Upper-Ocean Response to the Super Tropical Cyclone Phailin (2013) over the Freshwater Region of the Bay of Bengal

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ABSTRACT

This study investigates the impact of salinity stratification on the upper-ocean response to a category 5 tropical cyclone, Phailin, that crossed the northern Bay of Bengal (BOB) from 8 to 13 October 2013. A drastic increase of up to 5.0 psu in sea surface salinity (SSS) was observed after Phailin's passage, whereas a weak drop of below 0.5°C was observed in sea surface temperature (SST). Rightward biases were apparent in surface current and SSS but not evident in SST. Phailin-induced SST variations can be divided into the warming and cooling stages, corresponding to the existence of the thick barrier layer (BL) and temperature inversion before and erosion after Phailin's passage, respectively. During the warming stage, SST increased due to strong entrainment of warmer water from the BL, which overcame the cooling induced by surface heat fluxes and horizontal advection. During the cooling stage, the entrainment and upwelling dominated the SST decrease. The preexistence of the BL, which reduced entrainment cooling by $\sim 1.09^{\circ}\text{C day}^{-1}$, significantly weakened the overall Phailin-induced SST cooling. The Hybrid Coordinate Ocean Model (HYCOM) experiments confirm the crucial roles of entrainment and upwelling in the Phailin-induced dramatic SSS increase and weak SST decrease. Analyses of upper-ocean stratification associated with 16 super TCs that occurred in the BOB during 1980–2015 show that intensifications of 13 TCs were associated with a thick isothermal layer, and 5 out of the 13 were associated with a thick BL. The calculation of TC intensity with and without considering subsurface temperature demonstrates the importance of large upper-ocean heat storage in TC growth.

1. Introduction

A tropical cyclone (TC), referred to as a typhoon in the Pacific or a hurricane in the Atlantic, is an energetic weather system that originates over the warm waters of tropical oceans. It is one of the most destructive natural disasters and causes severe death and property loss in coastal areas around the world (Frank and Husain 1971;

Emanuel 2003). The passage of a TC has a large impact on various aspects of the upper ocean. Among them, TC-induced cooling has received much attention (Leipper 1967; Price 1981; Shang et al. 2008; Yan et al. 2017). Observations show that the magnitude of sea surface temperature (SST) cooling caused by TCs ranges from 1° to 11°C, depending strongly on the prestorm mixed layer depth (MLD) and upper-ocean stratification, in addition to the individual TC's characteristics, such as intensity and translation speed (Sakaida et al. 1998; Shang et al. 2008; Lin et al. 2009). The instantaneous SST cooling provides a negative feedback to TC strength and helps to reduce TC intensity (Vincent et al. 2014). Therefore, a careful characterization of the TC-induced SST cooling may contribute to improving our understanding and prediction of TC intensity.

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In some regions of the tropical oceans, abundant freshwater input by rainfall and river discharge near the surface enhances the upper-ocean stratification, thins the surface mixed layer (ML) and forms a barrier layer (BL) underneath. The BL is located within the isothermal layer (IL), between the depths of the ML above and the thermocline below (e.g., Fig. 1 in [de Boyer Montégut et al. 2007](#)). As such, the BL acts as a barrier to the turbulent entrainment of the cold thermocline water into the ML (e.g., [Lukas and Lindstrom 1991](#); [Sprintall and Tomczak 1992](#)), and it thus affects SST variability (e.g., [Han et al. 2001](#); [Cai et al. 2009](#); [Drushka et al. 2014](#)). A few recent studies have reported a significant role of the BL in weakening the TC-induced cooling in the western Pacific, the Amazon region in the Atlantic, and the Bay of Bengal (BOB) of the Indian Ocean (e.g., [Wang et al. 2011](#); [Grotsky et al. 2012](#); [Wang and Han 2014](#)). Consequently, the existence of a BL is favorable for TC intensification ([Wang et al. 2011](#)). [Balaguru et al. \(2012\)](#) found that TC intensification is up to $\sim 50\%$ higher in the presence of a BL. However, some recent studies demonstrated that the BL has little impact on SST and TC intensification ([Newinger and Toumi 2015](#); [Hernandez et al. 2016](#)). These controversial results are due to the complicated processes through which the BL affects TCs, and which depend on both the individual TC's characteristics and the upper-ocean stratification ([Yan et al. 2017](#)).

The upper-ocean response to TC forcing in the BOB has been well documented in many previous studies (e.g., [Murty 1983](#); [Gopalakrishna et al. 1993](#); [Murty et al. 1996](#); [Behera et al. 1998](#); [Chinthalu et al. 2001](#); [Wang et al. 2012a,b](#); [Vissa et al. 2013](#); [Li et al. 2013](#)). These studies noted that in regions of weaker upper-ocean stratification (e.g., the southern and western BOB), TC-induced SST cooling is around 2° – 6°C ([Gopalakrishna et al. 1993](#); [Rao et al. 2007](#); [Mahapatra et al. 2007](#)), while over the northern bay with stronger upper-ocean stratification due to low surface salinity induced by rainfall and river discharge, the SST decrease is below 1.5°C ([Murty et al. 1996](#); [Subrahmanyam et al. 2005](#); [Vissa et al. 2013](#)).

[Sengupta et al. \(2008\)](#) studied premonsoon and postmonsoon cyclones in the BOB, and found that cyclone-induced mixing does not significantly cool the sea surface during the postmonsoon season, mainly because of the presence of a deep IL or warm BL that embeds itself in the IL during the postmonsoon period. Using ocean general circulation model (OGCM) simulations, it has been suggested that in the BOB during the postmonsoon season, a very deep IL combined with considerable upper-ocean freshening strongly inhibits TC-induced surface cooling, and the salinity

stratification accounts for 40% of the cooling reduction ([Neetu et al. 2012](#)). Due to the existence of the BL and subsurface warm advection in the northwestern BOB, vertical mixing induces near-surface warming by entraining warmer water from the BL, which weakens the cooling induced by turbulent heat fluxes in the footprints of TCs ([Wang and Han 2014](#)). However, due to the lack of observations, an observational depiction of the effect of the BL, particularly with a preexisting temperature inversion between the BL and ML, on TC-induced SST cooling, is still lacking and the associated processes require in-depth investigation.

The satellite-observed *Aquarius/Satelite de Aplicaciones Cientificas-D* (SAC-D) SSS enables tracking the spatial variability of SSS induced by TCs in an unprecedented manner ([Grotsky et al. 2012](#)). Thus, in this paper, we combine the *Aquarius/SAC-D* SSS with other remote sensing data and in situ observations to explore spatial and temporal variations of SSS and SST associated with TC Phailin in the fall of 2013, the most intense cyclone that occurred in the BOB and made landfall on the east Indian coast since Super Cyclone Odisha of 1999. The effects of the BL, with a preexisting temperature inversion between the ML and BL, on Phailin-induced surface cooling are quantitatively assessed, using field observations and a diagnostic ML temperature equation together with a hierarchy of experiments using the Hybrid Coordinate Ocean Model (HYCOM). In addition to Phailin, the upper-ocean stratification associated with 16 super TCs (category ≥ 3) over the BOB from 1980 to 2015 is analyzed.

The rest of the paper is organized as follow. [Section 2](#) describes the datasets and methodologies used in the study. [Section 3](#) characterizes the observed upper-ocean response to Phailin. [Section 4](#) examines the impact of the BL on Phailin-induced sea surface cooling and the associated processes. [Section 5](#) analyzes the observed upper-ocean stratification associated with the evolution of the 16 BOB TCs, and [section 6](#) provides the discussion and conclusions.

2. Data and methods

a. Data

The tropical cyclone track dataset of the International Best Track Archive for Climate Stewardship (IBTrACS) released by National Oceanic and Atmospheric Administration (NOAA) is used to document the trajectories of the 16 super TCs including Phailin. IBTrACS is a new merged dataset widely used by the tropical cyclone community and is constructed by collecting the historical tropical cyclone best track data from

FIG. 1. (a) SSS (psu) distribution (color contours: *Aquarius* data; line contours: Argo monthly gridded data) for October 2013 in the BOB, with the freshwater region denoted by SSS values lower than 32.5 psu. The 6-hourly positions of TC Phailin are shown by circles that are color coded by category (based on the Saffir–Simpson hurricane wind scale). The star denotes the RAMA buoy location. (b) The positions of Argo profiles near four locations of Phailin shown in the rectangle of (a). Squares, triangles, and diamonds denote floats 2901335, 2901334, and 2901327, respectively, with sampling time shown by color scales from 4 to 16 Oct. The exact sampling times for each of the three Argo floats from 8 to 13 Oct and their distance to the TC center are presented in the inset of (b). (c) Time evolution of maximum wind speed and intensity of Phailin (every 6 h) from IBTrACS. (d) BLT (m) from Argo monthly gridded data (color contours) and monthly averages of daily HYCOM reanalysis data (line contours) for October 2013 overlaid by 6-hourly positions of TC Phailin and the positions of Argo profiles (black crosses).

all available Regional Specialized Meteorological Centres (RSMC) and other agencies (Knapp et al. 2010). It consists of 6-hourly position, minimum center pressure, and maximum sustained wind speed of each cyclone.

As Phailin passed a nearby buoy that was part of the Research Moored Array for African–Asian–Australian Monsoon Analysis and Prediction (RAMA; McPhaden et al. 2009), and located at 15°N, 90°E (Fig. 1a), hourly time series of sea level pressure, rainfall, SST, and SSS obtained by this buoy are used to examine the evolution of ocean surface and atmospheric conditions associated with Phailin. The water temperature closest to the surface at 1-m depth is considered as the SST. As salinity above 10 m, as well as surface wind, are not available during the TC's passage, we use the RAMA buoy salinity measured at 10-m depth as SSS, and daily Advanced Scatterometer (ASCAT) surface wind on

0.25° grids, produced by Remote Sensing Systems and sponsored by the NASA Ocean Vector Winds Science Team (Ricciardulli and Wentz 2016), to document surface wind evolution. The wind vectors at the locations of the RAMA mooring are from the ASCAT wind product.

There are three Argo floats with IDs 2901327, 2901334, and 2901335 close to Phailin's track. Temperature and salinity profiles from the Argo floats are used to identify the subsurface ocean response to Phailin. The locations of the three Argo floats are shown in Fig. 1b. The distances between the TC center and the three Argo floats are about 20 km, 74 km, and a range of ~10–400 km, respectively (Fig. 1b). Two floats (2901327 and 2901334) sampled at an interval of 4 days, and the third float, 2901335, which is a special float, rapidly measured profiles about every 2 h during Phailin's passage, providing an exceptional opportunity to gain

insights to upper-ocean response to a strong TC. After passing real-time quality control, the Argo data are subjected to additional quality checks for drift error and offset in salinity, using the calibration method developed by Wong et al. (2003).

The *Aquarius* level 3 standard version 2.0 monthly and weekly SSS with 1° resolution, provided by the Ocean Data Processing System (ODPS) of NASA Goddard Space Flight Center (GSFC) (Foti et al. 2013) are used to document the basin-scale SSS structure over the BOB. The monthly *Aquarius* SSS data are used to examine the SSS distribution for October 2013, the month when Phailin occurred (Fig. 1). The corresponding monthly Argo product provided by the Scripps Institution of Oceanography during the same period (Roemmich and Gilson 2009) is also analyzed and compared with the *Aquarius* SSS. The weekly *Aquarius* data are used to explore Phailin-induced SSS variations during its passage. Additionally, the 1/12°, daily HYCOM reanalysis product available for 1992–2012 (Chassignet et al. 2009) is used to compare with the *Aquarius* SSS and to explore Phailin-induced BL evolution.

To provide a basin-scale view of the upper ocean response, we also analyze the gridded products of ocean surface currents, sea surface height (SSH), SST, ocean surface wind, surface heat flux, and precipitation. The daily Geostrophic and Ekman Current Observatory 2 products (GEKCO2) on a 1/4° grid are developed by Joël Sudre at LEGOS, France (Sudre et al. 2013). Daily SSH data are from the AVISO merged altimeter product on a 1/4° grid (<http://www.marine.copernicus.eu>). Two SST datasets are utilized, one from the daily high-resolution microwave plus infrared Optimum Interpolation SST (OISST) product on a 1/4° grid and the other from daily 1/4°-grid SST mapped from ascending orbital data of the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI); both datasets are produced by Remote Sensing Systems and sponsored by the National Oceanographic Partnership Program (NOPP) and the NASA Earth Science Physical Oceanography Program (data are available at www.remss.com). The wind data are from the aforementioned ASCAT surface wind stress and wind velocity, together with the Cross-Calibrated, Multi-Platform satellite wind version 2.0 (CCMP2) with a spatial resolution of 1/4° and a temporal resolution of 6 h (Wentz et al. 2015). The surface wind stress for the CCMP2 product, which only provides surface wind velocity, is calculated by the bulk formula of Oey et al. (2006). Daily evaporation and turbulent heat fluxes are obtained from the Objective Analysis Flux (OAFux) product with a spatial resolution of 1° × 1° (Yu and

Weller 2007). The 3-hourly TRMM accumulated precipitation data come from NASA on a 1/4° grid (Huffman et al. 2007).

Finally, the recently released 5-day resolution Simple Ocean Data Assimilation (SODA 3.3.1) ocean reanalysis data for the period 1980–2014 (Carton et al. 2018) are used to examine the barrier layer thickness (BLT), isothermal layer depth (ILD), and SST near the tracks of 16 super TCs with intensity larger than category 3 for the period of 1980–2015. SODA 3.3.1 relies on the NOAA/Geophysical Fluid Dynamics Laboratory CM2.5 coupled model with nominal 1/4° horizontal resolution and 50 vertical levels.

b. Methods

1) BARRIER LAYER THICKNESS

The BL is the layer between the ML and the thermocline, and the BLT is defined as the difference between the ILD and MLD:

$$\text{BLT} = \text{ILD} - \text{MLD}. \quad (1)$$

Here, the ILD is the depth at which the temperature decreases by 0.5°C compared to the temperature at 1-m depth (Chu et al. 2002). The MLD is defined as the depth at which density increases ($\Delta\sigma_t$) by the amount that is equivalent to the temperature decrease by 0.5°C (i.e., $\Delta T = -0.5^\circ\text{C}$) relative to 1-m depth.

2) SURFACE MIXED LAYER TEMPERATURE EQUATION

To understand the impact of the BL on surface cooling induced by Phailin, we perform a budget analysis using the ML temperature equation (Elsberry et al. 1976; Jacob et al. 2000; Wang et al. 2011):

$$\frac{\partial T}{\partial t} = \frac{Q}{C_p \rho_s h} - \mathbf{V} \cdot \nabla T - \frac{W_e \Delta T}{h}. \quad (2)$$

The term on the left-hand side represents the ML temperature tendency, and the first term on the right-hand side represents forcing by surface heat flux, followed by the effects of horizontal heat advection and entrainment, respectively. Parameter T is the average temperature of the ML, representing SST in our study; Q is the net-surface heat flux, $\mathbf{V}$ is the horizontal velocity vector, h is the MLD, C_p is the specific heat of water, ρ_s is the ocean surface water density, and ΔT is the temperature difference between T and the temperature at the base of the ML [same for ΔS to be shown in Eq. (5)]. Parameter W_e is the vertical velocity, including the upwelling effect velocity W_{up} and entrainment velocity W_{ent} ,

$$W_e = W_{\text{up}} + W_{\text{ent}}. \quad (3)$$

Parameter W_{up} (i.e., the Ekman pumping velocity) is calculated by

$$W_{\text{up}} = \text{curl}(\tau/\rho_w f). \quad (4)$$

Here, ρ_w is the ocean water density and is set to 1025 kg m^{-3} , f is the Coriolis parameter, and τ is the surface wind stress vector. The entrainment velocity W_{ent} is specified based on the wind stress-induced turbulence and sea surface buoyancy flux. This scheme has been employed in several previous studies on the oceanic response to cyclones (e.g., [Elsberry et al. 1976](#); [Jacob et al. 2000](#)). These studies suggested that the BL affects entrainment cooling through its modulation of ΔT . In the presence of a BL, ΔT is small or even negative when temperature inversion occurs (e.g., [Wang et al. 2011](#); [Wang and Han 2014](#)); consequently, density stratification is determined primarily by salinity stratification. Therefore, W_{ent} can be parameterized by Eq. (5), in which the salinity change ΔS between the ML and BL determines the density change:

$$W_{\text{ent}} = \frac{\Lambda(C_1 u_*^3 - C_2 Bh)}{-\beta gh \Delta S}. \quad (5)$$

In contrast, without a BL, W_{ent} can be parameterized by Eq. (6), in which temperature change ΔT determines density change:

$$W_{\text{ent}} = \frac{\Lambda(C_1 u_*^3 - C_2 Bh)}{-\alpha gh \Delta T}. \quad (6)$$

Here $u_* = \sqrt{\tau/\rho_s}$ is the ocean surface friction velocity; C_1 and C_2 are empirical coefficients and are set to 2.5 and 0.4, respectively ([Elsberry et al. 1976](#); [Jacob et al. 2000](#)); g is the acceleration due to gravity, and B is the surface buoyancy flux, which is

$$B = \alpha(C_p \rho_s)^{-1} Q + \beta S_0(P - E). \quad (7)$$

In Eq. (7), α and β are the coefficients of thermal expansion and saline contraction, and are set to $0.2 \times 10^{-3} \text{ K}^{-1}$ and 0.8×10^{-3} , respectively; S_0 is the SSS; and P and E are precipitation and evaporation. The Λ symbol is the Heaviside step function of $(C_1 u_*^3 - C_2 Bh)$, defined as below:

$$\Lambda = \begin{cases} 1, & C_1 u_*^3 - C_2 Bh > 0 \\ 0, & C_1 u_*^3 - C_2 Bh < 0 \end{cases}. \quad (8)$$

When $(C_1 u_*^3 - C_2 Bh) > 0$, there is sufficient turbulent energy to entrain and mix water from subsurface.

The cyclone-derived horizontal velocity in the ML is estimated as

$$V = \frac{\tau R_{\text{max}}}{\rho_s h U_h}, \quad (9)$$

where R_{max} and U_h are the radius of maximum wind and the translation speed of the TC, respectively ([Price 1983](#)).

3) HYCOM EXPERIMENTS

The ocean general circulation model, HYCOM version 2.2.18 ([Halliwell 2004](#)), has been used to perform a hierarchy of experiments to assess various forcing and processes associated with atmospheric synoptic-to-intraseasonal variability ([Li et al. 2014](#); [Li et al. 2017a,b](#)). The non-local k -profile parameterization (KPP) is used for the boundary layer mixing scheme ([Large et al. 1994, 1997](#)). Background diffusivity for internal wave mixing is set to $5 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$ ([Gregg et al. 2003](#)), and viscosity is set to be an order of magnitude larger ($5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$; [Large et al. 1994](#)). The diapycnal mixing coefficient is set to $(1 \times 10^{-7} \text{ m}^2 \text{ s}^{-2})/N$, where N is the buoyancy frequency. Isopycnal momentum dissipation values are formulated as $u_d \Delta x$, where Δx is the local horizontal mesh size and u_d is set to be 0.015 m s^{-1} for Laplacian dissipation and 0.005 m s^{-1} for biharmonic dissipation. A similar method is used for temperature and salinity diffusion, with u_d set to be 0.001 m s^{-1} for Laplacian diffusion. Since this study focuses on the upper-ocean processes under TC Phailin forcing, we analyze the results from the following experiments ([Table 1](#)). First, we analyze the Main Run (MR), which includes all forcing fields and is the most complete solution in the hierarchy; the MR was forced by daily ASCAT wind, TMI precipitation, shortwave and longwave radiation (SWR and LWR) from the Clouds and the Earth's Radiant Energy System (CERES) product, and 2-m air temperature and humidity from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) products. For more details please see [Li et al. \(2017a\)](#). Then, we analyze the NoSubseason, NoSWR, NoWIND, NoSTRESS, NoRAIN, and NoSMR experiments, which used 105-day low-passed atmospheric fields, shortwave radiation, wind speed, wind stress, rainfall, and climatological freshwater flux for January–March, respectively, to isolate the effects of different forcing fields and the associated physical processes.

In our HYCOM setup, oceanic dynamical processes (i.e., upwelling and horizontal advection) and mixed layer entrainment cooling were directly determined by surface wind stress. Surface wind speed, air temperature,

TABLE 1. The suite of HYCOM experiments designed to isolate processes associated with subseasonal (i.e., synoptic to intraseasonal) atmospheric forcing. Note that due to the nonlinearity of the oceanic system, the sum of process solutions is close to, but not exactly the same as, the Main Run solution. For details please see [Li et al. \(2014\)](#).

Experiment	Forcing	Description
MR	Daily forcing	Complete run
NoSubseason	(105-day) low-passed forcing	Remove all subseasonal forcing
NoSWR	Low-passed SWR	Remove subseasonal SWR
NoWIND	Low-passed wind speed and stress	Remove subseasonal wind speed and stress
NoSTRESS	Low-passed wind stress	Remove subseasonal wind stress
NoRAIN	Low-passed rainfall	Remove subseasonal rainfall
NoISMR	Same daily forcing as MR except using climatological freshwater flux for January–March	Remove the effect of Indian summer monsoon (ISM) freshwater flux
(MR – NoSubseason)		Isolate effect of total subseasonal forcing
(MR – NoSWR)		Isolate subseasonal SWR effect
(NoSTRESS – NoWIND)		Isolate effect of latent and sensible heat flux associated with subseasonal wind speed
(MR – NoSTRESS)		Isolate effect of ocean dynamics (upwelling + advection) and entrainment associated with subseasonal wind stress
(MR – NoRAIN)		Isolate subseasonal rainfall effect
(MR – NoISMR)		Isolate ISM freshwater flux effect
ReconWIND	Same daily forcing as MR except using the reconstructed wind	Examine wind stress strength effect by compared with MR

specific humidity, and HYCOM SST were used to calculate turbulent (sensible and latent) heat fluxes. Since the 105-day low-pass filter, applied to atmospheric forcing fields, removes both synoptic and intraseasonal variability in each of the experiments described above, the differences between the Main Run and each of the experiments (MR – NoSubseason, MR – NoSWR, NoSTRESS – NoWIND, MR – NoSTRESS, MR – NoRAIN, and MR – NoISMR) estimate the effects of all subseasonal atmospheric forcing, shortwave radiative flux, wind speed that affects surface turbulent heat fluxes, wind stress as a major driver of ocean dynamical processes and mixed layer entrainment, rainfall, and summer monsoon freshwater flux effects on the oceanic variability primarily at synoptic-to-intraseasonal time scales, respectively (Table 1). For the short duration of Phailin's passage, because oceanic intraseasonal variability is much weaker than synoptic variability (figure not shown), the solution differences shown above primarily measure the synoptic variability induced by TC forcing. Since the daily ASCAT wind and reanalysis winds significantly underestimate Phailin's strength compared to in situ observations ([Wang et al. 2012a,b](#)), we also performed the ReconWIND experiment. ReconWIND is the same as the HYCOM MR except that it uses the reconstructed wind in place of the daily ASCAT wind during TC Phailin's passage from 8 to 13 October over the BOB. The reconstructed wind magnitude matches the observed TC wind, based on the modified Rankine vortex method of [Wang et al. \(2012a\)](#).

3. Observed upper-ocean response to TC Phailin

a. The evolution of TC Phailin

The category 5 severe TC Phailin generated in October 2013 ([Fig. 1a](#)) was the most intense TC in the BOB that hit the east Indian coast since the devastating 1999 Super Cyclone Odisha. On 4 October, it was first noted as a tropical depression (TD) within the Gulf of Thailand and entered the north Andaman Sea (12.0°N, 96.0°E) on 8 October ([Fig. 2a](#)). On 9 October ([Figs. 1a, 2b](#)), it entered the BOB as a weak tropical storm (TS). From there, it intensified into a category 1 TC (14.5°N, 91.3°E) on 10 October and quickly grew into a category 4 TC (15.6°N, 89.4°E) within 18 h ([Figs. 1a,c](#)). Phailin reached category 5 intensity at around 1200 UTC 11 October, with its maximum wind speed exceeding 70 m s^{-1} ; the category 5 intensity was maintained for 18 h. It then weakened and dissipated after landfall on the east coast of India on 13 October. [Figure 2](#) shows the evolution of surface wind and rainfall fields during the TC's passage. The wind speed on the right of the TC center was evidently stronger than that on the left, as the motion of storms also contribute to their swirling winds ([Price 1994](#)). We will show later that this wind speed difference induced a much stronger upper ocean response on the TC's right side. Intense rainfall events were observed by the RAMA buoy located at 15°N, 90°E: one occurred near the end of 8 October with a rain rate of $\sim 57 \text{ mm h}^{-1}$, and the other occurred at the beginning of 10 October

FIG. 2. Daily sea surface wind (vectors; m s^{-1}) at 10-m height from ASCAT and TRMM rainfall (color contours; mm h^{-1}) during Phailin's passage overlaid by 6-hourly positions of TC Phailin from 8 to 13 Oct: (a) 8, (b) 9, (c) 10, (d) 11, (e) 12, (f) 13 Oct.

when Phailin reached its peak (Figs. 3b, 1c). On 11 October, strong rainfall occurred over most regions of the northern BOB, as shown by the daily mean TRMM rainfall data (Fig. 2).

As the northern BOB receives a considerable amount of local rainfall and a large quantity of freshwater from river discharges during the summer season, it is characterized by eastern-bay and western-bay plumes in the northern BOB during different seasons (Wu et al. 2007). Figure 1a shows the SSS distribution of *Aquarius* and Argo for a comparison during October 2013. The *Aquarius* SSS shows systematically lower values over

the BOB compared to Argo data, with the largest difference of about -0.4 psu in the northern bay. The systematically lower SSS biases of *Aquarius* data in regions of low SSS under intensive rainfall conditions have been noted by previous studies (e.g., Du and Zhang 2015). Despite the differences, both datasets show similar SSS patterns, with SSS lower than 32.5 psu over the northern and eastern BOB, and the minimum SSS occurring near the bay rim. The entire trajectory of Phailin, including its generation, development and decay, passed through the fresher (<32.5 psu) region of the BOB (Fig. 1a). Over the fresher region, the BL occupied almost the entire

FIG. 3. Real-time hydrological and meteorological data from the RAMA buoy at 15°N , 90°E (star in Fig. 1a) and satellite data during Phailin: (a) SLP (hPa), (b) rainfall (mm h^{-1}), (c) 10-m surface wind (m s^{-1}) from ASCAT at the location of the RAMA buoy, (d) SST ($^{\circ}\text{C}$), and (e) SSS (measured at 10-m depth; psu). Tick marks denote beginning of each day. The distance for the location of the RAMA buoy relative to TC center with time is denoted by the red curve in (a). The weekly *Aquarius* SSS values at the buoy location are shown as red crosses in (e).

northern bay, especially in the northeastern bay on the right of Phailin's track, where a thick BL of up to 30 m was present in both the HYCOM reanalysis and Argo observations (Fig. 1d). In a large region on the left of Phailin's track, Argo data indicates that there was a much thicker BL (~ 25 m) than HYCOM reanalysis data (~ 10 m), likely due to the very sparse observations by Argo floats there (Fig. 1d), which resulted in large uncertainties in the BLT estimation. Since low SSS can cause BL formation, which is sometimes accompanied by a temperature inversion between the ML and BL (Girishkumar et al. 2013), in sections 3b and 4 we will explore the effect of strong salinity stratification on the upper-ocean response to Phailin.

b. Upper-ocean response to Phailin

The temporal evolution of major atmospheric and oceanic surface variables obtained from the RAMA buoy show that before the passage of Phailin (1–8 October),

typical postmonsoon conditions prevailed in the BOB. Sea level pressure was around 1010 hPa (lower during 1–2 October) with an apparent diurnal cycle and accompanied by moderate southwesterly winds of $3\text{--}9\text{ m s}^{-1}$, rare rainfall events, and very low intensity (Figs. 3a–c). The SST was $\sim 28.5^{\circ}\text{C}$ with an apparent diurnal cycle and SSS was ~ 32.0 psu (Figs. 3d–e). During 9–13 October, the buoy detected the effects of TC Phailin's passage. When Phailin approached the buoy on 9 October, it quickly intensified and upgraded to category 3 on 10 October (Figs. 1a, 2b, 2c, 3a), and sea level pressure dropped to 988 hPa. Copious rainfall (57 mm h^{-1}) was observed one day before Phailin's arrival, and surface wind dramatically intensified and changed direction from southwesterly to northeasterly.

As a response to the intensifying surface wind, SST decreased and its diurnal cycle essentially disappeared. The SST drop, however, was only $\sim 0.29^{\circ}\text{C}$ (from 28.77°C averaged for 6–8 October to 28.48°C averaged for 9–11

October), and the cooling was still weak ($\sim 0.43^{\circ}\text{C}$) after the seasonal warming trend was removed from the time series, albeit with the strong wind and precipitation remaining. In contrast, the response of SSS was far stronger than that of SST, decreasing first from 31.83 psu (averaged for 7–8 October) to 31.35 psu (averaged for 9–10 October) in response to the heavy rainfall, and then increasing dramatically by 1.12 psu to 32.47 (averaged for 11–12 October). This increase in salinity is also well captured in *Aquarius* weekly SSS data. The causes for the different responses of SST and SSS to Phailin are attributed to the preexisting BL and will be examined in detail in section 4.

To explore the spatial–temporal features of the surface ocean response to Phailin, we analyzed multiple satellite observations (Figs. 4, 5). Figure 4 illustrates the distributions of near-surface currents, sea level, and surface winds for the periods before, during, and after the Phailin passage, together with their differences between the first two periods. Before Phailin's arrival, the circulation over the bay was characterized by high SSH and alongshore current around the northwestern, northern, and eastern boundaries, and mesoscale cold and warm eddies (CEs and WEs, respectively) in the bay interior (Fig. 4a). In particular, a CE–WE pair was visible in the northwestern bay. The CE was on the left of Phailin's track (17.5°N , 86.5°E) and the WE was on the right (18.5°N , 89.0°E), and they were associated with negative ($< -0.2\text{ m}$) and positive ($> 0.15\text{ m}$) SSH anomalies, respectively. From Figs. 4b–d, it is evident that the TC induced two significant changes in the surface circulation. First, on the right side of Phailin's track, very strong northwestward currents ($\sim 0.2\text{ m s}^{-1}$) along the direction of the track were observed, while on the left no such currents were observed adjacent to the track (Fig. 4d).

This rightward bias of near-surface velocity in the Northern Hemisphere has been noted in the previous study (Chang et al. 2016), which is due to larger wind stress (Figs. 2, 4h) and the resonant response on the right side of a moving TC (Chang et al. 2013; Wang and Han 2014). Second, although the preexisting WE was not significantly affected by the TC, the preexisting CE was greatly intensified, and the enhanced alongshore currents can be seen on the left side of the track, which seems to be associated with the geostrophic flow of the CE. The CE's intensification cannot be simply explained by the local wind stress, because of the coexistence of positive (upward) and negative (downward) Ekman pumping velocity associated with Phailin over the CE region (Fig. 4h). The detailed dynamics for its intensification is beyond the scope of this paper, but will be explored in our future research.

The spatial pattern of SSS anomalies revealed by *Aquarius* data resembles that of HYCOM reanalysis, albeit with much stronger magnitude (Figs. 5a,b). A drastic increase ($> 1.0\text{ psu}$) in SSS appeared in almost the entire bay area on the right side of Phailin's track; however, on the left side, SSS increased mainly near the track and a patch of freshening was observed near 17°N , 86°E (Figs. 5b,c). Indeed, the large SSS increase during Phailin contributed to the seasonal salting of SSS in October and November, compared to the climatological SSS (Fig. S1 in the online supplemental material). By contrast, only a weak drop of SST ($< 0.5^{\circ}\text{C}$) was found over most regions of the bay to the right of the TC track (Figs. 5e–g), with strong cooling up to 2.5°C being observed over the CE area. The rightward bias observed in the surface current and SSS in a large region of the bay to the right side of Phailin's track was unexpectedly absent from the SST.

To further illustrate the aforementioned different responses of SSS and SST on both sides of Phailin's track, the satellite observed SSS and SST changes within the distance of 180 km from the TC track are shown in the scatterplots of Figs. 5d and 5h. The mean increase in SSS on right side of the storm was about 3 times greater than that on left side (0.99 vs 0.26 psu). In contrast, the average SST drop on both sides was much closer in magnitude (-0.70° vs -0.52°C) and a strong rightward bias in SST response (-1.50° vs 0.25°C) was only found in the CE area. The absence of rightward bias in SST in a large region of the northern bay implies that the SST cooling in the region was greatly reduced, as we shall see below.

4. Impact of the BL on Phailin-induced surface cooling

a. Existence of the BL revealed in HYCOM reanalysis and Argo

To understand the causes for the weak rightward bias in surface cooling, Fig. 6 shows the evolution of the BLT, MLD, and ILD during Phailin's passage. On 7 October (before Phailin's passage), a very thick BL (20–30 m) occurred in the northeastern bay, particularly in the WE region where the isothermal layer was deep and mixed layer was thin, resulting in a thicker BL ($\text{BLT} = \text{ILD} - \text{MLD}$). In contrast, in the CE region, the ILD was shallow and thus the BLT was thin ($\sim 10\text{ m}$). On 12 October (after Phailin's passage), the BL near the TC's track was completely eroded. Both the MLD and ILD were deepened, but the deepening of MLD was far larger than that of the ILD, resulting in the erosion of the BL.

FIG. 4. Surface currents (vectors; m s^{-1}) from GEKCO2 products and AVISO sea level anomaly (SLA; color contours; m) averaged for (a) pre-TG (2–7 Oct), (b) mid-TG (8–13 Oct), and (c) post-TG (14–19 Oct); (d) surface currents and SLA differences between (b) and (a), assessing Phailin's impacts; (e)–(h) As in (a)–(d), but for 10-m ASCAT wind (vectors; m s^{-1}) and the corresponding Ekman pumping velocity W_{up} (positive upward; m s^{-1} ; color contours). The 6-hourly position of TC Phailin is shown by circles. WE and CE in (a) denote the pair of eddies close to Phailin's track.

On the right side of the TC track near the locations of the two Argo floats (Fig. 6i), however, the ILD was shoaling somewhat, due to the positive Ekman pumping velocity (Fig. 4) overcoming the deepening effect by TC-induced entrainment.

Observational evidence comes from the three Argo floats located on the right side of the TC track (Fig. 1b). Both floats 2901327 and 2901334 show very thick (~ 50 m) pre-TG BLs and large temperature inversions of $\sim 0.6^\circ\text{C}$ between the ML and BL (Fig. 7).

FIG. 5. *Aquarius* (color shading) and HYCOM reanalysis (line contours) SSS averaged for the periods: (a) before Phailin (1–7 Oct), (b) during Phailin (8–14 Oct), (c) the difference for (b) minus (a), and (d) scatterplot of SSS anomalies (SSSA) shown in (c) vs distance from the cyclone track within 180 km; light blue lines represent averaged SSSA together with the 95% confidence interval for the mean value using a t test (shaded area). (e)–(h) As in (a)–(d), but for SSTA from daily OISST. Red lines and the shaded area in (h) denote averaged SSTA for both sides of the TC over the cold SSTA region [red box in (g)] and their 95% confidence interval. Since *Aquarius* data have weekly resolution, here we use the average of 1 and 7 Oct to represent the pre-TC period, and the average of 8 and 14 Oct for the “during TC” period.

FIG. 6. (left) Distribution of BLT (m) derived from HYCOM reanalysis (color contours) and AVISO SLA (m; line contours) for (a) 7 Oct before Phailin's passage, (d) 12 Oct after Phailin's passage, and (g) their difference, (d) minus (a). (center) As in the left panels, but for MLD. (right) As in the left panels, but for ILD. The locations of two Argo floats shown in Fig. 7 for 7 and 8 Oct, are overlaid as the blue diamond and triangle, respectively, in each panel.

After Phailin's passage, the BLs were eroded, primarily due to the deepening of the MLD for both floats. To a lesser degree, the slight shoaling of the ILD for float 2901327 also contributed to the BL's disappearance (Fig. 7a), consistent with the result from HYCOM reanalysis (Fig. 6i). While the BL was being eroded, the strong mixing entrained the considerably saltier water from the BL to the surface, and therefore significantly increased the SSS (~ 3.5 psu) at both floats' locations. In contrast to the dramatic increase in SSS, the decrease in SST was weaker, being 1.2° and 0.5°C for floats 2901327 and 2901334, respectively (Fig. 7). This weak surface cooling was associated with the temperature inversion within the IL, as is demonstrated in section 4b below.

As float 2901335 provided continuous temperature and salinity profiles during Phailin's passage (Figs. 8, 9), its data are analyzed to examine Phailin-induced upper-ocean variability (Fig. 8). Before Phailin's passage, a thick freshwater (<32.5 psu) layer of ~ 20 m was present; the strong salinity stratification resulted in a thin MLD of ~ 10 m and a thick BL of ~ 50 m (Fig. 8a). Between the ML and BL, there was a large temperature inversion of 0.8°C with higher upper-BL temperature than that at the surface. As the storm arrived, heat and freshwater were pumped down to ~ 190 m. When the TC reached

its peak during 10–11 October (Figs. 1–3), the BL was thoroughly eroded within about 3 h near the end of 10 October and beginning of 11 October, while the ML rapidly deepened and ILD sharply shoaled (Figs. 8a,b). As we shall see below, before and after the BL's erosion, SST variations showed distinct differences, based on the mixed layer temperature equation.

b. Diagnostic results

Using the data from float 2901335, Eq. (2) shows that the SST experienced a warming stage from 0200 to 1800 UTC 10 October (stage 1), the pre-TC period when the BL was thick and a temperature inversion was present (Fig. 8c). As heated by the subsurface warmer water, SST continued to increase from 28.71° to 28.89°C . From 1800 UTC 10 October to 1800 UTC 11 October, during and after the BL's rapid erosion (stages 2 and 3, the TC forced and relaxation periods, respectively), the sea surface began to cool with a total decrease of $\sim 1.8^\circ\text{C}$, accompanied by a drastic increase in SSS (~ 5.0 psu). The evolution of Ekman pumping velocity at the Argo float location (Figs. 8d, 9) suggests that downwelling and upwelling likely played an important role in the downward heat and freshwater pumping and subsequent shoaling of the ILD, SST cooling, and SSS increase (Wang and Han 2014; Cheng et al. 2015).

FIG. 7. Argo profiles of temperature (red curves) and salinity (blue curves) before (solid) and after (dashed) the passage of Phailin: (a) float 2901327 on 8 (12) Oct, located at 15.65°N, 89.84°E (15.65°N, 89.92°E) with BLT of 45 m (11 m) and (b) float 2901334 on 7 (12) Oct located at 16.28°N, 89.49°E (16.22°N, 89.39°E) with BLT of 55 m (8 m). The profile locations are shown in Figs. 1b and 6; the MLD and ILD are shown by the solid (dashed) green and purple lines, respectively, before Phailin on 8 Oct (after Phailin on 12 Oct). Note that $BLT = ILD - MLD$.

To explore the causes for the diametrically opposite SST variations associated with the TC during the three stages, the impacts of major processes on the SST changes are quantified using the diagnostic ML temperature equation [See Eq. (2) of section 2b(2)]. Two heat budget estimations were conducted in stages 1 and 2: one for the BL condition (HEBL) and the other with no BL condition (HEnoBL). Under these two conditions, the temperature and salinity profiles observed from the Argo float 2901335 and averaged during stages 1 and 2 are used for the calculations. The input parameters of the ML equation are shown in Table 2. The only difference between the two estimations is the initial salinity-induced stratification condition in the upper ocean: HEBL considered the initial salinity-induced stratification within the IL, while HEnoBL only considered thermal-induced stratification in the ML (no BL). Both were done by setting the salinity to a constant within the ML, using the salinity values

from the base of the ML (33.63 psu in stage 1 and 32.99 psu in stage 2; Figs. 10a,c and Table 2).

The results for HEBL and HEnoBL during stages 1 and 2 are shown in Figs. 10b and 10d, respectively. The surface warming rate observed by the Argo float during stage 1 is reasonably reproduced in HEBL at $0.70^{\circ}\text{C day}^{-1}$, compared to the mean warming rate of $0.38^{\circ}\text{C day}^{-1}$ observed by Argo for the same period (Figs. 8c and 10b). By contrast, a cooling rate of $-0.39^{\circ}\text{C day}^{-1}$ is obtained under HEnoBL. These results suggest that the preexisting BL with large temperature inversion was key to the surface warming during stage 1, due primarily to the entrainment of the warmer BL water into the ML with a warming rate of $0.87^{\circ}\text{C day}^{-1}$. This entrainment warming dominated over the cooling from the local surface heat fluxes ($-0.15^{\circ}\text{C day}^{-1}$) and horizontal heat advection ($-0.02^{\circ}\text{C day}^{-1}$). Without the BL, SST showed a cooling tendency ($-0.35^{\circ}\text{C day}^{-1}$), because of the entrainment of the colder, thermocline water into

FIG. 8. Argo profiles of (a) salinity and (b) temperature from 4 to 16 Oct observed by float 2901335. The profile locations are shown in Fig. 1b. Dark circles on the top indicate the dates of the Argo profiles. The black (white) solid curve denotes MLD (ILD), and the two curves are almost identical in the absence of salinity stratification. (c) Amplitude of the temperature inversion (red line), which is the difference between BL temperature and SST; SST tendency (blue line); and distances (dark line) from Argo profiles to the TC center. (d) As in (c), but for Ekman pumping velocity W_{up} (positive upward; purple curve) and surface wind speed (green curve) estimated from 6-hourly CCMP2 wind. The purple, blue, and green boxes in (a)–(d) mark stage 1 (pre-TC when SST was warming), stage 2 (the forced stage when SST began cooling), and stage 3 (the relaxation stage) during the passage of Phailin, spanning from 0200 to 1800 UTC 10 Oct, then to 0600 UTC 11 Oct, and finally to 1800 UTC 11 Oct, respectively.

FIG. 9. The 6-hourly Ekman pumping velocity (color contours; positive upward; m s^{-1}) derived from CCMP2 10-m sea surface winds, overlaid by CCMP2 winds (vectors; m s^{-1}) during Phailin's passage across the Argo float with ID number 2901335: (a) 0000 UTC 10 Oct, (b) 1200 UTC 10 Oct, (c) 0000 UTC 11 Oct, and (d) 1200 UTC 11 Oct. (e),(f) As in (a)–(d), but derived from daily ASCAT sea surface winds for 10 and 11 Oct, respectively. The corresponding location of the Argo float at each time is shown as a purple diamond in each panel.

the ML, with the other two terms [SST tendencies caused by net surface heat flux and horizontal heat advection (TQ and TH, respectively)] having far smaller contributions. This result agrees well with the previous studies (e.g., Jacob et al. 2000; Wang et al. 2011). As the time scale considered here is in the order of several hours, the role of the horizontal advection term on the heat budget is relatively small compared to the vertical mixing. A considerable difference in

the surface heat flux forcing term is seen between the BL condition ($-0.15^{\circ}\text{C day}^{-1}$) and no-BL condition ($-0.03^{\circ}\text{C day}^{-1}$), and it is largely attributed to the difference in MLD between the two conditions as shown in Table 2. Since Ekman pumping velocity W_{up} remained negative throughout stage 1 (Fig. 8d), it pushed the warm and fresh surface water downward to about 190 m (Figs. 8a,b). Because the BOB is not a mean upwelling zone (Varkey et al. 1996), the downwelling

TABLE 2. Input parameters for the ML temperature diagnostic equation at the location of Argo float 2901335 for stage 1 (pre-TC when SST was warming), stage 2 (the forced stage when SST began cooling), and stage 3 (the relaxation stage). Two calculations under conditions with BL and without BL (values in parentheses) were made during stage 1 and stage 2. See Eq. (2) of section 2b(2).

Date	Stage 1 (i.e., 0200–1800 UTC 10 Oct)	Stage 2 (i.e., 1800 UTC 10 Oct–0600 UTC 11 Oct)	Stage 3 (i.e., 0600 UTC 11 Oct–1800 UTC 11 Oct)
U_h (m s^{-1}) ^a	3.5806	3.5079	3.6806
Rmax (km) ^a	27.78	27.78	27.78
MLD (m) ^b	11.59 (60.49)	32.55 (43.83)	28.56
ΔT ($^{\circ}\text{C}$) ^c	−0.11 (0.71)	0.11 (0.44)	0.40
ΔS (psu) ^c	−0.3112	−0.1412	−0.0807
S_0 ^d	30.2455 (33.6279)	32.5744 (32.9953)	33.6909
(τ_x, τ_y) ^d	(0.1137, 0.5274)	(0.1933, 0.4858)	(0.1557, 0.5519)
$(\tau_x, \tau_y)^{*e}$	(0.1272, 0.5904)	(0.5803, 1.4585)	(0.1743, 0.6179)
Rainfall (cm day^{-1}) ^f	12.48	12.99	11.93
Evaporation (cm day^{-1}) ^g	0.44	0.75	0.96
Turbulent heat flux (W m^{-2}) ^g	88.00	219.10	350.20

^a Phailin cyclone information for stages 1–3, including maximum wind radius R_{max} and translation speed U_h are from the Joint Typhoon Warning Center (data available at <http://www.metoc.navy.mil/jtwc/jtwc.html>).

^b MLD, ΔT , ΔS , and S_0 (near-surface salinity) are derived from the mean temperature and salinity profiles during each of these three stages observed by the ARGO float as shown in Figs. 10a, 10c, and 10e.

^c On average, the BL nearly disappeared ($\text{BLT} = 7.30 \text{ m}$) during stage 3, and consequently, the TC-induced strong entrainment could reach below the base of IL. Thus, ΔT (ΔS) during this stage was estimated using the temperature (salinity) difference between the average temperature (salinity) within ML and the temperature (salinity) at the base of the IL.

^d Grid-averaged daily ASCAT wind stress.

^e The $(\tau_x, \tau_y)^{*}$ is the scaled ASCAT wind stress to match the observed wind speed from IBTrACS.

^f Grid-averaged TRMM precipitation averaged during each stage using 3-hourly TRMM data.

^g Grid-averaged information from daily OAFLUX for each stage, and Turbulent heat flux = latent + sensible heat flux from OAFLUX.

induced by surface Ekman convergence associated with the TC did not reduce upwelling and warm the SST. Thus, upwelling cooling (Tup) is zero for both HEBL and HEnoBL for stage 1.

As for stage 1, the total surface cooling ($-0.26^{\circ}\text{C day}^{-1}$, Fig. 10d) for stage 2 (from 1800 UTC 10 October to 0600 UTC 11 October, the forced stage) was due primarily to entrainment ($-0.23^{\circ}\text{C day}^{-1}$), while surface heat flux forcing enhanced the cooling ($-0.14^{\circ}\text{C day}^{-1}$). The effect of horizontal advection partly offset the cooling ($0.11^{\circ}\text{C day}^{-1}$). Without the temperature inversion due to the BL, entrainment cooling was much stronger. Even though apparent shoaling of the ILD occurred during this period, which indicates the role of TC-induced upwelling, it did not contribute to the surface cooling because of the warm BL with thickness of 11.28 m (Fig. 10c). Thus, Tup was ~ 0 in stage 2. However, the largest cooling ($-3.33^{\circ}\text{C day}^{-1}$) of the TC period was observed by the Argo float during stage 2, and was poorly represented by HEBL and even by HEnoBL ($-0.66^{\circ}\text{C day}^{-1}$). This large discrepancy between the diagnostic results and Argo observations (Fig. 10d) results mainly from the weaker ASCAT wind compared to the in situ observed wind at the core of TC (Figs. 1c, 2c,d; Table 2) during the forced stage, which cannot generate sufficient cooling via entrainment and surface heat flux in the diagnostic equation.

Similarly, a heat budget is calculated for stage 3 (0600–1800 UTC 11 October), using the mean temperature and salinity profiles of the Argo float 2901335 during this period, together with the parameters shown in Table 2. The SST tendency of $-1.74^{\circ}\text{C day}^{-1}$ from the sum of the four terms (net surface heat flux, entrainment, horizontal heat advection, and Ekman pumping) agrees very well with the SST cooling tendency of $-1.62^{\circ}\text{C day}^{-1}$ observed by Argo during this stage. Unlike stages 1 and 2, the upwelling cooling term ($-0.42^{\circ}\text{C day}^{-1}$) associated with the positive Ekman pumping velocity was comparable to the entrainment term ($-1.17^{\circ}\text{C day}^{-1}$), and the two together contributed up to 91% of the total surface cooling during stage 3. Due to the nonexistence of a BL, entrainment and upwelling transported the colder water from the thermocline, instead of the warmer water from the BL, into the surface ML. The effects of net surface heat flux and horizontal advection were much weaker than entrainment and upwelling on SST cooling.

To demonstrate that the weaker ASCAT wind was indeed the major cause for the weaker surface cooling in the diagnostic equation, we carried out a heat budget analysis using the scaled ASCAT wind with magnitude matching the in situ observed wind. As shown in Table 2, the ASCAT wind stress was comparable to the real wind stress with slightly weaker amplitude during stage 1

FIG. 10. Upper-ocean temperature (red curve) and salinity (blue curve) averaged from the profiles observed by Argo float 2901335 during the three stages as shown in Fig. 8, respectively: (a) averaged for stage 1 (0200–1800 UTC 10 Oct) with BLT of 48.90 m and MLD of 11.59 m; (c) averaged for stage 2 (1800 UTC 10 Oct–0600 UTC 11 Oct) with BL of 11.28 m and MLD of 32.55 m; (e) averaged for stage 3 (0600 UTC 11 Oct–1800 UTC 11 Oct) with highly diminished BL (BLT = 7.30 m) and MLD of 28.56 m. The MLD (green line) and ILD (purple line) are also shown in (a), (c), and (e). For the hypothetical situation with no BL during stage 1 in (a) and stage 2 in (c), only thermal-induced stratification is considered, thus, the salinity values in (a) and (c) are set to be constant (dark dashed curve) within the IL using the salinity values [33.63 in (a) and 32.99 in (c)] at the base of the IL; for this case, MLD = ILD. (b) Diagnostic calculation using Eq. (2) for the BL condition (yellow bars) and for the hypothetical situation with no BL (blue bars) during stage 1. TQ, TE, and TH are the SST tendencies caused by net surface heat flux, entrainment flux, and horizontal heat advection, respectively. Upwelling cooling is estimated by Tup; it is zero for stage 1 because Ekman pumping velocity is negative. The SSTA is the sum of TQ, TE, TH, and Tup, together with Argo_SSTA, which is the SST tendency observed by Argo float 2901335. (d),(f) As in (b), but for stages 2 and 3, respectively. TE*, TH*, Tup*, and SSTA* are the same as TE, TH, Tup, and SSTA, respectively, but derived from the scaled ASCAT wind stress as shown in Table 2.

(pre-TC) and stage 3 (the relaxation period); however, the ASCAT wind dramatically underestimated the in situ wind during stage 2, when the storm's eyewall reached the Argo location. Therefore, there were no obvious differences in the heat budget estimations between the ASCAT and scaled wind forcing during stages 1 and 3 (Figs. 10b,f); but large differences occurred during stage 2. A far stronger total cooling ($-1.03^{\circ}\text{C day}^{-1}$) was seen in HEBL using the scaled wind forcing (Fig. 10d), primarily due to the larger entrainment cooling ($-1.21^{\circ}\text{C day}^{-1}$). The observed strong cooling ($-3.33^{\circ}\text{C day}^{-1}$) however remained greatly underestimated even using the scaled wind forcing. This underestimation was likely due to the fact that the mean BLT was used in the budget calculation instead of the real time BLT, which ranged from ~ 40 to 10 m (Fig. 8b). Compared to the observed cooling tendency, one can see that a reasonable cooling ($-3.28^{\circ}\text{C day}^{-1}$) was reproduced in HEnoBL (Fig. 10d), further suggesting the sensitivity of the entrainment cooling to the BLT.

c. HYCOM experiments

The successful simulations of upper-ocean variability over the BOB, including the MLD, BLT, ILD, SST, and SSS, have been demonstrated in Li et al. (2017a,b). Here, we only examine the SSS and SST related to TC Phailin (Figs. 11, 12). The HYCOM MR can better reproduce changes in the spatial pattern and temporal variation of SSS seen in the *Aquarius* data during Phailin's passage, compared to the HYCOM reanalysis (Fig. 11). For SST, even though HYCOM reasonably simulated the SST pattern on 9 October, when Phailin was still in the Andaman Sea (Fig. 2), the warming and cooling signals were underestimated in our model (cf. top row of Fig. 12). In particular, the TC-induced surface cooling in the northwestern bay on 13 October after Phailin's passage was missing in the HYCOM MR, as compared to TRMM and the HYCOM reanalysis (Fig. 12, bottom row).

Note that strong salinity stratification (i.e., vertical gradient) was very shallow, occurring between approximately 5 and 15 m, within the ML or at the top of the BL (Figs. 7, 8). This enabled HYCOM to reproduce the observed SSS increase by entraining higher salinity at a shallow depth, despite the weaker wind from ASCAT that was used to force HYCOM. The daily ASCAT wind had a maximum speed of only 35.29 m s^{-1} on 12 October, only half of the maximum wind of 70 m s^{-1} on 11 October associated with Phailin (Fig. 1c; also see Wang and Han 2014). In contrast, the strong temperature stratification was deep, located below the IL (i.e., $\sim 60\text{ m}$). As shown above, the strong winds during the

peak of Phailin eroded the BL within a day (Figs. 8a,b); yet, ASCAT winds were not able to do so due to their weaker magnitude. Despite the underestimated TC-induced cooling, the hierarchy of HYCOM simulations can be still used to assess the relative importance of major processes driving SSS and SST variability during the TC passage.

The HYCOM MR successfully simulated the dramatic SSS increase along, and to the right side of, Phailin's track, as well as the SSS reduction in the southwestern bay after Phailin's passage (cf. Figs. 13a and 5c). The good model/data agreements suggest that HYCOM has captured the fundamental processes that determine SSS variability associated with the TC, and therefore its solutions can be used for further investigation on the relevant processes. As shown in Fig. 13, the observed/simulated large-scale SSS changes were well reproduced by the subseasonal forcing (MR – NoSubseason) and were dominated by ocean dynamical processes, namely upwelling, horizontal advection, and entrainment associated with the surface wind stress effect (cf. Figs. 13a,b,e).

Other forcing factors, including SWR, evaporation associated with the wind speed effect, and precipitation, overall contributed to the SSS freshening in the southwestern bay, and they weakened the SSS increase somewhat in most areas of the northern bay. This is likely because the reduced SWR associated with Phailin cooled the SST and thus reduced evaporation and SSS, and strong rainfall associated with Phailin also decreased SSS. Regarding the effect of wind speed, while strong winds tend to increase evaporation and thus increase SSS, they also increase latent heat loss and therefore reduce SST, which in turn weakens evaporation and thus tends to reduce SSS. The total result is the competition of the two effects.

The HYCOM MR produced cooling in most regions of the bay but produced warming in an area of the northern bay (Fig. 13g). Although HYCOM simulated the TC-associated cooling and some regional warming, the patterns of SST variations differed from the observation (Fig. 5g), because of the aforementioned far weaker ASCAT daily wind compared to the real wind during the TC passage. Wind-induced upward turbulent heat flux played a dominant role in causing surface cooling. The SWR forcing contributed to the basinwide surface cooling as well as the regional warming (Fig. 13i). The cooling effect from surface heat fluxes, however, was greatly reduced by the strong warming from the wind stress effect due to the entrainment heating with the BL warm water, resulting in the weak surface cooling around the TC's track, consistent with

FIG. 11. (a)–(c) SSS averaged for the periods before Phailin (1–7 Oct), during Phailin (8–14 Oct), and after Phailin (15–21 Oct), respectively, from the forward model experiment, which is the HYCOM MR. (d)–(f), (g)–(i) As in (a)–(c), but where SSS is from the daily HYCOM reanalysis and weekly *Aquarius* satellite products, respectively. As in Fig. 5, since the *Aquarius* data have weekly resolution, we use 1–7, 8–14, and 15–21 Oct to represent the periods before, during, and after the TC, respectively.

our diagnostic results discussed above. The warming over the northern bay due to SWR was enhanced by the entrainment/upwelling effect, and the effect of precipitation was weak.

To understand the effect of salinity stratification on variability of SST, an additional HYCOM experiment, NoISMR, was performed (Li et al. 2017a,b). The NoISMR experiment removed the effect of Indian summer monsoon rainfall and river (ISMR) freshwater flux, by using the climatological mean precipitation and river discharge from January to March to force HYCOM, with other forcing fields remaining the same as those of the HYCOM MR. As the freshwater flux for January–March only accounts

for a very small portion (<20%) of the total annual freshwater input (Li et al. 2017b), the effect of summer monsoon freshwater flux on the ocean is removed in NoISMR. The solution difference, MR – NoISMR, quantifies the effect of salinity stratification due to monsoon freshwater fluxes on the SST response to Phailin.

As shown in Fig. 14, very weak or approximately zero SSS change is visible over the BOB in NoISMR (Fig. 14c), which is in marked contrast with the large SSS increase in the MR (Fig. 14a). SST cooling is apparently strengthened in NoISMR compared to the MR (Figs. 14b,d). These results clearly demonstrate that the enormous freshwater flux of the Indian summer monsoon

FIG. 12. (a)–(c) SST for 9, 11, and 13 Oct, respectively, from the forward model experiment HYCOM MR. (d)–(f), (g)–(i) As in (a)–(c), but where SST is from the daily HYCOM reanalysis and daily TMI satellite products, respectively.

freshens the SSS, increases the vertical salinity gradients, and results in the formation of the BL. The high SSS in NoISMR reduces the vertical salinity gradient that exists in the MR. Thus, the strong SSS increase due to entrainment and upwelling in the MR disappears in NoISMR (Figs. 14a–c). The salinity stratification and BL in the MR reduced the SST cooling by up to $\sim 1^{\circ}\text{C}$ around Phailin's track during its passage (Fig. 14f), which confirms the important role of salinity stratification in suppressing the TC-induced surface cooling discussed above (Figs. 10b,d).

The relative importance of major processes driving SST variability in the HYCOM model, as shown above, is consistent with the ML heat budget results during stage 1 (Fig. 10b), both of which reveal that

wind-stress-driven entrainment warming can greatly reduce surface cooling by surface heat fluxes. In our case, it can even overcome the surface cooling and cause surface warming, as suggested by observations. However, the strong cooling in the surface layer induced predominantly by entrainment and upwelling and, to a lesser degree, by surface heat fluxes, during stages 2 and 3, is absent in the HYCOM simulation (figure not shown). This is because the ASCAT winds used to force HYCOM are too weak to completely erode the thick BL, while in reality the TC's winds are strong enough to do so (Fig. 8). As shown in Fig. 15, when forced by the reconstructed wind with magnitude comparable to the observed TC wind, HYCOM is able to simulate the observed cooling.

FIG. 13. (a) SSS difference (psu) between the period during Phailin (8–14 Oct mean) and the period before Phailin (1–7 Oct mean) from the HYCOM MR solution, based on HYCOM results with the 19-day running average removed, in order to remove the influence of intraseasonal and longer time-scale variability. SSS differences from (b) the HYCOM experiment runs that isolate the effects of total TC forcing, (c) SWR forcing, (d) the evaporative and turbulent heat fluxes associated wind speed forcing, (e) upwelling + advection + entrainment associated with wind stress, and (f) precipitation forcing. (g)–(l) As in (a)–(f), but for SST instead of SSS.

The failure of the HYCOM MR to produce the TC-induced strong surface cooling during stages 2 and 3 underpins the importance of realistically representing the strong TC winds in satellite products under strong precipitation conditions, which remains a challenge for the satellite wind community (Weissman et al. 2012). It also indicates the importance of realistic simulations of the formation and erosion of the BL over strong salinity stratification regions of the tropical oceans in state-of-the-art climate models; this BL formation and erosion, in addition to the well-known efforts to improve grid resolution and parameterizations of air–sea coupled processes in climate models, plays a crucial role in affecting SST and its atmospheric response.

5. Relationship between the upper-ocean stratification and 16 super TCs

To explore the relationship between upper-ocean stratification (including the presence of a salinity-induced BL

and TC intensity, we examined the occurrence and paths of all the recorded TCs with intensity of category 3 and above over the BOB for the 36-yr historical period from 1980 to 2015 (Fig. 16). Among the 16 super TCs that met that criterion, 9 occurred during the fall season (October–November), and 7 occurred during the spring season (April–May). Note that 12 out of the 16 (75%) super TCs experienced intensification when they passed through the BOB region with salinity lower than 32.5 psu. However, the remaining 4 TCs intensified in the southern bay where SSS is higher. Previous studies (e.g., Lin et al. 2005) found that the presence of a WE can suppress typhoon-induced SST cooling owing to the deep IL over a WE preventing the cold thermocline water from being entrained into the ML. Whether or not similar processes are at work in the southwestern bay remains to be explored.

After analyzing the upper-ocean stratification and evolution for each TC (i.e., MLD, BLT, ILD, and SST), we categorize the TCs into three groups (Table 3). Group 1 represents the condition of both a thick BL

FIG. 14. (a) SSS difference (psu) for 12 Oct minus 8 Oct from the HYCOM MR solution; (c) as in (a), but from HYCOM NoISMR, which suppressed the effects of freshwater fluxes from summer monsoon rainfall and river discharges into the BOB; (e) SSS difference between HYCOM MR and NoISMR (MR – NoISMR), which isolates the effects of freshwater forcing (i.e., salinity stratification). (b), (d), (f) As in (a), (c), and (e), but for SST instead of SSS.

(BLT ≥ 10 m) and a thick IL (ILD ≥ 20 m) in regions where five TCs intensified; group 2 represents a thin BL but a thick IL in regions where eight TCs enhanced; group 3 represents regions with both a thin BL and a thin IL but with remarkably high SST ($\geq 29.8^\circ\text{C}$), where three TCs intensified. The composites of BLT, ILD, and SST for each group are shown in Fig. 17. As expected, while a preexisting thick BL is important for the intensification of TCs, a thick ILD is crucial and appears more important for the development of super TCs compared to BLT. This is because, while 13 out of 16 ($\sim 81\%$) TCs intensified under thick ILT conditions (five

in group 1 and eight in group 2; Figs. 17a,b and 17d,e), only five of the group 1 TCs were under thick BL conditions (Figs. 17a,b). For these two groups, SST was relatively low ($< 29^\circ\text{C}$; Figs. 17c,f). Under both thin BLT and ILD conditions (group 3), all three TCs occurred during April and May under extremely high SST condition ($\geq 29.8^\circ\text{C}$; Figs. 17g–i), suggesting that very high SST is required to fuel the TCs under thin ILD conditions. As suggested by a recent study, SST is one of the most important parameters for TC intensification, and it can explain up to 23% of the variance in TC intensification in the eastern Pacific (Foltz et al. 2018).

FIG. 15. (a) Sea surface wind vectors (arrows; m s^{-1}) and wind speed (color contours; m s^{-1}) at 10-m height for 11 Oct from the ASCAT product; (b) as in (a), but for the reconstructed wind with magnitude scaled to the observed TC wind, using the modified Rankine vortex method of Wang et al. (2012a). (c) SST difference (color contours) for 12 Oct minus 8 Oct from the HYCOM MR; (d) SST difference for 12 Oct minus 8 Oct from the HYCOM ReconWIND run forced by the reconstructed TC wind in the BOB during the Phailin period of 8–13 Oct (color contours) and from TMI observations (line contours).

The TC intensity with and without considering the subsurface temperature is further calculated, by using the potential intensity (PI) index (Fig. S2), which is widely used as a TC maximum intensity predictor representing atmospheric and oceanic conditions (Emanuel 1995; Bister and Emanuel 2002a,b). Although PI represents a theoretical intensity value that a TC may reach, higher PI (with all other parameters remain the same) indicates that the ocean can provide more thermal energy for its growth and thus a higher potential category. Therefore, in a region with higher PI, the ocean will also provide energy and favor TC growth.

Among the five category 3–5 TCs of group 1 (Fig. S3), four were generated in the southeast bay, where they grow to category 3 TCs. The other one was generated over land but also intensified over the southeast bay. The southeast bay region exhibits largest PI based on the ocean temperature averaged over the top 80 m and the scaled ocean temperature (i.e., T80 and Tm; their details are shown in the caption of Fig. S2), whereas based on SST the PI is relatively low in the Andaman Sea (Fig. S2). Among the eight

TCs in group 2, seven were generated and developed in the southeast bay or southern bay where all PIs (i.e., PI_T80, PI_Tm, and PI_SST) attain their maximum magnitudes. One came from land but also intensified over the high PI region. For the three TCs in Group 3, all were generated in the southeast bay and then became category 3 in the area (90°–95°E, 5°–10°N). In this area, PI_80 and PI_Tm are the highest but PI_SST is a “low region.” For this group, both ILD and BLT are relatively thin; consequently, SST is very sensitive to surface forcing and therefore warms/cools rapidly. The high ocean heat content (OHC) however, changes slowly and provides energy for the TCs’ growth. These results further support the strong relationship between the 16 category 3–5 TCs and upper-ocean OHC.

Here, our analysis results suggest that among other causes, the heat storage in the upper ocean is important for TC intensification. The high thermal energy can be stored in the thick ILD, and/or thick BLT, or extremely high SST, which provides fuel for TC growth in the BOB. Classification of the similarities and differences of their impacts on TC intensity requires further investigation.

FIG. 16. Distribution of the number (color shading) of TCs on a 1° grid that occurred in the BOB during 1980–2015, and the distribution of sea surface salinity with values lower than 32.5 psu (black line contours). The super TC (category ≥ 3) tracks during this period are also shown in blue lines with colored dots denoting TC intensity.

6. Conclusions

In this paper, we provide a detailed examination of the processes that control the upper-ocean response to the passage of a category 5 TC, Phailin, over the low-SSS region of the BOB (Figs. 1, 2, 3, 9), using both in situ and spaceborne observations combined with diagnostic analysis using the mixed layer temperature equation

and a suite of HYCOM experiments. In addition, the upper-ocean conditions associated with the development of all 16 BOB super TCs that occurred during the 30-yr period from 1980 to 2015 are also analyzed (Fig. 16). In particular, the *Aquarius* salinity data provide an unprecedented spatial coverage for exploring the spatial pattern and temporal evolution of SSS related to TC–upper ocean interactions.

A drastic increase of up to 5.0 psu in SSS appeared in almost the entire northern bay associated with Phailin's passage, with larger values occurring on the right side of the TC track (Figs. 3, 5, 7). However, only a weak drop in SST of $\sim 0.5^\circ\text{C}$ was present, and the apparent rightward bias observed in a large region of the northern bay in the changes of SSS, and to a lesser degree, in surface current, was not evident in SST (Figs. 4, 5). The increase of SSS was dominated by mixed layer entrainment and ocean dynamical processes (i.e., upwelling and horizontal advection) driven by TC winds, which transported saltier water from below (Figs. 6–8, 11, 13).

The weak cooling in a large region of the northern bay to the right side of the TC was caused by the preexisting BL over the freshwater region of the BOB. Argo observations revealed strong temperature inversions between the BL and ML, which significantly modified the SST variations associated with Phailin. Specifically, the SST variations were characterized by three distinctive stages: the warming in stage 1 due to heating by the entrainment of the warmer water from the BL into the

TABLE 3. Names, dates, and upper-ocean stratification conditions for the 16 super TCs with intensity of category 3 and above, based on the Saffir–Simpson hurricane wind scale, and that occurred in the BOB since 1980 (see Fig. 16). Among these, nine occurred during the fall season (October–November) and seven occurred during the spring season (April–May). Based on the upper-ocean stratification condition during the TC intensification period, the TCs are divided into three groups: 1) passing over a thick barrier layer (BLT ≥ 10 m) with a thick isothermal layer (G1; ILD ≥ 20 m), 2) passing over a thin barrier layer (BLT < 10 m) but with a thick isothermal layer (G2; ILD ≥ 20 m), and 3) passing over a thin barrier layer (BLT < 10 m) and a thin isothermal layer (G3; ILD < 20 m). BLT, ILD, and SST are averaged within 100 km along a TC track.

No.	TC name	Passing date	Intensity (category)	BLT (m), group No.	ILD (m)	SST ($^\circ\text{C}$)
1	—	30 Apr–5 May 1982	4	11.14, G1	30.11	28.61
2	—	21–30 Nov 1988	3	3.78, G2	28.34	27.57
3	—	3–11 May 1990	4	4.98, G2	34.92	29.52
4	—	22–30 Apr 1991	5	7.01, G2	26.96	28.72
5	Forrest	8–22 Nov 1992	4	4.51, G2	47.48	27.88
6	—	26 Apr–3 May 1994	4	2.11, G3	11.72	29.81
7	—	18–25 Nov 1995	3	5.67, G2	34.86	28.25
8	—	1–7 Nov 1996	4	15.51, G1	31.14	28.88
9	—	13–20 May 1997	4	3.16, G3	11.80	29.86
10	—	25 Oct–3 Nov 1999	5	11.37, G1	22.87	28.44
11	—	26 Nov–6 Dec 2000	3	13.67, G1	41.43	28.11
12	Mala	24–29 Apr 2006	4	2.90, G3	8.89	30.09
13	Sidr	10–16 Nov 2007	5	5.75, G2	31.77	29.03
14	Nargis	25 Apr–4 May 2008	4	8.17, G2	22.90	29.49
15	Phailin	7–14 Oct 2013	5	16.17, G1	34.43	28.95
16	Hudhud	6–14 Oct 2014	4	6.14, G2	29.18	28.72

FIG. 17. Composite (a) BLT, (b) ILD, and (c) SST from the SODA pentad data along TC tracks (circles) during the TC intensification periods for five TC cases with both thick BL and thick ILD (TC numbers 1, 8, 10, 11, 15 of Table 3); (d)–(f) As in (a)–(c), but for eight TC cases with thin BL but thick ILD (TC numbers 2, 3, 4, 5, 7, 13, 14, 16 of Table 3). (g)–(i) As in (a)–(c), but for three TC cases with both thin BL and thin ILD (TC numbers 6, 9, 12 of Table 3). The TC intensity is indicated by the color code and size of the circles, as in Fig. 1a.

ML, the cooling in stage 2 (the TC forcing period), and the cooling in stage 3 (the relaxation period) owing primarily to entrainment and upwelling of the colder thermocline water into the ML after the BL was eroded (Figs. 6–8, 10, 12–15). To a lesser degree, the net surface heat flux also contributed to the surface cooling. The preexisting BL reduced the entrainment cooling by $\sim 1.09^{\circ}\text{C day}^{-1}$ using ASCAT winds, which significantly underestimated the TC wind strength, and $1.72^{\circ}\text{C day}^{-1}$ using the reconstructed TC winds. The latter were

constructed with their magnitude matching the in situ observed TC Phailin winds during Phailin's passage, based on the diagnostic calculation using the ML temperature equation (section 4b). Both observation-based diagnostic equations and OGCM experiments using HYCOM clearly demonstrate that realistic representation of TC wind magnitude is crucial for realistic simulation of TC impacts on the upper ocean, and therefore its feedback to TCs, as well as seasonal and longer-term climate.

When considering the subsurface temperature, which is an indicator of upper-ocean heat content, a more reasonable potential intensity (PI) index of the 16 super TCs is produced, compared with the PI index without subsurface temperature (Figs. S2 and S3). This further confirms that the heat stored in the upper ocean, either within the thick ILD, the thick BLT or the remarkably warm ML (Fig. 17), likely provides a favorable environment for the development and intensification of TCs, even though many other factors (e.g., atmospheric wind shear) are important in affecting TCs. Therefore, in order to achieve a better forecast for TC intensification, the contributions of the upper-ocean to TC growth should be thoroughly investigated. In situ observations (e.g., RAMA buoys and Argo data) along with satellite-based data (e.g., SSS from *Aquarius*, surface ocean vector wind products from ASCAT and CCMP2, and SST), combined with reanalysis data and model experiments, are critical for better understanding and forecasting TCs.

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