly SSS dataset from May to August, obtained from the Japan Oceanographic Data Center (JODC), was constructed for comparison with the monthly averaged model results. The individual measurements of SSS, derived from the JODC, were averaged over 1° × 1° boxes using a simple averaging method. The JODC dataset is freely available from the JODC website (http://www.jodc.go.jp/index.html). The climatological values are depicted in Figs. 3a–d.

To validate the accuracy of the modeled SST, we used the Multiscale Ultrahigh Resolution Sea Surface Temperature (MURSST) dataset provided by the Jet Propulsion Laboratory (http://www.jpl.nasa.gov). This is a multisatellite-derived SST dataset constructed using various techniques, as described in Chin et al. (2010). To construct the MURSST dataset, satellite SST data derived from the Moderate Resolution Imaging Spectroradiometer (MODIS) Terra, MODIS Aqua, the Advanced Very High Resolution Radiometer, and Aqua Advanced Microwave Scanning Radiometer for Earth Observing System are used. The MURSST dataset comprises daily mean SSTs produced on a grid with resolution of about 0.011° × 0.011°, and it is available for the period 18 June 2002 to the present. This daily dataset from 2003 to 2010 was interpolated linearly to the model’s spatial grid and used for comparison with the modeled SSTs.

#### b. Hydrodynamic model

The hydrodynamic model used in this study was the same as that of Hirose et al. (2013), constructed based on the Research Institute for Applied Mechanics Ocean model (Lee et al. 2003) to develop assimilation and prediction systems for the East Asian marginal seas and named the Data Assimilation Research of the East Asian Marine System (DREAMS) model. It should be noted that although data assimilation techniques such as Kalman filtering were not applied in the numerical experiments of this study, the prediction system model used in this study is hereinafter referred to as DREAMS. Because the DREAMS model used in this study is the same as that employed in Hirose (2011) and Hirose et al. (2013), only a brief description of it is provided here.
The DREAMS model comprises a large-scale parent model (DR_B) and a nested fine-resolution child model (DR_M; Fig. 1), which were developed by Hirose (2011) and Hirose et al. (2013), respectively. The DR_B model has a grid spacing of $\frac{1}{4}^\circ$ longitude and $\frac{1}{8}^\circ$ latitude but with the same vertical resolution as the DR_M model. The DR_M model covers only the East Asian marginal seas with a grid spacing of $\frac{1}{12}^\circ$ longitude and $\frac{1}{12}^\circ$ latitude but with the same vertical resolution. In these two models, surface values such as SST and currents are estimated using linear extrapolation methods.

The DR_M computation was conducted for the period January 1981 to December 2010 (30 years). The DR_B model outputs at the end of 1980 were linearly interpolated to the DR_M spatial and vertical model grid and used as initial conditions. The Japanese 25-year Reanalysis dataset (JRA-25; Onogi et al. 2007) from 1981 to 2007 and the Grid Point Value (GPV) meteorological data after 2007 derived from the Japan Meteorological Agency (http://database.rish.kyoto-u.ac.jp/arch/jmadata/data/gpv/original/) were used as the surface meteorological conditions in the DR_M computation. Bulk formulas (Kondo 1975; Hirose et al. 1999) were used in conjunction with the above reanalysis products to estimate the fluxes of surface heat, freshwater, and momentum. In the DREAMS computation, to adjust the modeled temperature and salinity to the observations, the magnitude of total wind stress was reduced by 18%, in line with a preliminary optimization proposed by Hirose (2011). The validity of this reduction was evaluated by comparison of the variation and volume transport of the Kuroshio between the model and observations. Because the physical reasoning for the wind stress reduction has been addressed in Hirose (2011), we do not discuss it further here. Solar insolation was concentrated to a 6-h period each day to approximate diurnal variation. In the present application, daily mean shortwave radiation data were increased fourfold over 6 h during the daytime (Hirose 2011). The tidal phase and amplitude derived from the global ocean tide model NAO.99b (Matsumoto et al. 2000) were added at the open boundary of DR_M after optimization using the model Green functions proposed by Moon et al. (2012).

The monthly mean river discharge derived from the Chinese Ministry of Water Resources (MWR 2011) was used for the Changjiang River. In our model, the Changjiang diluted water was discharged into the uppermost layer in the upstream region of the Changjiang River with 0 psu. The Changjiang diluted water temperature was assumed to be the same as the air temperature derived from the reanalysis data at the nearest grid point because there were no observational data and thus no way of knowing the real Changjiang diluted water temperature (this assumption is justified in section 3). Although there is a method that sets the river temperature to that of the SST at the nearest grid point to the river mouth, this adjustment makes the river water too warm in reproducing the CRD that includes meltwater in spring. To estimate the land water discharge from other rivers such as the Huanghe River (Yellow River), we used the parameterization based on coastal precipitation within 78 km landward from the coastline (Hirose 2011). Note that DREAMS is a forced ocean model in which the SST was computed using the prescribed fluxes of momentum, heat, and freshwater based on the bulk formulas described above.

To assess the importance of CRD in determining SST distribution, two simple model experiments were performed using DREAMS: one with and the other without CRD. Hereinafter, for convenience, the model run with CRD is called the control run and the model run without CRD is called the nonriver run.

3. Results

a. Comparison of modeled SSS/SST with observations on a seasonal time scale

To validate the accuracy of the DREAMS output, we compared the model results with observational data. Note that only the monthly data were used for the seasonal-scale comparison. Figure 3 shows the horizontal distributions of climatological monthly SSS derived from the JODC data and from DREAMS; the areas with climatological values derived from the JODC data do not extend close enough to the Chinese coast because of the lack of observational data. The outer boundaries of the Changjiang diluted water can be identified by the 32-psu isohaline (Su and Weang 1994). Figure 3 demonstrates that the modeled SSS is higher than that from the JODC data over the ECS. A possible cause of this overestimation is related to the underestimation of climatological monthly SSS derived from the JODC data. Meanwhile, a comparison of the simulated SSS with reanalysis products, such as the Hybrid Coordinate Ocean Model and Simple Ocean Data Assimilation data, is also difficult because of the lack of observational data used in the data assimilation. Therefore, in this analysis, the horizontal distribution of SSS and its behavior are both discussed only in a qualitative sense.
The horizontal distribution of modeled SSS is qualitatively similar to that derived from the JODC. For example, a plume with salinity of $<32$ psu near the Changjiang River mouth in May gradually extends eastward with time, finally reaching the boundary at Jeju and Tsushima islands in August. These results are consistent with those of previous studies (e.g., Chang and Isobe 2003). The seasonal variability exhibited by the modeled SSS is considered in good agreement with that derived from the JODC and presented in previous studies.

Here, the results of the comparison between the modeled and the satellite-derived SSTs during the period from January 2003 to December 2010 are shown. Figure 4a shows the averaged difference between two model runs (overestimation by the model is indicated by the positive values), and it indicates remarkable negative differences to the north of the Kuroshio, although the differences in the Changjiang River mouth and over the entire area of the continental shelf are relatively small (i.e., $<1^\circ$C). The small biases around the Changjiang River mouth and the continental shelf indicate that the CRD temperature data used in this study (air temperature derived from reanalysis data; see section 2b) are relatively reasonable. The large underestimation to the north of the Kuroshio suggests that the heat losses to the atmosphere, derived from the modeled SST and reanalysis data using the bulk formulas (see section 2), have been overestimated in this region. Explicitly, turbulent heat flux and the entrainment velocity might be overestimated in the DREAMS model.

However, an underestimation of SST by about $3^\circ$C over the broad area north of the Kuroshio is unlikely to be an essential flaw in the present study. This is because the region of large bias is to the east of our main feature of interest, and the effect of the CRD is evaluated by the difference between the SSTs obtained in the two model experiments described in section 2b. It was found that the root-mean-square (RMS) differences are $<1.0^\circ$C, except for the coastal regions (Fig. 4b), and that the correlation coefficients are significant at a 95% confidence level over the ECS (Fig. 4c). Thus, the seasonal variations and their amplitudes in the MURSST are well reproduced over the ECS by the DREAMS model.

b. Effect of the CRD on the distribution and variation of SST in the ECS

To assess the importance of CRD in determining SST distribution, comparisons between the two simple model experiments described in section 2b were performed. In the nonriver run, the eastward extension of the low-salinity area from May to August, observed in the SSS field derived from the JODC and in the control run (Fig. 3), is missing from the SSS distributions (Fig. 5). This result also indicates that CRD plays an important role in the determination of SSS distributions in summer.

Figure 6 shows the difference in SST from May to August between the control and the nonriver runs. Positive values indicate that the SST from the control run is higher than that from the nonriver run. The regions where the difference is significant at the 90% confidence level are indicated by the black contours. A significant increase in SST when CRD is included in the model can be seen near the mouth of the Changjiang River from May to August and along the Chinese coast from July to August. The maximal SST increase of $0.5^\circ$C in August is in close agreement with Park et al. (2011); they focused on the CRD effect only in August. However, our new
finding is that SST is lowered significantly by the cold CRD in the central region of the ECS from May to August.

To investigate the possible causes of the increase/decrease of SST in the control run, we computed the SSS difference and the difference of net heat flux in the ML (i.e., surface thermal forcing) between the two model runs (Fig. 7). The ML heat budget can be expressed as follows (Qiu and Kelly 1993; Kako and Kubota 2009):

\[
\frac{\partial T_m}{\partial t} = -\mathbf{U} \cdot \nabla T_m + \frac{q_{net} - q_d}{\rho_0 c h_m} + \text{residual,} 
\]

where \( T_m \) and \( h_m \) denote ML temperature and depth, respectively; \( q_{net} \) is the net heat flux from the atmosphere to the ocean; and \( \rho_0 \) and \( q_d \) are the ML density and the downward radiative flux at the bottom of the ML, respectively. Vector \( \mathbf{U} = (u, v) \) is the current vector in the ML. The specific heat of seawater is denoted by \( c \) (\( =3990 \text{ kg}^{-1} \text{ J}^{-1} \)), and the third term on the right-hand side of the above equation represents the residual heat flux that includes other factors such as latent heat and sensible heat.
side denotes the residual including diffusion and entrainment velocity. The second term on the right-hand side is the surface thermal forcing, which was computed in each model run. To compute the ML depth in Eq. (1), the values of temperature $T$ and salinity $S$ (i.e., $\sigma_0$), derived from the two model experiments, were linearly interpolated to values at intervals of 1 m depth.

In Fig. 7, positive values indicate that the SSS/surface thermal forcing derived from the control run is higher/greater than from the nonriver run. Hereinafter, the ML (IL) depth is defined as the depth at which $\sigma_0 (T)$ increases (decreases) by 0.125$\sigma_0 (1.0^\circ C)$ from the sea surface. In the control run, a decrease in SSS over the ECS from May to August can be found (Figs. 7a–d). This indicates that the surface stratification in the control run is enhanced by the Changjiang diluted water over the ECS. In fact, this enhanced stratification contributes to the greater net heat flux in the ML in the control run (Figs. 7e–h). Thus, although it might be expected that the one-dimensional ML and BL processes mentioned...
in the introduction would cause the increase in the SST, the actual reduction in the SST cannot be explained by such processes.

The vertical profiles of climatological $T$ and $S$ derived from both model experiments are shown in Fig. 8a, while those of climatological $T$ and density $\sigma_T$ near the Changjiang River mouth (30°N, 123°E) in August are shown in Fig. 8b. These figures demonstrate that the density profile in the control run is determined mainly by the $S$ profile, and that CRD contributes to the shallower ML formation by inducing a strong density difference between the bottom of the ML and the layer below the ML. The pronounced Changjiang diluted water in the control run induces both the ML (4 m) and the IL (17 m), which results in the formation of the BL. Meanwhile, the difference between the depths of the ML (12 m) and IL (17 m) in the nonriver run is thinner compared with the control run. Although the vertical resolution of DREAMS might be somewhat coarse to reproduce the BL processes precisely, it can be proposed that the enhanced stratification due to the Changjiang diluted water prevents heat exchange between the ML and the layer below the ML in the control run. As a result, the formation of the ML and BL induced by the CRD near the Changjiang River mouth contributes to SST warming in this region. These results are relatively consistent with Park et al. (2011).

Meanwhile, the cooling of SST due to the Changjiang diluted water in the control run over a large area of the ECS cannot be explained by such one-dimensional processes, because heat input into the ML in summer in the control run is greater than in the nonriver run, which is accompanied by significant surface freshening due to the Changjiang diluted water (Fig. 7). Thus, in the control run, it is expected that the relatively cold water mass, compared with the nonriver run, is advected into the ECS from another region.

c. Possible causes of SST cooling in summer

To identify the source region of the cold water mass, an experiment using a backward particle-tracking model (PTM) was conducted. The PTM used by Isobe et al. (2009) was also used in this study. Ambient surface currents carry the modeled particles in the PTM, computed in the control run, in conjunction with a random walk process. The description and the capability of the PTM in
the present study have already been presented in Isobe et al. (2009) and Kako et al. (2010) and will not be discussed further here.

In the backward-in-time PTM experiments, a thousand particles were placed at each of the 10 points in the center of the ECS (Fig. 9a) on 15 July, that is, when the relatively cold water was distributed in the center of the ECS (Fig. 6). The directions of modeled surface currents are reversed in sign for both horizontal current components, and these currents carry the modeled particles. Vertical motion was not considered in the PTM, because our focus was on the behavior of the water mass related to CRD (i.e., the motion of very light water associated with the Changjiang diluted water). In this experiment, climatological daily mean values were used. Figures 9b–d show the modeled particle distribution from May to January (2–6 months previously). The percentages of the numbers of modeled particles in each model grid to the total number of modeled particles (10,000) computed from the backward-in-time PTM are indicated by colors. Note that these percentages (showing the seawater source) do not necessarily relate directly to the rate of contribution to the reduction of SST in summer in the ECS. It is demonstrated that most particles were positioned near the Chinese coast (26°–32°N, 120°–123°E) on 15 January; 80% of the modeled particles reached this region. This result might indicate that the relatively cold water mass was transported into the ECS from the Chinese coast, suggesting that the cold water mass was formed along the Chinese coast in winter.

To investigate the effect of CRD on the SSS/SST along the Chinese coast in winter, the SSS/SST distributions derived from the two model runs in January were compared. Figures 10a and 10b show the horizontal distributions of climatological SSS in both model runs, which demonstrate that a narrow lower-salinity band is formed along the Chinese coasts in the control run, while the southwestward alongshore extension of less saline water disappears in the nonriver run. In contrast to the summer season, the less saline water related to the CRD flows southwestward along the Chinese coast in winter and its behavior is associated with the northeasterly monsoonal winds (Chang and Isobe 2003). This computed SSS pattern in the control run is in good agreement with the results of previous studies (e.g., Chang and Isobe 2003), and thus, it is expected that CRD would contribute to upper-ocean stratification along the Chinese coast in winter. Figure 10c shows the difference in SST between the two model runs in January, and significant SST reduction due to surface freshening in the control run can be seen along the Chinese coast. Considerable cooling occurs along the Chinese coast, and the maximum SST difference between the two model runs reaches about 3.0°C. This indicates that CRD contributes significantly to the reduction of SST in winter along the Chinese coast.

Figures 11 and 12 illustrate the influence of the Changjiang diluted water on the upper-ocean stratification. Figure 11 shows the monthly mean seasonal variations of salinity and temperature against depth at 28°N, 122°E in the two model runs. Figure 12 shows the monthly mean seasonal variation of each term in the ML salinity budget in the two model runs at the same location. The ML salinity budget can be computed as follows (Kako and Kubota 2009):

\[
\frac{\partial S_m}{\partial t} = -\mathbf{U} \cdot \nabla S_m + \frac{(E - P)}{h_m} S_m + \text{residual},
\]

where \(S_m\) is ML salinity; \(h_m\) is ML depth; evaporation and precipitation are denoted by \(E\) and \(P\), respectively; and the third term on the right-hand side denotes the residual, including diffusion and entrainment velocity. We computed the climatological value of each term using data derived from JRA-25, GPV, and the two model runs.
In the control run, at the onset of the northeasterly monsoon from late summer, the intrusion of less saline water (i.e., Changjiang diluted water) in the upper-20-m layer can be found (Fig. 11a). In fact, Fig. 12 demonstrates that the salinity tendency term changes from positive to negative at the end of the summer monsoon season, in accordance with the horizontal salinity advection in the control run, whereas the salinity tendency is nearly equal to zero in the nonriver run in this month because of the lesser effect of horizontal salinity advection. Therefore, a strong halocline is formed in the salinity profile of the control run, especially after the end of summer monsoon season, while this salinity profile disappears in the nonriver run (Fig. 11). Thus, the determination of the upper-ocean

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**Fig. 9.** Results of backward-in-time PTM experiment: (a) initial positions of modeled particles (15 July), (b) 2 months previously (15 May), (c) 4 months previously (15 March), and (d) 6 months previously (15 January). Colors indicate the percentages of the numbers of modeled particles in each model grid to the total number of modeled particles (10,000). Number of particles > 0.2% is indicated by the same color (red).
stratification in winter along the Chinese coast is largely affected by the less saline water related to the Changjiang diluted water.

Although horizontal salinity advection in winter acts to increase the salinity tendency term (Fig. 12a) related to the decreased amount of CRD (Fig. 2), the effect of lower salinity associated with horizontal advection in August and September is continued through to about March. For example, the integrated value of horizontal salinity advection in August and September is $-18.1 \times 10^{-7} \text{ psu s}^{-1}$, whereas from October to March it is $19.9 \times 10^{-7} \text{ psu s}^{-1}$.

In fact, the existence of less saline water in the control run in this region in winter can be found in Fig. 10, and it contributes to the formation of enhanced stratification at the bottom of the ML in the control run, compared with that in the nonriver run (see Figs. 11, 13a).

It is worth noting that the temperature inversion layer is induced by the CRD. Thus, a vertical thermal structure with cooler ML water, in comparison with the water below the ML, is enhanced in association with the intrusion.
of the Changjiang diluted water. In Fig. 11, the formation of the temperature inversion layer can be found from the end of the summer monsoon season, and it can also be found that the temperature difference between the sea surface and the depth of 50 m (i.e., the sea floor in this region) eventually reaches a maximum of 2.5°C (Fig. 11b) from January to February. Figure 13a also demonstrates that the temperature inversion layer comprises less saline cold water in the ML and relatively warm saline water below the ML. The formation regions of the temperature inversion layer in the control run are shown in Fig. 13b. Figure 13b demonstrates that a strong temperature inversion layer is formed along the southeastern Chinese coast. This result is consistent with that derived from the analyses in Hao et al. (2010). Using historical data from 1930 through 2003, they indicated the formation of a temperature inversion layer near the southeastern Chinese coast during October–May.

The temperature inversion layer in winter is formed by the same mechanism as in summer, where the intrusion of less saline water associated with the Changjiang diluted water leads to SST warming near the mouth of the Changjiang River in summer. In winter, although the ML temperature is decreased by heat losses from the ocean to the atmosphere and monsoon wind, the increase of the ML density, associated with the temperature decrease, is negated by the effect of the less saline water such that vertical static instability cannot be generated. Thus, the enhanced stratification generated by the CRD at the ML bottom inhibits heat exchange between the ML and the layer below the ML. As a result, since the effects of heat losses and wind mixing are limited to within the ML, the layer below the ML is less subject to the effects of surface cooling. Consequently, cooling of the ML is accelerated and a temperature inversion layer is formed.

From the results discussed in this section, it can be considered that the cold water mass, formed along the Chinese coast in winter by the effects of CRD, is advected toward the northeast by ocean surface currents associated with the summer monsoon, and that it contributes to the determination of the SST distribution in the ECS. Therefore, the cold water mass formed along the Chinese coast in winter, associated with the intrusion of Changjiang diluted water, plays an important role in the reduction of summertime SST in the ECS.

4. Summary and discussion

This study investigated the effects of Changjiang diluted water on the distribution and variation of SST in
summer in the ECS using the DREAMS high-resolution ocean general circulation model. The model results were established as being in reasonable agreement with in situ and satellite observations, and thus a comparison of two DREAMS simulations with and without the CRD was undertaken.

The initial hypothesis that motivated this study was that the intrusion of Changjiang diluted water into the surface layer of the ECS in summer would contribute to SST warming through the influence of the pronounced salinity-induced ML, which inhibits heat exchange between the ML and the layer below the ML. Although, in winter, an upward heat loss of the ML through the sea surface results in the reduction of SST across a broad region, the water exchange through the bottom of the ML is relatively suppressed by the robust stratification. Consequently, since the layer below the ML is less subject to the effects of surface cooling, a temperature inversion layer is formed. During the southwesterly monsoon season, a cold water mass, generated within the temperature inversion layer along the Chinese coast, is advected into the ECS, and it contributes to the reduction of summertime SST over a large area of the ECS.

As mentioned in the introduction, the robust stratification associated with the CRD-induced ML and BL mostly confines the effects of the momentum and heat fluxes to within the shallow ML. Thus, it is to be expected that the CRD-induced ML and BL would affect not only the upper-ocean stratification but also the ocean surface currents. To justify this hypothesis, we compared the ocean surface currents between the two model runs. Note that the ML depth in the control run is shallower than that in the nonriver run because of the intrusion of less saline water in the surface (Fig. 7).

Figures 14a and 14b show the differences in ocean surface currents between the two model runs in January and June. Positive values indicate that the ocean surface current in the control run is faster than in the nonriver run. The overlaid black vectors were generated using the differences between the zonal and the meridional ocean surface current components of the two model runs. Thus, the differences in the ocean surface current between the two model runs are induced by the predominant ocean surface current in the direction of the arrow in the control run.

In winter, significant enhancement of the ocean surface current in the control run can be seen from the Changjiang River mouth, southwest along the Chinese coast. Along the Chinese coast in the control run, in addition to the influence of the ML and BL, the large horizontal density gradient associated with the intrusion of the Changjiang diluted water contributes to the enhancement of the southwestward ocean surface current. Moreover, it seems likely that part of the northeastward Taiwan Warm Current might be weakened by the southwestward current associated with this horizontal density gradient (Fig. 14a). In summer, the difference between the ocean surface currents becomes negative in the center of the ECS (Fig. 14b), as does the difference of SST shown in Fig. 6, while it remains positive around the Changjiang River mouth. The enhancement of the ocean surface current around the...
river mouth could be explained by the influence of the BL, ML, and CRD, as mentioned above.

These results indicate that the CRD plays important roles in determining the seasonal variation of both ocean surface currents and upper-ocean stratification. However, the possible causes of the negative differences in the region north of Taiwan and the continental shelf remain unknown. Although we do not investigate

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**FIG. 13.** (a) Vertical profiles of temperature and salinity derived from the two model runs at 28°N, 122°E in January. Horizontal dashed line indicates the bottom of the mixed layer for the control run. (b) Horizontal distribution of temperature difference between the surface and the bottom in the control run (surface to bottom). Negative temperature differences are shown by color shading, and those lower than −0.5°C are contoured.

**FIG. 14.** Horizontal distribution of the ocean surface current difference between the two model runs in (a) January and (b) June. Regions where the difference is statistically significant at the 90% confidence level are shown by color shading. The overlaid black vectors were generated using the differences between the zonal and the meridional ocean surface current components of the two model runs.
this mechanism here, an in-depth examination of the influence of the Changjiang diluted water on ocean surface currents, using high-resolution ocean general circulation models, should be conducted in future research.

Some previous studies have demonstrated that the CRD-induced ML and BL contribute to SST warming in the ECS in summer. However, this study found that the intrusion of less saline water along the Chinese coast in winter has a considerable effect on the reduction of SST in the ECS in the following summer. This is the first work to recognize the impact of CRD on the reduction of SST in the ECS in summer. Through air–sea interactions, the variation of SST due to the intrusion of Changjiang diluted water could affect the regional climate. For example, enhancement/reduction of baroclinicity induced by the increase/decrease of SST related to the intrusion of Changjiang diluted water might affect the enhancement/reduction of the strength of extratropical cyclones and typhoons, although the degree of the contribution of the CRD remains unknown. A twin-model experiment (i.e., with and without CRD) that allows air–sea interactions would greatly improve our understanding of the potential role of CRD on climatic variability.

Acknowledgments. This work was supported by the Japan Society for the Promotion of Science KAKENHI Grant 15K17763. We are grateful to the Japan Oceanographic Data Center (JODC), Jet Propulsion Laboratory (JPL), and Japan Meteorological Agency (JMA) for providing the in situ observations, satellite observations, and data from the Japan Ocean Polar Research Institute. We are grateful to the Japan Oceanographic Data Center for the data on SST, SSH, and TOM. We are also grateful to the Japan Ocean Polar Research Institute, the Japan Oceanographic Data Center, and the Japan Marine Inspection Agency for the data on SST and SSH. We are grateful to the China Oceanographic Data Center for the data on SST and SSH. We are also grateful to the Japan Oceanographic Data Center and the Japan Ocean Polar Research Institute for the data on SST and SSH. We are also grateful to the Japan Oceanographic Data Center and the Japan Ocean Polar Research Institute for the data on SST and SSH. We are also grateful to the Japan Oceanographic Data Center and the Japan Ocean Polar Research Institute for the data on SST and SSH.

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