Response of the Salinity-Stratified Bay of Bengal to Cyclone Phailin

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ABSTRACT

Cyclone Phailin, which developed over the Bay of Bengal in October 2013, was one of the strongest tropical cyclones to make landfall in India. We study the response of the salinity-stratified north Bay of Bengal to Cyclone Phailin with the help of hourly observations from three open-ocean moorings 200 km from the cyclone track, a mooring close to the cyclone track, daily sea surface salinity (SSS) from *Aquarius*, and a one-dimensional model. Before the arrival of Phailin, moored observations showed a shallow layer of low-salinity water lying above a deep, warm “barrier” layer. As the winds strengthened, upper-ocean mixing due to enhanced vertical shear of storm-generated currents led to a rapid increase of near-surface salinity. Sea surface temperature (SST) cooled very little, however, because the prestorm subsurface ocean was warm. *Aquarius* SSS increased by 1.5–3 psu over an area of nearly one million square kilometers in the north Bay of Bengal. A one-dimensional model, with initial conditions and surface forcing based on moored observations, shows that cyclone winds rapidly eroded the shallow, salinity-dominated density stratification and mixed the upper ocean to 40–50-m depth, consistent with observations. Model sensitivity experiments indicate that changes in ocean mixed layer temperature in response to Cyclone Phailin are small. A nearly isothermal, salinity-stratified barrier layer in the prestorm upper ocean has two effects. First, near-surface density stratification reduces the depth of vertical mixing. Second, mixing is confined to the nearly isothermal layer, resulting in little or no SST cooling.

1. Introduction

Enthalpy flux from the warm ocean to the atmosphere sustains the heavy rainfall and destructive winds associated with tropical cyclones. Cooling of sea surface temperature (SST) due to storm-induced mixing with deeper ocean water inhibits cyclone intensity (e.g., Price 1981;

Emanuel 1999; Schade and Emanuel 1999; Zedler et al. 2002; D'Asaro 2003; Emanuel 2003; Lloyd et al. 2011), while rapid intensification of cyclones has been observed in regions with anomalously warm subsurface ocean temperatures (e.g., Lin et al. 2009; Jaimes and Shay 2009, 2015; Yablonsky and Ginis 2013; Girishkumar and Ravichandran 2012; Lin et al. 2014; Balaguru et al. 2018). Some recent studies address the interaction between tropical cyclones and the ocean in the presence of a “barrier layer,” that is, a deep, warm isothermal layer lying beneath a shallow, salinity-stratified layer (Sengupta et al. 2008; Balaguru et al. 2012, 2014; Neetu et al. 2012; Vincent et al. 2014). These studies indicate that (i) the added stability of the salinity-dominated

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density stratification reduces the total depth of vertical mixing; (ii) acting in concert with the deep isothermal layer, this reduces storm-induced SST cooling (or even makes it insignificant), and (iii) muted SST cooling under the track favors cyclone intensification.

The Bay of Bengal (BoB) is infamous for destructive tropical cyclones, which have killed nearly half a million people since 1970 alone (Frank and Husain 1971; Webster 2008). Communities living around the large river deltas of India, Bangladesh, and Myanmar are particularly vulnerable to storm surge and flooding from cyclones (e.g., De et al. 2005; Raghavan and Rajesh 2003; Adler 2005; Lin et al. 2009). The BoB experiences an average of three to four tropical cyclones per year; there are two cyclone seasons, April–June and late September–December (premonsoon and postmonsoon seasons). The temperature–salinity structure of the upper ocean in the north BoB is quite different in these two seasons. In the premonsoon season, the upper 50 m of the ocean is stratified mainly by temperature, and also by salinity. Summer monsoon rainfall and Ganges–Brahmaputra–Meghna river discharge (Sengupta et al. 2006; Papa et al. 2012; Chaitanya et al. 2014) forms a very low-salinity mixed layer with a halocline at 5–20 m, and a deep barrier layer, in the postmonsoon season. The shallow salinity stratification of the north BoB at this time is nearly 10 times larger than in the west Pacific. Temperature inversions develop through autumn and winter (Thadathil et al. 2016) as the sea surface temperature cools, while penetration of shortwave radiation below the shallow mixed layer continues to warm the subsurface ocean.

Satellite microwave SST measurements (Wentz et al. 2000) have been widely used to study the spatial structure and time evolution of SST cooling due to vertical mixing induced by tropical cyclones (e.g., D'Asaro et al. 2007; Sanford et al. 2007; Foltz et al. 2018). In a salinity-stratified ocean, sea surface salinity (SSS) is expected to rise due to mixing with saltier subsurface water (Sengupta et al. 2008). The impact of Bay of Bengal cyclones on upper-ocean salinity and temperature stratification has been previously reported from moored observations (Venkatesan et al. 2014) and Argo data (McPhaden et al. 2009; Lin et al. 2009; Maneesha et al. 2012). The influence of tropical cyclones on the space–time evolution of ocean salinity was not well documented until recently. With the launch of the *Aquarius/Satlite de Aplicaciones Cientficas-D* (SAC-D) (Lagerloef 2012) and the Soil Moisture and Ocean Salinity (SMOS) satellite missions (Reul et al. 2012), it became possible to observe changes in surface salinity over extended regions during tropical cyclone passage. For example, Grodsky et al. (2012) use SMOS SSS data

to study the interaction of the Amazon–Orinoco river plume with Hurricane Katia.

In this paper we investigate the response of the Bay of Bengal to Tropical Cyclone Phailin with the help of hourly surface meteorology and subsurface ocean temperature, salinity, and velocity from moorings, daily SSS data from the *Aquarius/SAC-D* satellite, and a one-dimensional model forced by hourly surface fluxes from moorings. We use data from the north BoB “observatory” (NBoB in Figs. 1a,b), a cluster of three closely spaced moorings located at 18°N, 89.5°E, about 200 km to the right of the cyclone track at closest approach, and mooring BD10 at 16.5°N, 88°E, almost directly under the cyclone track (Figs. 1a,b). As far as we know, this is the first detailed observational study of the space–time evolution of upper-ocean salinity and temperature in response to a powerful cyclone. The paper is organized as follows. We present the main characteristics of Phailin in section 2. Section 3 describes the data and methods used for analysis. In sections 4 and 5 we present results on the response of the upper ocean to the passage of Phailin, based on mooring and satellite data. In section 6, we study how the ocean’s response depends on (i) near-surface salinity stratification (i.e., salinity drop across the shallow halocline, at 5–20-m depth), and (ii) a preexisting deep isothermal layer, using a one-dimensional model (Price et al. 1986). We end with a discussion and summary of our main results in section 7.

2. Tropical Cyclone Phailin

Cyclone Phailin traversed the BoB during 8–12 October 2013 (Fig. 1a) [we use best track and other data from the India Meteorological Department (IMD)]. It was the strongest cyclone to make landfall on India’s east coast since the category 5 Odisha cyclone of October 1999. A low pressure system in the Andaman Sea became a very severe cyclonic storm (VSCS) with estimated 3-min maximum sustained winds of 34 m s^{-1} and quickly intensified to reach maximum sustained winds of 60 m s^{-1} on 10 October. IMD classified Phailin as a rapidly intensifying storm, that is, the maximum sustained wind increased by 15 m s^{-1} or more in 24 h (Kotal and Roy Bhowmik 2013). Cyclone Phailin made landfall at Gopalpur, Odisha state (19.2°N, 84.9°E), at 1700 UTC 12 October, subsequently weakening into a depression on 14 October (Kotal et al. 2013). Various aspects of winds, rainfall, and storm surge associated with Cyclone Phailin, and the impact of Phailin on temperature and chlorophyll in the upper ocean, have been reported in Kotal et al. (2014), Venkatesan et al. (2014), Murty et al. (2014), Lotliker et al. (2014), and in the modeling study of Prakash and Pant (2017).

FIG. 1. Map showing the track of Cyclone Phailin and mooring locations. (a) The 24-h accumulated TRMM rain on 11 Oct 2013 (shaded; cm). Three-hourly locations of Cyclone Phailin (IMD track data) are marked by open circles of different sizes: D, DD, CS, and VSCS stand for depression, deep depression, cyclonic storm, and very severe cyclonic storm. The filled triangle and star represent the moorings at 16.5°N and 18°N. (b) The North Bay of Bengal observatory (NBoB) consists of three moorings: INCOIS, NIOT BD08, and NIOT BD09, located 28–33 km apart. (c) Sea level pressure (hPa) and maximum sustained wind speed (km h^{-1}) measured by IMD at Gopalpur [white circle in (a)].

Price (1981) and Price et al. (1994) list the most important factors that determine the response of the upper ocean to hurricanes in terms of a few relevant parameters and nondimensional numbers. With the help of the IMD reports and the observations from mooring BD10, we reconstruct these parameters and numbers for Cyclone Phailin (Table 1). In this study, we estimate the nondimensional storm speed S_{trans} , defined as

$$S_{\text{trans}} = U_h / 2f_c R_m, \quad (1)$$

where f_c is the local Coriolis frequency, U_h is the translation speed of the storm, and R_m is the radius of maximum wind. If the value of S_{trans} is $O(1)$, the storm has resonant and asymmetric effects on the ocean on either side of the track and generates swift near-inertial currents in the ocean. During the time of our observations, Phailin has a translation speed of 4.2 m s^{-1} , giving a value of S_{trans} equal to 0.9 (Table 1) typical of many tropical cyclones. As expected, the observations described here show strong near-inertial resonance. However, the focus here is on the unique features of this storm that result from the unusual temperature and salinity stratification of the Bay of Bengal.

3. Data and methods

We use IMD 3-hourly best track positions (which are consistent with JTWC positions to within 10 km),

intensities, and other parameters of Phailin, including IMD ground station data on sea level pressure and maximum sustained wind speed at Gopalpur, in Odisha.

a. Moorings and satellite data

The National Institute of Ocean Technology (NIOT), Chennai, maintains the Ocean Moored Buoy Network for the Northern Indian Ocean (OMNI) observing system, which consists of six moorings with surface meteorological and subsurface ocean sensors in the Bay of Bengal (Venkatesan et al. 2013). In this study,

TABLE 1. Basic characteristics of Cyclone Phailin. Maximum sustained wind speed, translation speed of Cyclone Phailin, and radius of maximum sustained winds are from IMD estimates. Closest approach is from IMD's best track estimates of the storm center to the mooring BD10. Surface pressure is from direct observation at mooring BD10.

Category of the cyclone	Very severe cyclonic storm (wind speed $> 32 \text{ m s}^{-1}$)	
Distance of the center of the Cyclone Phailin from BD10	1.5 km	
Time and date of closest approach	0800 UTC 11 Oct 2013	
Maximum sustained wind speed	U_m	60 m s^{-1}
Radius of maximum winds	R_m	56 km
Pressure	P_m	920 hPa
Translation speed	U_h	4.2 m s^{-1}
Nondimensional storm speed	S	0.9

TABLE 2. List of all PWP model experiments at mooring BD09. Each model simulation or sensitivity experiment has an abbreviation: B indicates the basic (or control) run, and S stands for the sensitivity experiment with zero salinity stratification. We refer to experiments at mooring BD09 as 09—thus B09a is the control run for BD09. F09a, F09b, F09c and F09d refer to different experiments where the letter a indicates only one mixing parameter, b is without wind forcing, c is without heat flux, and d is with zero r_d .

Experiment	Forcing	Initial conditions
B09a (control run: ocean response at mooring BD09)	Estimated from hourly moored observations and COARE 3.0 algorithm at BD09	Observed T and S from mooring BD09
B09b,c (sensitivity run with enhanced/reduced surface fluxes)	Enhanced/reduced by 5%–10% relative to the control run	Identical to B09a
S09a (sensitivity run with zero salinity stratification)	Identical to B09a	Observed T profiles, and constant S equal to observed S at 100 m (34.5 psu at BD09)
S09b,c (sensitivity run with enhanced/reduced surface fluxes and zero salinity stratification)	Identical to B09b,c	Identical to S09a
F09a (sensitivity run with only mixing parameterization based on bulk Richardson number)	Identical to B09a	Identical to B09a
F09b,c (sensitivity run without wind forcing/heat fluxes)	Identical to B09a except for wind stress/heat fluxes, which are set to zero	Identical to B09a
F09d (sensitivity run with zero r_d)	Identical to B09a	Identical to B09a

we use hourly observations from four moorings: a mooring deployed by the Indian National Centre for Ocean Information Services (INCOIS) at 18°N, 89.5°E, and three OMNI moorings (BD08, located at 18.2°N, 89.67°E; BD09, 17.88°N, 89.67°E; and BD10, 16.5°N, 88°E). Meteorological sensors for air temperature, surface pressure, relative humidity, wind speed and direction, and downwelling longwave and shortwave radiation are mounted at 2–3-m height; subsurface conductivity and temperature (CT; Seabird/MicroCAT SBE37) sensors are at 5, 10, 15, 20, 30, 50, 75, 100, 200, and 500-m depth. The moorings are also equipped with an acoustic current meter (Teledyne RDI) at 1-m depth and a downward looking 150 kHz Acoustic Doppler Current Profiler (ADCP; Teledyne RDI) at 6.5 m, which measures ocean currents from 16- to 111-m depth with 5-m resolution. We use only hourly surface and subsurface data directly observed at these moorings in our analyses and one-dimensional model experiments (see below). In particular, the NIOT moorings measure 10-min averaged winds every hour. The INCOIS mooring (INC, Fig. 1b) at 18°N, 89.5°E is equipped with CT sensors at depths of 1, 4, 7, 15, 20, 50, and 100 m, and two current meters (RDI Instruments Doppler Volume Samplers, DVS) at 5- and 30-m depth.

We also examine the response of SSS to Cyclone Phailin using *Aquarius* Level 3 V4.0 daily data (Lagerloef 2012). The spatial resolution of this dataset is $1^\circ \times 1^\circ$. *Aquarius* SSS retrieval is based on L band microwave measurements, which are sensitive to surface roughness (appendix B); the measurements are therefore unreliable in the presence of high winds

and heavy rainfall associated with tropical cyclones (Reul et al. 2012). Therefore, we do not use satellite SSS data during the period when the cyclone is active, but only examine the difference between prestorm and poststorm SSS.

b. One-dimensional ocean model

We use the one-dimensional Price–Weller–Pinkel (PWP) ocean model (Price et al. 1986) to simulate the ocean response to Cyclone Phailin. Model control runs at the location of mooring BD09 (200 km to the right of the track) and BD10 (near the track) employ temperature and salinity initial conditions constructed from the mooring data interpolated in the vertical. Model vertical resolution is 0.25 m, and the time step is 1 h. Surface forcing is based on observed hourly incoming shortwave and longwave radiation, and turbulent fluxes are estimated from hourly moored measurements of air temperature, surface pressure, sea surface temperature, relative humidity, and wind using the COARE 3.0 bulk flux algorithm (Fairall et al. 2003); outgoing surface longwave radiation is estimated from the observed temperature at 5-m depth as a proxy for SST. Enthalpy fluxes are calculated according to Black et al. (2007), and wind stress is estimated following Powell et al. (2003). Several sensitivity runs are carried out to obtain uncertainty estimates, and to study the contribution of salinity-mediated density stratification to the overall response of the upper ocean. The experiments are summarized in Table 2 (and Tables S1 and S2 in the online supplemental material), and results are described in section 6.

FIG. 2. Hourly observations from moorings (left) BD10 and (right) BD09. (a),(d) Sea level pressure (hPa); (b),(e) wind speed (m s^{-1}) at 10-m height (U_{10}); and (c),(f) vector winds based on hourly observations from moorings BD10 and BD09. The wind sensor on mooring BD10 did not work after 15 Oct 2013.

4. Response of the ocean away from the cyclone track

a. Observed upper-ocean response at moorings

The observatory NBoB is located nearly 200 km to the right of the cyclone track at the time of closest approach. The maximum 10-min average wind speed measured at mooring BD09 (extrapolated to 10-m height) is 18.78 m s^{-1} at 0000 UTC 11 October; at this time, sea level pressure has fallen to 999 hPa from a prestorm value of 1010 hPa (Fig. 2). The wind speed is low immediately before and after the storm (2 m s^{-1} or less on 8 and 15 October).

The response of the upper ocean to Phailin is seen in the time evolution of potential temperature, salinity, and potential density at different depths from the BD09 hourly observations (Figs. 3a–d). Before the storm, the ocean has a warm, nearly isothermal layer to at least 50-m depth (the next sensor is at 75 m). The mean salinity difference across the upper 50 m from 1 to 8 October is 1.67 psu. Density stratification in the upper 50 m is determined mainly by the salinity gradient—the prestorm potential density gradient across the top 50 m is 1.24 kg m^{-3} . In response to the storm, upper-ocean temperature, salinity, and density are mixed down to 50 m. Mixing begins in the near-surface ocean and reaches 50-m depth at 0000 UTC 11 October, at about the same time as the wind stress reaches its maximum value. Subsequently, salinity in the upper 50 m remains nearly uniform for about 10 days. The observations show near-inertial oscillations in temperature (maximum amplitude 1.2°C), salinity (up to 1.5 psu) and potential density (up to 1.24 kg m^{-3}) at 75 m during 12–22 October—we discuss

the role of the near-inertial oscillations later in the section.

Before the storm, all three moorings in the observatory show significant vertical salinity gradients at shallow depths (Figs. 3c,e,f), with prestorm salinity differences between the surface and 50 m exceeding 1.5 psu. Although there are some differences in salinity between moorings, mixing induced by Cyclone Phailin makes the upper-ocean salinity nearly uniform down to at least 50-m depth at all three moorings. Surface salinity at the moorings increases by 1.2–1.8 psu within 12 h. Salinity remains well-mixed up to 50 m from 12 to 20 October (Figs. 3c,e,f).

Restratification of the upper ocean after storm-induced mixing has been reported in many previous studies based on both observations and models (Price et al. 2008; Mei and Pasquero 2012; Haney et al. 2012; Mrvaljevic et al. 2013). Many processes can contribute to restratification, including air–sea heat flux, Ekman buoyancy flux, geostrophic adjustment, and mixed layer eddies. Restrartification begins between 30- and 50-m depth, and somewhat later at shallower depths. For example, the INCOIS mooring (INC) shows that 50-m salinity reaches close to the prestorm values on 25 October, whereas salinity stratification in the upper 20 m becomes significant only after 27 October (Fig. 3f). The recovery of the shallowest measured potential temperature (5-m depth at BD09; Fig. 3b) begins immediately after the passage of Cyclone Phailin. The observed diurnal cycle of 5-m temperature is a response to diurnally varying surface heat flux (not shown). Unlike temperature, salinity remains nearly uniform in the upper 50 m for about

FIG. 3. Hourly observations of (a) wind stress (N m^{-2}), (b) potential temperature ($^{\circ}\text{C}$), (c) salinity (psu), and (d) potential density (kg m^{-3}) from mooring BD09 at different depths; (e), (f) salinity (psu) from moorings BD08 and INCOIS at different depths. Sensor depths for BD08/BD09 and INCOIS are mentioned in (d) and (f). Standard error (red bars) estimates of wind stress are marked in (a). All three moorings show that cyclone-induced vertical mixing extended to at least 50-m depth.

10 days, as noted earlier. The very low-salinity water in the upper 10 m seen at all three moorings before 9 October, when Cyclone Phailin begins to mix the upper ocean, is of riverine origin (Sengupta et al. 2016). Near-surface salinity falls gradually from 20 October onward, and more sharply from 25 or 26 October, but the very fresh prestorm values are not seen again until 30 October (Figs. 3c,e,f).

b. Observed upper-ocean salinity and vertical mixing

In a stratified ocean, the vertical shear of horizontal velocity can generate three-dimensional turbulence, while density stratification tends to quench turbulence. The square of the Brunt–Väisälä frequency $N^2 = -(g/\sigma_\theta)(d\sigma_\theta/dz)$ is used as a measure of density stratification, where σ_θ is potential density and z is depth; squared shear is defined as $S^2 = (dU/dz)^2 + (dV/dz)^2$, where U and V are eastward and northward components of velocity. The stability of shear flow in a stratified fluid is determined by the nondimensional Richardson number (Ri), defined as the ratio of N^2 to S^2 . Linear theory and laboratory measurements show that shear-induced turbulent mixing occurs when $Ri = N^2/S^2 < 0.25$. In the absence

of turbulence measurements, the depth at which reduced shear $S^2 - 4N^2$ goes to zero is often used as a proxy to demarcate the depth of the turbulent layer in the upper-ocean and laminar regions (Kunze et al. 1990; Sanford et al. 2011).

The high values of N^2 at mooring BD09 indicate very stable near-surface stratification in the prestorm north Bay of Bengal (Fig. 4b). The value of N^2 between 5 and 10 m is $0.7 \times 10^{-3} \text{ s}^{-2}$ (the vertical resolution of the temperature and salinity data is coarse, as mentioned earlier). The high near-surface values of N^2 are no longer seen after 11 October. In response to the increased surface wind stress associated with the cyclone, the depth of maximum S^2 at the base of the mixed layer increases monotonically, as the mixed layer deepens (Fig. 4c). Across the mixed layer base, $S^2 - 4N^2$ is positive, indicating that shear generated turbulence is likely responsible for deepening the mixed layer. This is confirmed by modeling in section 6.

Mixed layer depth is defined as the depth at which potential density exceeds its surface value by 0.05 kg m^{-3} . The deepening of MLD happens in 10–11 h (0000–1100 UTC

FIG. 4. Hourly (a) wind stress (N m^{-2}), (b) Brunt-Väisälä frequency-squared ($N^2; \text{s}^{-2}$), (c) shear-squared ($S^2; \text{s}^{-2}$), and (d) reduced shear ($S^2 - 4N^2; \text{s}^{-2}$) based on observations at mooring BD09. The gray line in (b) and (c) marks the mixed layer depth (m), defined as the depth at which potential density exceeds its surface value by 0.05 kg m^{-3} ; the black contour in (d) marks $S^2 - 4N^2 = 0$. The N^2 , S^2 , and reduced shear have been smoothed by a 3-h moving average. The deepening of the mixed layer (11 Oct) is due to shear-generated turbulence. The black horizontal bar above (b) marks the 11-h period of largest mixed layer deepening.

11 October), approximately one-fourth of the inertial period. The moored observations broadly agree with the predictions of Pollard et al. (1972), who showed that near-inertial shear deepens the mixed layer to 85% of its maximum value within one-fourth of an inertial period. After a half-inertial period, mixed layer deepening is arrested by rotation at a value about 95% of its maximum depth. The zero contour of reduced shear and the mixed layer depth track each other (Figs. 4c,d), indicating that the mixed layer deepens under the action of shear-generated turbulence. The high values of S^2 at subsurface depths below 60 m after 14 October are generated by inertial wave propagation (see section 4c).

SST cooling due to vertical mixing by postmonsoon cyclones in the north Bay of Bengal is generally minimal because a strong, shallow halocline lies above a deep, warm layer (Sengupta et al. 2008). In the case of Cyclone Phailin, moored observations show a distinct shallow halocline between 5- and 20-m depth; the prestorm salinity difference is nearly 2 psu between 5- and 50-m

depth. A 50 m deep warm subsurface layer is present, with a modest temperature inversion. In spite of storm-induced vertical mixing extending to 50-m depth, SST cooling at the mooring is only about -0.1°C (Fig. 5a). However, storm-induced vertical mixing of the salt-stratified upper ocean leads to a surface salinity increase of 1.6 psu (Fig. 5b). Changes in surface salinity and evolution of subsurface salinity are comparable at the three moorings INC, BD08, and BD09 in the cluster NBoB (Figs. 3c,e,f), as a result of shear-induced mixing.

The net change in sea surface salinity due to Cyclone Phailin as seen from the *Aquarius* satellite is consistent with the moored observations. Before the storm, the *Aquarius* Level 3 gridded SSS product shows salinities lower than 32 psu to the north of 15°N (Fig. 5d). After the storm has passed, SSS is seen to increase over the entire northern Bay of Bengal, an area of nearly one million square kilometers. The SSS increase is largest in the northwest Bay of Bengal, where prestorm SSS is lowest (Fig. 5f); the maximum net SSS

FIG. 5. Mean prestorm (blue) and poststorm (red) profiles of (a) potential temperature ($^{\circ}\text{C}$), (b) salinity (psu), and (c) potential density (kg m^{-3}). (d) Prestorm *Aquarius* SSS (psu). (e) Poststorm minus prestorm SSS (ΔS); prestorm SSS contours (contour interval is 0.8 psu) are overlaid. The diamond and square represent the positions of the moorings at 16.5° and 18°N . *Aquarius* SSS increases by 1–4 psu, with the largest increase in the region of lowest prestorm SSS. We define 7–8 Oct and 12–13 Oct as the pre- and poststorm periods for SST assessment.

change exceeds 5 psu in a small region close to the cyclone track off the Odisha coast. Unfortunately, there are no Argo float observations in this area. At NBoB, the value of poststorm minus prestorm *Aquarius* SSS is nearly 1.6 psu, in close agreement with moored observations.

There are few previous reports of an increase of SSS due to storm-induced mixing. A 2.62-psu rise of near-surface salinity has been reported by Vissa et al. (2013) near 17°N in the Bay of Bengal during Cyclone Sidr from Argo float measurements. Domingues et al. (2015) measure 0.6-psu change in 10-m salinity from glider measurement north of Puerto Rico due to Hurricane Gonzalo in October 2014. Grodsky et al. (2012) report a rise of SSS by 1–2 psu in the Amazon/Orinoco plume due to the passage of Hurricane Katia in August–September 2011. The *Aquarius* data record SSS changes due to storm-induced mixing as well as advection by surface currents associated with mesoscale eddies and basin-scale circulation (Sengupta et al. 2016). The East India Coastal Current (EICC) is usually set up in late October or November each year (Shankar et al. 1996). This is a time of rapid change in the basin-scale

circulation of the Bay of Bengal. The decrease in SSS (negative values of poststorm minus prestorm SSS) in the eastern and western basin near 15°N are likely to be due to basin-scale circulation.

c. Inertial motions

Inertial oscillations play an important role in the response of the ocean to tropical cyclones. Mooring BD09 lies to the right of the storm track, where the near-inertial response of the ocean is enhanced because the surface wind vector turns in the same direction as the inertial current (Price 1981). Hourly observations of horizontal velocity from the moored ADCP shows that the storm generated near-inertial currents (Figs. 6a,b) from 9 to 22 October. The inertial period at 18°N is 39.5 h; we use a 30–46-h Butterworth filter to extract the near-inertial signal in currents, temperature and salinity. As near-inertial energy propagates out of the mixed layer to deeper levels, surface mixed layer currents gradually diminish. Note that the subsurface near-inertial signal below 70 m or so has higher amplitudes than those in the mixed layer (Figs. 6a–d). The upward phase propagation in the zonal and

FIG. 6. Hourly (a),(b) zonal velocity U and meridional velocity V (m s^{-1}) from the ADCP on mooring BD09 and (c),(d) bandpassed (30–46 h) near-inertial U and V . (e) Unfiltered (blue) and 12-h smoothed (red) pycnoline displacement (m) based on observations at 75-m depth and (f) bandpassed near-inertial pycnoline displacement at 75-m depth. The subsurface near-inertial signal has higher amplitudes than those in the mixed layer.

meridional components of velocity (U and V) indicates that near-inertial waves carry energy out of the mixed layer to the subsurface ocean (Leaman and Sanford 1975; Gill 1984; D'Asaro et al. 1995). Here, the near-inertial signal at subsurface depths persists for at least five inertial periods; further, most of the enhanced vertical shear of raw, unfiltered velocity (Fig. 4c) is contributed by near-inertial currents. The sustained shear at subsurface depths is consistent with previous observational studies, such as the ocean's response to Typhoon Fanapi in the northeast Pacific (Hormann et al. 2014). Near-inertial currents in the upper mixed layer induce horizontal divergence and pressure differences in the subsurface ocean, resulting in heaving of subsurface isopycnals at the local inertial frequency. The vertical displacement of isopycnals η can be estimated from

$$\eta = \frac{\Delta\rho}{\frac{d\rho}{dz}}, \quad (2)$$

where $\Delta\rho$ is the amplitude of density oscillations. We infer peak-to-peak vertical isopycnal displacements of about 20 m at 75-m depth (Figs. 6e,f). These are also dominantly of near-inertial frequency below 60-m depth (Figs. 3b,c) and thus are also due to near-inertial waves (Gill 1984) as has been reported in several previous studies (e.g., Black and Dickey

2008; Oey et al. 2007; Sanford et al. 2011; Rayson et al. 2015).

5. Response of the ocean close to the cyclone track

a. Description of response

Measurements from mooring BD10, which lies under the storm track, show maximum 10-min average wind speed of 39 m s^{-1} and minimum sea level pressure of 920.6 hPa at 0800 UTC 11 October; at this time the pressure drop at the center of the cyclone is at least 80 hPa (Fig. 2a). Note that wind speed is not zero at the time of minimum surface pressure (Figs. 2a,b); the detailed evolution of hourly surface pressure and wind vectors indicate that the storm track passed 1–3 km to the right of the mooring.

Upper-ocean potential temperature, salinity, and potential density at different depths measured at BD10 are shown in Fig. 7. The prestorm salinity at 10-m depth (the CT sensors at 5, 20, and 30 m failed) is 30 psu at BD10, nearly 2 psu lower than at 18°N (i.e., at moorings BD08, BD09, and INC). In response to the storm, salinity changes by nearly 3 psu and potential density by nearly 2 kg m^{-3} in the upper 15 m of the ocean. Upper-ocean restratification appears to begin almost immediately after the peak wind stress passes the mooring, whereas restratification begins after about 10 days

FIG. 7. Hourly (a) wind stress (N m^{-2}), (b) potential temperature ($^{\circ}\text{C}$), (c) salinity (psu), and (d) potential density (kg m^{-3}) from mooring BD10 at different depths. Standard error (red bars) estimates of wind stress are marked in (a).

at 18°N (see Fig. 3). Prestorm potential temperature and density at 75- and 100-m depth show a signature of internal waves with the period of the semidiurnal tide. The vertical spacing of working CT sensors is too coarse to estimate mixed layer depth. Although maximum wind stress at BD10 exceeds 4 N m^{-2} (nearly 5 times larger than that at the 18°N moorings), data from BD10 suggests that mixed layer deepening is short-lived, and the net increase of MLD is smaller than at 18°N .

b. Currents and shear

The local inertial period at BD10 is about 42 h. The hourly velocity observations suggest strong near-inertial currents in the mixed layer (Figs. 8c,d), as well as more persistent near-inertial oscillations at depths below 30 m (peak-to-peak amplitudes exceed 1.5 m s^{-1} at the surface and 0.75 m s^{-1} at 75-m depth). We use a broadband (36–60 h) filter to extract the near-inertial signal at this location (Figs. 8e,f) to minimize spectral leakage. The observations show a rapid decay of near-inertial currents in the upper mixed layer and suggest that mixed-layer energy is carried to subsurface by large-amplitude near-inertial waves (Figs. 8c–f). Near-inertial velocities below the mixed layer are nearly 1.5 times larger at

mooring BD10 (0.75 m s^{-1}) than at BD09 mooring (0.5 m s^{-1} , see Fig. 6). Maximum values of shear-squared S^2 at BD10 are 2 times as large as maximum S^2 at the 18°N moorings, but the vertical resolution of CT data is not adequate to estimate N^2 or reduced shear. In the next section, we use one-dimensional model simulations to aid further analysis of the ocean response away from the storm track and directly under the storm track.

6. One-dimensional model

a. Model setup and forcing

The one-dimensional PWP model (Price et al. 1986) is used to simulate the response of the ocean mixed layer to Cyclone Phailin at moorings BD09 and BD10. The PWP model has three mixing parameterizations based on static stability, bulk Richardson number, and gradient Richardson number that determine vertical mixing in response to surface momentum, heat, and freshwater fluxes. However, the model cannot simulate divergence of mixed layer currents and inertial pumping, or the energy transfer from the ocean

FIG. 8. Hourly (a) wind stress (N m^{-2}), (b) shear-squared S^2 (smoothed by 3-h moving average on both time and depth). The black contour marks $S^2 = 1 \times 10^{-3} \text{ s}^{-2}$. The 12-h moving average of (c) zonal velocity (m s^{-1}) and (d) meridional velocity (m s^{-1}) at mooring BD10. Bandpassed (36–60 h) near-inertial (e) zonal current and (f) meridional current from mooring BD10.

mixed layer to subsurface depths by near-inertial waves (e.g., Pollard and Millard 1970; D'Asaro 1985). Instead, this decay is modeled by an empirical decay constant r_d that is chosen to match the observed decay rate. Previous observations suggest values from a few days to 3 weeks (Van Meurs 1998; Park et al. 2009).

The PWP simulations start at 0000 UTC 7 October from observed temperature and salinity profiles and zero initial velocity and end at 2300 UTC 23 October. Surface forcing is derived from hourly moored observations with the COARE 3.0 algorithm (Fairall et al. 2003). COARE 3.0 assumes wind stress uncertainty of 5% for wind speeds of $0\text{--}10 \text{ m s}^{-1}$ and 10% for wind speeds between 10 and 20 m s^{-1} . At wind speed higher than 20 m s^{-1} , we follow the prescription of Powell et al. (2003). The intensity of a tropical cyclone depends strongly on its wind stress as it can limit the intensity by extracting energy and momentum from the storm. So the accuracy at higher wind speeds is essential for both forecasting and predicting the oceanic response of the tropical cyclones. Model simulations are sensitive to the value r_d . Values of 5 day^{-1} at BD09 (away from the storm track), and 1.5 day^{-1} at BD10 (close to the storm track) are chosen to best fit the data. Possible reasons for the difference between decay time scales away from the

track of Phailin and close to the track are discussed in appendix A. Sensitivity of the mixing to these choices is discussed in section 6c.

We compare the time evolution of hourly temperature T , salinity S and zonal and meridional velocity (U and V) from the PWP model with observations from moorings BD09 and BD10. The observations occasionally have a prominent semidiurnal variability of temperature and salinity (e.g., Fig. 7) and in velocity (not shown). In addition, a slowly varying background flow is evident at all depths in the observed velocity field, for instance the background flow at BD10 is northwestward and has a typical speed of 24 cm s^{-1} (see Figs. 8c,d). Therefore, (i) we smooth the hourly T , S , U , and V observations by a 15-h running average to suppress internal waves with semidiurnal (or shorter) period, and (ii) we subtract the 7–23 October time mean of 90–100-m depth-averaged currents from U and V at all depths to suppress the slowly varying background flow associated with mesoscale eddies. After removal of the semidiurnal tide and the slowly varying flow from observations, we present model–data comparisons as (i) time–depth diagrams and (ii) line plots of T , S , U , and V at selected depths. Net uncertainties arising from surface forcing are indicated by error bars in all figures.

FIG. 9. Hourly BD09 observations (blue) and model simulated (red) (a)–(e) temperature ($^{\circ}\text{C}$), (f)–(j) salinity (psu), (k)–(o) zonal velocity (m s^{-1}), and (p)–(t) meridional velocity (m s^{-1}) at selected depths. All quantities are smoothed by a 12-h running average. Standard error bars indicate uncertainty arising from surface forcing.

b. Ocean response away from the cyclone track

At the mooring BD09, there is good agreement between U and V from the PWP model and the observations during the initial storm-forced period. From the surface to 30-m depth, both model and observations show near-inertial currents with maximum speeds that can exceed 0.5 m s^{-1} (Figs. 9k–m, p–r). The model underestimates near-inertial U and V at 50 m (Figs. 9n, s), which is close to the maximum mixed layer depth (see below). The PWP model suggests that vertical shear of storm-generated currents is responsible for rapid deepening of the mixed layer (Price et al. 1986). Sensitivity experiments to isolate the relative contribution of the different mixing processes parameterized in the PWP model indicate that the initial deepening of the mixed layer is mainly due to mixing determined by the bulk Richardson number criterion (R_b mixing): the mixed layer depth reaches 95% of its maximum value when the model is run with R_b mixing alone (Simulation F09a).

The time evolution of model and observed mixed layer T and S are broadly in agreement. The effects of storm-induced mixing are most clearly seen in the increase in near-surface salinity and a 0.4–0.5-psu drop in subsurface salinity at 30- and 50-m depth (Figs. 9f–i). The one-dimensional model is not expected to capture the near-inertial T and S response at 75- and 100-m depth in the moored observations (Figs. 9e, j), which are due to inertial pumping. We carried out sensitivity experiments (Simulations B09b, B09c) with observed wind stress reduced and enhanced by 5%–10% relative to the control run (Simulation B09a). The experiments indicate that the large uncertainty of simulated temperature and salinity at 50 m (see error bars in Figs. 9d, i) arises from the uncertainty in estimates of surface fluxes and wind stress (Fairall et al. 2003). The net SST change (poststorm minus prestorm SST) in the model is $-0.14^{\circ} \pm 0.03^{\circ}\text{C}$, whereas the net change in model SSS is 1.43 ± 0.03 psu. We note that the net change in SST and SSS in response to storm-induced mixing agrees with the

FIG. 10. As in Fig. 10, but for BD10.

moored observations. Observed and model temperature at 5 and 15 m both warm gradually after 12 October (Fig. 9a). The model does not show the increase of mixed layer salinity after 15 October and the subsequent freshening (Fig. 9f).

c. Ocean response close to the cyclone track

We depict the time evolution of upper-ocean T , S , U , and V from observations at mooring BD10 and the PWP model in the same manner as above. The vertical spacing of functioning CT sensors on this mooring is too coarse for reliable MLD estimates, as noted earlier (Fig. 10). Since the shallowest CT sensor is at 10-m depth, model initial conditions assume uniform T and S in the uppermost 10 m. The BD10 observations suggest that the upper-ocean response to Cyclone Phailin is only one-dimensional for a short time during the storm passage as the three-dimensional effect becomes important immediately thereafter. The time evolution of model temperature at 10- and 15-m depth agrees with observations until 12 October (Figs. 10a,b). The rise of 10-m salinity due to vertical mixing in the model is somewhat smaller than the observed rise of nearly 3 psu on 10 and 11 October (Figs. 10e,f).

To assess net change due to cyclone passage, we consider the SST (or SSS) difference between the time when model MLD is deepest (12–14 October) and

the prestorm value of SST (or SSS). The observed net change in 10-m temperature and salinity at mooring BD10 is -0.77°C and 2.93 psu, respectively, whereas the net change in model SST and SSS is $-0.73^{\circ} \pm 0.04^{\circ}\text{C}$ and 2.02 ± 0.04 psu, respectively. In the storm-forced phase, moored observations and the model suggest vertical mixing of the prestorm warm, low-salinity layer to 40-m depth due to vertical shear of storm-induced currents. There is limited agreement between the model and observed currents in the upper 15 m. The observed disappearance of near-inertial currents in 2–3 inertial periods is the most important dynamical difference from the response at the 18°N moorings. A sensitivity experiment (Simulation F10d) with zero decay rate indicates (not shown) the deepening of the mixed layer depth (MLD) to 50-m depth instead of 40-m depth in the control run (Simulation B10a). The rapid decay of mixed layer near-inertial currents may be related to lateral vorticity gradients on 100-km scales (see appendix A). To further assess the contribution of heat fluxes and wind stress in storm-induced SST cooling, we have conducted experiments by setting (i) all air–sea heat fluxes to zero (Simulation F10c) and (ii) wind stress to zero (Simulation F10b). The results (not shown) agree with previous studies (e.g., D’Asaro et al. 2007), which show that nearly 85% of the SST cooling under tropical cyclones is due to mixing produced by wind stress.

FIG. 11. (a) Observed initial conditions for the PWP model: observed initial salinity (red), constant salinity initial condition (red dotted), and observed temperature (blue) profiles from BD09; (b) wind stress (N m^{-2}); (c) mixed layer depth (blue); and model mixed layer depth calculated with constant salinity initial condition (orange) and observed salinity initial condition (gray). Standard error bars indicate uncertainty arising from surface forcing. The depths of measurements are marked by black dots on the y axis of (a). The depth of vertical mixing increases by nearly 55% in the absence of salinity stratification.

d. Effect of near-surface salinity stratification

The response of the upper ocean to cyclones is sensitive to the vertical structure of prestorm salinity and temperature. To assess the effect of prestorm salinity stratification on ocean response away from the storm track (mooring BD09) and close to the storm track (mooring BD10), we performed additional numerical experiments: In the control simulations (Simulations B09a and B10a), initial T and S profiles are based on the moored observations. In the sensitivity experiments (Simulations S09a and S10a), we initialize the model with vertically uniform salinity of 34.5 psu at BD09 and 35 psu at BD10. The initial temperature profiles and surface forcing are identical to the control runs (B09a and B10a).

With no initial salinity stratification in the upper ocean (Simulation S09a) model MLD reaches 70 ± 3 m in response to storm-induced mixing, as compared to about 45–50 m in the control run (B09a) or in the BD09 observations (Fig. 11). The shallow stratification is sufficiently strong that it prevents nighttime deepening of the mixed layer during the prestorm period 7–10 October, when the sky is clear and wind speed is low (Fig. 11c). In the absence of salinity stratification, however, MLD responds to diurnal variation of surface fluxes (not shown), deepening from 5 m or less in the daytime to nearly 50 m at about 0000 UTC (local dawn). In both the control simulation

(Simulation B09a) and experiment (Simulation S09a), storm-induced mixing begins in the early hours of 10 October, and the deepest MLD is achieved on 13 October (Fig. 11c). Net SST cooling (poststorm minus prestorm SST) away from the cyclone track in the sensitivity experiment (Simulation S09a) is $-0.16 \pm 0.04^\circ\text{C}$, which is only 0.02°C greater than in the control run (Simulation B09a). Although the depth of vertical mixing increases by nearly 55% in the absence of salinity stratification, storm-induced SST cooling remains modest because the prestorm ocean has a deep, warm subsurface layer.

A similar sensitivity experiment at mooring BD10 (Simulation S10a) indicates that prestorm salinity stratification has significant influence on SST cooling under the cyclone track. At this location, the prestorm upper ocean is stratified by both salinity and temperature. The salinity, potential temperature, and potential density difference between 10- and 75-m depth are nearly 4 psu, 4°C , and 5 kg m^{-3} , respectively. In response to the cyclone, model MLD at mooring BD10 in the control run (Simulation B10a) and the sensitivity experiment (Simulation S10a) reach maximum depths of about 40 ± 5 m and 60 ± 5 m, respectively, due to deeper mixing in the absence of salinity stratification. Net storm-induced SST cooling is -0.77°C in the moored observations and in the control run (Simulation B10a), whereas it is nearly $-1.4 \pm 0.12^\circ\text{C}$ in the sensitivity experiment (Simulation S10a; Fig. 12d). Net changes

FIG. 12. (a) Observed initial conditions for the PWP model for observed initial salinity (red), constant salinity initial condition (black), and observed temperature (blue) profile from BD10; (b) observed wind stress (N m^{-2}); (c) mixed layer depth (m) calculated from the PWP model with constant salinity initial condition (orange) and observed salinity initial condition (gray) from BD10. (d) Observed prestorm temperature (blue) and model poststorm temperature profiles for observed salinity initial condition (green) and constant salinity initial condition (black) BD10. Standard error bars indicate uncertainty arising from surface forcing. The depths of measurements are marked by black dots on the y axis of (a). The existence of prestorm salinity stratification at 5–20-m depth reduces ΔSST by 0.7°C , i.e., half of the observed value of ΔSST (see Table 3).

in observed and model SST and SSS at moorings BD09 and BD10 in response to the cyclone are listed in Table 3.

e. The Arabian Sea profile

To further assess the influence of the strong salinity stratification of the Bay of Bengal on cyclone-induced mixing we performed a sensitivity experiment (Simulation SARA) with initial T and S profiles from the Arabian Sea. Several prestorm T and S profiles from Argo float 2901432 during 15–30 September 2013 were chosen and averaged, from a $2^\circ \times 2^\circ$ box centered at 15.5°N , 66.6°E in the Arabian Sea. Unlike the Bay of Bengal, upper-ocean salinity stratification in the Arabian Sea is very small at this time, and potential density is determined almost entirely by temperature (Tomiczak and Godfrey 2003). Prestorm salinity and temperature differences between 5- and 75-m depth are 0 psu and 2.7°C , respectively, for the Arabian Sea profile (Fig. 13a). Surface forcing for the sensitivity experiments is identical to that at Bay of Bengal mooring BD10 (Simulation B10a). The experiments (SARA) show that starting from a prestorm MLD of 15 m, model MLD reaches 75 ± 3 m. The absence of prestorm salinity stratification and significant vertical

temperature gradients results in net storm-induced SST cooling of $-1.67^\circ \pm 0.06^\circ\text{C}$ (Fig. 13d), which is nearly 2.5 times larger than in the Bay of Bengal.

7. Discussion and conclusions

Tropical Cyclone Phailin developed from a low pressure system in the Andaman Sea and intensified to a category 5 storm before making landfall in Gopalpur on 12 October 2013, with a minimum surface pressure of 920 hPa and maximum sustained wind speed of 60 m s^{-1} . We study the response of the salinity-stratified Bay of Bengal to the passage of Phailin.

Our work is divided into three parts. In the first part, we study the response of the upper ocean with the help of in situ observations from moorings and *Aquarius* sea surface salinity data. Moored observations show that net SST cooling in response to Cyclone Phailin is small or modest (-0.10°C near 18°N , 89.5°E , 200 km to the right of the storm track, and -0.77°C at mooring BD10 located at 16.5°N , 86.5°E , almost directly under the track), due to the presence of a shallow halocline and a deep barrier layer in the prestorm upper ocean. Three moorings near 18°N showed a net sea surface salinity change of 1.4–1.6 psu due to cyclone-induced mixing.

TABLE 3. Comparison between net change (poststorm minus prestorm) of SST ($^{\circ}\text{C}$), SSS (psu), and MLD (m) due to Cyclone Phailin from observations and model at moorings BD09 and BD10.

	BD09			BD10		
	ΔSST ($^{\circ}\text{C}$)	ΔSSS (psu)	Prestorm (poststorm) MLD (m)	ΔSST ($^{\circ}\text{C}$)	ΔSSS (psu)	Prestorm (poststorm) MLD (m)
Observed	-0.1	1.46	5 (50)	-0.77	2.93	10 (15) ^a
B09a, B10a	-0.14 ± 0.03	1.43 ± 0.03	5 (45 ± 3)	-0.73 ± 0.04	2.02 ± 0.05	10 (40 ± 5)
S09a, S10a	-0.16 ± 0.04	0.05 ± 0.01	10 (70 ± 4)	-1.4 ± 0.12	0.1 ± 0.01	15 (60 ± 4)

^a The vertical spacing of functioning CT sensors on the mooring BD10 is too coarse for reliable MLD estimates.

Surface salinity increased by 2.93 psu at mooring BD10, close to the storm track, due to storm-induced vertical mixing. The poststorm minus prestorm SSS from the *Aquarius* satellite is in good agreement with mooring measurements at 18°N , but not at 16.5°N . *Aquarius* data show that sea surface salinity increased throughout the northern Bay of Bengal in response to the storm, indicating widespread vertical mixing of the upper ocean over an area of nearly a million square kilometers.

The largest-amplitude inertial currents (1.5 m s^{-1}) are observed directly under the cyclone track. After generation, they disperse from the mixed layer to the thermocline with a decay time scale of about 1.5 days at 16.5°N and 5 days at 18°N . Estimates of Richardson number from moored density stratification and velocity

data suggest that shear-induced mixing associated with storm-generated currents deepens the mixed layer from prestorm values of less than 10 m to poststorm values of about 50 m at the 18°N moorings. Observations at BD10 under the storm track show the signature of short-lived mixing down to nearly 100-m depth.

The second part of our work deals mainly with modeling the ocean response to Cyclone Phailin. We perform experiments with the one-dimensional PWP model to simulate the response of the ocean mixed layer at moorings BD09 (north of the storm track) and BD10 (directly beneath the storm track). Both forcing and initial conditions are taken from observations. The decay rate of inertial motions is chosen to best fit the observations 5 day^{-1} at BD09 and 1.5 day^{-1} at BD10.

FIG. 13. (a) Initial profiles of salinity (red) and temperature (blue) taken from Argo float 2901432 in the eastern Arabian sea. (b) Wind stress (N m^{-2}) from mooring BD10, (c) mixed layer depth (m) calculated from the PWP model starting from Argo initial conditions and forced by BD10 fluxes (gray). (d) Observed prestorm temperature profile (blue) and model poststorm temperature profile (green). The depths of T and S are marked by black dots on the y axis of (a). The initial conditions have (i) substantial vertical temperature gradients in the upper ocean and (ii) insignificant salinity stratification up to 75-m depth. The PWP model shows net storm-induced SST cooling is $-1.67^{\circ} \pm 0.06^{\circ}\text{C}$, nearly 2.5 times larger than in the Bay of Bengal.

Away from the cyclone track (at BD09), simulated mixed layer T , S , and horizontal velocities agree well with observations. However, beneath the track (at BD10), model results differ from observations, except during a short period of intense mixing as the storm approaches the mooring. We propose that both the poor simulation and the rapid decay of inertial motions is due to the presence of mesoscale vorticity gradients near BD10, which is not included in the simulation. Sensitivity experiments (Simulations F09b, F09c, F10b, and F10c) to quantify the relative effect of surface heat flux and wind stress forcing indicate that wind stress is the primary driver of mixing. The PWP model indicates that preexisting strong salinity stratification at 5–20-m depth reduces the maximum depth of storm-induced vertical mixing by about 50%.

In the final part, we briefly address an important question. What is unique about the Bay of Bengal response to tropical cyclones? Copious summer monsoon rainfall and discharge from the Ganges–Brahmaputra–Meghna river create a shallow, fresh mixed layer in the north Bay of Bengal, which caps a deep warm layer in the postmonsoon season. Several previous studies (e.g., Sengupta et al. 2008; Neetu et al. 2012; Balaguru et al. 2012) suggest that two mechanisms related to Bay of Bengal stratification can impact cyclone intensification: (i) strong salinity stratification reduces the depth of cyclone-induced mixing and (ii) a weak vertical temperature gradient in the upper ocean reduces the net SST change (ΔSST) due to vertical mixing. We carry out model sensitivity experiments (Simulations S10a and SARA), and find that i and ii have comparable effects on SST cooling: near-surface salinity stratification reduces ΔSST by a factor of 2, and the presence of a deep warm barrier layer reduces ΔSST by a factor of 2.5. These experiments suggest that the absence of SST cooling in the open ocean may contribute to the rapid intensification of postmonsoon (October–November) Bay of Bengal cyclones.

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APPENDIX A

Dispersion of Near-Inertial Current

Gill (1984) explained the dispersion of near-inertial currents from the mixed layer to subsurface depths with the help of modal decomposition. He showed that the rotation of a given vertical mode depends on the mode number. The estimated decay scale t_n for the n th mode to become 90° out of phase with a wave rotating at the inertial frequency is

$$t_n = \pi f_0 / (k^2 + l^2) c_n^2 \quad (\text{A1})$$

where f_0 is local inertial frequency, (k, l) are the horizontal wavenumber and c_n is the phase speed of the n th mode. D'Asaro (1989) showed that in the presence of the planetary vorticity gradient β , the decay time scale of mixed layer currents is reduced significantly due to radiation of near-inertial waves to the subsurface ocean. Subsequently, Van Meurs (1998) established that in the presence of mesoscale eddies, the local gradient of relative vorticity can be as important as the planetary vorticity gradient. The characteristic time scale of decay can be estimated from

$$t_n^3 = 3\pi / f_0 \nabla \zeta \cdot \nabla \zeta R_n^2 \quad (\text{A2})$$

where f_0 is local inertial frequency, $\nabla \zeta$ is the gradient of relative vorticity, R_n the Rossby radius of deformation for the n th vertical mode, and t_n the decay scale for the n th mode. From AVISO sea surface dynamic topography and geostrophic velocity, we find that mooring BD10 laid between two counterrotating mesoscale eddies, during the passage of Cyclone Phailin (Fig. S2). The estimated relative vorticity gradient $\nabla \zeta$ at the mooring is $-2 \times 10^{-10} \text{ s}^{-1} \text{ m}^{-1}$, nearly 10 times the planetary vorticity gradient β . Considering only the planetary vorticity gradient (i.e., $\nabla \zeta = \beta$), Eq. (A2) gives a decay scale t_1 of about 5 days for the first baroclinic Rossby radius $R_1 = 6 \times 10^4 \text{ m}$ (Chelton et al. 1998); however, if we account for the observed relative vorticity gradient, t_1 reduces to 1.3 days. Note that these values represent lower limits for the decay scale, since R_n decreases with increasing mode number. We propose that the spatial gradient in relative vorticity is the main reason for the observed rapid dispersion of mixed layer near-inertial currents to the subsurface at mooring BD10.

APPENDIX B

Comparison of *Aquarius* L3 SSS and INCOIS Mooring

Comparison of *Aquarius* Level 3 (L3) gridded SSS with salinity measurements at 1-m depth from a mooring deployed by the INCOIS shows that (i) the mean overall bias of *Aquarius* SSS from January to November 2013 is less than 0.1 psu, but that the satellite SSS is generally biased high when salinity is lower than 30 psu (Sengupta et al. 2016), and (ii) *Aquarius* estimates can have substantial bias during the passage of Cyclone Phailin (Fig. S1).

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