Quick transport of primary produced organic carbon to the ocean interior

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1. Introduction

To better understand the oceans’ ability to uptake atmospheric CO₂, the downward transport of carbon via the “biological pump” mechanism needs to be quantified [e.g., Volk and Hoffert, 1985]. Time-series data collected using sediment trap has been effectively used for this purpose [e.g., Honjo et al., 1999]. However, sediment traps have usually been deployed in the deep sea (>1000 m). Therefore, concern has been expressed about how well deep sediment trap data represents the biological activity in the upper layer. Interestingly, progress in satellite oceanography has enabled us to learn about spatial and temporal variation of biological activity [e.g., Banse and English, 1999]. Unfortunately, ocean color data provides only near surface information and the existence of clouds often hampers acquisition of satellite color data, especially for the coastal and high latitude region. Thus, we chose to simultaneously observe and quantify biological activity in the euphotic layer using in situ optical sensors and particle fluxes just below the winter mixed layer with a sediment trap. These concurrent observations have enabled the establishment of linkages and correlations between upper and deeper ocean biological and biogeochemical processes.

2. Experimental Procedure and Methods

A bottom tethered mooring system was deployed at station K2 (47°N, 160°E, water depth 5280 m) in the Western Subarctic Gyre (WSG). The WSG has large seasonal variability in physical, chemical and biological parameters [Harrison et al., 2004]. Seasonal surface water temperature and mixed layer range from ca. 1°C and ca. 100 m in winter to >10°C and ca. 30 m in summer, respectively. In late spring, primary productivity increases with temperature, light intensity and water stratification. Seasonal drawdown of nutrients from late winter to early autumn is larger than that in other oceans. The euphotic layer is relatively constant (ca. 50 m) and diatom numerically predominates the phytoplankton assemblage year round. A time-series sediment trap (McLane Mark7G) with 21 collecting cups was installed at 150 m that is ca. 50 m below the late winter mixed layer at station K2 (M. C. Honda, unpublished data, 2005). Sampling of settling particles began on 20 March 2005 and ceased on 11 September 2005. The particle collecting interval was 14 days for the first 4 cups and 7 days for the remaining 17 collecting cups. Collected samples were preserved with seawater based buffered 5% formalin. After pretreatment of samples, the concentrations of organic carbon were measured with an elemental analyzer and concentrations of Si, Ca and Al were measured with ICP-AES, and converted to organic materials,opal, CaCO₃ and lithogenic materials, respectively following Honda et al. [2002]. In order to estimate trapping efficiency, ²³⁰Th and ³²²Th was measured using the α counting method [Anderson and Fleer, 1982]. The Bio-optical Long-term Optical Ocean Measuring System (BLOOMS [Dickey et al., 2003]) which consists of a Satlantic Inc. spectral radiometer [OCR-504-ICWS] with data acquisition/storage systems was installed for measurement of downwelling spectral irradiance (Ed) at 4 wavelengths (412, 443, 490, and 555 nm). The depth of BLOOMS measured with a depth sensor (RIGO DP1158) was stable during the experimental period (37.5 ± 2.0 m on average). The optical system was kept free of biofouling by copper shutters [Manov et al., 2004]. The values of Ed at 4 wavelengths were measured every hour during the local daytime period (19:00–7:00 UTC).
increase of light attenuation or increase of turbidity in the water column above BLOOMS.

4. Discussion

4.1. Comparison of Optical Signals and Organic Carbon Fluxes

[6] It has been verified that Ed at ca. 440 nm is preferentially absorbed by pigments (chlorophyll) of principal phytoplankton as opposed to Ed at ca. 550 nm [Kirk, 1994]. Thus the ratio of Ed at 555 nm to that at 443 nm (Ed_{555}/Ed_{443}) serves as a good index of Chl-a [e.g., Loisel and Morel, 1998]. Using BLOOMS data, seasonal variability in the Ed_{555}/Ed_{443} as proxies for Chl-a was evident (Figure 2). The Ed_{555}/Ed_{443} began to increase in late-April and the maximum values were observed in late-June to early-July. This seasonal pattern correlated well with that of the surface PAR/in situ QI ratio (R = 0.94) and, moreover, that of the SeaWiFS-derived surface Chl-a in the vicinity of station K2 (R = 0.78, P < 0.001). This suggests that the increase of turbidity between the late-April and mid-August with maximum in July is likely attributable to increases of phytoplankton. As shown in Figure 2, seasonal variability in organic carbon flux (OCF) follows the same pattern as optical variability: OCF began to increase in late-April and increased more largely between the late-June and August. The correlation coefficient between OCF and the Ed_{555}/Ed_{443} was found to be statistically significant (R = 0.62, P < 0.005). Based on good synchronization between seasonal patterns of OCF and optical signals, it is concluded that primary produced organic carbon in the euphotic layer was likely quickly transported to the ocean interior with time lag <1 week.

4.2. Estimation of Primary Productivity and Export Ratio

[7] Based on previous observations in the northern North Pacific (Figure 3), integrated Chl- a abundance (Chl-a_{int}) upper 40 m (mg m^{-2}) can be expressed as logarithmic function of the Ed_{555}/Ed_{443} at 40 m:

\[
\text{Chl-a}_{\text{int}} = 29.89 + 33.375 \times \log(\text{Ed}_{555}/\text{Ed}_{443}) \text{ratio}
\]

Using the following empirical equation, depth-integrated production in the upper 40 m (PP) can then be determined using equation 1 and daily surface PAR.

\[
PP = \text{Chl-a}_{\text{int}} \times \text{surface PAR} \times \Psi
\]

Figure 2. The Ed_{555}/Ed_{443} (open circles), surface PAR/ in situ QI ratios (closed circles), and Organic Carbon Flux (OCF) at 150 m (bar graphs). Squares are SeaWiFS derived surface Chl-a.
where $\Psi$ is water column light utilization index [e.g., Platt et al., 1988]. In this study, a constant value of $\Psi$ (0.3 ± 0.1) reported by Imai et al. [2002] for the WSG was applied.

[8] Seasonal variability in estimated PP also shows similar pattern of the $Ed_{555}/Ed_{443}$ and Chl-$a_{(int)}$ resulting good synchronization of OCF (Figure 4). Minimum and maximum of PP were estimated to be ca. 50 mgC m$^{-2}$ day$^{-1}$ in late-March and ca. 1200 mgC m$^{-2}$ day$^{-1}$ in July, respectively. Mean PP was estimated to be 297 ± 99 mgC m$^{-2}$ day$^{-1}$ [Imai et al., 2002]. Figure 4 also shows PP estimated with satellite data and algorithm proposed by Kameda and Ishizaka [2005] (PP$_{(KJ)}$) based on VGPM algorithm [Behrenfeld and Falkowski, 1997]. Though PP estimated in this study tends to be lower than PP$_{(KJ)}$ after late-July and other factor such as temperature might be considered in equation 2, both estimates generally coincided well. Thus, the mean export ratio (ratio of mean OCF at 150 m, 13 mgC m$^{-2}$ day$^{-1}$, to PP) was estimated to be 4 ± 1%. This export ratio was considerably smaller than the annual average of export ratio (ca. 45%) estimated using the seasonal amplitude of nutrients (new production) and the observed PP for the WSG [Honda, 2003]. However if trapping efficiency of our sediment trap was ca. 20% as described before, actual OCF at 150 m should be 5 times larger than the observed OCF, resulting in an export ratio of 22 ± 7%. Further, the export ratio of 45% by Honda [2003] was an estimate at 100 m. It has been reported that OCF decreases with depth [Martin et al., 1987]. If decrease of OCF with depth can be formulated as a power function as proposed by them, OCF at 100 m becomes 1.4 times larger than OCF at 150 m. If this is the case, the OCF at 100 m is estimated to be ca. 90 mgC m$^{-2}$ day$^{-1}$ and becomes comparable to seasonal new production for the WSG (ca. 90 mgC m$^{-2}$ day$^{-1}$ [Midorikawa et al., 2002]). Consequently, the export ratio is estimated to be ca. 31 ± 10%. This export ratio is significantly higher than that reported for other oceans (< 10% [Buesseler, 1998]). Although many uncertainties remain, high export flux and ratio values in the WSG are supported by previous reports: export fluxes and ratios are high in the productive and diatom-dominant areas [Buesseler, 1998].

5. Concluding Remarks

[9] It was verified that measurement of optical fields in the water column was useful for the estimation of PP as a substitute for satellite data, and there is good evidence of quick transport of organic carbon assimilated in the euphotic layer to the ocean interior (at least to 150 m). For improved quantification, more data need to be accumulated; in particular, at least for one year’s worth of data are needed especially for estimation of trapping efficiency. In addition, more precise values of $\Psi$ are needed.

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