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The large variation in organic carbon consumption in spring in the East China Sea

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Abstract

A tremendous amount of organic carbon respired by planktonic communities has been found in summer in the East China Sea (ECS), and this rate has been significantly correlated with fluvial discharge from the Changjiang River. However, data related to this issue in other seasons have rarely been collected. To evaluate and reveal the potential controlling mechanism of organic carbon consumption in spring in the ECS, research using stations covering almost the entire ECS shelf was conducted in the spring of 2009 and 2010. During both periods, the fluvial discharges were similar, and these rates were comparable to high riverine flow in summer. Interestingly, planktonic community respiration (CR) varied widely in both springs; in 2009, the level of CR was double that of 2010, with mean (±SD) values of 111.7 (±76.3) and 50.7 (±62.9) $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$, respectively. The CR was positively linearly regressed with concentrations of particulate organic carbon and/or chlorophyll *a* (Chl *a*) in 2009 (all $p < 0.01$). These results suggest that the rate was dependent on planktonic activities, especially that of phytoplankton, in 2009. During this period, phytoplankton growth flourished due to allochthonous nutrients discharged from the Changjiang River. Furthermore, higher phytoplankton growth leaded to the absorption of an enormous amount of fugacity of CO_2 (fCO_2) in the surface waters, even with a significant amount of inorganic carbon regenerated via CR. In 2010, there were even more riverine runoff nutrients into the ECS than in 2009. Surprisingly, the growth of phytoplankton in 2010 was not stimulated by enriched nutrients, and its growth was likely limited by low water temperature and/or low light intensity. Low temperature might also suppress planktonic metabolism, and this could explain why the CR was lower in 2010. During this period, lower surface water fCO_2 might have mainly been driven by physical process(es). To conclude, these results indicate that organic carbon consumption (i.e. CR) in the ECS in spring might be controlled by the magnitude of planktonic activities and physical factor (e.g. temperature), and that the latter is especially important during a cold spring season. This further suggests

that the high intraseasonal variability of organic carbon consumption needs to be kept in mind when budgeting the annual carbon balance.

1 Introduction

The annual carbon budget is crucial when it comes to determining whether a system is a carbon sink or source. This issue is especially important on high productivity coastal and shelf ecosystems, since it accounts for one-fifth to one-third of global marine primary productivity (e.g. Walsh, 1991; Wollast, 1998). However, there is still controversy over carbon sinks and sources in coastal and shelf ecosystems due to the complexity of their physical and biogeochemical processes (e.g. de Haas et al., 2002; Duarte and Agustí, 1998; Walsh et al., 1981). Estimation of the annual carbon budget will depend on the production and consumption of organic carbon. Organic carbon production, i.e. primary productivity, has been widely measured at a spatiotemporal scale around the world oceans (e.g. Falkowski and Woodhead, 1992; Thornton, 2012 and citation therein). However, there are relatively few data on the decomposition of organic carbon by planktonic respiration (e.g. del Giorgio and Duarte, 2002; del Giorgio and Williams, 2005). Furthermore, highly seasonal and annual variability of organic carbon consumption make it even more difficult to budget the carbon balance (e.g. Chen et al., 2009; Smith and Kemp, 1995).

Contradictory results regarding carbon sinks and sources have also been found in the East China Sea (ECS) ecosystem, one of the largest continental shelves in the world. By using the fugacity of CO_2 (fCO_2), previous studies showed that remarkable atmospheric CO_2 was a drawdown in the surface waters of the ECS (Peng et al., 1999; Tsunogai et al., 1999; Wang et al., 2000). A similar result was also observed in summer, with flourishing planktonic activities; however, supersaturated CO_2 has regenerated in the subsurface waters below pycnocline in the ECS (Chen et al., 2006; Chou et al., 2009). Therefore, as Chen et al. (2006) propose, whether the ECS shelf acts as a carbon sink or source might depend on the amount of regenerated inorganic carbon in the subsurface waters releasing through prevailing physical forces. Furthermore, it has also been validated that the organic carbon consumption rate correlates to planktonic activities, which is in turn proportional to the fluvial discharge rate (e.g. Chen et al., 2009). These results also indicate that organic carbon respired by the planktonic community serves as an important factor in carbon cycling processes, but may vary seasonally under different physical and biogeochemical conditions. Since most studies on organic carbon consumption in the ECS were performed in summer (Chen et al., 2003, 2006, 2009), studies on other seasons are needed to conclusively reveal the annual carbon budget.

To explore this phenomenon further, data from two spring seasons were used to examine the roles and the potential controlling mechanisms of organic carbon consumption. Planktonic community respiration (CR) was taken as the rate of organic carbon consumption in this study. Interestingly, even with similar fluvial discharge, the CR was an order of magnitude different between the two spring seasons, with high spatial variation. To evaluate this difference, physical and chemical hydrographs and biological activities were analyzed and compared to elucidate the potential causes for the spatial and intraseasonal variations. In addition, the relationship between CR and $f\mathrm{CO}_2$ was examined to reveal the role of planktonic activity in the carbon balance in spring.

2 Materials and methods

2.1 Study area and sampling

This study is part of the Long-term Observation and Research of the East China Sea (LORECS) program. Samples were collected on board the *R/V Ocean Researcher I*, Taiwan, in the spring of 2009 (29 April–10 May) and 2010 (11 April–22 April), with a total of 32 and 28 stations, respectively, in the ECS (Fig. 1). Water samples were collected using Teflon-coated Go-Flo bottles (20 l, General Oceanics Inc., USA) mounted on a General Oceanic rosette assembly (Model 1015, General Oceanics Inc., USA). There were 6 to 9 sampling depths at intervals of 3 to 50 m, depending on the water column depth at each station. Subsamples were taken immediately for further analyses (i.e. nitrate, chlorophyll *a*, and particulate organic carbon) and on-board incubation (i.e. primary production and plankton community respiration).

2.2 Hydrographic and optical measurements

Conductivity, temperature, and depth were recorded throughout the water column with a CTD (SBE9/11 plus, Seabird Inc., USA). Photosynthetically active radiation (PAR) was measured throughout the water column with an irradiance sensor (4π; QSP-200L). The depth of the euphotic zone (Z_E) was taken as the depth of 1 % surface light penetration. The mixed layer depth (M_D) was based on a 0.125 unit potential density criterion (Levitus, 1982).

2.3 Nitrate, chlorophyll *a*, particulate organic carbon, and the fugacity of CO_2 (fCO_2)

Water samples for nitrate (NO_3^-) were collected from every sampling depth with 100 ml polypropylene bottles and were frozen immediately with liquid nitrogen. A custom-made flow-injection analyzer was used for nitrate analysis with a detection limit of 0.15 µM (Gong et al., 2003). Integrated values for nitrate and other variables in the water column above the Z_E were estimated by trapezoidal method, whereby depth-weighted means were computed from vertical profiles and then multiplied by Z_E (e.g. Smith and Kemp, 1995). Average nitrate concentration over the Z_E was estimated from the vertically integrated value divided by Z_E, and this calculation was also adopted with other variables.

Chlorophyll *a* (Chl *a*) concentration was measured with a Sea Tech fluorometer attached to the SeaBird CTD for a continuous profile of in vivo fluorescence and calibrated by in vitro fluorometry. Water samples for in vitro Chl *a* measurement were immediately filtered through GF/F filter paper (Whatman, 47 mm) and stored in liquid nitrogen The Chl *a* retained on the GF/F filters was determined fluorometrically (Turner Design 10-AU-005, Parsons et al., 1984). If applicable, Chl *a* was converted to carbon units using a C : Chl value of 52.9, estimated from the shelf waters of the ECS (Chang et al., 2003).

At selected stations, samples for particulate organic carbon (POC) were filtered through a Whatman 25 mm GF/F filter, wrapped in aluminum foil, and then stored at 4 °C. Both the filter and aluminum foil were prebaked at 500 °C for 2 h. After being dried and acid fumed, POC samples were measured by the combustion method using a HORIBA EMIA-510 analyzer (Shiah et al., 2000b).

The fugacity of CO_2 in the surface waters was calculated from dissolved inorganic carbon (DIC) and total alkalinity (TA) data using the designed program (Lewis and Wallace, 1998). For details on TA and DIC measurements, please refer to Chou et al. (2007).

2.4 Primary production

Primary production was measured by the ^{14}C assimilation method (Gong et al., 1999; Parsons et al., 1984). Water samples were collected from three depths within Z_E at selected stations. Samples were prescreened through 200 µm woven mesh (Spectrum), and inoculated with $H^{14}CO_3^-$ (final conc. 10 µCi ml^{-1}) in 250 ml clean polycarbonate bottles (Nalgene). Samples were incubated on board for 2–4 h in chambers filled with running surface seawater and illuminated by fluorescent bulbs with a light intensity corresponding to the in situ irradiance levels (Gong et al., 1999). Following retrieval, samples were filtered on GF/F filters and acidified overnight after the addition of 0.5 ml 2N HCl. Following immersion in 10 ml of scintillation cocktail (Ultima Gold, Packard), total activity on the filter was counted in a liquid scintillation counter (Packard 2700TR). To estimate the euphotic zone-integrated primary production at stations where incubation was not performed, an empirical function was applied (please refer to Gong and Liu (2003) for details).

2.5 Plankton community respiration (CR)

The CR was measured as the decrease in dissolved oxygen (O_2) during dark incubation (Gaarder and Grann, 1927). Incubation was conducted at most of the stations in the ECS, with duplicate samples taken from 2 to 6 discrete depths within Z_E at each station. Treatment samples were siphoned into 300 ml biological oxygen demand bottles. The treatment involved incubating bottles for 24 h in a dark chamber filled with running surface water which have maximum temperature changes (mean ± SD) of $1.89 \pm 1.70\,^{\circ}$C and $3.33 \pm 2.52\,^{\circ}$C during each incubations in 2009 and 2010, respectively. Temperature differences between top and bottom of Z_E in all incubation stations were also small with mean (±SD) values of 1.58 (±1.30) °C and 0.91 (±1.20) °C in 2009 and 2010, respectively. The difference in O_2 concentration between initial and dark treatment was used to compute the CR (see Chen et al. (2003) for details). To convert respiration from oxygen to carbon units, a respiration quotient (RQ) of 1 was assumed (Hopkinson, 1985; Parsons et al., 1984).

3 Results and discussion

3.1 Hydrographic patterns in the East China Sea (ECS) in spring

The ECS shelf ecosystem has been strongly influenced by complex and dynamics physical forcing including, coastal river discharge, intrusions of the Yellow Sea waters, the Taiwan Strait waters, and the Kuroshio waters, as well as the alternating monsoons (e.g. Liu et al., 2003). Regardless of its complexity, there is a general physical distribution pattern in the ECS, i.e. both sea surface temperature (SST) and salinity (SSS) increase from the inner shelf toward the slope (Gong et al., 1996; Tseng et al., 2000). Spatially, the deviation of SST and SSS is predominantly contributed by the coastal river discharge, especially from the Changjiang (aka Yangtze River).

The largest variation was consistent with high river flow periods in late spring to early summer in the ECS (Chen et al., 1994; Tseng et al., 2000). High fluvial discharge was also observed in both spring seasons included in this study, and this can be demonstrated in the large area of the Changjiang Diluted Water (CDW; SSS $\leqq 31.0$ psu), which is assumed to be an index of riverine runoff (Chen et al., 2009). The area of CDW for 2009 (23 638 km^2) and 2010 (19 907 km^2) in this study were larger than the mean area of CDW (15 604 km^2) in summer observed in a previous study (Chen et al., 2009). As expected, the increasing trend of SST and SSS from the inner shelf to the slope was also found in both spring seasons included this study, with the lowest values observed in the Changjiang plume region (Fig. 2). This phenomenon was caused by tremendous fluvial discharge from the Changjiang River, and it can be verified from the positively linear relationship between SST and SSS in both study periods (all $p < 0.001$).

Even with similar a distribution pattern, larger spatial deviations for SST and SSS were observed in 2010 than for 2009 (Table 1; Fig. 2). The values of SST and SSS in 2009 were in the range of 15.7–25.3 °C (mean ± SD = 19.4 ± 2.3 °C) and 27.93–34.53 psu (mean ± SD = 32.61 ± 1.93 psu), respectively (Table 1; Fig. 2a, b). Their values in 2010 were in the range of 10.6–25.3 °C (mean ± SD = 16.9 ± 4.7 °C) and 18.35–34.76 psu (mean ± SD = 32.20 ± 3.46 psu), and SST was significantly lower in 2010 than in 2009 (Table 1; Fig. 2c, d). These results suggest that huge amounts of dilute waters were discharged from China's coast, especially the Changjiang, onto the ECS shelf, and the discharged riverine water was colder during the 2010 study period.

During riverine runoff, large amounts of dissolved inorganic nutrients usually accompany and discharge onto the shelf ecosystem. In this study, this can be validated from the surface contours of nitrate, which revealed a similar trend to SSS distribution during both periods, i.e. the nitrate concentration in the surface waters was higher along the inner shelf and nearly depleted in the slope (Fig. 3a, c). A similar spatial pattern was also found for phosphate and silicate in the surface water in this (data not shown) and previous studies (e.g. Gong et al., 1996, 2003). Further analyses show that nitrate concentrations in the surface waters were negatively linear regressed with SSS in both periods (Fig. 4a; all $p < 0.001$). These inverse relationships also hold true in terms of SSS vs. phosphate or silicate in the surface waters (data not shown; all $p < 0.001$). These results indicate that nutrients followed a dilution pattern with riverine runoff as a major nutrient source in the ECS surface plume ecosystem.

Comparing between two spring periods, the surface water nitrate was significantly lower in 2009 than in 2010, with values in the range of 0.0–25.3 µM (mean ± SD value = 3.1 ± 5.8 µM) and 0.0–61.3 µM (mean ± SD value = 10.5 ± 14.4 µM), respectively ($p < 0.005$; Fig. 3a, c). The average nitrate concentration over the euphotic zone was also lower in 2009 (mean value = 3.6 µM) than in 2010 (mean value = 10.1 µM), but this result was not statistically significant ($p = 0.14$; Table 1). Interestingly, a previous study showed that the nutrient concentration was positively related to the area of CDW in summer (Chen et al., 2009). Nitrate concentration, however, was relatively lower in 2009, even with the larger CDW area, when compared to 2010. This implies that sequential biological responses may behave differently following the intrusion of coastal diluted waters enriched with nutrients and organic matter.

3.2 Responses of planktonic community in spring

To explore the sequential biological response to fluvial nutrient input, the planktonic communities (i.e. biomass and rates) were compared between the spring of 2009 and 2010. In 2009, the mean values of mixed-layer depth (M_D) and euphotic depth (Z_E) were 20.6 ± 17.7 m and 35.8 ± 16.6 m, respectively (Table 1). The mean value of M_D was smaller than Z_E in 2009, but the difference was not statistically significant. The M_D was slightly deeper than Z_E in 2010 with mean ± SD values of 31.0 ± 24.2 m and 28.3 ± 20.5 m, respectively (Table 1). Overall, the M_D was shallower than or close to Z_E in this study. For comparison, the biomass and rate of planktonic communities were therefore integrated over Z_E, since most of them were measured within Z_E.

Phytoplankton is one of the major planktonic communities, and responds instantaneously following intrusion of enriched nutrient diluted water. In the ECS, previous studies found a higher biomass of phytoplankton in the CDW during high riverine flow periods in summer, with a mean value of 3.3 mg Chl m^{-3} (Gong and Liu, 2003; Gong et al., 2003). The surface water Chl *a* showed a similar trend, with the highest value found in the CDW in 2009 of this study (Fig. 3b). The values were in the range of 0.20–10.96 mg Chl m^{-3} with a mean ± SD value of 1.81 ± 2.36 mg Chl m^{-3} during this period.

The mean Chl *a* value over the Z_E was almost the same as that of surface water (Table 1). Even though the mean value was not as high as 3.3 mg Chl m^{-3} observed during the peak summer season, the highest value (10.96 mg Chl m^{-3}) was comparable to that in summer (Gong and Liu, 2003; Gong et al., 2003). Moreover, the Chl *a* spatial trend was similar to surface water nitrate distribution in the spring of 2009 (Fig. 3a, b). One would therefore expect that the phytoplankton biomass would be associated with available nutrients during this study period. This assumption was confirmed by the positively linear relationship found between Chl *a* and nitrate concentrations in the surface waters in 2009 ($p < 0.001$). It was also supported by linear correlations observed between Chl *a* vs. phosphate or silicate concentrations of the surface waters (data not shown; $p < 0.001$). In addition to surface water variables, linear regressions were also statistically significant between the averaged values over Z_E of Chl *a* vs. nitrate or silicate concentrations in the spring of 2009 ($p < 0.001$). These results all show that the vigorous growth of phytoplankton was enhanced by dissolved inorganic nutrients enriched from the intrusion of riverine water into the ECS in this period (Fig. 4).

The distribution pattern was similar in SSS and nitrate in the spring of 2009 and 2010, and a similar pattern was also found in surface water Chl *a* in 2009. Surprisingly, the spatial trend was somewhat different in surface water Chl *a* in 2010, and the higher value was observed in the middle shelf (Fig. 3d). In the spring of 2010, the surface water Chl *a* value was in the range of 0.03–2.48 mg Chl m^{-3} with a mean ± SD value of 1.03 ± 0.72 mg Chl m^{-3}. Amazingly, even nitrate concentrations were higher in 2010 than that in 2009; both mean Chl *a* values in the surface waters and averaged over Z_E were relatively low in 2010 compared to 2009, but this difference was statistically insignificant (Fig. 3b, d; Table 1). This suggests that growth of phytoplankton might not have been limited by nitrate in 2010, especially in the CDW regions. This could also explain why even surface water nitrate was linear regressed with SSS, but surface water Chl *a* was not related to SSS in 2010 (Fig. 4).

Besides available dissolved inorganic nitrogen, phytoplankton growth limited by phosphate has been observed in the ECS, especially in the CDW (Chen et al., 2004; Gong et al., 1996; Harrison et al., 1990). In 2010, the mean molar ratio of N/P in the surface waters was 17.5 for the stations around the CDW regions (e.g. Sts. 19 and 21). These results indicate that dissolved inorganic nutrients may not limited growth of phytoplankton in terms of either availability or N/P molar ratio, especially in the CDW. Beyond nutrients, phytoplankton growth may have been regulated by temperature, light intensity, and/or herbivorous grazers. Among the analyzed variables, SST was significantly lower in 2010 (mean = 16.9 °C) than that in 2009 (mean = 19.4 °C; Table 1). Moreover, the mean SST was much lower in the CDW region in 2010 (mean = 12.2 °C) than in 2009 (mean = 18.0 °C; Fig. 2a, c).

A temperature limit on the primary production and growth of phytoplankton has been observed in many estuarine and marine ecosystems (e.g. Malone et al., 1988; Behrenfeld et al., 2006). To explore this further, Chl *a* concentrations in the surface waters were positively and negatively correlated, respectively, with SST $< 15\,^{\circ}\mathrm{C}$ and $\geqq 15\,^{\circ}\mathrm{C}$ (Fig. 5). The positive linear regression found in SST $< 15\,^{\circ}\mathrm{C}$ indirectly evidenced that the growth of phytoplankton was mostly limited by cold temperature, and not nutrient availability (mean value = 20.9 µM). In addition to temperature, the inverse relationship for SST $\geqq 15\,^{\circ}\mathrm{C}$ suggested that phytoplankton growth might also be regulated by other factors, and the amount of available nutrients might be the potential variable. Similar combined effects of temperature and substrate limiting bacterioplanktonic production have been observed in the ECS (Shiah et al., 1999). This assumption can be verified through significant multiple linear regression between Chl *a* vs. SST and surface water nitrate in SST $\geqq 15\,^{\circ}\mathrm{C}$ in this period ($p < 0.05$). To briefly conclude, the growth of phytoplankton was mostly limited by SST in the inner shelf when SST $< 15\,^{\circ}\mathrm{C}$, and it was regulated by a combined effect of SST and nutrient availability for the middle to outer shelves in the spring of 2010.

In addition to SST and inorganic nutrients, light intensity might be another important factor regulating phytoplankton growth. The mean value of PAR in the water surface was about two times lower in 2010 (34.4 E m^{-2} d^{-1}) than in 2009 (60.9 E m^{-2} d^{-1}; Table 1; $p < 0.001$). This suggests that the lower Chl *a* in 2010 might be strongly associated with the lower light intensity during this period. The effect of light intensity on primary production (PP) in aquatic ecosystems is well known, and PP usually increases hyperbolically with an increase in light intensity (Kirk, 1994). As stated above, a previous study showed that the highest PP, seasonally, was observed in summer with mean values of integrated PP and PAR of about 1000 mg C m^{-2} d^{-1} and 79.9 E m^{-2} d^{-1}, respectively (Gong et al., 2003). The mean (±SD) values of integrated PP and averaged PP over Z_E were 485.9 (±571.9) mg C m^{-2} d^{-1} and 17.8 (±15.9) mg C m^{-3} d^{-1}, respectively, in the spring of 2010 (Table 1). This mean integrated PP value was only about half of that observed during the high production season in summer (Gong et al., 2003).

The averaged PP in 2010 were two orders of magnitude lower than the highest value (i.e. 747.2 mg C m^{-3} d^{-1}) observed during a phytoplankton bloom in the ECS (Chen et al., 2004), and most of them were at the lower end of the reported values in coastal ecosystems (Table 1; Duarte and Agustí, 1998 and citations therein). Unfortunately, PP was not measured in this study in 2009. However, the lower PP observed in 2010 suggests that growth of phytoplankton might have been limited by low light intensity (mean PAR value = 34.4 E m^{-2} d^{-1}) in this period (Table 1). Light limitation on phytoplankton growth was even more critical due to deeper M_D (31.0 m) and shallower Z_E (28.3 m) in the spring of 2010 (Table 1). To conclude, these results imply that the growth of phytoplankton was limited by both the cold water temperature and low light intensity in spring of 2010. This can explain why the phytoplankton biomass (i.e. Chl *a*) was lower in 2010 than in 2009 although there were more available nutrients in the spring of 2010 (Table 1).

3.3 High organic carbon consumption in spring 2009

Planktonic CR provides an integrated rate measurement of biotic organic carbon consumption, integrating components including bacterioplankton, phytoplankton, planktonic protozoa, and zooplankton in aquatic ecosystems (Calbet and Landry, 2004; del Giorgio and Duarte, 2002; Hernández-León and Ikeda, 2005; Hopkinson et al., 1989). Previous studies have shown that the CR was mostly attributed to bacterioplankton and/or phytoplankton, and the contribution by planktonic protozoa was trivial in the ECS (Chen et al., 2003, 2006, 2009). This further demonstrates the CR rate was positively related to total planktonic biomass in the ECS (Chen et al., 2006).

As stated above, the mean Chl *a* values were relatively higher in 2009 than in 2010, both in the surface waters and in the averaged value over Z_E (Fig. 3b, d; Table 1). Higher CR could therefore be expected for spring 2009 with the higher phytoplankton biomass. Indeed, the mean value of averaged CR over Z_E was double in 2009 than that in 2010 with mean (±SD) values of 111.7 (±76.3) and 50.7 (±62.9) $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$, respectively (Table 1; $p < 0.001$). The mean averaged CR value in 2009 was comparable to the high rate (i.e. 114 $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$) observed during peak planktonic growth in the ECS in summer (Chen et al., 2009). This indicates that a huge amount of organic carbon was respired by planktonic communities in the spring of 2009.

The CR contour demonstrates spatial patterns of biotic organic consumption across the entire shelf of the ECS. The CR in the surface waters ranged from 10.1 to 458.6 $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$ with a mean (±SD) value of 132.0 (±95.6) $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$ in 2009 (Fig. 6a). Spatially, in 2009, the higher CR values were mostly along the coast, and surprisingly, in part of outer shelf of the ECS (e.g. Sts. 9, 10, and 12). The higher CR can be attributed to the higher planktonic biomass; a positive correspondence between CR and phytoplankton biomass or primary production has been observed widely in marine ecosystems (e.g. Jensen et al., 1990; Robinson et al., 2002; Smith and Kemp, 1995). This linear relationship was also evidenced between CR and Chl *a* either in surface waters ($r^2 = 0.34$; $p < 0.001$) or using averaged values over Z_E ($r^2 = 0.25$; $p < 0.01$)

in this period. The result still holds true when using Chl *a* expressed in carbon units; the coefficients of linear slope for the surface waters and the averaged value were 0.60 and 0.42, respectively. This suggests about half of the CR was contributed by phytoplankton, and other half might be attributed to others planktons.

Among other planktonic communities, bacterioplankton has been recognized as an important component of organic carbon consumption in many aquatic ecosystems (e.g. Chin-Leo and Benner, 1992; del Giorgio and Duarte, 2002; Griffith et al., 1990). Shiah et al. (2000b) estimated that in the ECS, heterotrophic bacterioplankton has completely consumed in situ particulate organic carbon production. Unfortunately, this was not measured in the present study. However, to examine how CR relates to planktonic communities, POC can be assumed to be an indicator of total planktonic biomass. The averaged POC value over Z_E was in the range of 82.2–613.8 mg C m^{-3}, with a mean (±SD) value of 227.5 (±141.9) mg C m^{-3} in 2009 (Table 1). Phytoplankton biomass accounted for 42 % of the mean POC if Chl *a* was expressed per carbon units. Although the mean Chl *a* value (1.8 mg Chl m^{-3}) in 2009 was not as high as the mean value (3.3 mg Chl m^{-3}) observed in summer in previous studies (Table 1; Gong and Liu, 2003; Gong et al., 2003), the POC values in this study were higher than those found in a previous study in summer in the ECS (Chen et al., 2009). Moreover, the high CR rate was linear regressed with POC both in the surface waters ($r^2 = 0.38$; $p < 0.001$) and using averaged values over Z_E ($r^2 = 0.31$; $p < 0.01$) in this period. These results imply that high rates of CR in 2009 were associated with a higher planktonic biomass. The high POC suggested that, besides phytoplankton, bacterioplankton might be served as another important component contributing to the CR, in this period.

3.4 Low organic carbon consumption in the spring of 2010

It was reasonable to expect a lower CR observed in 2010, since the phytoplankton biomass and SST were lower in this period. Indeed, the rates of CR were significantly lower in 2010 compared to those in 2009 ($p < 0.001$; Table 1). The averaged CR values over Z_E ranged from 4.2 to 242.6 mg C m^{-3} d^{-1} with a mean (±SD) value of 50.7

(±62.9) mg C m^{-3} d^{-1} in the spring of 2010 (Table 1). This mean value was at the lower end of the reported CR values in the ECS and in the coastal shelf, as well as in the slope regions (i.e. 2.3–485.3 mg C m^{-3} d^{-1} if assuming RQ = 1; Biddanda et al., 1994; Williams, 1984; Chen et al., 2003, 2006, 2009). As stated above, one of reasons for the lower CR might be the lower phytoplankton biomass observed in this period. This can be confirmed from the significant linear relationship found between integrated values over Z_E of CR and Chl *a* in 2010 (data not shown; $r^2 = 0.29$; $p < 0.01$), but not that in the surface waters and in the averaged values over Z_E. In addition, the linear relationship between the integrated values of CR and PP was also proven for this period (data not shown; $r^2 = 0.89$; $p < 0.001$). This further supports the assumption that lower CR observed in 2010 might be related to the low biomass and production of phytoplankton.

Besides phytoplankton, the CR might also have been limited by the lower temperature in this period. A limitation of temperature on the growth of bacteria has been found in the ECS, particularly for water temperatures below 20 °C (Shiah et al., 1999, 2000a). The mean value of SST was 16.9 °C in the spring of 2010 (Table 1). This indicates that bacterial growth, as well as CR, was potentially confined by the low water temperature. Even though the low temperature limitation on bacterial biomass and production could not be verified, since it was not measured in this study, the temperature limitation on phytoplankton could be indirectly validated from the positively linear relationship between Chl *a* concentrations and SST < 15 °C in the surface waters ($p \leqq 0.001$; Fig. 5). Moreover, the integrated CR values were positively linearly regressed with the averaged temperature over Z_E in 2010 ($r^2 = 0.27$; $p < 0.01$). These results suggest that the lower CR in 2010 might be due to the low planktonic biomass and the low temperature restriction on the metabolic rates of planktonic communities.

To elucidate whether an ecosystem is autotrophic or heterotrophic, the ratio of primary production to respiration (P/R ratio) is applied. In this study, the ratio of integrated values (mg C m^{-2} d^{-1}) of primary production to CR was used to explore organic carbon utilization between production and consumption in the spring of 2010 in the ECS. The results showed that the P/R ratios were in the range of 0.06 to 2.30 with a mean (±SD) value of 0.85 (±0.71). There were five stations with a P/R ratio $\geqq 1$ (mean value = 1.47), and they were all in the inner shelf (i.e. Sts. 5, 21, 28, 29, and 30; Fig. 1). Interestingly, the lower P/R ratios (< 1) were observed mostly in the middle to outer shelves in this period, except for St. 19A. These results suggest that, in the ECS, the inner shelf ecosystem was autotrophic, but the middle to outer shelves were more heterotrophic in the spring of 2010. Previous studies have also shown that the ECS shelf was a net heterotrophic ecosystem in other seasons (Chen et al., 2003, 2006), except for the inner shelf with high primary productivity ($> 3500\,\mathrm{mg\,C\,m^{-2}\,d^{-1}}$). To support the claim that the ECS is net heterotrophic, high bacterial consumption has been suggested and supported by the observation of a tremendous amount of organic carbon discharged from coastal rivers, especially the Changjiang River (Cauwet and Mackenzie, 1993; Chen et al., 2003; Chen and Wang, 1996). However, high CR was not observed during the high riverine discharge period of 2010 (Table 1; Fig. 2c, d).

Bacterial activity might be limited by lower water temperature (mean value = 12.2 °C in the CDW region) even with an enormous amount of organic carbon in the inner shelf. This can further explain why an autotrophic ecosystem was still found in this low primary production region. The low P/R ratio (i.e. 0.85) in this period was also consistent with previous studies, where such a low ratio (i.e. < 1) has been widely observed in coastal regions with low primary productivity (e.g. del Giorgio et al., 1997; Duarte and Agustí, 1998).

3.5 Effect of planktonic community respiration on $f\mathrm{CO_2}$

To further understand how organic carbon consumption effects $f\mathrm{CO_2}$ dissolved in the surface waters, the relationship between CR and $f\mathrm{CO_2}$ was evaluated. In this study, $f\mathrm{CO_2}$ was in the range of 130.9–363.4 µatm (mean ± SD = 269.6 ± 55.3 µatm) and 199.7–400.3 µatm (mean ± SD = 293.6 ± 58.4 µatm) in the spring of 2009 and 2010, respectively (Table 1). The mean value of $f\mathrm{CO_2}$ was slightly higher in 2010 than in 2009, even though the SST was higher in 2009 (Table 1). Exploring the absorption of atmospheric $\mathrm{CO_2}$ by surface waters in the ECS, Tsunogai et al. (1999) found that $f\mathrm{CO_2}$

was positively multiple linear regressed with SST, SSS, and phosphate. Similar multiple regression was also significantly proven for 2009 ($r^2 = 0.74$; $p < 0.001$); however, a significant relationship was not found for 2010. These results indicate that to estimate fCO_2, the empirical function suggested by Tsunogai et al. (1999) might need to be modified, especially during the cold period in the ECS. Furthermore, fCO_2 is temperature dependent, and should decrease with decreasing temperature (e.g. Goyet et al., 1993; Tsunogai et al., 1999). The mean value of fCO_2, however, was higher in the lower SST of 2010 than that in the higher SST of 2009 (Table 1). This result suggests that the lower fCO_2 observed in 2009 might have resulted from strong absorption by photosynthesis. Although primary production was not measured in 2009, this assumption could still be indirectly evaluated from relationships between fCO_2 and Chl *a*, POC, or CR – particularly CR, which is an integrated response of planktonic activities. To verify this, it was found that fCO_2 was indeed negatively linearly regressed against CR in the surface waters in 2009 ($p < 0.05$; Fig. 7). A similar relationship still held true between fCO_2 and the averaged CR over Z_E in 2009 ($p < 0.01$). The higher CR indicates that planktonic activities were vigorous. The lower fCO_2 observed in 2009 implies that more CO_2 was absorbed via photosynthesis than that regenerated from CR in regions with higher planktonic activities. This was also be supported by the negatively linear relationship found between fCO_2 and POC in the surface waters of 2009 ($r^2 = 0.43$; $p < 0.001$). A negatively linear regression was also found between fCO_2 and Chl *a* in the surface waters from 2009, but at the margin of statistical significance ($p = 0.07$). A similar relationship between fCO_2 and Chl *a* was evident at the level of significance in a previous study conducted in the ECS (Chen et al., 2006). This implies that the huge deficiency between surface water fCO_2 (269.6 µatm) in 2009 and atmospheric CO_2 (approx. 360 µatm) could mostly be attributed to vigorous planktonic activities. These results also suggest that there was a strong carbon sink during this study period, especially with the help of biological uptake.

There was no significant relationship observed between fCO_2 and CR or Chl *a* in surface water for 2010 (Fig. 7). This suggests that effect of biological activity on fCO_2

might be trivial, especially during a cold period with low primary production. The $f\mathrm{CO}_2$ might therefore have been driven mostly by the physical properties of seawater during this period. This can further explain why a heterotrophic ecosystem with a low P/R ratio (0.85) would still have a small mean value of $f\mathrm{CO}_2$ (293.6 µatm). Overall, these results suggest that the relative contribution of planktonic activities and physical processes to shelf carbon cycling in spring might depend on the magnitude of planktonic growth. That is, planktonic communities could have a significant impact on surface water $f\mathrm{CO}_2$ during flourishing planktonic growth. However, surface water $f\mathrm{CO}_2$ might be mostly controlled by physical factor(s) and less influenced by planktonic activities in a cold spring period.

4 Conclusions

Previous studies on planktonic community respiration (CR) in the East China Sea (ECS) have mostly been conducted in summer, and the rate of CR is significantly correlated with the fluvial discharge rate (e.g. Chen et al., 2009). To fill the seasonal gap, the present study on organic carbon consumption (e.g. CR) was performed in the spring of 2009 and 2010, with stations covering almost the entire ECS shelf, when riverine flows were similar to summer with high fluvial discharge. The hydrographic results showed that spatial distribution patterns in temperature (SST) and salinity (SSS) of the surface waters were similar in both springs, i.e. both SST and SSS increased from the inner shelf toward the slope. However, mean value of SST was significantly lower in 2010 (16.9 °C) than that in 2009 (19.4 °C).

Spatially, nitrate concentration in the surface waters also revealed a similar trend to SSS distribution in both periods, but the mean value of nitrate was statistically higher in 2010 (10.5 µM) than in 2009 (3.1 µM). These results also suggest that nutrients follow a dilution pattern with riverine runoff as a major nutrient source in the ECS surface plume. Surprisingly, even with higher nitrate concentration, the mean Chl *a* value in surface waters was lower in 2010 ($1.03\,\mathrm{mg\,Chl\,m^{-3}}$) compared to 2009 ($1.81\,\mathrm{mg\,Chl\,m^{-3}}$).

Further analyses showed that the growth of phytoplankton in spring of 2010 might be limited by low water temperature and light intensity, but not nutrients.

Even though the phytoplankton biomass was slightly higher in 2009, the CR in this period was double (mean value = 111.7 $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$) that of 2010 (mean value = 50.7 $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$). This CR rate in 2009 was comparable to the highest rate observed in summer in the ECS (Chen et al., 2006, 2009). Spatially, the higher rates of CR were mostly observed along the coast, and this corresponded to the higher Chl *a* concentration, suggesting that a remarkable amount of organic carbon consumption was respired by flourishing planktonic communities in 2009. Although no other planktonic biomass was measured in this study, its contribution to CR can be indirectly evidenced from the high particulate organic carbon (mean value = 227.5 $\mathrm{mg\,C\,m^{-3}}$) observed in 2009.

Even with the high CR, a huge amount of fugacity of CO_2 (fCO_2) in the surface waters was still drawn down due to vigorous phytoplankton activity in this period. This also can explain why mean fCO_2 during the warm spring of 2009 (mean value = 269.6 µatm) was even lower than that in the cold spring of 2010 (mean value = 293.6 µatm). In contrast, the mean CR value in 2010 was at the lower end of the reported CRs in the ECS and in the coastal shelf, as well as slope regions (e.g. Biddanda et al., 1994; Williams, 1984; Chen et al., 2003, 2006, 2009). The results suggest that this might have been caused by the low biomass and metabolism of plankton, which was suppressed by the low water temperature. During this period, the fCO_2 of the surface waters might mainly have been driven by physical processes, and the effect of biological activity on fCO_2 could be trivial.

To conclude, these results indicate that the contribution of planktonic communities to organic carbon consumption and fCO_2 in surface waters could have high intraseasonal variability. While planktonic growth is flourishing, the production and consumption of organic carbon by planktonic communities could have a significant impact on surface water fCO_2. However, surface water fCO_2 might be mainly controlled by physical factor(s) and less influenced by planktonic activities in cold periods. The results also suggest that special attention needs to be given when budgeting annual carbon balances due to this intraseasonal variability.

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Table 1. Range of different variables with mean ± standard deviation (in parentheses) in surface water (2–3 m) and averaged values over euphotic depth (Z_E; m) in the ECS in the spring of 2009 and 2010. Surface water variables include water temperature (SST; °C), salinity (SSS), and fugacity of CO_2 (fCO_2; µatm). Variables of averaged value include nitrate (NO_3^-; µM), chlorophyll *a* (Chl *a*; $mg\,Chl\,m^{-3}$), particulate organic carbon (POC; $mg\,C\,m^{-3}$), primary production (PP; $mg\,C\,m^{-3}\,d^{-1}$), and planktonic community respiration (CR; $mg\,C\,m^{-3}\,d^{-1}$). Values of Z_E and mixed layer depth (M_D; m) are also shown. In addition, photosynthetically active radiation (PAR; $E\,m^{-2}\,d^{-1}$) is provided for reference. Results of the comparison between 2009 and 2010 are also indicated.

Months Variables	2009	2010
PAR	20.6–73.1 (60.9 ± 13.9)	7.7–58.3 (34.4 ± 18.2)^b
Z_E	9.0–66.0 (35.8 ± 16.6)	1.0–70.7 (28.3 ± 20.5)
M_D	5.0–74.0 (20.6 ± 17.7)	5.0–81.0 (31.0 ± 24.2)
SST	15.7–25.3 (19.4 ± 2.3)	10.6–25.3 (16.9 ± 4.7)^a
SSS	27.93–34.53 (32.61 ± 1.93)	18.35–34.76 (32.20 ± 3.46)
fCO_2	130.9–363.4 (269.6 ± 55.3)	199.7–400.3 (293.6 ± 58.4)
NO_3^-	0.0–25.6 (3.6 ± 5.4)	0.1–59.5 (10.1 ± 13.3)
Chl *a*	0.3–7.9 (1.8 ± 2.0)	0.1–5.9 (1.3 ± 1.1)
POC	83.2–613.8 (227.5 ± 141.9)	–
PP	–	5.9–62.7 (17.8 ± 15.9)
CR	15.2–307.3 (111.7 ± 76.3)	4.2–242.6 (50.7 ± 62.9)^b

–: no data;
[a]: $p < 0.05$;
[b]: $p < 0.001$

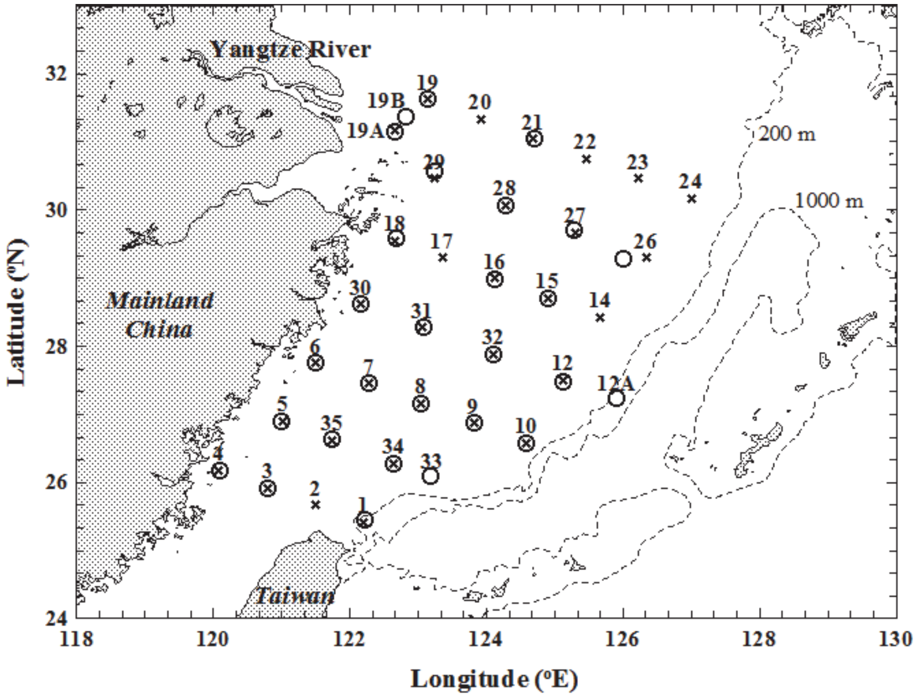

Fig. 1. Map of stations in the spring of 2009 (×) and 2010 (∘) in the East China Sea (ECS) with the station number above the mark. Bottom depth contours (dashed lines; 200 and 1000 m) are also shown; this is also the case in Figs. 2, 3, and 6.

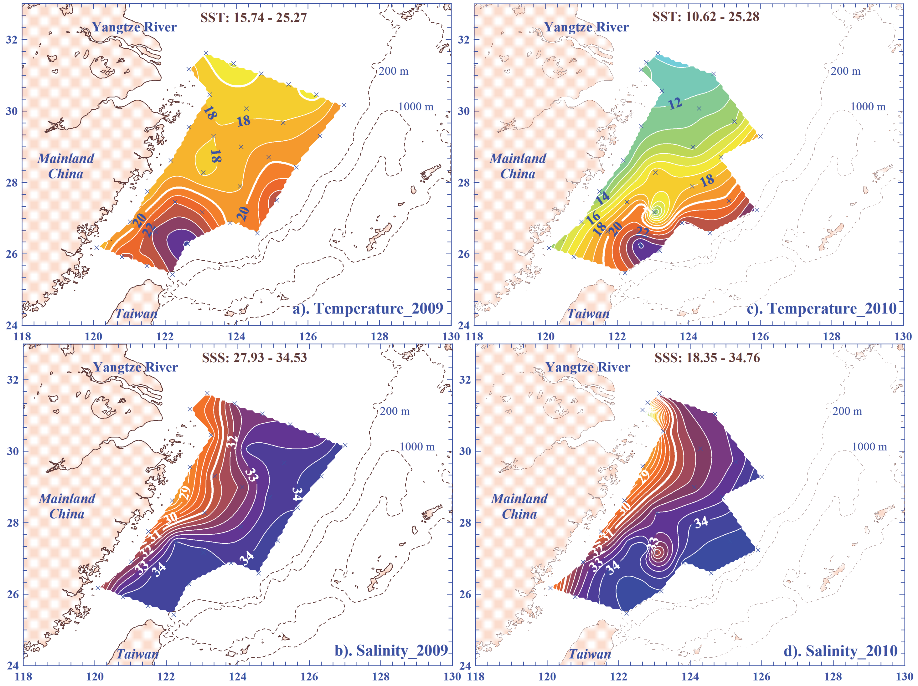

Fig. 2. Contour plots of surface water temperature (SST) and salinity (SSS) of the ECS in 2009 **(a, b)** and 2010 **(c, d)**, with contour lines of SST = 20 °C and SSS = 31 were bolded for reference. Contour intervals of temperature and salinity are 1 °C and 0.5, respectively.

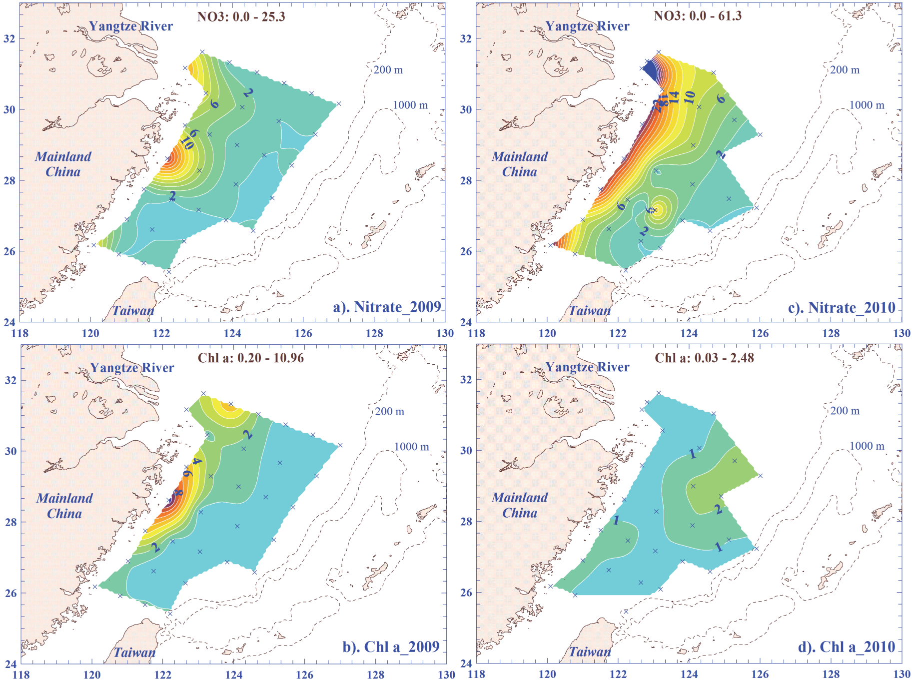

Fig. 3. Contour plots of nitrate and chlorophyll *a* (Chl *a*) in the surface waters of the ECS in 2009 **(a, b)** and 2010 **(c, d)**. The contour intervals of nitrate and Chl *a* are 2 µM and 1 mg Chl m^{-3}, respectively.

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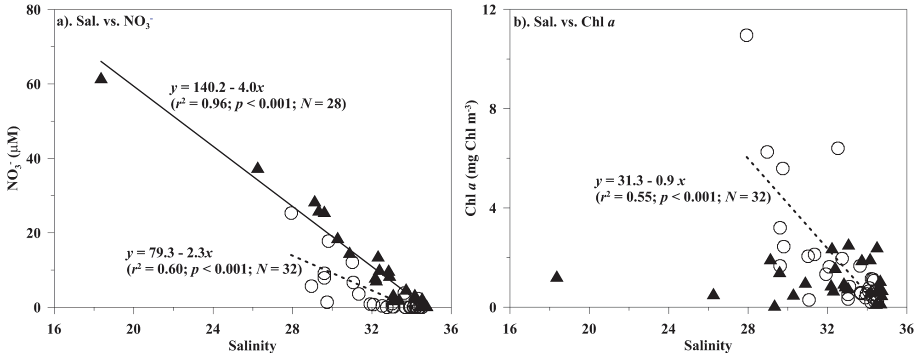

Fig. 4. Relationships between **(a)** nitrate (NO_3^-) and **(b)** chlorophyll *a* (Chl *a*) vs. salinity of the surface waters in the spring of 2009 (∘) and 2010 (▲) of the ECS. Both p and r^2 values of linear regression are also shown if statistical significance was evidenced.

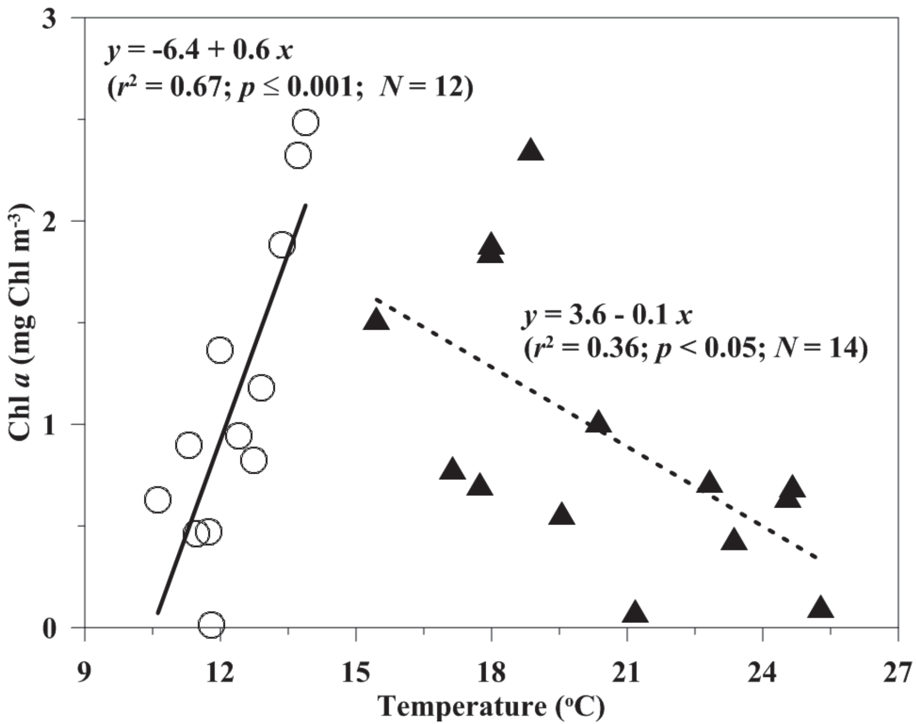

Fig. 5. Relationship between Chl *a* and temperature in the surface waters in the spring of 2010. Linear regressions between Chl *a* vs. temperature < 15 °C (∘; solid line) or temperature ≧ 15 °C (▲; dashed line) with r^2 and p are also shown.

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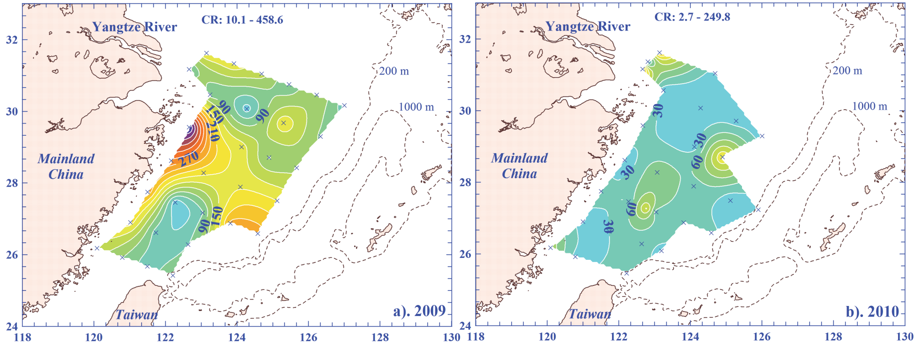

Fig. 6. Contour plots of planktonic community respiration (CR) in the surface waters of the ECS in **(a)** 2009 and **(b)** 2010, with a contour interval of 30 $\mathrm{mg\,C\,m^{-3}\,d^{-1}}$.

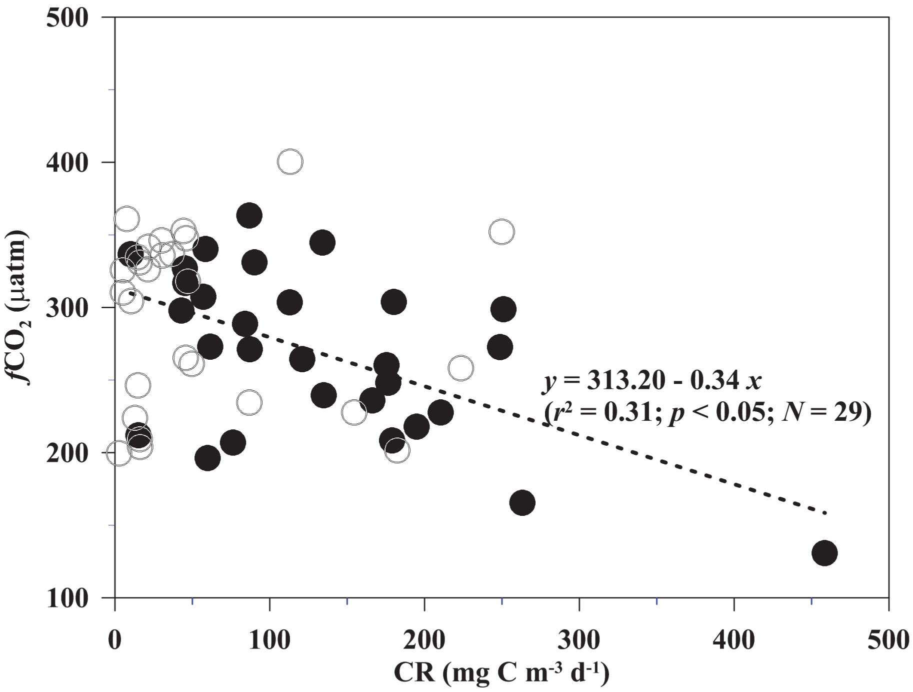

Fig. 7. Relationship between fugacity of CO_2 (fCO_2) and planktonic community respiration (CR) in the surface waters in the spring of 2009 (•) and 2010 (◦). Linear regression (dashed line) with r^2 and p values for 2009 is also shown.