High-frequency time-series autonomous observations of sea surface $p$CO$_2$ and pH

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Abstract

Carbon dioxide partial pressure ($p$CO$_2$) in surface water was continuously measured every 3 h from July 2012 to June 2013 using an autonomous $p$CO$_2$ system (MAPCO$_2$) deployed on a moored buoy on the East China Sea shelf (31°N, 124.5°E). Sea surface $p$CO$_2$ and pH had the largest variations in summer, ranging from 215 to 470 μatm, and 7.941 to 8.263 (averagely 8.084 ± 0.080), respectively. They varied little in winter, ranging from 328 to 395 μatm, and 8.003 to 8.074 (averagely 8.052 ± 0.010), respectively. The seasonal average sea surface $p$CO$_2$ was respectively 335 ± 70 μatm, 422 ± 43 μatm, 362 ± 11 μatm, and 311 ± 59 μatm in summer, autumn, winter, and spring, and was overall undersaturated with respect to atmosphere on a yearly basis. Although the average sea surface $p$CO$_2$ in summer was below the atmospheric level, the net CO$_2$ flux has suggested a CO$_2$ source status due to the influence of typhoon. Our observation thus demonstrated the significant, even dominant impact of episodic typhoon events on surface ocean CO$_2$ chemistry and air–sea CO$_2$ gas exchange, which would be impossible to capture by shipboard observation. The high wind stress and curl associated with the northward movement of typhoon induced complex sea surface water movement, vertical mixing, and subsequent biological drawdown, which differed in pre-, onset, and post-typhoon stages. Based on our estimates, the degassing fluxes during typhoon reached as high as 82 mmol m$^{-2}$ CO$_2$ and 242 mmol m$^{-2}$ CO$_2$ in summer and autumn, respectively, accounting for twice as large as the summer CO$_2$ sink during non-typhoon period, and 28% of the total CO$_2$ source in autumn.

Despite having a relatively modest surface area, the coastal ocean plays a considerable role in the uptake of atmospheric CO$_2$ (0.2–0.5 Pg C yr$^{-1}$, 1 Pg = 10$^{15}$ g) (Bauer et al. 2013; Dai et al. 2013; Laruelle et al. 2014). The large uncertainties associated with estimations of coastal air–sea CO$_2$ fluxes, which often possess highly dynamic ranges, are largely due to the poor spatial, and especially the temporal, coverage of field observations and measurements (Gruber 2015). As a consequence, the current accuracy of coastal ocean carbon budgets remains insufficient to be included in the earth system model for reliably predicting future climate changes (Cai 2011; Gruber 2015).

Both physical and biological processes are known to affect CO$_2$ dynamics, including the partial pressure of CO$_2$ ($p$CO$_2$) in coastal waters over various time scales spanning from diurnal to seasonal changes. These processes can be further divided into those which lower sea surface $p$CO$_2$, such as the biological removal of carbon fueled by riverine or subsurface
inputs of nutrients (Chen et al. 2012; Wang et al. 2017a,b; Li et al. 2018), and those which increase pCO$_2$, such as vertical mixing driven by episodic strong winds (Bates et al. 1998; Nemoto et al. 2009) or the seasonal collapse of surface stratification (Zhai and Dai 2009) resulting in the supply of subsurface high-CO$_2$ waters into the surface mixed layer. In addition, temperature also exerts a dominant control on sea surface pCO$_2$ variability by affecting CO$_2$ solubility (Weiss 1974).

The East China Sea is characterized by significant terrestrial inputs from one of the world’s largest rivers, the Changjiang, and dynamic exchanges with complex water masses (Chen 2009), for example, from the Yellow Sea and the Kuroshio Current (Chen and Wang 1999). Located in the temperate zone of the western Pacific Ocean, the East China Sea is regularly impacted by western Pacific tropical cyclones during warm seasons (D’Asaro et al. 2011). The East China Sea and its estuarine and nearshore regions have been a focus of coastal air–sea CO$_2$ exchange studies in the past decades (e.g., Zhai and Dai 2009; Guo et al. 2015; Chou et al. 2017). Subject to large spatial and temporal variabilities, this coastal region has distinct seasonal patterns, with a warm and productive summer and a cold and less productive winter. In the East China Sea, areas located between 28.5–33.0°N and 122.0–126.0° E exhibit the largest seasonal variations in sea surface temperature, salinity, and pCO$_2$, as a consequence of biological CO$_2$ uptake during warm seasons, ventilation in autumn, and cooling in winter (Guo et al. 2015).

Most of the previous studies in the East China Sea are based on shipboard observations (e.g., underway or discrete measurements), with a focus on mapping the spatial dynamics of pCO$_2$ (Zhai and Dai 2009; Tseng et al. 2011; Guo et al. 2015). However, these studies are limited in terms of temporal resolution, leading to large uncertainties in estimations of air–sea CO$_2$ fluxes and an incomplete understanding of their underlying processes (Chou et al. 2009b; Nemoto et al. 2009). For instance, Li et al. (2018) investigated the dependence of air–sea CO$_2$ fluxes on sampling frequency (e.g., every 3 d, weekly, and monthly) on the inner shelf of the East China Sea, showing that the potential error in estimated fluxes increases sharply with decreasing sampling frequency. Even weekly sampling is capable of introducing flux biases of up to ±63% (Li et al. 2018). Time-series studies in other coastal waters around the world have also demonstrated the importance of high-frequency observations to accurately estimate air–sea CO$_2$ fluxes (Dai et al. 2009). High-frequency pCO$_2$ values recorded in the Bay of Brest (Bozec et al. 2011) suggested that sampling only during the day or night could introduce an 8–36% bias on air–sea CO$_2$ monthly estimates. Adding in additional complexity, episodic events such as typhoons can promote vertical mixing (e.g., Price 1981; Nemoto et al. 2009), thereby introducing CO$_2$-rich subsurface water into the surface layer. Continuous observations by Nemoto et al. (2009) have successfully captured the dynamics of sea surface pCO$_2$ during the passage of three typhoons across the East China Sea. The efflux of CO$_2$ induced by the severe weather increased by 150% over normal warm season conditions (Nemoto et al. 2009). Similar to Nemoto et al. (2009), most of the previous continuous observations (e.g., Bond et al. 2011; Wada et al. 2011; Sun et al. 2014) only focused on the uplifting process during typhoon, without paying enough attention to the typhoon-induced phytoplankton blooms (Hung and Gong 2011; Lin 2012; Ye et al. 2013). Moreover, although efforts have been put into understanding the biogeochemical responses of phytoplankton flourished after the passage of typhoon, few of them have concentrated on the interaction between wind, sea surface pCO$_2$, air–sea CO$_2$ flux, and chlorophyll $a$ (Chl $a$), not to mention the evolution of upper ocean response throughout the typhoon period (i.e., pre-typhoon, during typhoon, and post-typhoon). As a consequence, for regions vulnerable to typhoons such as the East China Sea, the demand for buoy-based time-series observations has increased.

In this study, we report here the first year-round high-frequency (3-h intervals) data set with a combination of physical, biological, and chemical parameters (such as sea surface temperature, sea surface salinity, surface wind velocity, Chl $a$, pCO$_2$, and pH), from 28 July 2012 to 01 June 2013 based on a buoy system deployed on the East China Sea shelf. We aim to investigate the seasonal evolution of sea surface pCO$_2$ and pH along with other associated parameters. In particular, we focused on the temporal variability of sea surface pCO$_2$ and air–sea CO$_2$ gas exchange during episodic events (e.g., typhoon, bloom), in order to fill the knowledge gaps from traditional ship-based surveys. A novel conclusion of this study is that three stages of typhoon are identified, during which the surface wind field interacts with the observing system, resulting in different responses of surface CO$_2$ chemistry.

**Materials and methods**

**Study area**

The East China Sea is one of the major marginal seas of the Pacific Ocean, and one of the most productive shelf regions in the world (Gong et al. 2003; Hung et al. 2003). It is characterized by significant riverine inputs from the Changjiang River, which ranks 4th globally in terms of freshwater discharge (Dai and Trenberth 2002), as well as exchange with the Kuroshio along the eastern boundary of the outer shelf (Chen and Wang 1999). The major surface water masses in the East China Sea (Chen 2009) include (1) the Changjiang River plume, also known as Changjiang Diluted Water, which flows northeastward in summer and southwestward in winter (Lee and Chao 2003; Bai et al. 2014), (2) East China Sea Coastal Water, which flows southward during winter monsoons and northward during summer monsoons, (3) the Taiwan Warm Current, which flows northward all year round, (4) Yellow Sea Coastal Water, which flows generally southeastward; and (5) the Kuroshio Current, which generally flows along the 100–200 m isobaths in the East China Sea and leaves the continental slope at 30°N and 128–129°E (Fig. 1).
Previous studies on the East China Sea carbon cycle have suggested that the East China Sea acts as an annual net CO$_2$ sink, yet with remarkable seasonal and spatial variability (Guo et al. 2015, and references therein). Across the entire East China Sea shelf, the core area of the plume-influenced region in the outer Changjiang estuary (28.5–33.0°N, 122.0–126.0°E) in summer has the lowest sea surface pCO$_2$ and the largest variability in air–sea CO$_2$ fluxes (ranging from 1.6 to 10.2 mmol m$^{-2}$ d$^{-1}$; Guo et al. 2015). The highly productive Changjiang River plume, especially in summer, can promote photosynthesis in the outer estuary (Gong et al. 2003; Tseng et al. 2011; Chen et al. 2012) where our buoy is located (31°N, 124.5°E, ~300 km east of Shanghai, with a water depth of ~60 m; Fig. 1). The CO$_2$ fluxes over the rest of the East China Sea shelf are less variable (Guo et al. 2015).

**Measurements and equipment**

Time-series data were collected from 28 July 2012 to 01 June 2013 by sensors (monitoring systems) deployed on a buoy constructed by Xiamen University. Sensors were mounted at depth of 0.5 m, including a Battelle pCO$_2$ monitoring system with 3-h sampling intervals, a SeaFET Ocean pH sensor with 1-h sampling intervals, a SBE 37-SI conductivity and temperature sensor with 10-min sampling intervals, and a WET Labs fluorescence sensor to measure turbidity (nephelometric turbidity units [NTU]) and Chl $a$ at 1-h intervals. Wind velocity and barometric pressure were obtained by a weather station (Model 05106 and 61302V, Young Company, U.S.A.) equipped on the buoy at ~10 m above the sea surface. Sensors were connected to the data acquisition system on the buoy, and data were sent to the land-based laboratory in real-time by satellite transmission.

**The Battelle CO$_2$ system**

A Battelle CO$_2$ monitoring system was deployed for measuring air and surface seawater pCO$_2$ values, which were determined by a Li-Cor 820 nondispersive infrared (NDIR) gas analyzer contained in the Electronics Assembly. The Battelle CO$_2$ system is the same one as the widely used Moored Autonomous pCO$_2$ (MAPCO2) system, as described in Sutton et al. (2014$b$). The pCO$_2$ monitor cycles a closed circuit of air through the surface seawater via an “h” shaped tube assembly to reach equilibrium of CO$_2$ with the air, then the equilibrated air is analyzed to determine the CO$_2$ mole fraction (xCO$_2$).

![Fig. 1. Maps of the East China Sea. The red star shows the location of the buoy site at 31°N, 124.5°E. The main currents in the East China Sea in summer and winter are shown; the Kuroshio, Taiwan Warm Current (TWC), East China Sea Coastal Water (ECSCoW), Changjiang River plume (CRP), and Yellow Sea Coastal Water (YSCW). The solid arrows represent currents in summer, and the dashed arrows represent currents in winter, except for the Kuroshio and TWC which flow northward all year round. Note that the colormaps in (a) and (b) are in different scales.](image-url)
After the seawater $x_{\text{CO}_2}$ measurement, air $x_{\text{CO}_2}$ is measured by directly drawing an air sample from above the sea surface (see Sutton et al. 2014b for more detail). Since the “h” shaped pipe is the only part that contacts seawater and is made of a copper nickel alloy, the CO$_2$ monitor has a good performance in resisting biological fouling. The CO$_2$ measurements are quality-controlled by a two-point auto-calibration prior to every sample using a zero reference (ambient air that has passed through a soda lime canister to remove all CO$_2$) and a span reference (stored dry air with a precisely known xCO$_2$). The CO$_2$ span reference we used in this study had a value of 458.81 μmol mol$^{-1}$, of World Meteorological Organization level. The uncertainty of the calculated $p$CO$_2$ (Eq. 1) by the MAPCO2 system is reported to be $\leq 2$ μatm (Sutton et al. 2014b; with a precision of $< 0.6$ μmol mol$^{-1}$ for the measured xCO$_2$ values).

Water $p$CO$_2$ was calculated from the xCO$_2$ value in dry air equilibrated with surface water and the barometric pressure ($P$) after correcting for the vapor pressure ($P_{120}$) of water at 100% relative humidity (Weiss and Price 1980):

$$p\text{CO}_2 = (P - P_{120}) \times x\text{CO}_2$$

(1)

$x$CO$_2$ in the air was calculated similarly based on measured xCO$_2$ values in the air and the barometric pressure, using a formula similar to Eq. 1.

The pH sensor

The SeaFET Ocean pH sensor is an autonomous data logger based on a Honeywell Durafet® pH sensor (Martz et al. 2010). The primary element of SeaFET is an ion-sensitive field effect transistor (ISFET), which produces robust, stable, and precise measurements. Martz et al. (2010) evaluated the performance of the ISFET sensor in a number of different situations (e.g., a temperature-controlled calibration vessel, the Monterey Bay Aquarium Research Institute [MBARI] test tank, shipboard underway mapping and a surface mooring): it showed a short-term precision of ± 0.0005 over periods of several hours and a stability of better than ± 0.005 over periods of weeks to months. A comparison of high-resolution time-series data of upper ocean pH using the SeaFET pH sensors was provided by Hofmann et al. (2011). The experiments were conducted across a variety of ecosystems ranging from polar to tropical, open-ocean to coastal, and kelp forest to coral reef. It was suggested that the pH uncertainty is higher at dynamic sites due to the intense gradients, where the observed sampling errors ranged from ± 0.0007 to ± 0.015. However, in many instances, the uncertainty was expected to be better than ± 0.01. In addition, biofouling was believed to compromise the data quality during deployments over 2 months in high-fouling environments, therefore, we protected the sensor with a perforated copper anti-fouling guard. pH values in this study are all on total scale at in situ temperature.

Data quality assessment by shipboard measurements

To assess the data quality of the MAPCO2 system, we compared buoy and shipboard measurements of sea surface $p$CO$_2$ and other parameters (Table 1) at the beginning and end of the observation, respectively. The shipboard measurements in July 2012 collected discrete samples for total alkalinity (TA), dissolved inorganic carbon (DIC), and Chl $a$ (Table 1). TA and DIC samples were measured to calculate $p$CO$_2$ and pH using the Excel program “CO2SYS” (Pierrot et al. 2006). The dissociation constants for carbonic acid were taken from Mehrbach (1973) as refitted by Dickson and Millero (1987), the dissociation constant for sulfate was taken from Dickson (1990), the total boron-salinity relationship was from Lee et al. (2010), and pH values are given on the total hydrogen scale at in situ temperature. TA was measured using an automated Gran acidimetric titration using an Apollo Alkalinity Titrator (Model AS-ALK 1+), with a precision of ± 0.1% (Cai et al. 2004). DIC was measured using an Apollo DIC Analyzer (Model AS-C3) with a precision of ± 0.1% (Cai et al. 2004). Reference materials from Andrew G. Dickson’s laboratory at Scripps Institution of Oceanography were used to calibrate the system to an accuracy of ± 2 μmol kg$^{-1}$.

The shipboard measurements in July 2013 were made using the underway $p$CO$_2$ system with a shower-head water-air equilibrator and a NDIR detector as described in detail in Zhai et al. (2005). The NDIR detector (Li-Cor 7000) were calibrated every 8 h using a wide range of CO$_2$ gas standards that covered the $p$CO$_2$ variability during the cruises. The overall uncertainty of the CO$_2$ measurements was < 1% as constrained by the standard gases. Sea surface temperature and salinity were determined by a Yellow Springs Instrument meter (YSI 6600), which was combined with the continuous $p$CO$_2$ system. Due to the severe weather conditions, the comparison was made at nearshore site outside Qushan Island (at 30.44°N, 122.28°E) when the buoy was dragged back, where relatively higher $p$CO$_2$ values (≈ 800 μatm) were found (Table 1). Shipboard measurements of $p$CO$_2$ in July 2013 were performed over the same site (30.44°N, 122.28°E) at 0.5-h intervals.

The overall offset of the buoy measurements from the shipboard measurements was $-0.013 \pm 0.010$ for pH, 0.045 ± 0.075 μg L$^{-1}$ for Chl $a$, and < 2% for $p$CO$_2$ (Table 1). It is also noteworthy that the average offsets in $p$CO$_2$ in 2012 and 2013 were 2.1% and 1.6%, respectively, showing a good consistency with respect to time (i.e., no significant drift). This indicates a reasonably good accuracy of the buoy measurements. A good agreement of $p$CO$_2$ between the MAPCO2 system and shipboard measurements was also found by Liu et al. (2019) in the adjacent area to our study site.

Tropical depression data acquisition

The tropical depression (typhoon) data in this study corresponds to track data from Unisys Weather (http://weather.
Table 1. Comparison of surface seawater $p$CO$_2$, pH, and Chl between buoy and shipboard measurements during the deployment (at the study site 31°44'N, 122°28'E in 2012) and the recovery of buoy (at 30°44'N, 122°28'E in 2013). Date format: year/month/day.

| Date and time | Buoy observations | Shipboard observations |
|---------------|-------------------|------------------------|
|               | pCO$_2$ (μatm)    | pCO$_2$ (μatm)         |
|               | pH                | pH                     |
|               | Chl a (μg L$^{-1}$) | Chl a (μg L$^{-1}$)    |
|               | SST ($^\circ$C)   | SST ($^\circ$C)         |
|               | DIC (μmol kg$^{-1}$) | DIC (μmol kg$^{-1}$)  |
|               | TA (μmol kg$^{-1}$) | TA (μmol kg$^{-1}$)    |
|               | SSS ($^\circ$C)   | SSS ($^\circ$C)         |
|               | SSS (PSS)         | SSS (PSS)              |

For the data in year 2012, we used the relationships predicted by Takashita et al. (1993), assuming that the temperature dependence of $p$CO$_2$ for a parcel of seawater with a constant chemical composition be expressed as $\partial \ln p$CO$_2/\partial T = 0.0423^\circ$C$^{-1}$. To remove the temperature effect, the observed $p$CO$_2$ value was normalized to an average temperature ($Np$CO$_2$) as:

$$Np$CO$_2 = (p$CO$_2)_{obs} \times \exp[0.0423 \times (T_{ave} - T_{obs})]$$

where the subscripts “ave” and “obs” represent the average and observed values, respectively. $T$ denotes temperature.

The net air–sea $CO_2$ flux is estimated using the air–sea $p$CO$_2$ difference and the gas transfer coefficient:

$$F_{CO_2} = k \times \alpha \times (p$CO$_2,sw - p$CO$_2,air)$$

where $k$ is the CO$_2$ gas transfer velocity, $\alpha$ is the solubility of CO$_2$ in seawater (Weiss 1974), and subscripts “sw” and “air” denote $p$CO$_2$ of seawater and atmosphere, respectively. A positive flux indicates a net CO$_2$ flux from seawater to atmosphere, while a negative flux indicates a net CO$_2$ exchange from atmosphere to seawater. The gas transfer velocity was estimated using the wind velocity relationship from Sweeney et al. (2007), expressed as:

$$k = 0.27 \times (u_{10})^2 \times (Sc/660)^{-1/2}$$

where $u_{10}$ is the wind velocity at 10 m above the sea surface; and the Schmidt number (Sc) is temperature-dependent for CO$_2$ in seawater, computed from in situ temperature (Wanninkhof 1992).

On the basis of Eq. 3, we adapted it to examine how the variations in different terms, that is, gas transfer coefficient, sea surface $p$CO$_2$, and atmospheric $p$CO$_2$, would affect the $CO_2$ flux estimation:

$$F_{CO_2} = K \times (p$CO$_2,sw - p$CO$_2,air)$$

where $K$ is the CO$_2$ gas transfer coefficient, $K = k \times \alpha$.

The differential of the $CO_2$ flux, with respect to time, can be written as:

$$\frac{\partial F}{\partial t} = \frac{\partial K}{\partial t} \times \Delta p$CO$_2 + \frac{\partial p$CO$_2,sw}{\partial t} \times K - \frac{\partial p$CO$_2,air}{\partial t} \times K$$

where the first term on the right-hand side of Eq. 6 represents the variation in $CO_2$ fluxes attributed to changes in the gas transfer coefficient (wind velocity dominated), and the rest two terms represent $CO_2$ flux variations attributed to changes in $p$CO$_2,sw$ and $p$CO$_2,air$. 

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unisys.com/hurricane/) based on the U.S. Joint Typhoon Warning Center. The track data consist of the location, intensity, and status of the typhoon center at 6-h intervals.
Results

Hydrological settings

The time-series observation (Fig. 2a,b) shows the strong seasonal variations in sea surface temperature and salinity (hereafter referred to simply as temperature and salinity) at the study site. Seasons are defined as each consisting of 3 months, for instance, summer from June to August. During the observation, temperature decreased from 28.4 ± 1.3°C in summer to 22.6 ± 2.1°C in autumn, to the lowest 13.0 ± 2.5°C in winter, and increased to 13.5 ± 3.5°C in spring. Salinity changes also indicated the seasonal pattern. The study site is located in a river-dominated system (Tseng et al. 2011), subjected to seasonally variable freshwater discharge from the Changjiang (Zhai et al. 2007), thus resulting in a large fluctuation in salinity in summer. The study area is also subjected to significant surface stratification in spring and summer and well-mixed condition in seasons after (Zhai and Dai 2009). The average summer salinity was 31.8 ± 1.4 and reached a low value of 28.5, comparable to the previous observation (salinity of around 32.3 in summer in Chou et al. 2009a) identifying the signal of Changjiang River plume. Salinity variations in autumn and spring (Fig. 2b) also implied the influence of freshwater runoff, with average values of 32.7 ± 0.6 and 32.3 ± 0.9, respectively. Salinity in winter was relatively constant and higher, and the winter average value (32.7 ± 0.3) ranked the highest among the seasons.

It is noteworthy that the observations in this study started in late July 2012 and ended in early June 2013, so data in June and July were missing during the “continuously” observed annual cycle. This may potentially influence the understanding of the seasonal cycle of the surface seawater $pCO_2$ and subsequent estimation of the annual air–sea $CO_2$ flux (more details in “Air–sea $CO_2$ fluxes” section).

Biogeochemical parameters

During the observation period, sea surface $pCO_2$ ($pCO_{2,sw}$) exhibited significantly large variations (Fig. 2c; Table 2). It showed an overall increasing trend to above the atmospheric level from summer to mid-autumn, despite being characterized by dramatic fluctuations, followed by a transitional period from $CO_2$ supersaturation to undersaturation in late autumn. Afterward, it stayed fairly constant until late winter. There was a sharp decrease in $pCO_{2,sw}$ in spring, where it reached the lowest value throughout the observation period. The highest $pCO_{2,sw}$ value of 513 μatm was observed in late October 2012, while its lowest value of 167 μatm was observed in late April 2013.

In contrast to sea surface $pCO_2$, atmospheric $pCO_2$ ($pCO_{2,air}$) had its maxima in winter and minima in summer (Fig. 2c), behaving much more stable than $pCO_{2,sw}$, with an annual average value of 380 μatm.

$pH$ closely mirrored $pCO_{2,sw}$ (Fig. 2d), with the lowest values of 7.976 ± 0.026 observed in October and the highest values of 8.070 ± 0.043 in April (Table 2). Turbidity (Fig. 2f) had several peaks during summer and autumn coinciding with the peaks in wind velocity (Fig. 2e): however, the timing of Chl $a$ peaks (Fig. 2f) did not correspond exactly to those in turbidity or wind velocity, often having a lag of a few days.

Dynamics of sea surface $pCO_2$ and $pH$ over different timescales

On the seasonal timescale, the average $pCO_{2,sw}$ in summer was 335 ± 70 μatm, relatively lower than the average atmospheric $pCO_2$ ($pCO_{2,air}$) value of 369 ± 4 μatm, with considerably large variations. In autumn, the study site became a $CO_2$ source. $pCO_{2,sw}$ exceeded the atmospheric level, and reached its maximum at 422 ± 43 μatm. In winter, $pCO_{2,sw}$ decreased sharply to 362 ± 11 μatm, and the study site turned into a $CO_2$ sink. In spring, due to the sudden drop in values starting in May, $pCO_{2,sw}$ was reduced to 311 ± 59 μatm, the lowest seasonal average value. In contrast to $pCO_{2,sw}$, $pH$ exhibited an opposite seasonal trend: it was highest in spring ($pH = 8.097 ± 0.076$), followed by summer ($pH = 8.084 ± 0.080$) and then winter ($pH = 8.052 ± 0.010$), and reached its minimum in autumn ($pH = 7.998 ± 0.039$). Surface water pH showed the largest seasonal fluctuations in spring and summer, and remained fairly constant in autumn and winter.

In addition to the large seasonal variability, $pCO_{2,sw}$ and $pH$ were also highly variable at the intraseasonal timescale, especially during warm seasons (Table 2). Both $pCO_{2,sw}$ and $pH$ had the largest diurnal variations in August and September of 2012 and May of 2013, when the seasonal fluctuations of these parameters were also the largest.

Dynamics of sea surface $pCO_2$ and associated parameters during typhoon

During the observation, a total of five typhoons (Danney, Haikui, Bolaven, Tembin, Sanba) passed by the buoy site in summer and autumn 2012 with a distance as close as 50 km (i.e., Tembin, see Fig. 3). All typhoons, except Haikui which was 350 km far away from the buoy (Fig. 3), exerted great impacts on the study site as seen from sea level pressure (hereafter referred to simply as pressure unless otherwise specified, measured by weather station at the buoy), wind velocity and turbidity (Figs. 4–6). Danney was formed on 28 July, and traveled across the study site from its east on 02 August. Tembin and Bolaven were formed on 19 August and 20 August, respectively. However, Bolaven arrived at the buoy location approximately 2 d earlier than Tembin, and had a larger wind radius. Bolaven passed 110 km northeast of the buoy site at a translation speed (describing how fast a typhoon was moving) of 8.3 m s$^{-1}$, and Tembin passed 50 km east of the buoy site with a translation speed of 11.7 m s$^{-1}$. Sanba formed on 10 September, and passed by the buoy on 17 September from its southeast with a translation speed of 9.0 m s$^{-1}$. The sensors on the buoy kept working (Figs. 4–6) without significant
interruptions during the typhoon’s passages, which further demonstrated the high performance of the buoy system under severe weather conditions.

The occurrence of Damrey has caused a drop in pressure by around 1 kPa and an increase in turbidity by 2.5 NTU (Fig. 4a,d), whereas the pressure drop during Haikui was
comparable to that during Damrey with no obvious change in turbidity was found. This might suggest the limited impact of Haikui on the study site and therefore it will not be discussed further. During the impact of Damrey, there was no clear evidence of deep mixing as temperature and salinity seemed to only have some random covariations (Fig. 4c). The increase in turbidity was accompanied by a sharp increase in Chl $a$ as large as 16 $\mu$gL$^{-1}$ within 6 h. Sea surface $p$CO$_2$ was undersaturated from 01 to 03 August, while it became slightly over-saturated during the post-typhoon period (Fig. 4e).

Bolaven and Tembin came by the study site one after another within 4 d. Pressure decreased by 4 kPa and 2 kPa during their occurrences, with an increase in turbidity up to 9 NTU and 5 NTU, respectively. Both pressure and wind velocity reached their extreme values when these two typhoons were closest to the buoy. During Bolaven (27–29 August), temperature decreased along with increases in salinity, $p$CO$_2$ (Fig. 5c,e), and temperature-normalized $p$CO$_2$ ($Np$CO$_2$, not shown). The largest response in temperature was about $-3.0^\circ$C, occurring 6 h after the pressure minimum, during which sea surface $p$CO$_2$ and $Np$CO$_2$ increased rapidly by 200 $\mu$atm and 242 $\mu$atm, respectively. The associated variations in temperature (decreased by $3.0^\circ$C), salinity (increased by 2.0), pH (decreased by 0.22), and turbidity (increased by 9 NTU, indicating bottom sediment resuspension), along with the increase in $Np$CO$_2$, suggested strong vertical mixing driven by the intense winds, which entrained the CO$_2$-rich deep water (Chou et al. 2009b) into the surface layer. During Tembin (29–31 August), the responses of these parameters were also conspicuous although the variations were much weaker (Fig. 5). An increasing trend in temperature was also observed: the largest amplitudes for variations in $p$CO$_2$,sw, $p$CO$_2$,air, sea surface temperature (SST), salinity (SSS), pH, and Chl $a$ at the study site.

Table 2. Summary of the monthly average value and averaged monthly diurnal range of surface $p$CO$_2$,sw, $p$CO$_2$,air, sea surface temperature (SST), salinity (SSS), pH, and Chl $a$ at the study site.

| Month     | Monthly average value (average ± standard deviation) | Average monthly diurnal range |
|-----------|------------------------------------------------------|-------------------------------|
|           | $p$CO$_2$,sw ($\mu$atm) | $p$CO$_2$,air ($\mu$atm) | SST (°C) | SSS | pH | Chl $a$ ($\mu$gL$^{-1}$) | $p$CO$_2$,sw ($\mu$atm) | $p$CO$_2$,air ($\mu$atm) | SST (°C) | SSS | pH | Chl $a$ ($\mu$gL$^{-1}$) |
| 2012/08   | 348±67        | 369±4        | 28.1±1.3 | 32.1±1.3 | 8.073±0.076 | 0.71±0.80 | 55 | 5.3 | 0.78 | 1.02 | 0.06 | 1.4 |
| 2012/09   | 404±50        | 375±3        | 24.7±0.6 | 32.0±0.6 | 8.016±0.049 | 2.11±1.74 | 51 | 4.8 | 0.68 | 0.54 | 0.07 | 2.4 |
| 2012/10   | 448±30        | 382±4        | 23.0±0.7 | 32.9±0.2 | 7.976±0.026 | 0.90±0.37 | 29 | 8.0 | 0.24 | 0.23 | 0.03 | 0.6 |
| 2012/11   | 415±32        | 391±4        | 20.1±1.1 | 33.2±0.2 | 8.003±0.028 | 0.58±0.10 | 20 | 7.2 | 0.18 | 0.28 | 0.02 | 0.3 |
| 2012/12   | 371±9         | 398±5        | 15.9±1.3 | 33.0±0.2 | 8.043±0.010 | 0.51±0.07 | 12 | 7.7 | 0.21 | 0.28 | 0.01 | 0.2 |
| 2013/01   | 360±8         | 404±7        | 11.8±0.7 | 32.5±0.1 | 8.052±0.007 | 0.45±0.12 | 12 | 8.5 | 0.35 | 0.19 | 0.01 | 0.3 |
| 2013/02   | 351±7         | 403±5        | 10.6±0.5 | 32.6±0.2 | 8.063±0.003 | 0.56±0.11 | 14 | 6.5 | 0.30 | 0.24 | 0.01 | 0.2 |
| 2013/03   | 351±7         | 403±7        | 10.6±0.5 | 32.7±0.4 | 8.059±0.010 | n.m. | 16 | 9.2 | 0.67 | 0.56 | 0.01 | n.m. |
| 2013/04   | 344±31        | 400±5        | 12.2±1.0 | 32.9±0.3 | 8.070±0.043 | n.m. | 26 | 7.7 | 0.80 | 0.32 | 0.04 | n.m. |
| 2013/05   | 233±29        | 394±6        | 17.8±2.2 | 31.4±0.9 | 8.163±0.099 | n.m. | 38 | 6.5 | 1.02 | 0.48 | 0.21 | n.m. |

n.m., not measured.
An increase in Chl a appeared 2 d after the passage of Tembin, lasting from 02 September to 07 September, corresponding to the decrease in pCO2. Due to the fact that these two typhoons passed the study site with an overlap, we did not completely separate one from another and we combined them as one event in the following sections.

Sanba went by the study site parallelly to the track of Tembin but much further away. Despite that it had less impact on pressure (a drop by 1 kPa), temperature (no significant decrease was observed apart from the diurnal signal), and salinity (increased by 1.5), the wind velocity and turbidity stimulated by Sanba were comparable to those
during Bolaven and Tembin. Surface $p$CO$_2$ increased by 120 $\mu$atm with the study site turning from neutral to a CO$_2$ source, and lasting for around 10 d until gradual increase in Chl $a$ was observed.

Throughout the passages of these typhoons, the covariations in $p$CO$_2$,sw and pH were strongly correlated, with the Pearson correlation coefficient of $-0.99$. In addition, we observed the changes in the wind vector field according to the movement of each individual typhoon (figs. 5b, 6b): prior to the minimum pressure (indicating the closet approach of typhoon), it was usually east wind prevailed, whereas after the minimum pressure it was west wind prevailed.
Discussion
Seasonal evolution of sea surface $\text{pCO}_2$ and its controlling mechanisms

No conspicuous relationship between sea surface $\text{pCO}_2$ and temperature at a yearly timescale was found during the observation. However, for some certain periods in late autumn and early winter, sea surface $\text{pCO}_2$ followed the temperature-driven (thermodynamically expected) trend (gray curves in Fig. 7a). Data falling beyond the lower/upper limits (Fig. 7a) implied the nonthermal removal/addition of $\text{CO}_2$. The lower part in Fig. 7a (below the gray curves) consists of data mostly from the entire summer, and some from autumn and spring, respectively. The temperature-normalized $\text{pCO}_2$ ($N\text{pCO}_2$, Fig. 7b) values showed a positive relationship with salinity, with the lower end representing fresher Changjiang River plume and

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Fig. 6. Influence of typhoon Sanba on $\Delta\text{pCO}_2$ and other related parameters (a: SLP and wind velocity, b: wind vector, c: SST and SSS, d: Chl $a$ and Turbidity, e: $\Delta\text{pCO}_2$ and pH, and f: air-sea $\text{CO}_2$ flux) measured at the buoy site. The pink shading indicates the Sanba-influenced period. $\Delta\text{pCO}_2$, difference between $\text{pCO}_2,\text{sw}$ and $\text{pCO}_2,\text{air}$; SLP, sea level pressure; SSS, sea surface salinity; SST, sea surface temperature.
upper end representing saltier subsurface water. In contrast, the upper part in Fig. 7a (above the gray curves) consists of data from mid-winter to early spring. The sea surface NpCO2 exhibited a negative relationship with salinity, implying the mixing of different source of waters as discussed below.

From August to October, pCO2,sw was positively correlated with salinity (Pearson correlation coefficient of 0.73) and negatively correlated with temperature (Pearson correlation coefficient of −0.73). The low pCO2,sw along with low salinity and high temperature is attributed to the influence of the Changjiang River plume. Since the Changjiang River plume carries a large amount of organic carbon and nutrients, the upper estuary is usually supersaturated in CO2 with respect to the atmosphere, whereas the mid and lower estuary is usually undersaturated in CO2 due to the enhanced photosynthesis (Chen et al. 2012). On the contrary, the high pCO2,sw along with high salinity and low temperature is associated with the vertical mixing during typhoons, when CO2-rich subsurface waters were entrained into the surface.

From October to the end of November (i.e., 02 October to 30 November), a continuous efflux of CO2 was observed (Figs. 2c, 8a). During this period, salinity increased after the collapse of summer stratification (as described by Zhai and Dai 2009; see also Fig. 2b and Table 2 for the monthly variation in salinity), pCO2,sw reached its maximum value of 513 μatm on 24 October (sharp peak in Fig. 7a). Afterward, pCO2,sw gradually decreased and re-equilibrated with the atmosphere.

From December on, pCO2,sw was fairly constant and was lower than the atmospheric level (Fig. 2c). It decreased slightly along with the decreasing temperature (Fig. 7a). The ratio between the natural logarithm of pCO2,sw (lnpCO2) and temperature (∂lnpCO2/∂T) was only 0.0096 °C−1, much lower than the expected dependence (i.e., 0.0423 °C−1, Takahashi et al. 1993), suggesting that the existence of nonthermal processes affecting pCO2,sw partly counteracted the temperature effect. This is also supported by increasing NpCO2 with decreasing salinity (Fig. 7b). Biological activity (e.g., respiration) seems implausible because it is limited by low temperatures in winter on the East China Sea shelf (Gong et al. 2003). Air–sea CO2 exchange only contributes minorly (Li et al. 2018) because it requires a much longer timescale (from months to a year, Jones et al. 2014). Lateral mixing with Yellow Sea Water (lower salinity, lower temperature, and higher pCO2 than East China Sea offshore water; Zhai et al. 2014) is therefore the most likely explanation for the increase in pCO2 and NpCO2. After the onset of spring bloom in May 2013, corresponding with the significant decrease in salinity (Fig. 7b) which was driven by the freshen Changjiang River plume input, pCO2,sw began to fluctuate again.

**Air–sea CO2 fluxes**

The study site overall acted as a net CO2 sink throughout the observation (−1.48 ± 7.16 mmol m⁻² d⁻¹), however, with significant temporal variability (Fig. 8). In summer, the study site served as a source of CO2 to the atmosphere, with a seasonal average flux of 1.69 ± 10.26 mmol m⁻² d⁻¹. This is contrary to the general consensus that the Changjiang plume is a CO2 sink in summer due to the large nutrient inputs and subsequent biological CO2 uptake (Chou et al. 2009a; Zhai and Dai 2009; Guo et al. 2015). There are three possible explanations for this discrepancy. First, it could be due to the lack of data in June and July, when the East China Sea shelf would generally absorb CO2 from the atmosphere if not perturbed by typhoons, when the CO2 influx could reach as high as −6.5 ± 10.7 mmol m⁻² d⁻¹ (Guo et al. 2015). Second, the influence of typhoon Bolaven and Tembin in August (Damrey was not taken into account due to the lack of wind velocity data) should have played a predominant role in shifting the direction of CO2 flux. The CO2 flux reached as high as 32 mmol m⁻² d⁻¹, accompanied by a wind velocity of 25 m s⁻¹. The average CO2 flux during typhoon period was 13.91 ± 10.92 mmol m⁻² d⁻¹; whereas it was −0.79 ± 7.44 mmol m⁻² d⁻¹ in the absence of typhoon influence (Fig. 8b). It is therefore presumable to state that the study site could be a CO2 sink if no severe weather events happened. Third, the study site was located at the edge of the most productive area, which resulted in a weaker CO2 sink than the average of the Changjiang River plume (Guo et al. 2015).

In autumn, the CO2 flux estimate was 4.59 ± 7.83 mmol m⁻² d⁻¹, to some extent higher than previous estimates (Zhai and Dai 2009; Guo et al. 2015) even if the same gas transfer velocity parameterization (Wanninkhof 1992) was applied. Similar to summer, this difference was partly attributed to the passage of typhoon Sanba on 16–18 September, when the average CO2 flux reached 31.01 ± 19.72 mmol m⁻² d⁻¹. In addition, the collapse of summer stratification and the resulting CO2 degassing also contributed to the enhanced CO2 flux, as previously reported by Zhai and Dai (2009).

In winter and spring, the study site turned into a moderate CO2 sink, with less temporal variability. The fluxes were −5.06 ± 3.50 mmol m⁻² d⁻¹ and −4.15 ± 3.35 mmol m⁻² d⁻¹ for the two seasons, respectively. These estimates were in reasonable agreement with previous studies (−5 to −14 mmol m⁻² d⁻¹ in winter by Guo et al. 2015, and Zhai and Dai 2009; −2 to −18 mmol m⁻² d⁻¹ in spring by Guo et al. 2015, and Zhai and Dai 2009 in the adjacent area) than those in summer and autumn which were characterized by large variability.

**Impact of typhoon on pCO2 and associated parameters**

In general, we observed three evolutional stages during typhoons for those approaching the study site from the south (Fig. 9). At pre-typhoon stage (prior to colored shadings in Figs. 4–6; stage I in Fig. 9a), east wind prevailed as the typhoon getting toward the buoy since tropical depressions spin cyclonically in the northern hemisphere (Figs. 5b, 6b).
Such east wind drives offshore water masses toward nearshore, which resulted in an increase in salinity from the very beginning of typhoon (prior to the colored shadings in Figs. 4c, 5c, 6c), and it was usually not necessarily accompanied by an increase in $pCO_2$ (e.g., Fig. 6c,e). For the observed typhoons (Figs. 4–6), the average salinity increase during this stage was 2.03. At the onset of typhoon (colored shadings in Figs. 5c, 6c; stage II in Fig. 9b), strong winds always stimulated vertical mixing and sediment resuspension and hence increased turbidity, and surface $pCO_2$ was therefore elevated (Figs. 5e, 6e). For the observed typhoons, the $pCO_{2,sw}$ increase during this stage reached as large as 187 μatm (Bolaven and Tembin) and 52 μatm (Sanba), respectively. At post-typhoon stage (following the colored shadings in Figs. 4–6; stage III in Fig. 9c), west wind prevailed after the passage of typhoon (Figs. 5b, 6b), which was capable of blowing the nearshore water eastward and offshore. In addition, the surface ocean condition would normally favor phytoplankton blooms in several days with the supply of nutrients from subsurface waters (Chen et al. 2017; Wang et al. 2017a,b). The $pCO_2$ would consequently go down as observed (Figs. 5d, 6d). For the observed typhoons, the $pCO_{2,sw}$ decrease during this stage reached as large as 166 μatm (Bolaven and Tembin) and 169 μatm (Sanba), respectively.

Figures 4–6 demonstrate, however, the different response in Chl $a$ to typhoons during the observation. Generally, Chl $a$ showed no evident changes during the first few days of typhoon, except Damrey, because strong turbulence under high winds do not favor phytoplankton blooms (Wang et al. 2017a,b). In the case of Damrey (Fig. 4), turbidity only increased by 3 NTU (no wind velocity data available) as the typhoon approaching the study site, suggesting the minor impact of Damrey on the observing system. Given that the large fluctuation in Chl $a$ corresponded to the changes in temperature and salinity (Fig. 4c,d), we suggest that it was likely related to the cross-shelf export of coastal waters with extremely high Chl $a$. In the case of Bolaven and Tembin (Fig. 5), a sudden increase in Chl $a$ started from 01 September, reaching up to 7 μg L$^{-1}$. Chl $a$ then gradually decayed and persisted for around 1 week, coinciding with the decreases in salinity and $pCO_{2,sw}$. In contrast, in the case of Sanba (Fig. 6), surprisingly, a slight increasing trend in salinity (by 0.5) was found to accompany the gradual increase in Chl $a$ which started from 25 September, with the highest Chl $a$ concentration reaching 10 μg L$^{-1}$. Despite the different response in Chl $a$, $\Delta pCO_2$ has overall the similar natures: the study site was turned into a CO$_2$ source during typhoon due to the strong vertical mixing indicated by the elevation in turbidity (stage II in Fig. 9) but recovered following phytoplankton blooms (stage III in Fig. 9).

The above phenomenon reveals two different mechanisms controlling the variation of $pCO_2$ at stage III. For Bolaven and Tembin, we suggest that the increase in Chl $a$ and recovery of $pCO_{2,sw}$ were caused by the movement of the Changjiang River plume (low salinity, low $pCO_{2,sw}$ but productive) induced by wind (i.e., west wind) because salinity was observed to decrease. Chl $a$ also evolved from thriving to decayed. This could happen immediately after the passage of typhoon, therefore the time lag in responses of Chl $a$ and salinity was only 2 d. For Sanba, we suggest that the increase
in Chl $a$ and recovery of $pCO_{2,sw}$ were due to in situ growth of phytoplankton because salinity increase was observed since 22 September (although salinity dropped slightly during 18–22 September with insignificant change in Chl $a$). Chl $a$ also evolved from negligible to thriving. Therefore, although Chl $a$ and CO$_2$ drawdown were eventually found in these typhoons, they were, however, experiencing different processes. As most previous studies only focused on the biogeochemical response of $pCO_{2,sw}$ to typhoon due to the latter process (e.g., Hung and Gong 2011; Lin 2012), our study shows the possibility of typhoon-induced movement of water masses which has not previously been appreciated. This finding also emphasizes the more complex and wider impact of typhoon on CO$_2$ chemistry in coastal seas with large spatial variability than in the open ocean.

**Impact of typhoons on air–sea CO$_2$ fluxes**

We examined the contributions of different factors that influenced the temporal variability of air–sea CO$_2$ flux during typhoons based on Eq. 6. The first term on the right-hand side of Eq. 6 is a function of wind velocity square (Eq. 4), therefore wind velocity exerts the greatest impact on the gas transfer coefficient term.

Figure 10 shows the contributions of different terms. We do not consider the influence of Damrey and Haikui due to the lack of wind velocity data, and we combine Bolaven and Tembin given the short interval between them. For both events (i.e., Bolaven & Tembin and Sanba), $pCO_{2,air}$ generally played a negligible role throughout the observation. Prior to the pressure minima (indicating the closest approach of typhoon), the increase in CO$_2$ efflux was due to the combination of increase in wind velocity and supply of CO$_2$-rich subsurface water; after the pressure minima, the decrease in CO$_2$ efflux was dominated by the decreasing wind velocity, expect for some certain periods during Sanba, when the diurnal signal of increasing $pCO_{2,sw}$ (corresponding to semidiurnal tide-driven variations) seemed to compete with wind velocity (Fig. 10b). Specifically, the temporal variability in CO$_2$ flux before pressure minimum during Sanba was more complex than that during Bolaven & Tembin. Unlike the monotonically increasing trend in wind velocity, $pCO_{2,sw}$ exhibited more pronounced diurnal cycle, as indicated by the abrupt peaks and declines (Fig. 10b). As a result, the contributions from wind velocity and $pCO_{2,sw}$ tended to offset when their variations were at negative phase due to diurnal variations or decrease in $pCO_{2,sw}$ with lower temperature; whereas tended to accumulate at positive phase.
The air–sea CO2 flux in both summer and autumn was influenced significantly by the passage of typhoons. In summer with no perturbation from typhoons, the study site was a weak sink (with average air–sea CO2 flux of \(-0.79 \pm 7.44 \text{ mmol m}^{-2} \text{ d}^{-1}\)). However, the study site was converted to a strong CO2 source of \(13.91 \pm 10.92 \text{ mmol m}^{-2} \text{ d}^{-1}\) by typhoon, the strength of which was \(\sim 18\)-folds of the CO2 sink in the periods without typhoon. Therefore, although the period of typhoon perturbation was short (\(\sim 4 \text{ d}\)), typhoons completely dominated the seasonal CO2 sink/source status. In summer, the study site accumulatively absorbed 42 mmol m\(^{-2}\) CO2 in the absence of typhoon (we have only 1-month data in summer, therefore this magnitude is small). However, Bolaven (27–29 August) and Tembin (29–31 August) contributed 52 mmol m\(^{-2}\) and 30 mmol m\(^{-2}\) CO2 release, respectively, ending up with the accumulative flux of 41 mmol m\(^{-2}\) CO2 into the atmosphere in summer. The amount of CO2 emitted from the sea surface during typhoon was twice as much as the CO2 uptake at the other time in summer. Therefore, in summer with no perturbation from typhoons, the study site was a moderate CO2 source of 4.14 mmol m\(^{-2}\) d\(^{-1}\). The CO2 source was enhanced by \(\sim 8\)-folds to be 31.01 ± 19.72 mmol m\(^{-2}\) d\(^{-1}\) during the passage of typhoon Sanba. The study site accumulatively released 242 mmol m\(^{-2}\) of CO2 to the atmosphere, mostly attributed to the collapse of summer stratification. Although typhoon Sanba perturbation was only 2 d, it additionally contributed 93 mmol m\(^{-2}\) CO2 emission, accounting for 28% of the total CO2 emission in autumn.

Despite the fact that blooms drawdown pCO2 by a large amount (around 100 μatm) at post-typhoon stage (Figs. 5d, 6d), it was not strengthening the CO2 sink (due to low wind velocity; Figs. 5f, 6f) but just accelerating the re-equilibration/recovery of pCO2. Therefore, we emphasize that the typhoon-induced CO2 emission and its potential in shifting the CO2 sink-source status has an irreversible impact on air–sea CO2 flux, especially for such river-dominated margin system. We also argue that the previous understanding of sea surface CO2 chemistry and air–sea CO2 flux in summer and autumn based on shipboard observations (e.g., Guo et al. 2015) might need revision to take into account of the typhoon impact.

**Concluding remarks**

Time-series observation of sea surface pCO2 and pH data, along with other related parameters, were obtained from July 2012 to June 2013 with an autonomous buoy monitoring system deployed on the East China Sea shelf. On annual scale, the study site was overall a CO2 sink with respect to the atmosphere. On seasonal time scale, the study site was a CO2 source in summer and autumn due to the mixing of CO2-rich subsurface water, while it converted to a CO2 sink in winter and spring mainly due to low temperature in winter and strong biological CO2 uptake in spring. The passages of typhoon exerted significant influence on sea surface pCO2, pH, and air–sea CO2 flux. In summer, typhoons converted the study site from CO2 sink to source although the average sea surface pCO2 was below the atmospheric level; in autumn, the typhoon strengthened the CO2 source by a quarter. We have suggested a novel concept of three different stages of the northward movement of typhoon involving wind-induced
biogeochemical response of surface $pCO_2$. The lateral surface seawater movement was significant during pre- and post-typhoon periods when east and west wind prevailed, respectively, and vertical mixing was significant at the onset of typhoon. The variation of wind field (direction) were of great importance in elevating salinity prior to typhoon, as well as in triggering the rapid growth of Chl $a$ along with decreasing $pCO_2$ after the passage of typhoon. Given the great impact of typhoon on air–sea $CO_2$ exchange, we suggested the necessity of considering typhoon impact the synthesized studies, especially for marginal seas, which are subjected to large spatial variability.

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Acknowledgments
This study was jointly funded by the Ocean Public Welfare Scientific Research Project, State Oceanic Administration of China through grant 201505003 (subtask no. 201505003-3) and National Basic Research Program of China through grants 2015CB954001 (CHOICE-C I). We acknowledge Shanghai Marine Meteorological Center for their efforts and assistance in deploying and maintaining the buoy. Tao Huang, Liguo Guo, Yan Li, Yanping Xu, Junhui Chen, and Zhe Wang are appreciated for their help during the cruise preparation. We also thank Zhiqiang Liu at the Southern University of Science and Technology for providing suggestions on typhoon impact from the perspective of physical oceanography.

Conflict of Interest
None declared.