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The effect of typhoon on particulate organic carbon flux in the southern East China Sea

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The effect of typhoon on POC flux

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Abstract

Severe tropical storms play an important role in triggering phytoplankton blooms, but the extent to which such storms influence carbon flux from the euphotic zone is unclear. In 2008, typhoon Fengwong provided a unique opportunity to study the in situ biological responses including phytoplankton blooms and particulate organic carbon fluxes associated with a severe storm in the southern East China Sea (SECS). After passage of the typhoon, the sea surface temperature (SST) in the SECS was markedly cooler (~25 to 26 °C) than before typhoon passage (~28 to 29 °C). The POC flux 5 days after passage of the typhoon was 265 ± 14 mg-C m^{-2} d^{-1}, which was ~1.7-fold that (140–180 mg-C m^{-2} d^{-1}) recorded during a period (June–August, 2007) when no typhoons occurred. A somewhat smaller but nevertheless significant increase in POC flux (224–265 mg-C m^{-2} d^{-1}) was detected following typhoon Sinlaku which occurred approximately 1 month after typhoon Fengwong, indicating that typhoon events can increase biogenic carbon flux efficiency in the SECS. Remarkably, phytoplankton uptake accounted for only about 5% of the nitrate injected into the euphotic zone by typhoon Fengwong and it is likely that phytoplankton population growth was presumably constrained by a combination of light limitation and grazing pressure. Modeled estimates of new/export production were remarkably consistent with the average of new and export production following typhoon Fengwong. The same model suggested that during non-typhoon conditions approximately half of the export of organic carbon occurs via convective mixing of dissolved organic carbon, a conclusion consistent with earlier work at comparable latitudes in the open ocean.

1 Introduction

Recent research has shown that typhoons (severe tropical storms also referred to as tropical cyclones or hurricanes) can cause marked cooling of the sea surface, enhance nutrient pumping and result in phytoplankton blooms along storm paths of the typhoon

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and in adjacent areas (Babin, et al., 2004; Bates, et al., 1998; Chang, et al., 1996, 2008; Lin, et al., 2003; Shang, et al., 2008; Walker, et al., 2005; Zheng and Tang, 2007). Previous studies of typhoon induced phytoplankton blooms have been based largely on satellite remote sensing data, but it can often be difficult to obtain clear ocean color images during or shortly after the typhoon passages because of extensive cloud cover. Sea-based measurements of pre- and post-typhoon phytoplankton biomass and biological activity are also limited because of the dangers inherent in field work associated with typhoon events. Consequently, in situ field observations on the effects of typhoons on biological processes are rare (Chang, et al., 1996; Chen, et al., 2003; Shiah, et al., 2000; Zheng and Tang, 2007) and no reports of the effects of typhoon passage on ocean biogeochemistry, especially carbon export fluxes measured using floating sediment traps, have appeared.

The upwelling of subsurface water on the shelf-break of the SECS has long been observed (Chen, et al., 1990; Gong, et al., 1995; Liu, et al., 1992). The wax and wane of the outcrop of cool upwelled water were attributed to inflow of the Taiwan Strait water Gong, et al., 1995, or an influence of the strong northeast monsoon in winter (Chuang and Liang, 1994; Gong, et al., 1995). Recently, Wu et al. (2008) and Chang et al. (2009) reported a remarkable seasonal variability of this upwelling at depths of 100 m or more below the surface in summer but such upwelling was weak in winter. In addition to such factors, typhoons may also enhance upwelling in the SECS following passage of a typhoon through an adjacent region (Chang, et al., 2008; Tsai, et al., 2008). Tsai et al. (2008) reported that a typhoon induced significant upper water column cooling off northeastern Taiwan). Chang et al. (2008) reported that upwelled water persisted in the SECS for more than a week after a typhoon resulting in a phytoplankton bloom in the shelf region.

Several studies (Bates, et al., 1998; Lin, et al., 2003; Walker, et al., 2005; Zhao, et al., 2008; Zheng and Tang, 2007) have inferred that large quantities of nutrients were carried to the ocean surface resulting in enhanced phytoplankton biomass (even phytoplankton blooms) in the open ocean and coastal waters several days after a

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typhoon event. However, the cited reports were based on study of satellite images, and the reported phytoplankton biomass (e.g. usually determined from chlorophyll absorption spectra) has seldom been confirmed by in situ measurements. Thus, Shang et al. (2008) indicated with caution that dissolved organic matter and detritus could significantly affect detected chlorophyll (chl *a*) values near the coast of the northern South China Sea. The study area of Shang et al. (2008) was adjacent to the mouth of the Mekong River, which may discharge runoffs containing a high concentration of chromophoric dissolved organic matter. However, no large river was located near the our study area and the satellite ocean color data are thus likely to accurately represent of the sea conditions, being unaffected by influx of terrestrial materials. Besides, they are supported by ship-based measurements and, therefore, are included in this study as an important source of information. Most importantly, the flux of particulate organic carbon (POC) from the

euphotic zone is unknown because typhoons form in remote regions of the open ocean and sea-based experiments are difficult to conduct due to strong winds and high waves. Typhoons Fengwong passed through the SECS on 28 July 2008, and in the following 4–5 days we had a unique opportunity to conduct sea-based investigations in the affected area of the southern ECS. The main objective of the study was to conduct in-situ experiments to examine the physical and biological responses of the water column to typhoon forcing. We used hydrographic data (including information on temperature, salinity, density, light intensity and nitrate concentration) in efforts to interpret possible mechanisms affecting biogeochemical responses (chl *a* concentration and POC flux).

2 Materials and methods

Sampling was conducted in the euphotic zone (depth of 1% surface light penetration) of the SECS, in an area approximately 70 km northeast of Taiwan (25.40° N, 122.45° E, Fig. 1), from the R/V Ocean Researcher II (OR II). Three cruises were conducted: one

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each in June and August 2007 (non-typhoon periods) and one in August 2008 (2–3 August, 2008) during the passage of typhoon Fengwong. Data on wind speed during the typhoon were provided by the Central Weather Bureau of Taiwan. Water temperature was recorded using a SeaBird model SBE9/11 plus conductivity/temperature/depth (CTD) recorder. The concentrations of nitrate (NO_3) and chlorophyll *a* (chl *a*) at depths of 0, 10, 25, 50, 75, 100 and 125 m were determined according to Gong et al. (2000). Sinking particles were collected using a drifting sediment trap array, which consisted of six cylindrical plastic core tubes (6.8 cm diameter) with honeycomb baffles covering the trap mouths (Hung, et al., 2004; Hung and Gong, 2007). The array was attached to an electric surface buoy with a global positioning system (GPS) antenna (TGB-500, TAIYO, Japan). The trap tubes, filled with filtered seawater (< 0.5 µm), were deployed daily for 4 to 6 h at a depth of 70 m. Aliquots from sediment traps were filtered through 1 µm quartz filters (Whatman QMA) as described in Hung et al. (2009). The swimmers evident with a microscope on the filters were carefully removed using forceps. After fuming the filters with HCl, the POC on the filters was measured using an elemental analyzer (Elementa, Vario EL-III, Germany).

The sea surface temperatures (SST) from 25.2–25.7° N and 122.1 to 122.6° E were estimated (resolution 1.1 km) were estimated before and after passage of typhoons Fengwong and Sinlaku using AVHRR (Advanced Very High Resolution Radiometer) infrared sensors under cloud-free conditions. The AVHRR data were processed at the National Taiwan Ocean University ground station. SST estimates and atmospheric attenuation correction were based on the multi-channel SST (MCSST) algorithm (McClain, et al., 1985). Comparisons of the accuracy of AVHRR-derived SSTs with in situ data were based on the method of Lee et al. (2005). The root mean square errors and biases for satellite-derived SSTs were 0.65 °C and 0.01 °C, respectively. The derived surface chl *a* concentrations by MODIS (Moderate Resolution Imaging Spectroradiometer) were compared to chl *a* values measured in the field. Primary productivity (PP) was derived using an empirical MODIS chl *a*-temperature algorithm based on a function of maximum carbon fixation within a water column, sea surface

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photosynthetically active radiation, euphotic zone depth, chlorophyll concentration and photoperiod (Behrenfeld and Falkowski, 1997) or by the ^{14}C assimilation method (Parsons et al., 1984; Gong et al., 1999). New production (NP) was derived by the model of Laws et al. (2000) based on ef (= export flux (or NP)/PP) ratios and a function of temperature, euphotic zone depth and primary production (the detailed information on the model can be found at http://usjgofs.whoi.edu/mzweb/syndata.html#Biogeochemical).

3 Results

3.1 Variation in surface hydrographic settings before and after typhoon passage

Typhoon Fengwong was a category 2 typhoon (sustained winds = 43 m s^{-1}, wind gusts = 53 m s^{-1}) that traveled at a speed of 4.4–4.7 m s^{-1} on 26–27 July 2008 and 3.3–3.8 m s^{-1} on 28–29 July 2008 (Fig. 1). The typhoon made landfall on the eastern side of Taiwan on 28 July 2008 and affected the island for 10 h. The AVHRR-derived SSTs in the study region before (24–26 July) and after (1–3 August) the typhoon are shown in Fig 1. The daily average SST value for the study area (25.2–25.7° N and 122.1 to 122.6° E) is shown in Fig. 2a. It is notable that the average SST continually decreased from 28.9 °C on 24 July (before the typhoon) to 25.7 °C on 4 August, 2008 (after the typhoon) and then increased to 29 °C on 15 August (Fig. 2a). Furthermore, the area of the cold water patch (SST <27 °C) increased gradually from 118 km^2 on 24 July to 14 379 km^2 on 5 August (Fig. 2b). Similar surface water cooling phenomena were also found after the passage of typhoon Sinlaku, a category 3 typhoon with a sustained wind speed of 51 m s^{-1} which made landfall on Taiwan on 13–14 September 2008. The average daily SST decreased continuously from 28.6 °C on 7 September to 26.5 °C on 22 September, 2008 (Fig. 2a). The maximum area (17 876 km^2) of the cold water patch (SST <27 °C) caused by typhoon Sinlaku was greater than that resulting from typhoon Fengwong (Fig. 2b). The SST values of the SECS nonetheless showed water cooling after each typhoon.

As it is difficult to obtain measured vertical hydrographic data in the field prior to a typhoon event because of the dangerous and unpredictable sea conditions of the typhoon, temperature-salinity (*T–S*) diagrams (Fig. 3a) during non-typhoon (in June and August 2007) and typhoon periods (August 2008) were used to demonstrate the vertical mixing vs. upwelling process in the study area. A typical *T–S* diagram is shown in Fig. 3a, a red curve which shows source water of the Kuroshio subsurface upwelling in the SECS. During a non-typhoon period, weak upwelling resulting from subsurface water of the Kuroshio was detected in June (a blue curve) and August (a purple curve) 2007, respectively., Upwelled cold water consequently mixed with warm surface water of the SECS, resulting in slight changes in the *T–S* diagrams (Fig. 3a).Thus, variability in *T–S* association in our study area was not remarkable based on the many CTD hydrocasts (gray points in Fig. 3a) noted under non-typhoon conditions. As upwelling was week, the concentration of NO_3 in the surface layer was almost undetectable (e.g. ~0 µM) (Fig. 3b) and the surface chl *a* concentrations (Fig. 3c) were 0.23 mg m^{-3} in June 2007 and 0.41 mg m^{-3} in August 2007, respectively. However, it is evident that after passage of typhoon Fengwong (August 2008), a colder nutrient-rich water (a green curve in Fig. 3a and b) was brought to the surface from a deep water source derived from the Kuroshio upwelling, and this water mixed with surface water of the SECS resulting in a high surface chl *a* value (1.4 mg m^{-3}, Fig. 3c). In contrast, the source of upwelled water source 5 days after typhoon Fengwong was significantly deeper than that under non-typhoon conditions (Fig. 3a).

When non-typhoon conditions were prevalent (June and August 2007), the depths of the euphotic zone were 65 and 43 m (Table 1), respectively. However, after the Fengwong and Sinlaku typhoon events, the euphotic zones became shallower (34 m and 30–32 m, respectively, Table 1) and the depth was only about half that of the mean euphotic zone depth reported by Gong et al. (2001) for this region (60 ± 10 m). Fig. 4 shows the MODIS-derived average surface chl *a* concentrations before and after typhoons Fengwong and Sinlaku. The derived chl *a* values were much higher after typhoon Fengwong than before the typhoon (Fig. 4a and b), but the difference between derived chl *a* values pre-typhoon and post typhoon Sinlaku was less (Fig. 4c and d). MODIS-derived surface chl *a* values in the study area (25.40° N, 122.45° E) were 0.77 mg m^{-3} (24–26 July), 0.93 mg m^{-3} (31 July–2 August), 0.30 mg m^{-3} (8–10 September) and 0.70 mg m^{-3} (17–19 September), respectively. In comparison, the field-measured chl *a* value (1.4 mg m^{-3}) at 2 m on 2 August (after typhoon Fengwong) was 50% higher than the derived 3-day average chl *a* value (0.93 mg m^{-3}) determined using MODIS. In situ surface chl *a* concentrations on 19 and 21 September were 1.01 and 0.56 mg m^{-3}, respectively. Similarly, the in situ chl *a* value (1.01 mg m^{-3}) on 19 September was also higher than the derived 3-day mean chl *a* value (0.70 mg m^{-3}, 17–19 September) noted after passage of typhoon Sinlaku.

3.2 POC flux and vertical hydrographic values after passage of Fengwong and Sinlaku

Diel variations of POC flux were measured in the ECS in March/April of 2008 (Hung, et al., 2009, unpublished data). Daytime (8 am to 5 p.m.) POC fluxes at 120 m and 150 m were 45.9 ± 6.9 mg-C m^{-2} d^{-1}and 47.6 ± 7.6 mg-C m^{-2} d^{-1}, respectively. Night-time (12 am to 9 a.m.) POC fluxes at the same depths were 53.3 ± 8.5 mg-C m^{-2} d^{-1} and 43.8 ± 7.4 mg-C m^{-2} d^{-1}, respectively. Li (2009) investigated POC fluxes using a time-series (4, 8, 12 and 24 h) of trap deployments in the same region of the present study and and found no significant difference in POC flux between night and day. Thus, if diurnal variability in POC flux occurs, this may be small compared to other possible sources of error in our measurements. The trapping efficiency of floating sediment traps was 75% in the East China Sea (Li, 2009) and 80% in the oligotrophic water of the northwest Pacific Ocean (Hung and Gong, 2007), based on the ^{234}Th/^{238}U disequilibrium model of Buesseler et al. (1992). In addition to measurements associated with the passage of typhoon Fengwong, we also analyzed POC flux 5 and 7 days after typhoon Sinlaku passed over Taiwan on 13–14 September 2008, but obtained no vertical hydrographic data on either occasion.

The vertical distribution of temperature, NO_3 and chl *a* concentrations in the study area after the typhoon (2 August 2008) are shown in Fig. 3a. The temperature in the euphotic zone following the typhoon was significantly lower than that recorded during June and August, 2007 (Fig. 3a). The presence of colder water in the upper ~70 m indicates that cooling was quite intense. When averaged over the upper 75 m of the water column (Table 1), the increase in NO_3 concentration associated with this upwelling was 6.41 µm (=0.534–(0.046 + 0.060)/2 mol m^{-2})/75 m = 0.0064 mol m^{-3}). The level of nitrogen associated with the increase in POC must be added to this figure. The change in POC (Table 1) was 2.39 µmPOC (=5.5–(3.5+3.2)/2 g m^{-2})/75 m = 0.02867 gC m^{-3}). Assuming that Redfield stoichiometry was relevant (Redfield, et al., 1963), such a change in POC would be associated with an increase of 0.36 (=2.39/6.625) µmin nitrogen. Thus the net increase in N in the upper 75 m of the water column averaged 6.77 (=6.41+0.36) µM.

Areal chl *a* concentration peaked on 3 August 2008 (at 116 mg m^{-2}) and subsequently declined to 48 mg m^{-2} on 13 August 2008. This was similar to the average maximum integrated chl *a* concentration of 46 mg m^{-2} derived from extensive observations in a similar study region (25.5° N, 122.17° E) in good weather conditions (Gong, et al., 2000, 2001). The change in areal chl *a* concentration associated with typhoon Fengwong was therefore 72 mg m^{-2} (=116–(31+51+48+46)/4) and the associated change in areal POC level was 2.15 (=5.5–(3.5+3.2)/2) g m^{-2}, giving a ΔPOC/Δchl *a* ratio of 30 gC g^{-1} chl *a* (=2.15/0.072). This is similar to the geometric mean (36 gC g^{-1} chl *a*) of the range of carbon/chl *a* ratios reported for phytoplankton from eutrophic environments by Riemann et al. (1989), suggesting that most of the increase in POC following typhoon Fengwong was associated with an increase in phytoplankton biomass. Most previous studies using satellite surface color images suitable for large scale observations have focused on the surface layer, and hydrographic variation at depth has seldom been observed.

The POC flux at 70 m was approximately 265 ± 14 mg-C m^{-2} d^{-1} 5 days after the passage of typhoon Fengwong (Table 1). As we were unable to directly measure the

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effect of typhoon Fengwong on POC flux, we compared post-typhoon POC flux with fluxes measured during non-typhoon conditions in June and August 2007, which were 180 ± 10 and 140 ± 22 mg-C m^{-2} d^{-1}, respectively (Table 1 and Fig. 5). The elevated post-typhoon POC export flux is unlikely to have been caused by lateral transport of coastal waters or riverine input because the transmissometer profiles did not show any anomalies, and satellite color images did display any unusual terrestrial influence (Fig. 4). Enhanced POC export in the SECS was also observed after passage of typhoon Sinlaku in September 2008. Measured POC fluxes in the SECS study area 5 and 7 days after passage of this typhoon were 225 ± 34 and 224 ± 33 mg-C m^{-2} d^{-1}, respectively (Table 1).

The change in areal POC concentration following typhoon Fengwong (2.15 g C m^{-2}) can be equated to the difference between the integrals of new production and export production following the typhoon. Assuming that export production increased linearly from a baseline of 160 (=(180+140)/2) mg C m^{-2} d^{-1} to 265 ± 14 mg m^{-2} d^{-1} five days later, the integral of export production during that time interval is 1.06 gC m^{-2} (=160+265)×(5/2) = 1062.5 mg C m^{-2}). Addition of this value to the observed increase in POC yields 3.21 gC m^{-2} for the integral of new production, i.e., a new production rate of 0.642 (=3.21/5) gC m^{-2} d^{-1} following the typhoon.

4 Discussion

4.1 Hydrographic settings during non-typhoon and post-typhoon periods

Dramatic effects of typhoons on chl *a* levels, biological variables, and phytoplankton production have been reported several days following typhoon's passage based on data from satellite color images or field experiments (Babin, et al., 2004; Chang, et al., 2008; Lin, et al., 2003; Shiah, et al., 2000; Walker, et al., 2005; Siswanto et al., 2007; Zhao, et al., 2008; Zheng and Tang, 2007). The satellite-derived SST data used in the present study clearly showed upwelling after the passage of typhoons

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Fengwong and Sinlaku (Fig. 2a and b). Many previous reports have focused on variation in SST (e.g. entrainment of cold, nutrient-rich water brought to the surface), surface ocean color images, and sea level anomalies between pre- and post-typhoon periods (Chang, et al., 2008; Lin, et al., 2003; Zheng and Tang, 2007). Fig. 2b shows how SST in the area of the cold water patch changed over time following passage of typhoon Fengwong, with the lowest SST evident in the upwelling core. This demonstrates that satellite-derived information can be used to complement in situ sampling for assessment of the impact of cooling on POC production and flux (see discussion in Sect. 4.3). Fig. 6 shows a strong relationship between SST and NO_3 concentration in the study area. The decline in SST caused by typhoons could be caused by various processes including evaporative heat loss, wind enhanced vertical mixing, wind induced upwelling, and a reduction in solar insolation because of increased cloud cover, but delineation of the extent of involvement, and the contributions of different processes to observed cooling (e.g. SST change), is exceedingly complicated. The SST cooling factors are complicated, but they can be use to predict the nitrate concentration from the SST data after a typhoon if the relationship between temperature and nitrate in the open ocean is known. This is important when using an algorithm to estimate new production (NP) by means of remotely sensed ocean color to determine the nitrate concentration in the euphotic zone (e.g., Dugdale et al., 1989; Sathyendranath et al., 1991).

Chl *a* concentration in the euphotic zone (average value at 5 m) increased from about 0.6 mg m^{-3} in June and August 2007 to 1.4 mg m^{-3} on 2 August, 2008 (Fig. 3c), associated with the presence of cold nutrient-rich water. In white light, the chlorophyll-specific visible light attenuation coefficient (K_C) is ~14 m^2 g^{-1} chl *a* (Atlas and Bannister, 1980), but in the ocean K_C increases rapidly with depth because of an improved match between the spectral characteristics of submarine radiation and the principal absorption bands of photosynthetic pigments. At an irradiance equal to 33% of the surface level, K_C becomes higher than 30 m^2 g^{-1} chl *a*, and increases further to ~40 m^2 g^{-1} chl *a* at greater depths (Laws, et al., 1990). A linear plot of chl *a* concentration versus the visible light attenuation coefficient (=ln(100)/EZ from Table 1, EZ: euphotic zone), gives

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a slope of 37 m^2 g^{-1} chl *a* (Fig. 7, $r^2 = 0.976$), suggesting that virtually all of the decrease in EZ following typhoon Fengwong can be explained by a rise in phytoplankton biomass within the euphotic zone.

It is noteworthy that in situ concentrations of chl *a* (1.4 and 1.0 mg m^{-3} on 2 August and 19 September 19, respectively) were higher than satellite derived average chl *a* values (0.9 and 0.7 mg m^{-3} on 2 August and 17–19 September, respectively), suggesting that the derived chl *a* values might be underestimated by use of satellite color images following a typhoon. Although we recognize that the data are limited, we, we would suggest several possible explanations. (1) The measured chl *a* concentration was obtained at a specific location whereas the derived chl *a* value was a 3-day mean value derived from a large area (1×km^2); (2) The phytoplankton community may change after a typhoon (see detailed discussion in Sect. 4.3) and chl *a* may not be the principal pigment of some phytoplankton; (3) The presence of suspended particles and/or chromophoric dissolved organic matter may affect satellite derived chl *a* measurements (Hoge and Lyon, 2002; Shang et al., 2008); and (4) Sampling time differences between the two approaches could be influential if phytoplankton show diel variation. Clearly, evaluation of the relationship between in situ chl *a* and satellite derived chl *a* is warranted.

One obvious benefit of in situ sampling following typhoons is acquisition of information on vertical variation in nutrient concentration and phytoplankton biomass. Figs. 3B and 3C shows the vertical profiles of NO_3 and chl *a* concentrations after the passage of typhoon Fengwong. When the chl *a* concentrations are converted to phytoplankton nitrogen assuming a carbon/chl *a* ratio of 30 (see above) and relevance of the Redfield C/N (Redfield, et al., 1963), it shows that phytoplankton uptake five days after the typhoon had removed only 5.3% of NO_3 (=0.36/6.77) in the upper 75 m of the water column. Why was so little upwelled NO_3 assimilated by phytoplankton? With an initial areal chl *a* concentration of ~0.041 g m^{-2} (=(0.051+0.041)/2) (Table 1), and assuming a carbon/chl *a* ratio of 30 (see above), the biomass of phytoplankton just prior to typhoon Fengwong would have been 1.23 g C m^{-2}, thus approximately 37%

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(=1.23 g C m^{-2}/3.35 g C m^{-2}) of the POC in the water column. Areal chl *a* concentration rose from about 44 to 116 mg m^{-2}, thus showing a net rate of increase of approximately 0.19 d^{-1} (=(ln 116–ln 44)/5 day) or a doubling time of 3.6 (=(ln 2)/0.19) days. Assuming approximately one doubling each 3.6 days, phytoplankton biomass in the upper 75 m of the water column could have increased to 3.2 gC m^{-2} in five days, only equivalent to 0.53 µm(=3.2 gC m^{-2}/12 g mol^{-1}/75 m/6.6) phytoplankton nitrogen, thus significantly lower than the observed NO_3 pumping (6.77 µM). As these calculations are based on integration to a depth of 75 m, and because the 1% light level was substantially shallower than 75 m (31–44 m), the slow growth rate was probably attributable in part to considerable light limitation. An additional explanation for failure of the phytoplankton population to increase is grazing by protozoa, which are able to multiply rapidly and are among the dominant herbivores in tropical and subtropical open ocean food chains (Laws, et al., 2000).

4.2 POC fluxes following the passage of typhoons Fengwong and Sinlaku

Although typhoon Sinlaku was a category 3 typhoon, the lowest SST (26.5 °C) in the SECS caused by this typhoon was higher than that (25.7 °C) resulting from passage of typhoon Fengwong. Overall, the impact of both typhoons on SST levels in the SECS was similar to the effect of the category 5 typhoon Hai-Tang, which passed over Taiwan on 18–19 July 2005 (Chang, et al., 2008). On 22 July 2005, after the passage of typhoon Hai-Tang, the upwelling area (<26 °C) expanded rapidly to occupy a crescent shaped region of 9,146 km^2. This increased area of upwelling weakened gradually but persisted for more than 1 week. Additionally, Zheng and Tang (2007) found an initial decrease in SST (0.5–2 °C) associated with the passage of a typhoon in the South China Sea, whereas a large decrease in SST (4–5 °C) was noted 1 day after passage.

The measured POC fluxes in the SECS study area 5 and 7 days after passage of typhoon Sinlaku were 225 ± 34 and 224 ± 33 mg m^{-2} d^{-1}, respectively. These values are higher than the fluxes seen in June and August 2007, but less than the flux observed

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following the passage of typhoon Fengwong ($265 \pm 14\,mg\,m^{-2}\,d^{-1}$). As the sustained wind speeds ($51\,m\,s^{-1}$) during typhoon Sinlaku were significantly higher than those during typhoon Fengwong ($43\,m\,s^{-1}$), it is reasonable to assume that more nutrients were injected into the upper water column upon passage of the former typhoon. Other factors being equal, one might anticipate that POC fluxes would be greater 5–7 days after typhoon Sinlaku. Why was this evidently not the case?

As noted above, the amount of NO_3 injected into the upper 75 m of the water column after typhoon Fengwong had the potential to cause a much greater bloom of phytoplankton than was observed. The increase in phytoplankton biomass after a typhoon clearly depends on much more than physics. The greater level of wind mixing associated with typhoon Sinlaku likely increased the depth of the mixed layer to an extent greater than that associated with typhoon Fengwong. It is apparent from Table 1 and Fig. 3 that the mixed layer after passage of typhoon Fengwong was deeper than the euphotic zone. Further deepening of the mixed layer would therefore be associated with increased light limitation of net community production (Sverdrup, 1953). Since the NO_3 concentrations at most depths in the water column following typhoon Fengwong were more than adequate to saturate phytoplankton uptake kinetics (Eppley, et al., 1969), there is no reason to believe that additional mixing would have increased growth rates. Indeed, additional mixing and deepening of the mixed layer would have reduced growth rates due to light limitation. This scenario very likely explains the lower rates of POC export following typhoon Sinlaku, compared to those noted following the passage of typhoon Fengwong.

Other factors are worthy of consideration. Based on the recorded track (Fig. 4c and d), the center of typhoon Fengwong did not pass directly through the study area. Previous reports have indicated that upwelling and sea surface cooling are primarily observed along the typhoon track, or to the right of the typhoon center, suggesting that wind-enhanced eddy pumping and/or vertical mixing are major mechanisms involved in these phenomena (Lin, et al., 2003; Shang, et al., 2008; Walker, et al., 2005; Zheng and Tang, 2007). For example, enhanced phytoplankton growth to the

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right of a typhoon track has been reported in the Gulf of Mexico (Walker, et al., 2005) and the South China Sea (Zheng and Tang, 2007). Our study area was to the left of the track of typhoon Sinlaku's track but to the right of the track of typhoon Fengwong. Furthermore, photosynthetic rates are probably limited by light (controlled by cloud cover) and the size of the phytoplankton population after passage of typhoons. Typhoon Sinlaku occurred, approximately 7 weeks after typhoon Fengwong, and it appears that daily insolation fell by about 13% less after passage of the former typhoon (http://aom.giss.nasa.gov/srlocat.html).

4.3 Implications for new and export production after typhoon events

Boyd and Trull (2007) pointed out that many recent biogeochemical models may overestimate global POC export flux, and suggested that for global carbon modeling, more field observations were needed to validate modeled estimates (Laws, et al., 2000; Moore, et al., 2002). Many typhoons and hurricanes occurring in tropical open oceans each year (Webster, et al., 2005), and their collective effects on carbon export flux after typhoons remain unclear because of sampling difficulties. We used existing algorithms to estimate primary production (PP) (Behrenfeld and Falkowski, 1997) and new and export production (Laws, et al., 2000). The algorithm of Laws et al. (2000) is a steady state model in which new and export production are assumed to be the same. The estimated new/export production in the present study was 183–219 mg C m^{-2} day^{-1} during June and August 2007, 567 mg-C m^{-2} day^{-1} following typhoon Fengwong, and 277-359 mg C m^{-2} day^{-1} after typhoon Sinlaku. The modeled new/export production during June/August 2007 was slightly higher than the measured POC flux (Table 1). A likely explanation for this discrepancy is that some of carbon export from the open ocean at such latitudes may be attributable to convective mixing of dissolved organic carbon (Carlson, et al., 1994). Five days after typhoon Fengwong, the average of new/export production was 454 (=642+265)/2) mg-C m^{-2} day^{-1}, which is lower than the modeled new/export production of 567 mg-C m^{-2} day^{-1}. At 5–7 days after typhoon Sinlaku, the modeled average new/export production was 318 (= (277+359)/2) mg-C m^{-2} day^{-1}.

If this is likewise indicative of the average of new and export production at that time, the implied new production would be 412 (=318×2–224) mg-C m^{-2} day^{-1}, almost twice the measured POC export. This conclusion is roughly consistent with the behavior of new and export production five days following typhoon Fengwong, i.e., new production greater than POC export by a factor of 2.4 (=642 mg C m^{-2} day^{-1}/265 mg C m^{-2} day^{-1}).

Trap results (Table 1) suggest that the export ratio (POC flux/primary production) rises following typhoons. This probably reflects the fact that phytoplankton species composition changes from small-cell species to large diatoms due to the influence of nutrient-rich water. We did not investigate phytoplankton species composition in the present study, but we did examine phytoplankton assemblage changes (Hung et al., unpublished data) in the same study area between 5–20 August 2009, embracing the period before and after typhoon Morakot which passed over Taiwan on 8 August 2009. The major phytoplankton assemblages in surface waters before typhoon Morakot (August 5) were small *dinoflagellates* (cell size <10 µm), including *Gymnodinium spp.* The abundance of *Gymnodinium* was about 1500 cells L^{-1}. *Trichodesmium spp.* were also found in the study area implying that the surface waters were oligotrophic. However, the major phytoplankton assemblages 3 days (on 11 August) after typhoon Morakot, were composed of pennate diatoms including *Pseudonitzchia spp.* and *Nitzchia spp.* Seven days after passage of typhoon Morakot (14 August), the dominant phytoplankton species were large cell colony-formed diatoms including *Chaetoceros*, *Thalassionema*, and *Skeletonema costatum*. The highest abundance of these large diatoms (approximately 2×10^5 cells L^{-1}) occurred ten days after the typhoon. Following the typhoon, phytoplankton abundance decreased dramatically. Nutrient conditions were similar in the study area before and after typhoons Fengwong and Moraket. Analogous phytoplankton species changes have been reported in the East China Sea (Chang et al., 1996) and the Kuroshio (Chen et al., 2009). During non-typhoon conditions, a substantial proportion of export production may be associated with convective mixing of DOC (Carlson, et al., 1994). The model of Laws et al. (2000) suggests that the export ratio after typhoon is higher than that during non-typhoon conditions (Table 1). If this is true, this indicates that an important difference between typhoon and non-typhoon conditions is the mechanism of export: primarily POC sinking following typhoons and roughly equal contributions from POC sinking and convective mixing of DOC during non-typhoon conditions.

5 Conclusions

This study involved direct measurement of POC flux in the southern East China Sea several days after each of two typhoons. The POC flux after a typhoon was greater than that during non-typhoon periods indicating that biogenic carbon sinking induced by a typhoon is more efficient than that during non-typhoon conditions. The enhanced POC export flux noted following typhoon Fengwong was attributed to a phytoplankton bloom, as evidenced by a dramatic increase in areal chl *a* concentration. The failure of the phytoplankton community to assimilate more of the available nitrate may reflect the small initial size of the phytoplankton assemblages and the fact that the rate of increase in abundance of the phytoplankton community was constrained by a combination of light limitation and grazing.

POC export fluxes observed in the field were lower than those predicted by Laws et al. (2000) during non-typhoon conditions, probably because a substantial amount of the organic carbon export at such times is associated with convective mixing of DOC. Following typhoon Fengwong, there was excellent agreement between the model's estimate of new/export production and the average of estimated new production and POC export, suggesting that following typhoons most export occurs via sinking POC. Further studies are required to elucidate the mechanism of upwelling after typhoons, and to quantify the contribution of storm events to global POC export flux in other marine environments including marginal seas and tropical open oceans. The field observations of this study were conducted within a small area because of poor weather conditions. The availability of numerous moorings may be useful to help clarify physical mixing mechanisms and to enable a comprehensive understanding of biological processes operative during and after typhoons.

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Table 1. Data of mixed layer depth (MLD), euphotic zone (EZ: 1% of surface light intensity), integrated nitrate (I-NO3), chl *a* (IB), POC (I-POC) and POC flux from different cruises.

Date	TD	EZ	I-NO_3	IB	I-POC	C flux[a]	C flux[b]	I-PP	e[a]	e[b]
	(m)	(m)	mol m^{-2}	g m^{-2}	g m^{-2}	(trap)	(model)	(model)	(trap)	(model)
06/10/2007	16	65	0.046	0.031	3.2	180 ± 10	183	1111	0.16 ± 0.01	0.16
08/03/2007	12	43	0.060	0.051	3.5	140 ± 22	219	1283/1773*	0.11 ± 0.02	0.17
08/03/2008	11	34	0.534	0.116	5.5	265 ± 14	567	2775	0.10 ± 0.01	0.20
09/19/2008	n.a	32	n.a	n.a	n.a	225 ± 34	277	1384	0.16 ± 0.03	0.20
09/21/2008	n.a	30	n.a	n.a	n.a	224 ± 33	359	1711	0.13 ± 0.04	0.21

MLD: defined at depth which temperature increased by 2 °C from the temperature at 10 m.
NO_3, IB, I-POC: integrated NO_3, chl *a* , POC from 0 to 75 m.
C flux[a]: measured by sediment traps.
C flux[b]: estimated by model (=I-PP.×e ratio (i.e. e[b])) of Laws et al. (2000).
I-PP: estimated by the model (of Behrenfeld and Falkowski (1997).
e[a]: C flux[a]/I-PP; e[b]: C flux[b]/I-PP.
n.a.: data not available.
*: PP value was measured by C-14 incubation.
Data in 2007 from non-typhoon periods. Data in 2008 from post-typhoon periods.

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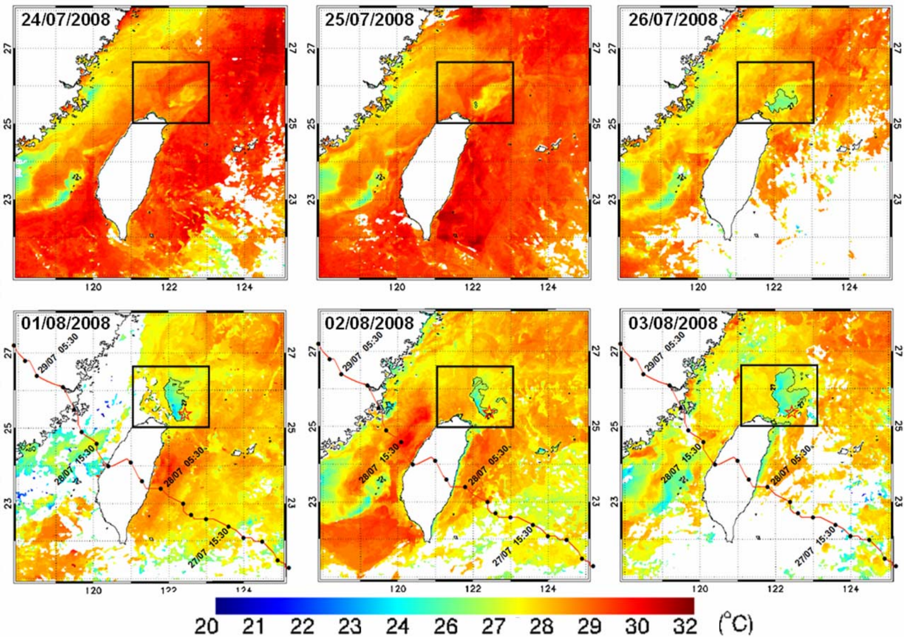

Fig. 1. Study area (blue box, sampling location: a red star) and track (red line in panel C) of typhoon Fengwong in the southern East China Sea. AVHRR satellite images on 23, 24 and 25 before the typhoon and 1, 2 and 3 August after the typhoon. The area of cold water patch (e.g. SST <27 °C) shown in the blue box on 23 July (A) is significantly smaller than that on 2 August (C). The red lines represent the typhoon moving track of Fengwong.

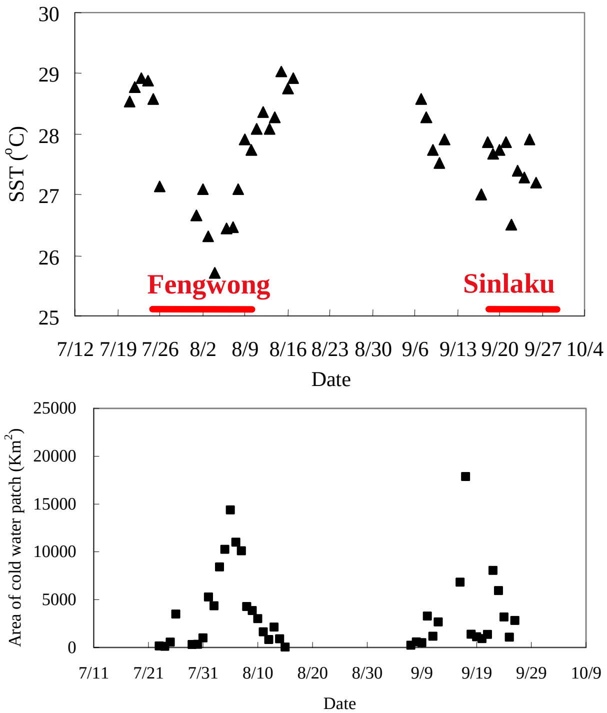

Fig. 2. (A) Sea surface temperature (SST) and (B) the area of cold water patch (e.g. SST <27 °C) in the southern East China Sea (between 25.2–25.7° N and 122.1 to 122.6° E. Each datum point represents the average daily value. The red line represents the influenced periods of pre- and post-typhoon.

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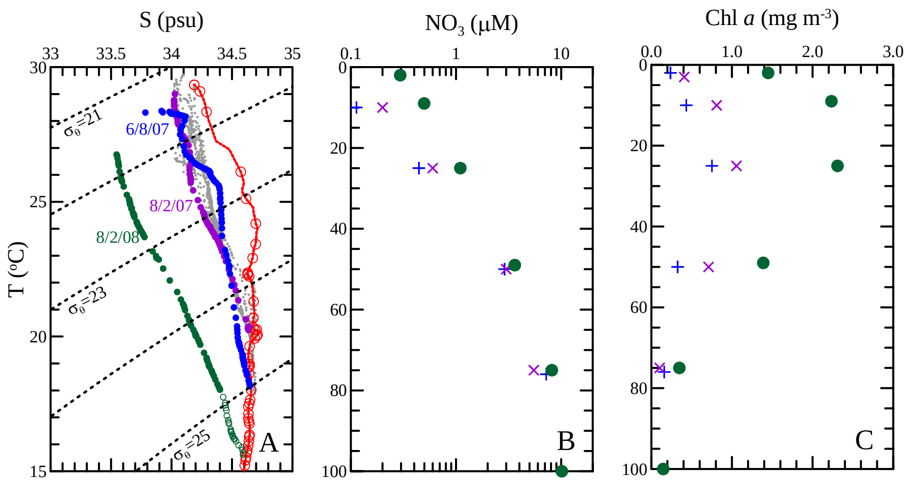

Fig. 3. (A) Diagrams of temperature and salinity (A: a red curve represents the characteristics of the Kuroshio Current; a green curve represents the nature of the upwelled water from the deeper water of the Kuroshio upwelling after a typhoon event; blue, purple, and gray curves represent the upwelled water from the shallow water of the Kuroshio Current under non-typhoon periods;). One can clearly see the colder water from deep water (after typhoon event was brought to the surface deeper than non-typhoon periods. Vertical distributions of nitrate concentration (B) and chlorophyll *a* (chl *a*) concentration (C) in the study area of the southern East China Sea during non-typhoon periods (blue, purple, and gray symbols) and a typhoon event (green symbols). Note: nitrate concentrations (below detection limit, ∼0 μM) in the surface water during non-typhoon periods did not appear because of log scale.

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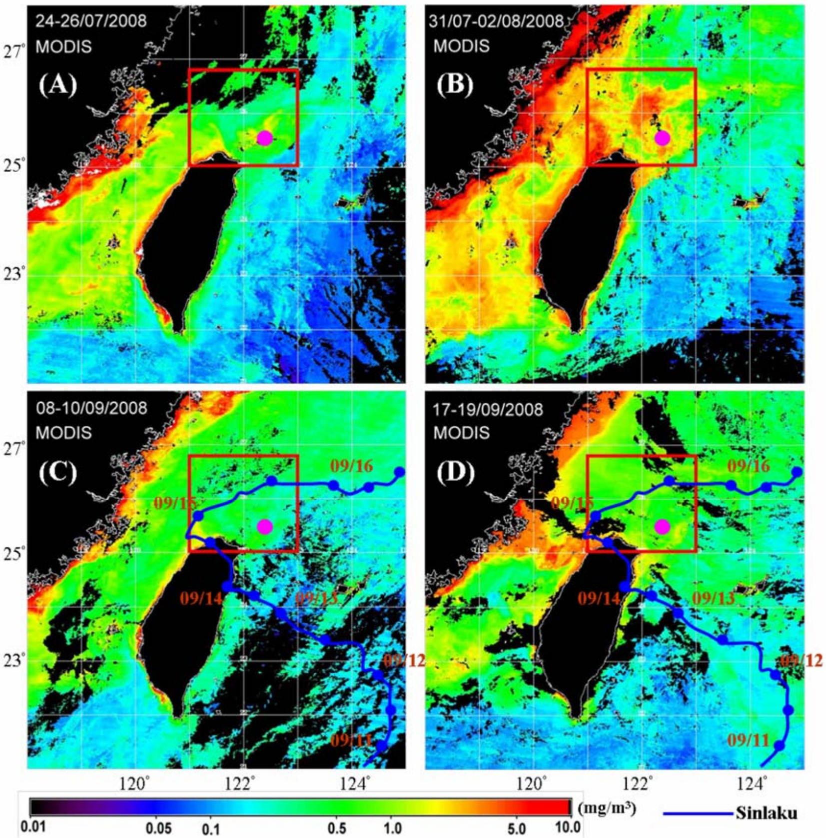

Fig. 4. The derived surface chl *a* concentrations by MODIS before and after typhoon Fengwong (panels A and B, from 24 July to 2 August 2008) and Sinlaku (panels C and D, from 8 September to 19 September 2008) respectively. A pink dot represents the sampling location and a blue line represents the track of typhoon Sinlaku.

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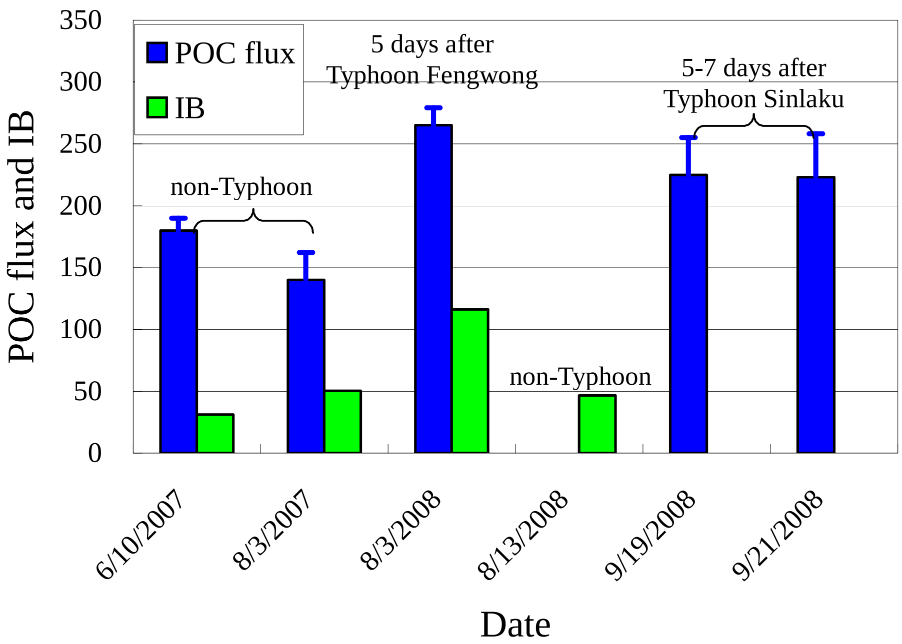

Fig. 5. POC export fluxes (blue bars, mg m^{-2} d^{-1}) and integrated chl *a* (IB, green bars: mg m^{-2}) in the study region on 6/10 in 2007, 8/3 in 2007, and 8/13 in 2008 (non-typhoon conditions), and during typhoon events in 2008 on 8/3 (typhoon Fengwong), and 9/19 and 9/21 in 2008 (typhoon Sinlaku).

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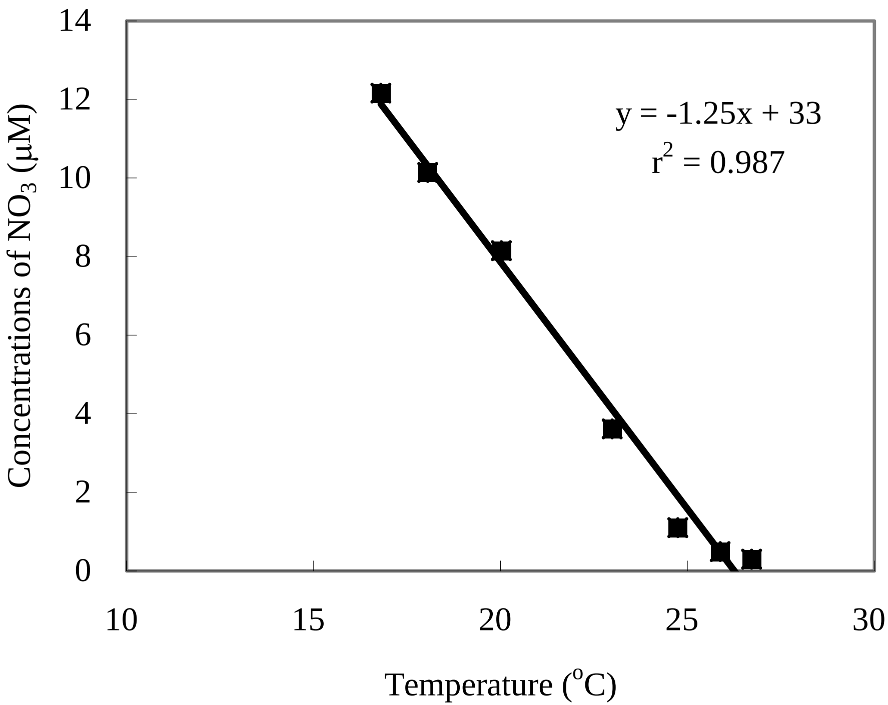

Fig. 6. Relationship between and sea surface temperature (SST) and nitrate concentration in the study area of the southern East China Sea.

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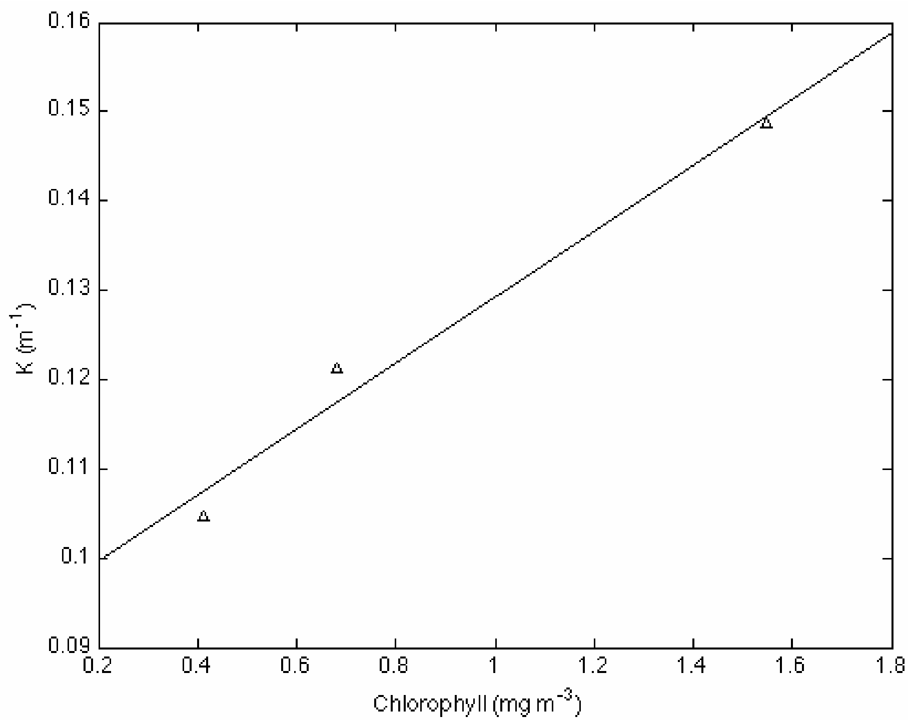

Fig. 7. A relationship between chlorophyll *a* concentration versus vertical light extinction coefficient.

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