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Primary Production in the Eastern East China Sea:
Modeling and Estimation of its Enhancement by
Tropical Cyclone

December 2005

Graduate School of Science and Technology
Nagasaki University

EKO SISWANTO
# Contents

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Abstract

The East China Sea (ECS) is one of the most productive waters in the world ocean. This study aims at modeling primary production in the ECS and its application to assess primary production enhancement induced by tropical cyclone (typhoon). The first part of this study (chapter 2) is constructing chlorophyll-a (Chl-a) vertical profile model and time integrated primary production model (TIM) to assess deep Chl-a maximum (DCM) contribution to integrated primary production (IPP) in the Kuroshio frontal region. The second part (chapter 3) is aimed at validating TIM with more IPP datasets collected in the eastern ECS, along with to compare with two previously established production models, empirical model (EM), and vertically generalized production model (VGPM). The last part (chapter 4) of this study is assessing IPP enhancement induced by typhoon.

In the Kuroshio frontal region, the types of Chl-a vertical profiles were associated with water mass features. Shelf water was commonly attributed with surface Chl-a maximum (SCM), while Kuroshio region was characterized by DCM. Frontal region however, showed high variability in depth of DCM. Carbon fixation rate at the DCM layer of the frontal region was higher than that in the Kuroshio region, as they potentially contributed within 30.9±9.1% and 20.9±5.4% to IPP in the frontal and Kuroshio regions, respectively, suggesting that phytoplankton biomass in the DCM layer of the frontal region is more important than that in the Kuroshio region. This was reflected by the higher IPP in the frontal region than that in the Kuroshio region. Therefore, these results imply that modeling IPP in the Kuroshio frontal region should consider depth dependent variation in Chl-a vertical profile.
From model validation, TIM and VGPM with their original optimum Chl-\(\alpha\) normalized carbon fixation rate (\(P_{\text{opt}}^B\)) models could not account for variation in measured IPP, while EM with its original \(P_{\text{opt}}^B\) models gave lower estimates than measured IPP. The discrepancies between their original \(P_{\text{opt}}^B\) models and \(P_{\text{opt}}^B\) expression from eastern ECS were responsible for these mismatches. Employing measured \(P_{\text{opt}}^B\), the three models could explain well IPP variance, but EM still showed lower estimates. The EM’s lower estimation seemed to be caused by employing constant factor (rather than as a light function) describing the depth of light-saturated production. This constant factor seems to be more appropriate for the region with SCM and surface production maximum (SPM). Whereas, this study region was more attributed to the DCM and deep production maximum (DPM). TIM and VGPM however agreed well with measured IPP, probably due to the facts that both models employ light as variable enables them to accommodate the depth of light-saturated production, and thus enables them to accommodate DPM.

Employing modeled \(P_{\text{opt}}^B\), TIM’s performance was reduced remarkably compare to VGPM’s performance. The remarkable reduce of TIM’s performance was caused by bias accumulation since more model inputs have to be estimated in TIM (\(P_{\text{opt}}^B\), Chl-\(\alpha\) vertical profile etc.) than in VGPM (\(P_{\text{opt}}^B\) only). Therefore, VGPM with adjusted \(P_{\text{opt}}^B\) model was considered as an optimal production model and selected to estimate IPP enhancement induced by typhoon using satellite data.

The study of typhoon effect on phytoplankton Chl-\(\alpha\) and IPP variations were divided into two parts, daily and interannual variations. The daily variation analysis was focused on typhoon Meari event that passed over the ECS from 25 to 28 September 2004 and the interannual variation analysis was focused on 11 typhoon passages during
the period of 7-year Sea-viewing Wide Field-of-view Sensor (SeaWiFS) mission (1998-2004). During two-day stay, Meari dropped SST by as much as 7°C, enhanced Chl-a and IPP by as much as 15 and 3 folds, respectively. We discerned that the high enhancements of Chl-a and IPP induced by Meari was caused by some factors as follows; 1) Meari might be a warm (El Nino) episode-affected intensive typhoon, 2) slow movement allowing intensive upwelling, 3) intensification of Meari as it traveled over the Kuroshio warm waters which gave more heat energy. Considering only the bloom area, Meari enhanced IPP by as much as 10 GgC during a single day, which is equal to the new production (INP) available to be exported from euphotic zone. The interannual variation analysis found that the enhancements of Chl-a and IPP positively correlated with wind speed, but negatively with typhoon speed and pressure. Great (less) enhancements were forced by typhoons born during warm years (cold years) or El Nino (La Nina) event, probably related to the facts that typhoons with high (low) intensity tend to form during warm years (cold years). Besides typhoon parameters, nitracline depth showed an important role in determining IPP enhancement in response to typhoon passages.

Employing multiple regression of IPP enhancement as a function of typhoon parameters and nitracline depth, we estimate IPP enhancement induced by all typhoons occurred during 1998~2004 period to see the interannual variation of typhoon contribution to annual and summer IPP, as well as INP. The least contribution revealed in 1998 due to only three typhoon passed over the investigated region, each with low intensity which is associated with La Nina episode. The greatest contribution observed in 2004 due to nine typhoons occurred, some of which are very intensive related to El
Nino episode. Depend on the typhoon intensity, the contribution of typhoon to summer IPP and INP were within the ranges of 1.0–8.6% and 4.0–35%, respectively.

The general conclusions which can be redrawn from this study are as follows; 1) efforts to model primary production in the ECS should consider Chl-$a$ vertical variation because DCM contributes significant amount to IPP, 2) for the reasons of simplicity however, VGPM with locally adjusted $P_{opt}$ model was likely an optimal model for the eastern ECS, 3) typhoon-induced IPP enhancements were related to typhoon parameters, as well as the nitracline depth, and showed a connection with El Nino/La Nina episodes. This study is thus the first study showing the probable remote forcing by global climate changes in driving biological production dynamics in the ECS through typhoon passages.
Abbreviations

ECS  East China Sea
SCS  South China Sea
Chl-$\alpha$ Chlorophyll $\alpha$ concentration (mg m$^{-3}$)
Chl$_0$ Sea surface Chl-$\alpha$ (mg m$^{-3}$)
Chl$_z$ Chl-$\alpha$ at depth $z$ (mg m$^{-3}$)
$E_0$ Surface irradiance (Einstein m$^{-2}$ d$^{-1}$)
$E_z$ Irradiance at depth $z$ (Einstein m$^{-2}$ d$^{-1}$)
SST Sea surface temperature ($^\circ$C)
$K_d$ Attenuation coefficient of downward irradiance ($m^{-1}$)
$z$ Depth (m)
$z_{eu}$ Euphotic depth (m)
INP Euphotic depth-integrated new production (mgC m$^{-2}$ d$^{-1}$)
IPP Euphotic depth-integrated primary production (mgC m$^{-2}$ d$^{-1}$)
IPP$_{uc}$ IPP with uniform Chl-$\alpha$ profile assumption (mgC m$^{-2}$ d$^{-1}$)
IPP$_{nc}$ IPP with depth dependent Chl-$\alpha$ profile (mgC m$^{-2}$ d$^{-1}$)
IPP$_{ms}$ Measured IPP (mgC m$^{-2}$ d$^{-1}$)
IPP$_{en}$ IPP enhancement by typhoon
$P_z$ Carbon fixation rate at depth $z$ (mgC m$^{-3}$ d$^{-1}$)
$P_{opt}^B$ the optimum Chl-$\alpha$ normalized primary production within water column (mgC mgChl-$\alpha^{-1}$ d$^{-1}$)
$P_z^B$ Chl-$\alpha$ normalized primary production at depth $z$ (mgC mgChl-$\alpha^{-1}$ d$^{-1}$)
$P_{opt}^B$ Chl-$\alpha$ normalized primary production (mgC mgChl-$\alpha^{-1}$ d$^{-1}$)
$E_{max}$ Irradiance at the inflection point between light-saturated and light-limited specific production (Einstein m$^{-2}$ d$^{-1}$)
$D_{nr}$ Daylength (h)
$C_0$ Shifted background of Chl-$\alpha$ profile (mg m$^{-3}$)
$C_p$ Peak concentration of Chl-$\alpha$ profile above $C_0$ in Chl-$\alpha$ profile with 4 parameters (mg m$^{-3}$)
$z_{m}$ Depth of Chl-$\alpha$ maximum (m)
$h$ Total biomass within Chl-$\alpha$ maximum layer or above $B_0$ (mg m$^{-2}$)
$\sigma$ Standard deviation controlling the thickness of Chl-$\alpha$ maximum layer (m)
$Chl_{max}$ Maximum Chl-$\alpha$ concentration (mg m$^{-3}$)
$C_p$ Peak of Chl-$\alpha$ profile above $C_0$ or $Chl_{max} - C_0$ (mg m$^{-3}$)
DCM Deep Chl-$\alpha$ maximum
SCM Surface Chl-$\alpha$ maximum
DPM Deep primary production maximum
SPM Surface production maximum
RMSE Root means squared error
Exp. Experiment
SOI Southern Oscillation Index
SeaWiFS Sea-viewing Wide Field-of-view Sensor
MODIS Moderate Resolution Imaging Spectroradiometer
TRMM/TMI Tropical Rainfall Measuring Mission/Microwave Imager
AVHRR Advanced Very High Resolution Radiometer
LAC Local area coverage
SMI Standard mapped image
<table>
<thead>
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<th>Abbreviation</th>
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<tr>
<td>TIM</td>
<td>Depth dependent, time Integrated Model</td>
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<tr>
<td>EM</td>
<td>Empirical Model</td>
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<tr>
<td>VGPM</td>
<td>Vertically Generalized Production Model</td>
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<tr>
<td>WS</td>
<td>Typhoon wind speed (knot)</td>
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<td>TP</td>
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<td>TS</td>
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Chapter 1

Introduction

1.1. Ocean Primary Production

Primary production is, in principle, the rate of photosynthetic fixation of carbon by chlorophyll-containing organisms utilizing energy from sunlight (Berger et al., 1987). In the ocean, phytoplankton is the main primary producer responsible for this process of sunlight-driven conversion of inorganic carbon into organic carbon via oxygenic photosynthesis and is the base of food web chain.

When the product of photosynthetic organic carbon is subtracted by the metabolic costs of all other metabolic processes by phytoplankton themselves, the remaining organic carbon becomes available to heterotrophs. This remaining organic carbon is called net primary production (NPP) (Lindeman, 1942). From biogeochemical and ecological perspectives, NPP is important because it represents the total flux of organic carbon (energy) that an ecosystem can utilize for all other metabolic processes.

It should be noted that NPP and photosynthesis are not synonymous. The former requires the inclusion of the respiratory term for the photoautotrophs (Williams, 1993). In reality, that term is extremely difficult to measure in natural water samples. Hence, NPP is generally approximated from measurements of photosynthetic rates integrated over some appropriate length of time (usually a day), and respiratory costs are either assumed or neglected.
1.2. Primary Production Measurement

The first quantitative measurements of phytoplankton productivity were made in the early 20th century. The researchers realized that oxygen could be used as a proxy for the synthesis of new organic material. However, while the oxygen method worked well in coastal waters, it gave ambiguous results in oligotrophic ocean waters because of its less sensitivity.

In 1952, Steemann Nielsen introduced the use of the radiotracer, $^{14}$C, to quantify the fixation of carbon by natural plankton assemblages in short-term (hours) incubations (Steemann Nielsen, 1952). The radiocarbon method was extremely more sensitive than the oxygen method. Within five years, the radiocarbon method had completely replaced the oxygen method for measuring oceanic primary productivity.

The other method is by using stable isotope of $^{13}$C which is firstly introduced by Slawyk (1979). This method had only been rarely applied because of its lower sensitivity than $^{14}$C method (Hama et al., 1993). Therefore to reach high accuracy when using $^{13}$C, more water volume is required. However, despite of $^{14}$C and $^{13}$C method limitation, these two methods agreed well as reported by Hama et al. (1997).

1.3. Satellite-based Primary Production Modeling

Primary production is a key component of the global carbon cycle. However, its spatial and temporal variability is poorly constrained observationally. It is impossible to quantify regional and basin scale variability from in situ measurements. Consequently, an extrapolation based on ship-based measurements is required to obtain primary production estimates for a large spatial scale.
Fortunately, satellite provides a solution (McClain et al., 1998). Ocean color sensors are presently used to estimate chlorophyll-\(a\) concentration (Chl-\(a\)) in the upper ocean. To convert Chl-\(a\) (a pool) to primary production (a rate) numerous models incorporating a time dependent variable (irradiance) have been established (e.g. Platt and Sathyendranath, 1993; Longhurst et al., 1995; Howard and Yoder, 1997; Antoine et al., 1996; Behrenfeld and Falkowski, 1997a). These models can be distinguished by the degree of depth and irradiance resolutions (see Behrenfeld and Falkowski (1997b)).

The complex models resolving production variation both vertically and spectrally, usually incorporate maximum Chl-\(a\) normalized carbon fixation rate \(P_{\text{max}}^B\) which describe a unique physiological parameter of phytoplankton assemblage. \(P_{\text{max}}^B\) is usually derived from the short-term incubation period (~2 hours) under constant irradiance. The complex models resolve vertical variation of \(P_{\text{max}}^B\) which is usually estimated from temperature. The simpler models, as will be investigated in the chapter 3, which do not resolve vertical and spectral variations usually employ an optimum Chl-\(a\) normalized carbon fixation rate within water column \(P_{\text{opt}}^B\). Unlike \(P_{\text{max}}^B\), \(P_{\text{opt}}^B\) is derived from the long-term incubation period (6–24 hours) under ambient light.

1.4. Modeling Primary Production in the East China Sea

Toward better estimations of the role of the ocean on global carbon cycle, more concerns on primary production estimation must be emphasized in the highly dynamic continental shelves (Chen et al., 2003). It is because even the world’s continental shelves represents less than 20% of the world’s oceanic areas, the high nutrient input such as river run-off and upwelling has made its primary production could be as important as that in the ocean interior (Walsh, 1991).
The East China Sea (ECS) with 70% of which is occupied by the continental shelf shallower than 200 m, is one of the most productive waters in the world’s oceans, maintained by two main nutrient sources, Yangtze River (Changjiang) discharge and Kuroshio subsurface upwelling. These two nutrient sources, together with hydrographic features that are strongly influenced by monsoon wind systems, modulate primary production in the ECS both spatially and temporally.

For the regional scale of the ECS, Gong and Liu (2003) constructed an empirical model (EM) from the most complete datasets from the western ECS covering coastal and continental shelf areas. EM can be classified as simple model because it does not resolve vertical and spectral variations. However, since primary production and its variable (e.g. Chl-α vertical profile) variation in the ECS varies spatially, model constructed from a certain part may not be applicable for the other part of the ECS.

As the main variable in estimating primary production using satellite production model, Chl-α in the ECS varies remarkably both spatially and temporally (Fei et al., 1992). Deep Chl-α maximum (DCM) is a common features of Chl-α vertical profile (hereafter Chl-α profile), often contributes significantly to primary production, especially when they are not very deep allowing light and nutrient to stimulate intensive phytoplankton growth. The DCM formation is often responsible for underestimation problem encountered by production model (e.g. Banse and Yong, 1990; Mouw and Yoder, 2005).

1.5. The Objectives of This Study

As an attempt to estimate primary production in the ECS, in chapter 2 we develop a depth dependent, time integrated production model (TIM), an analytical
model considering Chl-a profile. The main purpose of this work is to assess the DCM contribution to primary production as well as to clarify whether considering Chl-a profile in primary production modeling in the ECS is important.

For the purpose of providing optimal production model for larger spatial scale, in chapter 3 we further compare TIM with EM developed from the western ECS and globally developed vertically generalized production model (VGPM) proposed by Behrenfeld and Falkowski (1997a,b). This chapter validates these three models with primary production datasets collected from the eastern ECS. This model comparison aims essentially at selecting the optimal primary production model for the eastern ECS and being applied to assess primary production enhancement induced by tropical cyclone (typhoon) in chapter 4.

Typhoon is one of the atmospheric phenomena allowing nutrients to be pumped up to the euphotic zone enhancing primary production, and thus new production. However, typhoon’s severe weather condition and great variation in its trajectory hamper the ship-borne observation causing a severe paucity of information on phytoplankton dynamics in response to typhoon passages in the ECS. Thus, typhoon-enhanced new production has also been long-neglected from the annual new production estimation. Chapter 4 therefore presents the application of optimal production model resulted from the chapter 3 combining with satellite data to investigate phytoplankton dynamics in response to typhoon passages in the ECS in terms of both daily and interannual variations.
Chapter 2

Estimating Chlorophyll-α Vertical Profiles from Satellite Data and the Implication for Primary Production in the Kuroshio Front of the East China Sea

2.1. Introduction

The high biological productivity in the ECS is maintained mainly by two main nutrient sources, Changjiang discharge (horizontal flux) and Kuroshio upwelling (vertical flux). Phytoplankton Chl-α as a biological indicator of water productivity is sensitive to these nutrient fluxes, thus their distribution will vary both horizontally and vertically.

Vertically, DCM is a common feature observed in Chl-α profiles. However, the factors forming and maintaining the DCM can vary remarkably (Cullen, 1982). Numerous investigations have been conducted on Chl-α and primary production variations in the ECS (e.g., Fei et al., 1992; Hama et al., 1997; Guo, 1991; Gong et al., 1999; Gong et al., 2003). None of these previous studies, however, specifically addressed the dynamic of Chl-α profiles and their role in driving primary production in the ECS.

The Chl-α profile is essential, not only for estimating total phytoplankton biomass, but also as one of the principal inputs to analytical models of primary production using satellite remote sensing (Sathyendranath and Platt, 1993). As the ocean color sensor provides Chl-α data within a depth of one attenuation length or one-fifth of euphotic depth, transforming the Chl-α map into the map of primary production often assumes that Chl-α uniformly distributes throughout the water column (Platt, 1986; Ishizaka, 1998; Andre, 1992). However, as the DCM formation is a common
feature of biomass vertical profiles, ignoring it can lead to the underestimation problem in primary production estimation (Sathyendranath and Platt, 1993; Platt et al., 1991; Hidalgo-González and Alvarez-Borrego, 2000).

Furthermore, considering that the DCM in the ECS is not very deep, which enables light to stimulate intensive production in the deep layer (Fei et al., 1992), this study hypothesizes that the DCM will contribute significantly to the primary production within the euphotic zone. The Kuroshio upwelling that often occurs in the continental shelf edge region and bring nutrients from the deep layer to the euphotic zone may also be an important mechanism in driving photosynthesis at the DCM layer in this region.

This chapter will thus describe the study with the objectives are as follows;
1. constructing Chl-a profile and primary production models to estimate primary production specifically for the Kuroshio frontal region,
2. assessing the degree of which the DCM contribute to primary production in the Kuroshio frontal region,
3. clarifying whether considering Chl-a profile is important in estimating primary production in the ECS.

2.2. Materials and Methods

2.2.1. Study Area

Cruises were carried out to conduct primary production experiments and to collect discrete depth of Chl-a, photosynthetically available radiation (hereafter irradiance) vertical profiles, as well as other hydrographic data in the region mainly covering the Kuroshio frontal region (Fig.2-1). These data were used to construct Chl-a profile and primary production models specific for the Kuroshio frontal region.
Discrete depth of Chl-a measurements and profiling underwater irradiance were conducted during spring and autumn cruises. Primary production experiments, however, were only conducted during autumn. Therefore, the primary production model derived from these datasets is only representative for autumn conditions. The detail information on the cruises and data used to develop models are listed in Table 2-1.

2.2.2. Shipboard Measurements

Discrete measurement of Chl-a

For analyzing Chl-a profiles, discrete measurements of Chl-a samples were conducted at standard depths of 0, 10, 20, 30, 40, 50, 75, 100, 125, 150, 200 m using twelve 5-l Niskin bottles mounted on a Rosette sampler (General Oceanic, USA) with a CTD probe (Seabird SBE 9/11). Water samples were directly transferred from Niskin bottles into 100 ml polyethylene bottles and immediately filtered through a 25 mm glass fiber filter (Whatman GF/F). The Chl-a was measured with a Turner Design Model 10-AU fluorometer after extraction with N,N-dimethylformamide for a day (Suzuki and Ishimaru, 1990).

Primary production experiments

The rates of inorganic carbon fixation were determined by means of simulated in situ incubations with the stable isotope $^{13}$C. During T/V Nagasaki Maru cruise, water samples were collected using twelve 5-l Niskin bottles mounted on a Rosette sampler (General Oceanic, USA) from six depths corresponding to 100%, 50%, 25%, 10%, 5% and 1% of surface irradiance ($E_0$). Each depth was estimated from sea surface Chl-a ($Chl_0$) which was determined just before collecting the water samples.
before dawn). Daily ambient irradiance was recorded using a quantum scalar irradiance meter model QSL-2100 (Biospherical Inc.) mounted on an on-deck incubator.

Water samples were drained from Niskin bottles into 1000 ml polycarbonate bottles. We used four bottles for each depth, and after the addition of NaH\(^{13}\)CO\(_3\), three samples were incubated in an on-deck incubator filled with circulated sea surface water under the corresponding light intensity for 24 hours (from dawn) and one sample was immediately filtered through a 25 mm pre-combusted (450°C, 4 h) glass fiber filter (Whatman GF/F). All filter papers were then stored at −20°C until analysis.

The concentrations of particulate organic carbon and the isotopic ratio of \(^{13}\)C and \(^{12}\)C (\(^{13}\)C atomic %) of the samples were determined by a quadrupole mass spectrometer equipped with a combustion furnace (ANELBA TE-150), and the carbon fixation rate was calculated according to the method of Hama et al. (1993).

During R/V Yoko Maru cruises, water samples were collected using 10-l Go-Flo bottles mounted on a Rosette sampler (General Oceanic, USA) from seven depths corresponding to 100%, 50%, 25%, 10%, 2.5%, 1% and 0.5% of E\(_0\). Water samples were drained from Go-Flo bottles into 700 ml polycarbonate bottles and all incubations were conducted for 24 hours. Daily photosynthetic available irradiance was measured by an irradiance cosine collector (downward irradiance meter).

In this study, downward irradiance was used rather than scalar irradiance to ensure consistency with underwater irradiance measurement. To convert scalar irradiance measured during the T/V Nagasaki Maru cruise into downward irradiance, scalar irradiance values were simply multiplied by an average cosine value. Experimentally, the average cosine ranges between 0.7 – 0.9, increasing with solar
elevation and also depending upon the water type and wavelength. For practical purposes, the average cosine of 0.8 was used as the conversion factor (Hojerslev, 1975).

**Profiling underwater irradiance**

Underwater irradiance, used for constructing an empirical model of the underwater irradiance transmission parameter, was profiled with a Profiling Reflectance Radiometer, PRR-800 (Biospherical Inc.) during T/V Nagasaki and Kakuyo Maru cruises (Table 2-1). This irradiance vertical profiler measures downward irradiances at 13 wavelengths and photosynthetic available radiance (hereafter, irradiance). Euphotic depth was then defined as the depth corresponding to 1% of $E_0$.

### 2.2.3. Satellite Data

In order to understand variation in primary production in the Kuroshio frontal region synoptically, satellite data of sea surface temperature (SST), Chl$_0$ and $E_0$ were used. This study used the standard mapped image (SMI) of Sea viewing Wide Field-of-view Sensor (SeaWiFS) products Level 3 for Chl$_0$ and $E_0$ with 9 x 9 km$^2$ resolution, while for SST, descending passes of Advanced Very High Resolution Radiometer (AVHRR)-sensed SST version 4.1 with 9 x 9 km$^2$ resolution provided by Physical Oceanography-Distributed Active Archive Center (PO.DAAC) were used.

### 2.2.4. Primary Production Model

The euphotic depth-integrated primary production (IPP) was calculated using the depth-resolved, time-integrated production model as:

$$IPP = \int_{z=0}^{z=\alpha} \text{Chl}_z \times \left[ p_{\text{opt}} \times \frac{E_z}{E_{\text{max}}} \times \exp\left(1 - \frac{E_z}{E_{\text{max}}}\right)\right]$$  \hspace{1cm} (2-1)
where Chl_z, E_{max} and E_z are Chl-a at depth z (mg m^{-3}), irradiance value at the inflection point between light-limited and light-saturated phases (Einstein m^{-2} d^{-1}) and irradiance at depth z (Einstein m^{-2} d^{-1}), respectively. This model requires at least three inputs, Chl-a profile, photosynthetic parameter and underwater irradiance transmission parameter. The following descriptions explain the methods used to derive those three inputs.

Chl-a vertical profile parameters

A shifted, four-parameter Gaussian distribution was used to derive Chl-a profiles and to retrieve four profile parameters. The Gaussian distribution is sufficiently versatile to describe a large variety of profiles ranging from the coastal, upwelling, open oceans and Arctic waters, as long as the vertical profiles contain only a single peak (Platt et al., 1988; Lewis et al., 1983). The general shape of the shifted-Gaussian distribution is shown in Fig. 2-2 and has the equation:

\[
Chl_z = C_o + \frac{h}{\sigma \sqrt{2\pi}} \exp \left[ -\frac{(z-z_m)^2}{2\sigma^2} \right]
\]

(2-2)

where C_o is shifted background concentration (mg m^{-3}), z_m is Chl-a maximum depth (m), \sigma is standard deviation (m) that control the thickness of Chl-a maximum layer, and h is total biomass above the background concentration (mg m^{-2}).

In this study, cluster analysis was performed to see whether a type of Chl-a profile corresponds to a feature of water mass by means of JMP software (SAS Institute Inc.). The following description gives the method for cluster analysis. We extracted Gaussian-fitted Chl-a profile curves for each 10-m depth interval from the surface to 110 m depth, as the euphotic depth in this region is usually less than 110 m. We then treated these twelve extracted Chl-a values as variables in cluster analysis.
Before running the cluster analysis, we standardized each variable with the mean value and standard deviation of the total number of profiles to ensure comparability among profiles. The similarity index of two Chl-a profiles was calculated using the Euclidean distance, and hierarchically the similarity between established clusters was calculated with Ward’s clustering algorithm or Ward’s minimum variance method, as this clustering method is appropriate for a small number of dataset (less than a hundred observations). Obtaining the clusters of Chl-a profiles, Chl-a profile parameters for each identified cluster were related to SST and/or Chl0. The relationships were then applied to infer Chl-a profiles.

**Water column photosynthetic parameters**

To derive water column photosynthetic parameters of phytoplankton, we used Steele’s (1962) photosynthesis-irradiance model with only two photosynthetic parameters. This model can capture the three photosynthetic phases of light-limited photosynthesis, light-saturated photosynthesis and strong light photoinhibition. Photosynthetic parameters derived from this method do not describe unique physiological parameters from the same phytoplankton assemblages, but rather describe photosynthetic parameters within the euphotic zone. The original physiological parameters of $P_{\text{max}}^B$ and $E_k$ from photosynthesis-irradiance model are thus simply rewritten as $P_{\text{opt}}^B$ and $E_{\text{max}}$, which now describe photosynthetic parameters within the water column, as described by the following equation;

$$P^B = P_{\text{opt}}^B \times \frac{E}{E_{\text{max}}} \times \exp \left( 1 - \frac{E}{E_{\text{max}}} \right)$$  \hspace{1cm} (2-3)

where $P^B$ is Chl-a-normalized carbon fixation rate (mgC mgChl-a$^{-1}$ d$^{-1}$) and $E$ is irradiance (Einstein m$^{-2}$ d$^{-1}$).
It is well known that photosynthetic parameters such as $P_{\text{B, opt}}$ and $P_{\text{B, max}}$ are temperature dependent processes (e.g., Eppley, 1972; Behrenfeld and Falkowski, 1997a). Recently, it was shown that $P_{\text{B, opt}}$ is also influenced by Chl-a (Kameda, 2003; Kameda and Ishizaka, 2005). Based on these two facts, the model relating $P_{\text{B, opt}}$ to both SST and Chl$_0$ was constructed. Since $E_{\text{max}}$ demonstrated a strong correlation with $E_0$ (Behrenfeld and Falkowski, 1997a), we examined the empirical relationship between $E_{\text{max}}$ and $E_0$.

Attenuation coefficient

For each underwater irradiance vertical profile, the attenuation coefficient ($K_d$, m$^{-1}$) was derived by fitting underwater irradiance transmission equation as:

$$E_z = E_0 e^{-K_d z}$$  \hspace{1cm} (2-4).

We constructed an empirical model for estimating $K_d$ from Chl$_0$ and used the model to estimate underwater irradiance for IPP calculation.

2.3. Results

We analyzed the spring and autumn Chl-a profiles separately, since Chl-a distribution both horizontally and vertically are severely affected by monsoonal seasons. Moreover, SST and Chl$_0$ as independent variables for the model vary seasonally. Of 53 spring Chl-a profiles and 33 autumn profiles, we finally selected 42 and 31 plausible profiles according to the constraints suggested by Sathyendranath et al. (1995).

2.3.1. Spring Chl-a Vertical Profiles

Cluster analysis classified Chl-a profiles into three main clusters, marked by squares, closed circles and triangles in Fig.2-3. The square-marked cluster occupied the
northern part of the research area with low SST and high Chl$_0$. As these profiles had near surface or surface Chl-$a$ maximum (SCM) with the concentration ranging from 0.40 to 0.82 mg m$^{-3}$, and $z_m$ ranging from 5 to 40 m, we named the square-marked cluster as the shelf region profiles (Fig. 2-4).

Profiles in the cluster marked by closed circles had the DCM with the concentration ranged from 0.38 to 0.62 mg m$^{-3}$, and with $z_m$ ranged from 60 to 85 m. As this profile cluster had deep $z_m$ and was distributed in the southern part of the research area with high SST and low Chl$_0$, we named this cluster the Kuroshio region profile.

It was not obvious, however, that the triangle-marked cluster seemed to occupy the region between the shelf and Kuroshio regions. Profiles in this cluster had large variability in both the DCM concentration (ranged from 0.43 to > 1.10 mg m$^{-3}$) and $z_m$ (ranged from 18 to 58 m). We then intuitively named this cluster the frontal region profiles.

In several areas we found that Chl-$a$ profiles with deeper $z_m$ often occupied a shallower region, in contrast, profiles with shallower $z_m$ occupied a deeper region as denoted in Fig. 2-4 at the southern part of the research area. Such phenomena occur very often in the Kuroshio frontal region, as this region is usually registered as a region with very complex hydro-dynamical processes. Mixing processes, such as frontal eddies, which can frequently occur at the frontal region, seemed to be the responsible factors.

By plotting SST against Chl$_0$ (Fig. 2-5), it can clearly be observed that there is an obvious boundary (SST of 22.30°C) separating the shelf region profiles from other two profile types. Although there was a tendency that the Kuroshio profiles distributed
in the higher SST region, no clear, marked boundary between the Kuroshio and the frontal region profiles was observed.

Considering the Kuroshio and the frontal region profiles were indistinguishable and in regard to profile estimation in which SST and Chl$_0$ act as variables, both the Kuroshio and the frontal region profiles were estimated by applying the same empirical model. Expressing profile parameters as a function of SST or Chl$_0$, resulted in a low determination coefficient ($r^2$), with the exception of the parameter $z_m$, which could be well predicted without region separation (Table 2-2). The use of both SST and Chl$_0$ as variables gave an impressive increased in $r^2$, especially in the shelf region. An increasing $r^2$ could also be observed for the parameter $\sigma$ in the Kuroshio and frontal regions. This then led us to use both SST and Chl$_0$ for estimating Chl-$a$ profile parameters, except for the parameter $z_m$ (solely from SST) (Table 2-3).

2.3.2. Autumn Chl-$a$ Vertical Profiles

Two Chl-$a$ profile clusters were identified from cluster analysis, marked by closed circles and triangles in Fig.2-6. They were closely associated with the water mass feature surrogated by SST, and weakly by Chl$_0$ (Fig.2-7). As the closed circle-marked cluster with $z_m$ ranged from 70 to 90 m distributed in the high SST region, whereas the triangle-marked cluster with $z_m$ ranged from 35 to 70 m distributed in the low SST region, we named the closed circle-marked and triangle-marked clusters as the Kuroshio and the frontal region profiles, respectively. SST of 26.85 °C could be used as the boundary separating those two Chl-$a$ profile types (Fig. 2-8).

The correlations between profile parameters and sea surface variables were low, even when expressed as a function of both SST and Chl$_0$, except for parameter $C_0$. 
Like spring Chl-α profiles, $z_m$ could be well predicted from SST without region separation (Table 2-4). We used the mean values whenever parameters had no correlation with SST or/and Chl$_0$ (Table 2-5).

2.3.3. Water Column Photosynthetic Parameter and Attenuation Coefficient of Underwater Irradiance

With $P_{B_{\text{opt}}}$ ranged from 48.2 to 108.1 mgC mgChl-α$^{-1}$ d$^{-1}$ (mean 73.8 mgC mgChl-α$^{-1}$ d$^{-1}$), the best prediction for $P_{B_{\text{opt}}}$ could be obtained with the use of both SST and Chl$_0$ as variables. The form of the empirical model for estimating $P_{B_{\text{opt}}}$ is:

$$P_{B_{\text{opt}}} = 760.76 - 25.77 (\text{SST}) - 70.17 (\text{Chl}_0)$$

(2-5), with $r^2$ was 0.57 ($n = 13$, $p < 0.05$).

Unlike $P_{B_{\text{opt}}}$, $E_{\text{max}}$ demonstrated higher $r^2$ (0.65, $n = 13$, $p < 0.05$) when expressed as a three order polynomial function of $E_0$ with the empirical model as:

$$E_{\text{max}} = -11.04 + 3.07 (E_0) - 0.13 (E_0)^2 + 0.002 (E_0)^3$$

(2-6).

The underwater irradiance vertical profile analyzed to derive $K_d$ was that from the surface to the euphotic depth (1% of $E_0$). Thus, the $K_d$ obtained represented average $K_d$ within the euphotic zone (neither depth, nor spectral-dependent value). We found a strong correlation between $K_d$ and Chl$_0$ with $r^2 = 0.87$ ($n = 22$, $p < 0.001$) and its empirical model can be described as:

$$K_d = 0.047 + 0.063 (\text{Chl}_0)$$

(2-7).

2.3.4. Contribution of DCM to Phytoplankton Biomass

The performance of the Chl-α profile model was examined by comparing it to vertically uniform Chl-α distribution in terms of the average Chl-α within the upper and lower layers (hereafter, biomass). The upper and lower layers were defined by the
optical depth from 0.0 to 2.3 and from 2.3 to 4.6, respectively. Average biomass derived from measured Chl-a profiles was used as reference.

The spring Chl-a profile model gave better estimates of the average biomass than the vertically uniform Chl-a assumption (Fig.2-9A). At the lower layer of the frontal and the Kuroshio regions, biomass calculated using vertically uniform Chl-a profile underestimated measured biomass, while conversely, at the lower layer of the shelf region, it overestimated measured biomass. Use of the vertically uniform Chl-a profile underestimated measured biomass within the euphotic zone by as much as 48.7% and 19.8% for the Kuroshio and the frontal regions, respectively.

The autumn Chl-a profile model also gave better estimates of the average biomass than that estimated by means of a vertically uniform Chl-a profile (Fig.2-9B). Assuming a vertically uniform Chl-a profile throughout the euphotic zone led to an underestimation of biomass by as much as 35.9% and 29.9% for the Kuroshio and the frontal regions, respectively.

2.3.5. Contribution of DCM to Integrated Primary Production

Having empirical models for photosynthetic parameters and K_d, we calculated the IPP by incorporating measured Chl-a profiles and vertically uniform Chl-a profiles. We assessed the degree of the DCM contribution to the IPP at 9 of 13 production stations by comparing measured Chl-a profile based-production (IPPnc) and vertically uniform Chl-a profile-based production (IPPuc) referred to measured IPP (IPPms). We excluded the remaining four stations, since there is no pronounced DCM in their profiles.
Most of the probabilities of IPPuc/IPPms ratio fell below 100% and the peak of normal distribution was about $82\% \pm 26\%$ (Fig. 2-10A), while IPPnc was in good agreement with IPPms, displaying the highest probability of IPPnc/IPPms ratios and the peak of normal distribution fell around $99\% \pm 22\%$ (Fig. 2-10B). The difference between IPPuc/IPPms and IPPnc/IPPms ratios indicates that the DCM appears to give 17% of the IPP.

Encouraged by the results mentioned above, we then turned to autumn Chl-α profile stations consisting of the Kuroshio and the frontal region profiles, to estimate the magnitudes of the DCM contribution to the IPP in the Kuroshio and the frontal regions. In the following analysis, with the use of autumn empirical models of Chl-α profile, photosynthetic parameters and $K_d$, we assessed the DCM contribution to the IPP at autumn Chl-α profile stations, as photosynthetic parameters were only derived from the autumn production dataset. By applying all the empirical models, IPPnc and IPPuc were again calculated and the potential DCM contribution was estimated from their differences. Generally, the IPP estimates in the Kuroshio region were lower (mean $350 \text{ mgC m}^{-2} \text{ d}^{-1}$) than those in the frontal region (mean $897 \text{ mgC m}^{-2} \text{ d}^{-1}$) (Fig. 2-11). A remarkable phenomenon that should be highlighted here is that IPPuc underestimated IPPnc by as much as 12.2% ($\pm 8.9\%$) and 27.4% ($\pm 7.9\%$) for the Kuroshio and the frontal regions, respectively.

2.3.6. Satellite Data-based Integrated Primary Production During Autumn

For the purpose of displaying estimates of IPPnc, IPPuc and the DCM contribution to the IPP at a larger scale, we used satellite data involving SST, Chl$_0$ and E$_0$. Nevertheless, we did not calculate for the whole region of the ECS because of the
very limited dataset from which all of required empirical models derived. The ranges of SST and Chl$_b$ from which three empirical models derived were intersected to obtain the narrowest range of SST and Chl$_b$. The ranges of SST and Chl$_b$ applied for satellite data are 26.0 – 27.5°C and 0.20 – 0.42 mg m$^{-3}$, respectively.

We used satellite data of October 1999 (Fig.2-12A-C), as photosynthetic parameters derived only from the autumn dataset, and the 9 km resolution of AVHRR-sensed SST (version 4.1) is available up to 1999. Estimation of IPPnc, IPPuc and the DCM contribution to the production in the Kuroshio and the frontal regions were limited to the region approximately denoted by the white polygon in Fig.2-12D-F. The Kuroshio and the frontal regions at the northernmost part of the polygon could be distinguished by an SST of 26.77°C, while at the southernmost part by an SST of 27.20°C (not shown). The IPP in terms of IPPnc exhibited lower values in the Kuroshio region (mean 541 mgC m$^{-2}$ d$^{-1}$) than those in the frontal region (mean 953 mgC m$^{-2}$ d$^{-1}$) (Fig.2-12D). The same typical pattern was also true in terms of the mean IPPuc in which the Kuroshio and the frontal region’s mean IPPuc were 428 mgC m$^{-2}$ d$^{-1}$ and 658 mgC m$^{-2}$ d$^{-1}$, respectively (Fig.2-12E). Based on the maps of IPPnc and IPPuc, we calculated the underestimation of the IPP due to uniform Chl-$a$ profiles in the Kuroshio and the frontal regions (Fig.2-12F). Note that by replacing the negative sign of the label with the positive sign, Fig.2-12F also meant the DCM contribution to the IPP. We then assessed the mean DCM contribution to the IPP based on satellite data in the Kuroshio and the frontal regions were as much as 20.9% (± 5.4%) and 30.9% (± 9.1%), respectively.
2.4. Discussion

2.4.1. Variability in Chl-a Vertical Profile Parameters

We have described the empirical models for inferring the vertical structure of Chl-a within the water column by estimating four profile parameters ($C_0$, $h$, $\sigma$, and $z_m$) that determine the Chl-a profile shape. These parameters have been derived from SST and/or Chl$_0$, which also were used as variables for distinguishing water mass features. With the use of SST as a surrogate of the water mass feature, we have identified that Chl-a profiles in the shelf region were attributed to the SCM, while those in the Kuroshio and the frontal regions were attributed to the DCM.

Regardless of the season and water mass features, $z_m$ was the parameter that demonstrated a clear, gradual deepening from the shelf to the Kuroshio region. Such a across-shelf successive change in $z_m$ occurred concurrently with increasing SST and decreasing Chl$_0$, enabled $z_m$ to be well predicted from SST and/or Chl$_0$ and accounted for > 70% of the variation in measured $z_m$. In most studies on the prediction of Chl-a profiles parameters, $z_m$ seemed to be the most predictable parameter, because of its gradual variability along the latitudinal, across-shelf and from the coast to offshore directions (Sathyendranath et al., 1995; Watt et al., 1999; Richardson et al., 2003).

In the shelf region, $C_0$ was poorly predicted ($r^2 = 0.26$) from the sea surface variables, as the profiles were attributed to the SCM. This was simply caused by the large gap in the distance between predictor in the surface (e.g. Chl$_0$) and the deepest and lowest Chl-a value defined as $C_0$ by the Gaussian distribution model. This deepest and lowest Chl-a value would be extrapolated by the Gaussian distribution model to high variation in $C_0$ values that are uncoupled from surface variables, thus resulting in a lack of correlations between $C_0$ and surface variables. Considering the low prediction for $C_0$,
it seems to suggest that $C_0$ is not important determinant for predicting Chl-$a$ profiles with SCM in the shelf region. Since the remaining three parameters of the shelf region profile could be well predicted with $r^2$ of 0.548, 0.638 and 0.740 for $h$, $\sigma$ and $z_m$, respectively, it seems that the most reasonable way to improve Chl-$a$ profile estimation in the shelf region is by using a 3-parameter Gaussian distribution model (excluding $C_0$), rather than using a 4-parameter Gaussian distribution model.

In the Kuroshio frontal region, even $z_m$ could be well predicted, but the thickness of Chl-$a$ maximum itself ($\sigma$) was unpredictable ($r^2 < 0.32$). The greatest variability in $\sigma$ reflected the more factors govern the Chl-$a$ maximum layer in the Kuroshio frontal region than in the shelf region. The thickness of SCM ($\sigma$) in the shelf region was more or less governed by nutrient and light level in the upper layer, but in the Kuroshio frontal region, $\sigma$ seems to be the result of a multitude of factors, including physical and chemical (e.g. frontal upwelling, nutrient input and water column stability), and biological and physiological factors (e.g. variations in sinking rate, light-shade adaptation, and also species assemblages) (Ediger and Yilmaz, 1996; Murty et al., 2000; Cullen, 1982). In fact, Chl-$a$ distribution in the Kuroshio front of the ECS seems to be controlled by numerous processes (Ishizaka et al., 2001).

Parameter $C_0$ could be well predicted from SST and/or Chl$_0$ in the Kuroshio frontal region both during spring and autumn, accounting for $> 60\%$ of variability in measured $C_0$, while some previous studies seemed to have difficulties in predicting $C_0$ (Richardson et al., 2003; Siegel et al., 2001; Hidalgo-González and Alvarez-Borrego, 2000). Since Chl-$a$ profiles were closely associated with water mass features and the parameter $C_0$ did not seem to be the important determinant in the shelf region, but is
well predicted in the Kuroshio region, we therefore suggest that effort in Chl-\(a\) profile estimation must also consider water mass features besides considering regional division.

### 2.4.2. Variability in Water Column Photosynthetic Parameters

With the range of SST from 22.9 °C to 27.5°C, \(P_{\text{Bopt}}\) was negatively correlated with SST (Eq.2-5). There is a fact that in this SST range, a positive correlation between \(P_{\text{Bopt}}\) and SST cannot be observed. A positive correlation between \(P_{\text{Bopt}}\) and SST can be observed at SST lower than 20°C (Behrenfeld and Falkowski, 1997a; Kameda, 2003; Kameda and Ishizaka, 2005). In terms of \(P_{\text{Bmax}}\), a positive correlation can be observed at SST lower than 15°C (Balch and Byrne, 1994).

The lowest \(P_{\text{Bopt}}\) (3.02 mgC mgChl-\(a\)-\(^{-1}\) h\(^{-1}\)) in this study was observed in the Kuroshio region, being associated with the highest SST (27.5°C). This probably reflects the chronic nutrient limitation, as the Kuroshio waters are commonly depicted as low nutrient waters (Chang et al., 2003; Gong et al., 1996; Hama et al., 1997). The low \(P_{\text{Bopt}}\) of 5.0 mgC mgChl-\(a\)-\(^{-1}\) h\(^{-1}\) was also observed in the Kuroshio region during autumn 1999, which might be caused by low nitrate concentration (Shiah et al., 1996). We thus infer that the negative correlation between \(P_{\text{Bopt}}\) and SST can be more attributed to the secondary factors associated with elevated SST, such as low nutrient concentration as suggested by Behrenfeld and Falkowski (1997a).

In this study, \(P_{\text{Bopt}}\) also demonstrated a negative correlation with Chl\(_{o}\), even though it was not a remarkable relation (data not shown), which is probably caused by the limited range of Chl\(_{o}\) (0.16 – 1.45 mg m\(^{-3}\)). The following descriptions can probably explain the trend of increasing \(P_{\text{Bopt}}\) that was associated with decreasing Chl-\(a\). In the region with lower Chl-\(a\), small size phytoplankton will be more dominant in
phytoplankton communities. Chen (2000) found that, in the Kuroshio waters that registered as low Chl-a waters (oligothropic waters), phytoplankton communities are dominated by small size phytoplankton (picoplankton). Because small phytoplankton is more productive than large phytoplankton (Malone, 1980), low Chl-a waters with a high ratio of high productive small phytoplankton, will tend to have higher $P_{\text{opt}}^B$ (Kameda, 2003; Kameda and Ishizaka, 2005).

2.4.3. Underestimation of Primary Production as a Consequence of Vertically Uniform Chl-a Vertical Profile: Comparison of the Kuroshio and Frontal Regions

We have described that the Kuroshio and the frontal region Chl-a profiles had the same feature of DCM formation but were different in the magnitude of $z_m$. The Kuroshio region profiles were more attributed to the deeper DCM and the frontal region profiles were more attributed to the shallower DCM.

Applying the algorithm to a relatively small area covering the Chl-a profile autumn stations, a comparison of vertically non-uniform Chl-a with uniform Chl-a in terms of the IPP in the Kuroshio and the frontal regions reflects that the DCM did contribute to the IPP within 12.0 ± 8.9% and 27.4 ± 7.9% for the Kuroshio and the frontal regions, respectively. The same typical pattern, even at slightly higher magnitudes, also emerged when the estimation was performed using satellite data covering a larger spatial scale, in which the DCM contributed to the IPP within 20.9 ± 5.4% and 30.9 ± 9.1% in the Kuroshio and the frontal regions, respectively. It is somewhat surprising that these degrees of the DCM contribution agree with Platt and Sathyendranath’s (1988) result which find a maximum 40% contribution from the DCM to the IPP when the DCM occurs between approximately 2 and 3 optical depths.
This study has shown that the higher DCM contribution to the production in the frontal region reflects the greater importance of the DCM occurrence in the frontal region than its occurrence in the Kuroshio region. This is because, the DCM in the Kuroshio region generally occurred slightly above the base of the euphotic zone, whereas in the frontal region the DCM occurred between approximately 2 and 3 optical depths, which enables phytoplankton assemblages in the DCM layer to intensively absorb and utilize light for photosynthesis. In contrast, the importance of the DCM in the Kuroshio region was reduced because of low light level in the deep layer.

We have mentioned that the DCM contribution to the IPP was higher in the frontal region than in the Kuroshio region. However, in terms of the integrated biomass within the euphotic zone, the opposite was true. The DCM had lower relative biomass in the frontal region (29.9% and 19.8% for autumn and spring, respectively) than in the Kuroshio region (35.9% and 48.7% for autumn and spring, respectively). The question is, then, why the DCM contributions to the integrated biomass and production were not positively correlated. This phenomenon again probably reflects that the Chl-a maximum formed in the low-light, deep layer of the Kuroshio region did not arise from the intensive photosynthetic rate, but suggests that other mechanisms mediate the Chl-a enhancement in the deep layer. In contrast, in the frontal region, even the DCM contributed less biomass than in the Kuroshio region, but existed in a relatively higher light level, which then enabled phytoplankton assemblages to grow intensively, resulting in high production. These phenomena are then probably a reasonable explanation for the greater production observed in the frontal region than in the Kuroshio region.
2.5. Summary

Several significant findings from this study are as follows;

1. in the effort to infer Chl-\(a\) profiles, in which surface variables were used as estimator, the water mass features should be considered. SST might be a potential variable that can be used to distinguish Chl-\(a\) profile types from different water masses,

2. DCM did give a significant contribution to IPP in the Kuroshio front of the ECS. The presence of the DCM seems to be more important in the frontal region than in the Kuroshio region, as the DCM in the former region exists in a relatively shallower layer, allowing the phytoplankton cells to absorb and utilize adequate light. The deeper DCM in the Kuroshio region is probably detrimental to the phytoplankton cells, preventing them from being exposed to the optimum light for their growth, resulting in less DCM contribution, even if it is still a non-trivial amount,

3. the magnitude of the IPP in the Kuroshio front of the ECS was to some degree enhanced by the DCM, especially in the frontal region, as revealed by the higher IPP in the frontal region than in the Kuroshio one. Thus, we emphatically suggest that efforts in estimating primary production in the Kuroshio front of the ECS should consider vertical variation allowing the model to capture intensive production in the DCM.
Table 2-1. Cruise information and numbers of stations where data were collected.

<table>
<thead>
<tr>
<th>Cruise periods</th>
<th>Ship names</th>
<th>Data</th>
<th>Numbers of stations</th>
</tr>
</thead>
<tbody>
<tr>
<td>24 October 1999*</td>
<td>R/V Yoko Maru</td>
<td>Primary production</td>
<td>1</td>
</tr>
<tr>
<td>3 – 17 October 2000</td>
<td>R/V Yoko Maru</td>
<td>Primary production</td>
<td>10</td>
</tr>
<tr>
<td>15 – 23 May 2000</td>
<td>T/V Kakuyo Maru</td>
<td>Underwater irradiance</td>
<td>8</td>
</tr>
<tr>
<td>8 – 24 May 2001</td>
<td>T/V Nagasaki Maru and</td>
<td>1. Chl-a profile</td>
<td>42</td>
</tr>
<tr>
<td></td>
<td>Kakuyo Maru</td>
<td>2. Underwater irradiance</td>
<td>7</td>
</tr>
<tr>
<td>2 – 12 October 2001</td>
<td>T/V Nagasaki Maru</td>
<td>1. Chl-a profile</td>
<td>31</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2. Underwater irradiance</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3. Primary production</td>
<td>2</td>
</tr>
</tbody>
</table>

*day of primary production experiment
Table 2-2. Determination coefficients ($r^2$) obtained from the least-square method relating profile parameters to SST and/or Chl$_0$ for the spring Chl-$a$ profiles.

<table>
<thead>
<tr>
<th></th>
<th>Kuroshio &amp; frontal regions</th>
<th>Shelf region</th>
<th>Kuroshio &amp; shelf region</th>
<th>Chl$_0$ function</th>
<th>Shelf region</th>
<th>SST and Chl$_0$ function</th>
</tr>
</thead>
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<tr>
<td></td>
<td>$n$</td>
<td>SST function</td>
<td>Chl$_0$ function</td>
<td>SST function</td>
<td>Chl$_0$ function</td>
<td>SST and Chl$_0$ function</td>
</tr>
<tr>
<td>$C_0$</td>
<td>31</td>
<td>0.003</td>
<td>0.780</td>
<td>0.790***</td>
<td>11</td>
<td>0.062</td>
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<tr>
<td>$h$</td>
<td>31</td>
<td>0.004</td>
<td>0.510</td>
<td>0.514***</td>
<td>11</td>
<td>0.529</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>31</td>
<td>0.203</td>
<td>0.153</td>
<td>0.318**</td>
<td>11</td>
<td>0.589</td>
</tr>
<tr>
<td>$z_m$</td>
<td>31</td>
<td>0.217</td>
<td>0.112</td>
<td>0.301</td>
<td>11</td>
<td>0.310</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>$n$</th>
<th>SST function</th>
<th>Chl$_0$ function</th>
<th>SST function</th>
<th>Chl$_0$ function</th>
<th>SST and Chl$_0$ function</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$z_m$</td>
<td>SST function</td>
<td>Chl$_0$ function</td>
<td>SST function</td>
<td>Chl$_0$ function</td>
<td>SST and Chl$_0$ function</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kuroshio &amp;</td>
<td>Shelf region</td>
<td>Kuroshio &amp;</td>
<td>Shelf region</td>
<td>Kuroshio &amp;</td>
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<tr>
<td></td>
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</tr>
<tr>
<td></td>
<td></td>
<td>42</td>
<td>0.740***</td>
<td>0.626$^+$</td>
<td>0.708$^+$</td>
<td></td>
</tr>
</tbody>
</table>

Note: *** $p < 0.001$, ** $p < 0.01$, $^5 p > 0.05$, $^+$ statistical analysis performed for all dataset without regional separation.
Table 2-3. Empirical models used for estimating the spring Chl-$a$ profile parameters selected based on their best $r^2$ values.

<table>
<thead>
<tr>
<th>Region</th>
<th>Equation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kuroshio &amp; frontal regions</td>
<td>$C_0 = -0.033 + 0.003 \text{ (SST)} + 0.746(\text{Chl}_0)$</td>
</tr>
<tr>
<td></td>
<td>$h = 26.945 + 0.096(\text{SST}) - 49.862(\text{Chl}_0)$</td>
</tr>
<tr>
<td></td>
<td>$\sigma = -31.821 + 2.013(\text{SST}) - 25.087(\text{Chl}_0)$</td>
</tr>
<tr>
<td></td>
<td>$\ln(z_m) = -13.021 + 5.265 \text{ (SST)}$</td>
</tr>
<tr>
<td>Shelf region</td>
<td>$C_0 = 0.250 - 0.016 \text{ (SST)} + 0.230(\text{Chl}_0)$</td>
</tr>
<tr>
<td></td>
<td>$h = -102.830 + 6.411(\text{SST}) + 15.139(\text{Chl}_0)$</td>
</tr>
<tr>
<td></td>
<td>$\sigma = -51.800 + 4.043(\text{SST}) - 12.767(\text{Chl}_0)$</td>
</tr>
<tr>
<td></td>
<td>$\ln(z_m) = -13.021 + 5.265 \text{ (SST)}$</td>
</tr>
</tbody>
</table>
Table 2-4. Determination coefficients ($r^2$) obtained from the least-square method relating profile parameters to SST and/or Chl$_0$ for the autumn Chl-$a$ profiles.

<table>
<thead>
<tr>
<th></th>
<th>Kuroshio region</th>
<th>Frontal region</th>
</tr>
</thead>
<tbody>
<tr>
<td>$n$</td>
<td>SST function</td>
<td>Chl$_0$ function</td>
</tr>
<tr>
<td>$C_0$</td>
<td>12</td>
<td>0.076</td>
</tr>
<tr>
<td>$h$</td>
<td>12</td>
<td>0.076</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>12</td>
<td>0.001</td>
</tr>
<tr>
<td>$z_m$</td>
<td>12</td>
<td>0.106</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>SST function</th>
<th>Chl$_0$ function</th>
<th>SST and Chl$_0$ function</th>
</tr>
</thead>
<tbody>
<tr>
<td>$z_m$</td>
<td>31</td>
<td>0.704*</td>
<td>0.263*</td>
</tr>
</tbody>
</table>

Note: *** $p < 0.001$, $^*$ $p > 0.079$,
+ statistical analysis performed for all dataset without regional separation.
Table 2-5. Empirical models used for estimating the autumn Chl-α profile parameters selected based on their best $r^2$ values. Parameters $\sigma$ in both regions and $h$ in the frontal region for predicting Chl-α profiles were defined from their mean values.

<table>
<thead>
<tr>
<th>Region</th>
<th>Equation</th>
<th>Parameters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kuroshio region</td>
<td>$\ln(B_0) = -0.763 + 0.602 \ln(\text{Chl}_0)$</td>
<td>$B_0 = -0.432 + 0.015 \text{SST} + 0.961 \text{Chl}_0$</td>
</tr>
<tr>
<td></td>
<td>$h = -46.588 + 2.583 \text{SST} - 48.860 \text{Chl}_0$</td>
<td>$h = 15.242$</td>
</tr>
<tr>
<td></td>
<td>$\sigma = 14.072$</td>
<td>$\sigma = 12.907$</td>
</tr>
<tr>
<td></td>
<td>$z_m = -587.333 + 24.232 \text{SST}$</td>
<td></td>
</tr>
<tr>
<td>Frontal region</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>$B_0 = -0.432 + 0.015 \text{SST} + 0.961 \text{Chl}_0$</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$h = 15.242$</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$\sigma = 12.907$</td>
<td></td>
</tr>
<tr>
<td></td>
<td>$z_m = -587.333 + 24.232 \text{SST}$</td>
<td></td>
</tr>
</tbody>
</table>
Fig. 2-1. A: Location of stations where Chl-α profiles were collected and selected to establish an empirical model of Chl-α profile. Open and closed squares indicate the spring and autumn cruises, respectively. During the spring cruise, the southernmost transect denoted by KM-line was observed on four separate days of 17, 18, 20, and 21 May 2001. B: Location of stations where primary production experiments (open triangles) and profiling underwater irradiance (closed triangles) were conducted. Labels above triangles indicate station numbers.
Fig. 2-2. Shifted Gaussian distribution showing four parameters ($C_0$, $h$, $\sigma$ and $z_m$) used to describe Chl-$\alpha$ profiles.
Fig. 2-3. Top: total 42 Chl-α profiles collected during spring. Squares and curves are discrete measurements and Gaussian distribution-fitted profiles, respectively. Middle: dendrogram resulting from cluster analysis, showing three main clusters of Chl-α profiles marked by squares, closed circles, and triangles. Bottom: three patterns of Chl-α profiles corresponding to the three clusters.
Fig. 2-4. Spatial distribution of spring Chl-α profile clusters overlaid with SST (A) and Chl\(_0\) (B) maps. Squares, closed circles and triangles denote three clusters identified from cluster analysis. Additional panel at the bottom shows the results observed at KM-line for each observation day (17, 18, 20 and 21 May).
Fig. 2-5. Scatter plot of SST against Chl$_0$ for spring, showing SST of 22.3 °C as the boundary separating the low SST (shelf) region profiles from high SST region profiles consisting of the Kuroshio and the frontal region profiles.
Fig. 2-6. Top: total 31 Chl-a profiles collected during autumn. Squares and curves are discrete depth measurements and Gaussian distribution-fitted profiles, respectively. Middle: dendrogram resulting from cluster analysis, showing two clusters of Chl-a profiles marked by closed circles and triangles. Bottom: two patterns of Chl-a profiles corresponding to the two clusters.
Fig. 2-7. Spatial distribution of autumn Chl-a profile clusters overlaid with SST (A) and Chl$_0$ (B) maps. Triangles and closed circles denote two clusters identified from cluster analysis. Labels around the marks indicate numbers of profiles.
Fig. 2-8. Scatter plot of SST against Chl$_0$ for autumn, showing SST of 26.85 °C as the boundary separating the frontal and the Kuroshio region profiles.
Fig. 2-9. Bar graphs showing the average biomasses within the upper and lower layers calculated using vertically uniform Chl-a profiles (white bars), predicted Chl-a profiles (grey bars) and measured Chl-a profiles (black bars). A and B are for the spring and autumn cruises, respectively. Labels denote average biomass ± standard deviation.
Fig. 2-10. Bar graphs showing the probabilities of IPPuc/IPPms (A) and IPPnc/IPPms (B) ratios. Curves are normal distributions of probabilities.
Fig. 2-11. Euphotic depth-integrated primary production during autumn. Closed bars denote productions calculated with predicted vertically non-uniform Chl-α profiles (IPPnc); open bars are productions calculated with vertically uniform Chl-α profile (IPPuc).
Fig. 2-12. Top: October 1999 monthly composite images of SeaWiFS Chl₀ (A), AVHRR SST (B) and SeaWiFS E₀ (C). Bottom: integrated primary production estimated by predicted vertically non-uniform Chl-a profiles (D), estimated by vertically uniform Chl-a profiles (E), and percentage of the underestimation due to vertically uniform Chl-a profiles (F). White polygon denotes approximate area selected for determining the magnitudes of production in the Kuroshio and the frontal regions (see text).
Chapter 3

Optimal Primary Production Model and Parameterization in the Eastern East China Sea

3.1. Introduction

Quantification of primary production in the ocean is an important prerequisite when addressing the environmental issues like the role of ocean on carbon cycling. However, toward better estimation of global primary production, more concerns must be emphasized in the highly dynamic continental shelves (Chen et al., 2003). Even the world’s continental shelves represent less than 20% of the world’s oceanic areas, the high nutrient input, such as river run-off and upwelling has made its primary production could be as important as that in the open ocean (Walsh, 1991).

The ECS is one of the most productive waters in the world’s oceans. Attempt to develop primary production model in the ECS has been conducted by Gong and Liu (2003). They developed EM from the datasets collected from the regions covering coastal, upwelling, continental shelf and Changjiang adjacent waters of the ECS (hereafter western ECS). This region, especially in the northwestern part was characterized by high seasonal variation in primary production with the highest in summer (Gong et al., 2003). For more local scale, as described in chapter 2 we developed TIM for the Kuroshio frontal region. These two models incorporate $P_{\text{opt}}^B$, a key parameter in driving primary production (Behrenfeld and Falkowski, 1997a,b; Kameda and Ishizaka, 2005).

In the ECS, the common variables used in modeling primary production, such as SST and Chl$\alpha$, showed gradual changes from western to eastern ECS (Fei et al., 1992; Tseng et al., 2000; see also chapter 2). Such gradual changes forced us to verify
whether EM and TIM are also applicable in this study region spanning from continental shelf to the off-shelf deep waters of the ECS (hereafter eastern ECS) which presumably has different characteristic of Chl-a profile and optical property of water from the regions where TIM and EM were developed. This study region spans from the region with the bottom depth > 50 m to the off-shelf deep waters with the bottom depth > 1000 m, where the bottom sediment resuspension does not overestimate satellite-derived Chl-a (Kiyomoto et al., 2001).

We were also interested to investigate the performance of the VGPM proposed by Behrenfeld and Falkowski (1997a), as it is one of the most widely used production model. VGPM employs original $P_{opt}^{B}$ model (Behrenfeld and Falkowski, 1997a) and modified one proposed by Kameda and Ishizaka (2005). The original $P_{opt}^{B}$ model is a seventh-order polynomial function of SST, but one of Kameda and Ishizaka (2005) is a function of both SST and Chl$_{a}$.

This work, thus attempts to validate TIM, EM and VGPM with primary production datasets collected in the eastern ECS. We aim to identify the main factor(s) influencing primary production variation in the eastern ECS, as well as to investigate how the above three models explain the variation in measured primary production. The predominant factor(s) responsible for the discrepancies between calculated and measured productions, the advantages and limitations of each production model were assessed.
3.2. Materials and Methods

3.2.1. Primary Production Experiment

We conducted simulated *in situ* primary production measurements based on 24 h incubation periods at 30 stations in the eastern ECS (Fig.3-1) during the period 1999-2002 covering all seasons. At each station, water samples for primary production and Chl-α measurements were collected at six depths corresponding to 100%, 50%, 25%, 10%, 5%, and 1% of E₀, and sampled either using 10-l Go-Flo or 5-l Niskin bottle samplers. Sampling depths were estimated either from under water irradiance profiles recorded by means of a profiling reflectance radiometer, PRR-600 (Biospherical Inc.), or from the Kₐ estimated by means of empirical relationship between Chl₀ and Kₐ as described in chapter 2 (Eq.2-7). During the incubation period, incident irradiance were recorded either using Biospherical QSL-2100 scalar irradiance sensor or Li-Cor LI-190SB cosine corrected flat sensor (downward irradiance sensor). For consistency scalar irradiance was converted to downward irradiance. Scalar irradiance was simply multiplied by average cosine of 0.8, as it experimentally ranges between 0.7 and 0.9 depending upon solar elevation, water type and wavelength (Hojerslev, 1975).

The concentrations of particulate organic carbon and isotopic ratio of ¹³C and ¹²C (¹³C atomic %) of the samples were determined by a quadrupole mass spectrometer equipped with a combustion furnace (ANELBA TE-150). The carbon fixation rate was calculated according to the method of Hama et al. (1993), and the IPP was calculated using trapezoidal integration. Chl-α was measured with Turner Design Model 10-AU fluorometer after extraction with N,N-dimethylformamide (Suzuki and Ishimaru, 1990). The highest P^B within water column was set as measured P^B_{opt}. 
3.2.2. Validation of Primary Production Model

Three previously established models TIM, EM and VGPM were employed in the present study. The TIM was originally developed to assess the contribution of DCM to IPP in the Kuroshio frontal region of the ECS (see chapter 2). The EM was developed to provide production model for the ECS (Gong and Liu, 2003). The global VGPM was also validated for the reason of being most widely used and less complex compared to bio-optical models (e.g. Ishizaka, 1998; Morel and Berthon, 1989; Morel 1991; Platt and Sathyendranath, 1993).

The performances of production models in explaining measured IPP variance were investigated through three experiments (Exp.I, II and III) differing in P_{opt} parameterization. Exp.I employs P_{opt} model originally embedded in each production model, Exp.II incorporates measured P_{opt} and Exp.III applies modeled P_{opt} based on the datasets of this study (Table 3-1). Measured variable inputs of Chl_0, E_0, euphotic depth (z_{eu}) and daylength (D_{in}) were used for computing IPP. The E_z and K_d were estimated from z_{eu} (E_z = E_0 \exp(-z K_d) and K_d = 4.6/z_{eu}). In the Exp. III, Chl_z for TIM were estimated from Chl_z model derived from PN line datasets (period from 1981 through 1999, covering all seasons, Fig.3-1) provided by the Japan Oceanographic Data Center (JODC, http://www.jodc.go.jp/service.htm). The method to construct Chl_z model will be described in the following subsection.

3.2.3. Chl-a Vertical Profile Estimation for TIM Input

We fitted Gaussian equation to the discrete depths of Chl-a to determine Chl_z as well as their profile parameters (Lewis et al., 1983; Platt, 1986). We followed the constrains suggested by Sathyendranath et al. (1995) to obtain plausible profiles. In
the study, Chl \textsubscript{z} with DCM was well fitted with 4-parameter Gaussian model (Eq.3-1), whereas that with SCM was more satisfied with 3-parameter Gaussian model (Eq.3-2),

\[ \text{Chl}_z = C_0 + C_p \exp \left[ -\frac{(z-z_m)^2}{2\sigma^2} \right] \] (3-1)

and

\[ \text{Chl}_z = C_p \exp \left[ -\frac{(z-z_m)^2}{2\sigma^2} \right] \] (3-2)

where \( C_p \), is profile peak concentration above background (mg m\textsuperscript{-3}).

We obtained the relationships between profile parameters and surface variables and used these relationships to infer Chl\textsubscript{z}. Since many studies showed that \( z_m \) is predictable from surface parameter (e.g., Richardson, et al., 2003; see also chapter 2), the Chl\textsubscript{b} could provide a dependable estimate of \( z_m \) for both SCM and DCM formation (Table 3-2).

Following Ballestero (1999), we derived the slope and intercept of vertical maximum of Chl-a (Chl\textsubscript{max}) and Chl\textsubscript{b} relationship. The slope decreased from 1.0 (at \( z_m = 0 \) m) to 0 (at \( z_m = 40 \) m), following the fourth-order polynomial function of \( z_m \), while intercept was linearly correlated with \( z_m \) (Table 3-2). Therefore, the slope and intercept will be estimated from \( z_m \) in order to provide Chl\textsubscript{max} or \( C_p \) in inferring Chl\textsubscript{z} with SCM.

For profile with DCM, \( C_0 \) could be well predicted from Chl\textsubscript{b} and become an important determinant of Chl\textsubscript{z} (Table 3-2). Conversely, \( C_p \) was uncoupled from Chl\textsubscript{b}. Therefore, we used mean \( C_p \) for estimating Chl\textsubscript{z} with DCM (Table 3-2).

As a parameter determining the thickness of Chl\textsubscript{max}, \( \sigma \) was uncoupled from surface variables, but showed a seasonal variation. Therefore, we derived seasonal mean of \( \sigma \), and fitted it to the sinusoidal function to obtain its predictive equation (Table 3-2).
3.3. Results

3.3.1. Variation of Dataset

Temporal variations of measured $E_0$, SST, $Chl_0$, $P_{opt}$ and IPP were clearly emerged in this study (Fig.3-2). Large variation of $E_0$ was observed during fall (Fig.3-2A). Both the lowest (8.5 Einstein m$^{-2}$ d$^{-1}$) and the highest $E_0$ (53.4 Einstein m$^{-2}$ d$^{-1}$) were observed in September. SST exhibited the most distinct temporal variation with the lowest (14.5°C) and the highest (28.6°C) were in February and August, respectively (Fig.3-2B). $Chl_0$, the main variable of primary production models amenable to remote sensing, ranged from 0.09 mg m$^{-3}$ (September) to 1.45 mg m$^{-3}$ (October) with the mean value of 0.37 mg m$^{-3}$ (Fig.3-2C).

As a key parameter in satellite-based modeling primary production, $P_{opt}$ ranged from 31 mgC mgChl$^{-1}$ d$^{-1}$ (February) to 147 mgC mgChl$^{-1}$ d$^{-1}$ (August) with the mean value of 79 mgC mgChl$^{-1}$ d$^{-1}$ (Fig.3-2D). The lowest and the highest $P_{opt}$ were coincident with the lowest and the highest SST, but $P_{opt}$ showed greater monthly variation than SST. The lowest measured IPP (299 mgC m$^{-2}$ d$^{-1}$) was observed in September, while the highest one (2420 mgC m$^{-2}$ d$^{-1}$) was in October with the mean value of 855 mgC m$^{-2}$ d$^{-1}$ (Fig.3-2E).

3.3.2. Relationships Between Primary Production, Model Parameter and Surface Variables

Among the sea surface variables, $E_0$ showed a significant linear relationship ($r^2 = 0.19$) with measured IPP (Fig.3-3A). There was no significant relationship between measured IPP and SST (Fig.3-3B).

With all datasets, even the determination coefficient of the relationship between measured IPP and $Chl_0$ was higher ($r^2 = 0.25$, Fig.3-3C) than that with $E_0$, the
scatter plot actually showed a lack of linear relationship. The highest IPP measured in October 1999 that coincided with the highest Chl\textsubscript{0} seemed to be an outlier value from the seasonal variation (Fig.3-2E). These highest measured IPP and Chl\textsubscript{0} were probably affected by severe flood in the Changjiang valley occurred from June to August 1999 (Guo et al., 2004). Ishizaka (2000) has also remarked distribution of high satellite-based Chl\textsubscript{0} in 1999 in the ECS that may be associated with Changjiang plume variation. If excluding this highest IPP, Chl\textsubscript{0} could not explain measured IPP variance ($r^2 = 0.02$, Fig.3-3C).

Employing all datasets, P\textsuperscript{B\_opt} accounted for 20% of measured IPP variance (Fig.3-3D). Excluding the outlier value of measured IPP that may be mainly determined by Chl-a, P\textsuperscript{B\_opt} could account for 35% of variance (Fig.3-3D).

The significant exponential relationship was emerged between P\textsuperscript{B\_opt} and SST (Fig.3-4A). We thus employed this relationship to provide P\textsuperscript{B\_opt} estimates for calculating IPP using production models in the Exp.III. Despite lack of significant relationship, P\textsuperscript{B\_opt} tended to be reciprocally related with Chl\textsubscript{0} (Fig.3-4B). Photoadaptive parameter ($E_{\text{max}}$) was strongly correlated with $E_0$ (Fig.3-4C). Therefore, we used this relationship to provide $E_{\text{max}}$ for calculating IPP using TIM in the Exp.III.

3.3.3. Comparison of Calculated and Measured Production

From Exp.I, TIM with constant P\textsuperscript{B\_opt} of 73.8 mgC mgChl-a\textsuperscript{-1} d\textsuperscript{-1} and with Kuroshio frontal Chl\textsubscript{z} model (refer to second row, third column of Table 3-1 for detail parameterization) only accounted for 18% of the variation in measured IPP (Fig.3-5A). EM with originally embedded P\textsuperscript{B\_opt} models (refer to third row and third column of Table 3-1 for detail P\textsuperscript{B\_opt} equations), could explain measured IPP variation by as much as 50%
and 48%, but with remarkable low estimations as reflected by the low slopes (Fig.3-5B). VGPM with its originally embedded \( P_{\text{opt}}^{\text{B}} \) models (refer to fourth row and third column of Table 3-1 for detail \( P_{\text{opt}}^{\text{B}} \) formulations), explained variation in measured IPP by as much as 16% and 13% (Fig.3-5C).

Employing measured \( P_{\text{opt}}^{\text{B}} \) (Exp.II), TIM, EM and VGPM accounted for 93%, 84%, and 90% of the measured IPP variance, respectively (Fig.3-6A). However, while TIM and VGPM showed accurate predictions as their slopes were indistinguishable from the unity (Student’s \( t \)-test, \( p > 0.05 \)), EM still demonstrated low estimations with the slope significantly distinguishable from the unity (Student’s \( t \)-test, \( p < 0.001 \)) and with higher root mean square errors (RMSE) than those of TIM and VGPM. The outlier of highest measured IPP could also be predicted well by TIM and VGPM with the use of measured \( P_{\text{opt}}^{\text{B}} \) (Fig.3-6A).

In the Exp.III, we used modeled \( P_{\text{opt}}^{\text{B}} \) based on datasets of this study. \( F_{\text{max}} \) and Chl \( z \) for TIM were also estimated based on this study datasets. Because of low estimations encountered by EM in the Exp.II, we only run TIM and VGPM. The result of TIM, 60% of measured IPP variance could be accounted (Fig.3-6B), thus its capability in explaining measured IPP variance reduced by as much as 33% compared to its capability in the Exp.II (93%). VGPM accounted for 72% of variation in measured IPP (Fig.3-6B), thus the capability in explaining measured IPP variance reduced by as much as 18% from its capability in the Exp.II (90%).
3.4. Discussion

After excluding the highest measured IPP which was presumably an outlier, P^{Bopt} and E_0 were more important factors in determining variation in measured IPP than Chl_0. Many studies also showed that greater variance of measured IPP can usually be captured by including P^{Bopt} as a phytoplankton photoacclimation index, along with variables amenable to remote sensing (e.g., Balch et al., 1989; Banse and Postel, 2003). In the following subsections, we will elucidate the importance of P^{Bopt}, as well as E_0 in modeling primary production in this study region.

3.4.1. Difference of P^{Bopt} Estimation

The consistent increase in P^{Bopt} with increasing SST observed in this study, differs from both global seventh-order polynomial P^{Bopt} model (Behrenfeld and Falkowski, 1997a), and both SST and Chl_0 dependent P^{Bopt} model (Kameda and Ishizaka, 2005), which demonstrated P^{Bopt} decline at SST > 20°C (Fig.3-7). Such disparity was responsible for the inability of VGPM in explaining measured IPP variance in the Exp.1. In this study, P^{Bopt} increased consistently even at SST > 28°C, similar with that observed in Cariaco station, southeastern Caribbean Sea with consistent increase in P^{Bopt} even at SST as high as 29°C (Muller-Karger et al., 2004).

As noted by Behrenfeld and Falkowski (1997a), the decline of P^{Bopt} at SST > 20°C described by the global seventh-order polynomial equation, reflects a chronic nutrient limitation in the strongly stratified high SST regions. This disparity between global and local P^{Bopt} variations in response to SST is likely to reflect highly regional P^{Bopt} dynamics associated with chronic nutrient status defining phytoplankton growth (Kamykowski et al., 2002). Such a high P^{Bopt} variation was also responsible for the
imprecision of TIM estimation which employed constant $P_{\text{opt}}^B$ derived from Kuroshio frontal region (Exp.1).

The pattern of increasing $P_{\text{opt}}^B$ with SST observed in this study also resembled that estimated using third-order polynomial $P_{\text{opt}}^B$ model noted by Gong and Liu (2003) for the coastal and continental shelf regions of the ECS (Fig.3-7), except $P_{\text{opt}}^B$ of this study (mean 78.8 mgC mgChl-a$^{-1}$ d$^{-1}$) were higher than that of Gong and Liu (2003) (mean 41.8 mgC mgChl-a$^{-1}$ d$^{-1}$). This difference was responsible for the low IPP estimated using EM in the Exp.I.

Based on the fact that Chl$\_0$ of this study (mean 0.37 mg m$^{-3}$) were lower than that of Gong and Liu (2003) (mean 0.84 mg m$^{-3}$), the difference in $P_{\text{opt}}^B$ seemed to reflect the reciprocal relation between $P_{\text{opt}}^B$ and Chl$\_0$ as noted by Kameda and Ishizaka (2005). Despite lack of significant relationship, in fact $P_{\text{opt}}^B$ tended to be reciprocally related with Chl$\_0$ (Fig.3-4B). Kameda and Ishizaka (2005) postulated that the decline in $P_{\text{opt}}^B$ with the increase in Chl$\_0$ might be caused by the increase in the fraction of low-productive large size phytoplankton.

3.4.2. Light-Dependent Function

Incorporating measured $P_{\text{opt}}^B$, we further discerned that severe underestimation of EM-calculated IPP in the Exp.II, mostly arose from the data exhibiting deep primary production maximum (DPM) or deep high production at DCM layer. In fact, by removing these data, the slope of EM-based estimation increased from 0.65 (for all data, Fig.3-6A) to 0.75 reflecting that, to some degree EM was likely unable to accommodate intensive production at DCM, as many studies also found that the model underestimation problem is often caused by DCM presence (e.g., Banse and
Yong, 1990; Mouw and Yoder, 2005; Morel and Berthon, 1989; Ondrusek, et al., 2001). However, the slope of 0.75 still reflects a low estimation, as it was statistically distinguishable from the unity (Student’s $t$-test, $p < 0.05$), meaning that there is still another cause of EM low estimation. Using a constant factor (1.805, see Table 3-1), instead of $E_0$ dependent function may be the other probable cause of the low estimation encountered by EM.

$E_0$ dependent function is an important factor, as it describes the depth of light saturation of carbon fixation rate at depth $z$ ($P_z$) and thus the shape of $P_z$ (Banse and Yong, 1990; Behrenfeld and Falkowski, 1997b), which is influenced by Chl$_z$ and $E_z$. Changes in $E_0$ dependent light saturation depth can potentially cause IPP to vary by factor of $\sim$4 (Behrenfeld and Falkowski, 1997b). Previous studies showed that Chl-$a$ profile varied spatially in the ECS, and the farther the distance from the coast of China, the deeper the DCM would be (e.g., Fei et al., 1992; see also chapter 2). Many of datasets from which EM was developed were collected from coastal and Changjiang adjacent waters of the ECS (see Fig.3-1 in Gong and Liu (2003)) and with higher turbidity (mean $K_d=0.17 \text{ m}^{-1}$, compare to mean $K_d$ of this study $=0.08 \text{ m}^{-1}$) presumably contained many data with SCM and surface primary production maximum (SPM). About 28% and 18% of the datasets based on which EM was developed were the data with DCM and DPM, respectively (G.-C. Gong, personal communication). In contrast, 67% and 53% of this study datasets exhibited DCM and DPM, respectively (Fig.3-8). Therefore, EM with a constant factor, seemed to be inherently more adapted in the turbid coastal and continental shelf regions of the ECS with SCM and SPM, less appropriate for this study region characterized by DCM, DPM and low turbidity.
The problem of underestimation was not encountered by VGPM after locally tuning on $P_{\text{B_opt}}$ (Exp.II and III), even from the level of model complexity, VGPM is comparable to EM which can be classified as depth integrated models. This is due to the fact that, even VGPM does not explicitly resolve $P_z$, it employs $E_0$ dependent function as $0.66125 \frac{E_0}{(E_0+4.1)}$ (see Table 3-1). Employing $E_0$ enables VGPM to determine the depth of light saturation of $P_z$. The other probable explanation for the good performance of VGPM is that, based on Kiyomoto et al. (2001), the optical property of this study region resembles that of the domain from which VGPM was developed in which Chl-$\alpha$ is the main constituent determining optical property of water.

TIM as an analytical model explicitly uses $E_0$ to estimate $E_z$, and Chl$_z$ as variable inputs. This enables TIM to resolve $P_z$ (the product of Chl-$\alpha$ specific carbon fixation rate at depth $z$ ($P_{\text{B}_z}$) and Chl$_z$). Therefore, by incorporating measured $P_{\text{B_opt}}$ (Exp.II), TIM seemed to be very promising as it can also accommodate DPM. However, Chl$_z$ estimation itself, as well as other input variables, such as $E_{\text{max}}$ and $E_z$, seemed to constrain high accuracy of TIM (Exp.III). This is inevitable limitation in TIM to derive Chl$_z$, $E_z$, and $E_{\text{max}}$ solely based on sea surface variables.

From the above discussion, we therefore conclude that for the reason of simplicity, VGPM formulation with locally tuning on $P_{\text{B_opt}}$ seemed to be an effective model for estimating IPP in this study region. However, besides the problem of satellite Chl-$\alpha$ retrieval in case 2 waters, modeling IPP for the entire region of the ECS spanning from coastal turbid waters to off-shelf region, is also likely challenged by the highly dynamics of $P_{\text{B_opt}}$ itself.
3.5. Summary

From the above discussion, we draw some conclusions as follows:

1. $P_{opt}^B$ was a key parameter in driving IPP, and in our view, modeling $P_{opt}^B$ accurately is likely one of the main routes in estimating IPP even within the regional scale of the ECS,

2. TIM as an analytical model, retains useful information on $P_s$, as well as DPM. However, the high accuracy of TIM was constrained by its complexity in inferring vertical variations of its variable inputs,

3. EM as the simplest model, is indeed accurate for the western ECS, but it should be used with caution in the eastern ECS characterized by higher $P_{opt}^B$, deeper DCM, as well as lower turbidity,

4. VGPM, even as the global model, coupled with local tuned $P_{opt}^B$ seemed to be an optimal model for the eastern ECS.
Table 3-1. Primary production models and parameterizations applied in Exp.I, Exp.II and Exp.III. In the Exp.I, each model incorporates its originally embedded parameters, while in the Exp.II and III, the models employ measured and modeled parameters, respectively.

<table>
<thead>
<tr>
<th>Model</th>
<th>Model equation</th>
<th>Exp. I</th>
<th>Exp. II</th>
<th>Exp. III</th>
</tr>
</thead>
<tbody>
<tr>
<td>TIM</td>
<td>$IPP = \sum_{i=0}^{\infty} Chl_i \times \left[ P_{\text{opt}}^{i} \times \frac{E_x}{E_{\text{max}}} \times \exp \left( 1 - \frac{E_x}{E_{\text{max}}} \right) \right]$</td>
<td>$P_{\text{opt}}^{i} = 73.8 \text{ mgC mgChl}^{-1} d^{-1}$</td>
<td>Measured $P_{\text{opt}}^{i}$</td>
<td>$P_{\text{opt}}^{i} = 9.06 e^{0.06(SST)}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$E_{\text{max}} = 11.0 + 3.1(E_0) - 0.1(E_0)^2 + 0.002(E_0)^3$</td>
<td>Measured $E_{\text{max}}$</td>
<td>$E_{\text{max}} = 0.29(E_0) + 6.23$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Estimated $Chl_i$</td>
<td>Measured $Chl_i$</td>
<td>Estimated $Chl_i$</td>
</tr>
<tr>
<td>EM</td>
<td>$IPP = 1.805 [Chl_0 \times P_{\text{opt}}^{i} \times K_d^{i-1}]$</td>
<td>$P_{\text{opt}}^{i} = -286.17 + 49.166(SST)$</td>
<td>Measured $P_{\text{opt}}^{i}$</td>
<td>Excluded from exercise</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$-2.543(SST)^2 + 0.0435(SST)^3$</td>
<td><strong>$P_{\text{opt}}^{i} = -3.55 + 3.036(E_0) - 0.1243(E_0)^2$</strong></td>
<td><em><strong>$P_{\text{opt}}^{i} = 0.00213(E_0)^2 - 0.0000114(E_0)^4$</strong></em></td>
</tr>
<tr>
<td>VGPM</td>
<td>$IPP = 0.66125 \times P_{\text{opt}}^{i} \times \left[ \frac{E_x}{(E_0 + 4.1)} \right] \times z_{\text{so}} \times Chl_0 \times D_{\text{ir}}$</td>
<td>$P_{\text{opt}}^{i} = -3.27e^{-8(SST)} + 3.4132e^{-6(SST)^2}$</td>
<td>Measured $P_{\text{opt}}^{i}$</td>
<td>$P_{\text{opt}}^{i} = 9.06 e^{0.06(SST)}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$-1.348e^{-4(SST)^2} + 2.462e^{-3(SST)^3}$</td>
<td>$P_{\text{opt}}^{i} = 0.0205(SST)^2 + 0.0617(SST)^3$</td>
<td>$P_{\text{opt}}^{i} = 0.2749(SST)^2 + 1.2956$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$+ 0.2749(SST)^2 + 1.2956$</td>
<td>$P_{\text{opt}}^{i} = (0.071(SST) - 3.2e^{-3(SST)^2})$</td>
<td>$P_{\text{opt}}^{i} = 3.0e^{-5(SST)^2} / Chl_0$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$+ 3.0e^{-5(SST)^2} / Chl_0$</td>
<td>$P_{\text{opt}}^{i} = (1.0 + 0.17(SST) - 2.5e^{-3(SST)^2})$</td>
<td>$P_{\text{opt}}^{i} = 8.0e^{-5(SST)^2}$</td>
</tr>
</tbody>
</table>

*Mean value derived from Kuroshio frontal region (Siswanto et al., 2005)

#Model was derived from Kuroshio frontal region (Siswanto et al., 2005)

**Third-order polynomial function of SST (Gong and Liu, 2003)

***Fourth-order polynomial function of $E_0$ (Gong and Liu, 2003)

†Seven-order polynomial function of SST (Behrenfeld and Falkowski, 1997a)

++Function of both SST and Chl_0 (Kameda and Ishizaka, 2005)
Table 3-2. Equations used to derive Chl with DCM (C₀, Cₚ, σ, zₘ) and SCM (Chl_max, σ, zₘ). All of coefficient determinations \( r^2 \) are significant \( (p < 0.001) \).

<table>
<thead>
<tr>
<th>Equations for Chl with SCM (3-parameter Gaussian model)</th>
<th>Equations for Chl with DCM (4-parameter Gaussian model)</th>
</tr>
</thead>
<tbody>
<tr>
<td>( z_m = 12.37 \text{ (Chl}_o)^{-0.71} )</td>
<td>( z_m = 12.37 \text{ (Chl}_o)^{0.71}; )</td>
</tr>
<tr>
<td>( \sigma = 27.43 - 10.67 \sin((360/365<em>JD) - (0.50</em>180)) )</td>
<td>( \sigma = 27.43 - 10.67 \sin((360/365<em>JD) - (0.50</em>180)) )</td>
</tr>
<tr>
<td>( \text{Cₚ or Chl}_\text{max} = (\text{Chl}_o - \text{Intercept})/\text{Slope} )</td>
<td>( \text{C₀} = 0.66 \text{ (Chl}_o) + 0.03 )</td>
</tr>
<tr>
<td>Intercept = 0.01 zₘ</td>
<td>( \text{Cₚ} = 0.37 )</td>
</tr>
<tr>
<td>Slope = 1.0 - 3.4E⁻² zₘ + 2.9E⁻³ zₘ² - 1.0E⁻⁴ zₘ³ - 2.0E⁻⁶ zₘ⁴</td>
<td>( r^2 = 0.99 )</td>
</tr>
</tbody>
</table>
Fig. 3-1. Primary production experiment stations (circles) and PN-line transect (line) superimposed on the map of the ECS. The size of circle represents relative magnitudes of measured IPP (mgC m$^{-2}$ d$^{-1}$). Labels H and L indicate the highest and lowest measured IPP, respectively. Number 5 indicates 5 data at the same station. The 50, 200 and 1000 m isobaths are also shown.
Fig. 3-2. Temporal variations of $E_0$, SST, Chl$_0$, $P^0_{\text{opt}}$, and IPP observed in this study.
Fig. 3-3. Linear relationships (solid lines) for all datasets between measured IPP and $E_0$ (A: $r^2 = 0.19, p < 0.05$), SST (B: linear fit is not shown because $p > 0.05$), Chl (C: $r^2 = 0.25, p < 0.001$), and $p_{\text{opt}}$ (D: $r^2 = 0.35, p < 0.05$). Dashed lines in C (insignificant, $p > 0.05$) and D ($r^2 = 0.35, p < 0.05$) are linear fits when excluding outlier value (datum labeled by ol).
Fig.3-4. A and B are exponential relationship between $P_{\text{opt}}^B$ and SST ($P_{\text{opt}}^B = 9.06 e^{0.08\text{SST}}$, $r^2 = 0.67, p < 0.001$) and reciprocal relation between $P_{\text{opt}}^B$ and Chl$_0$ (fitting was not conducted as $p > 0.05$), respectively. C is a linear relationship between $E_{\text{max}}$ and $E_0$ ($E_{\text{max}} = 0.29(E_0) + 6.23, r^2 = 0.40, p < 0.001$), respectively.
Fig. 3-5. Relationships between calculated and measured IPP from Exp. I. Calculated IPP in A, B and C were calculated using TIM, EM and VGPM, respectively. In A, TIM incorporates constant $P_{opt}^B$. In B, EM incorporates its original SST (closed triangle) and $E_0$ (open triangle) dependent $P_{opt}^B$. In C, VGPM incorporates its original SST (closed circle) and both SST and Chl$_0$ (open circle) dependent $P_{opt}^B$ (refer to Table 3-1, third column for detail parameterizations). Fitted line, slope, $r^2$ for each model and 1:1 line are shown. RMSE is root mean square errors.
Fig. 3-6. Relationships between calculated and measured IPP from Exp. II (A) and III (B). In A, TIM, EM and VGPM incorporate measured $P^B_{\text{opt}}$. In B, TIM and VGPM incorporate modeled $P^B_{\text{opt}}$ (refer to Table 3-1, fifth column for detail parameterization). Fitted line, slope, $r^2$ for each model and 1:1 line are shown. RMSE is root mean square errors.
Fig. 3-7. Relationships between measured $P_{\text{B, opt}}$ and SST. Squares, circles, diamonds and triangles indicate $P_{\text{B, opt}}$ variations based on datasets of this study, model of Behrenfeld and Falkowski (1997), model of Kameda and Ishizaka (2005) and SST dependent $P_{\text{B, opt}}$ model of Gong and Liu (2003), respectively (refer to Table 3-1 for detail $P_{\text{B, opt}}$ formulations).
Fig. 3-8. Vertical profiles of Chl-a (A) and primary production (B) of this study datasets.
Chapter 4

Primary Production Enhancement Induced by Tropical Cyclone in the East China Sea Revealed by Satellite Remote Sensing

4.1. Introduction

One of the mechanisms relevant to drawdown of atmospheric CO₂ is the enhancement of new production due to nutrient pumping driven by typhoon. Quantitatively, new production is equivalent to the organic matter that can be exported from IPP out of euphotic zone, without the production system running down (Eppley and Peterson, 1979).

However, because of typhoon’s severe weather condition and great variation in trajectory, the ship-borne observation on prescribed schedules and cruise tracks is not feasible and highly risk. These constrains are the causes of the lack of data on typhoon effects on upper layer biological processes.

Satellite remote sensing, with its capability to observe over large domain synoptically, has the potential to overcome the problems mentioned above. This study therefore uses satellite data to investigate the effects of typhoon on the dynamics of phytoplankton biomass, primary production and new production in the ECS.

Daily satellite data were analyzed to see the days in which or during which ocean biological processes respond to typhoon passage. Seven-year SeaWiFS data were also used in this study in order to see the interannual variation of phytoplankton dynamics as responses to typhoon events and links to the global atmospheric anomalies (El Nino-Southern Oscillation/ENSO), as widely known that typhoon intensity is correlated to El Nino/La Nina events.
This chapter will thus study relevant topics as follows:

1. investigating the variations of Chl-\(a\) and IPP enhancements driven by typhoon within the scales of days and interannual variations,
2. investigating the relationships between typhoon variables (such as, typhoon speed, wind speed and pressure) and enhancements of Chl-\(a\) and IPP,
3. investigating whether nitracline depth also play an important role in determining IPP enhancement
4. linking typhoon-driven IPP enhancements to the climate changes of El Nino/La Nina events,
5. assessing typhoon contribution to annual and summer IPP and integrated new production (INP).

4.2. Materials and Methods

4.2.1. Satellite Data

Sea surface Chl-\(a\) and irradiance

Phytoplankton Chl-\(a\) which is commonly considered as a direct proxy for the magnitude of the surface ocean’s phytoplankton population was derived from SeaWiFS ocean color sensor. To investigate daily variations of phytoplankton dynamics driven by typhoon, SeaWiFS daily LAC (local area coverage) Chl-\(a\) with 1 x 1 km\(^2\) spatial resolution were used. We focused on typhoon Meari that traveled slowly over the ECS from 25 to 29 September 2004. Eight-day composite SMI of SeaWiFS Chl-\(a\) with 9 x 9 km\(^2\) spatial resolution were used to provide interannual variations in phytoplankton dynamic in response to typhoon events. We only emphasized variations during the period from early summer through early fall (24 June – 8 November) for the reasons
that this period is the most frequent typhoon passages over the ECS and has the minimum cloud coverage within a year.

For the purpose of calculating primary production using primary production model, irradiance data are required as one of the input variables. Like Chl-α, daily LAC (1 x 1 km² resolution) and 8-day composite SMI (9 x 9 km² resolution) irradiance data derived from SeaWiFS were used in this study.

**Sea surface temperature**

SST, the ocean’s surface geophysical variable which is commonly used as a proxy of environmental condition driving physiological state of phytoplankton community is also required for determining photosynthetic parameter. This photosynthetic parameter will be then embedded in primary production model. We used daily Moderate Resolution Imaging Spectroradiometer (MODIS) Aqua-retrieved SST level-2 for the purpose of elucidating daily variation in primary production as it has the same spatial resolution (1 x 1 km²) with daily LAC of SeaWiFS Chl-α data.

For the study of interannual variation in primary production in response to typhoon events, we used Tropical Rainfall Measuring Mission/Microwave Imager (TRMM/TMI)-retrieved SST data with spatial resolution of about 25 x 25 km². The advantage of TMI over SeaWiFS and MODIS is the capability to penetrate cloud cover as TMI uses microwave band, whereas MODIS (for SST) and SeaWiFS (for Chl-α) use infrared and visible bands, respectively. The cloud penetrating capability of TMI allows the SST in the area covered by cloud during typhoon event to be sensed. The use of TMI SST is also benefit in primary production calculation as it ensures that the cloud problem only arises from SeaWiFS datasets.
Since 8-day means of TMI SST data are not available, and in order to ascertain a consistency with SeaWiFS datasets, we averaged daily SST data for each 8-day period over the period from 24 June through 8 November to produce 8-day mean TMI SST data. These 8-day means of SST data were then remapped to match the SeaWiFS data dimension.

4.2.2. Primary and New Production Estimations

In chapter 3, for the reasons of fast computation and simplicity to provide reliable estimation of primary production in the ECS especially during summer and late summer and in the region less affected by tidal mixing, Changjiang discharge and bottom influences, VGPM developed by Behrenfeld and Falkowski (1997a) incorporated with $P_{opt}^R$ model derived in chapter 3 (refer to Table 3-1) was more appropriate. Therefore the optimal model derived from chapter 3 was used in this study.

VGPM also requires euphotic depth, the depth correspond to 1% of $E_0$, as one of the input variables for integrating primary production within euphotic zone. In this study we use the model relating $K_d$ to Chl$_0$ as described in chapter 2 (Eq. 2-7). This model was also derived from datasets mostly collected from the regions less affected by Changjiang discharge, tidal mixing and bottom sediment resuspension influences.

We are also interested to assess the typhoon contribution to annual and summer INP. Based on nitrate uptake experiments, the $f$-ratio (the ratio of INP and IPP) in the continental shelf of ECS is about 0.4 (Chen et al., 1999; Chen et al., 2001a; Chen and Chen, 2003). However, Chen (2003) suggests that this $f$-ratio (0.4) seems to be considerably higher than the value based on the carbon export experiment because of recycled nitrate in the shelf can reenter into the euphotic zone easily. As the INP is
quantitatively equivalent to the organic matter that can be exported from IPP out of
euphotic zone, without the production system running down (Dugdale and Goering,
1967; Eppley and Peterson, 1979), we therefore used carbon export experiment-based f-
ratio which is typically 0.15 (Chen 2003; Liu et al., 2003), rather than 0.4 derived from
nitrate uptake experiment, for estimating annual and summer INP in the continental
shelf region (depth < 200 m). But for the off shelf region (depth > 200 m) we used the
typical f-ratio for oligothropic waters (0.1) proposed by Eppley (1989). As the nutrient
upward flux by typhoon mainly because of stirred-upwelling process, we used f-ratio for
upwelling region for assessing INP induced by typhoon. Typically, the f-ratio in the
upwelling regions (e.g., northeast of Taiwan and Peruvian upwelling) ranges from 0.2 to
0.8 (Chen et al., 1999; Wilkerson and Dugdale, 1987). We used the f-ratio of 0.5 to
calculate post-typhoon INP.

4.2.3. Data Analysis

Interannual variation in phytoplankton biomass in the ECS

In this analysis we investigate the 8-day means of Chl-a over the region
bordered by Cheju island to the north, Goto island to the northeast, Ryukyu islands to
the east, Taiwan to the south and 50 m isodepth to the west (region-1, Fig.4-1). We
masked the area with the bottom depth less than 50 m for the west side due to the fact
that this area potentially has the overestimation problem in satellite-retrieved Chl-a
caused by high SS turbid waters (Kiyomoto et al., 2001). Although the extension of
turbid waters is expected to be variable with time, the use of such a constant mask
allows the spatial means to be computed over fixed areas.
Rather than interpolation to fill cloud-covered pixels, eight-day SeaWiFS Chl-α images with cloud coverage larger than 75% within interest region as mentioned in Fig. 4-1, were not analyzed in this spatial average analysis. This is to minimize the error estimations due to cloud problems.

**Interannual variations of typhoon-induced phytoplankton biomass and production**

Analysis of the effects of typhoon on phytoplankton dynamics focus only on the region considered to be less affected by nutrient loading from Changjiang discharge and bottom sediment resuspension influences. Optically, such region probably more depicts case 1 waters, meaning that phytoplankton and their biological products, rather than terrigenous materials, are primarily responsible for variations in optical properties. CDOM (also known as yellow substance) and SS make relatively small contributions to the optical properties in this area. The region of interest (later defined as region-2) is considered based on some facts as follows (detail for defining study area is described in section 4.3.1):

1. from the viewpoint of new production, upward flux of nutrient-rich deep water that occurs in the oligotrophic case 1 waters is more important than horizontal flux of nutrient-rich river discharge usually occurs in case 2 waters (e.g. Chen *et al.*, 1999) meaning the region in which phytoplankton is not the only determinant factor but also CDOM and/or SS responsible for variation in optical properties, due to the shallow depth, strong tidal mixing, and river discharges,

2. Satellite ocean color Chl-α retrieval algorithm has the overestimation problem in the region with high terrigenous SS and CDOM loading from river discharge. Overestimation in Chl-α retrieval will inevitably lead overestimation in primary production estimations in this region,
3. previously established or modified production models (see chapters 2 and 3) are likely only applicable for either region less affected or region strongly affected by Changjiang discharge, coastal and bottom influences (Gong and Liu, 2003), but not for both regions.

The effects of typhoon on phytoplankton biomass and primary production were determined in term of both spatial average and spatial integration pixel by pixel over the selected region. However, in most cases the discussion emphasizes the spatial integration values rather than spatial average, as we want to know the total of phytoplankton biomass enhanced and total carbon fixed over the entire selected region due to typhoon passages. To minimize cloud effects, we only analyze 8-day means of SeaWiFS Chl-α with percentage of free cloud pixel >75 % of total pixel within selected region.

**Typhoon parameters**

We investigates the enhancements of phytoplankton Chl-α and IPP induced by typhoons that passed over the selected region over the period from 24 June 1998 through 8 November 2004. Actually, there were 38 typhoons passed over the selected region during the period above, however, because of cloud problem, we could only analyzed 11 typhoons.

The upper ocean response to a particular tropical cyclone depends on several parameters involving atmospheric and oceanic variables (e.g., Babin et al., 2004; Price, 1981; Dickey et al., 1998b; Lin et al., 2003). Some of the important atmospheric variables include cyclone size (e.g., radius of storm force winds, radius of cyclone-force winds), strength (wind speed), and transit speed. In this study we thus investigate the relationship between atmospheric parameter of typhoons and Chl-α and IPP.
enhancements in the selected region. We also investigated whether nitracline depth as one of the oceanographic variable related to enhancements. To estimate nitracline depth along the typhoon tracks, we investigated the relationship between bottom depth and nitracline depth derived from the summer 1995 PN line dataset. We found a significant relationship between bottom depth and nitracline depth (Fig.4-2).

We used typhoon wind speed, pressure and transit speed as well as best track data from http://agora.ex.nii.ac.jp/digital-typhoon and http://www.solar.ifas.hawaii.edu/tropical/tropical.html. Since this study on the phytoplankton dynamics driven by typhoon focuses only on the selected region, the typhoon parameters were also averaged when the typhoon track passed over the selected region. The southern oscillation index (SOI) as an index of El Nino/La Nina episodes were extracted from NOAA Climate Prediction Center (http://www.cpc.ncep.noaa.gov/data/indices/).

4.3. Results

4.3.1. Interannual Variation in Summer Chl-α

The spatially and temporally averaged 8-day means Chl-α over the region depicted by Fig.4-1 and over the period from 26 June 1998 through 8 November 2004, gave the overall mean value of 0.50 ± 0.13 mg m⁻³. The summer average for each year of 7-year SeaWiFS operation were showed in Table 4-1.

The highest mean (0.68 ± 0.14 mg m⁻³) Chl-α was observed in 1999. Means of Chl-α in the years 1998 and 2003 also showed higher values (0.52 mg m⁻³) than overall mean value, however they showed noticeable difference in standard deviation. In year 1998, the magnitudes of mean Chl-α exhibited relatively constant variation throughout the summer period, but in the year 2003, it was high in early and late
summer, but low in mid summer. It was also noticeably observed that means of Chl-a in August 1998 and 1999 were higher than means of Chl-a in other years (Fig.4-3).

In general, variations higher than 0.50 mg m\(^{-3}\), resulted from one or more noticeable enhancement(s) of Chl-a in China coastal waters, Changjiang-influenced shelf waters and response to typhoon passages. For examples, Fig. 4-4A and Fig. 4-4B show images of the first 8-day of August 1999 and the third 8-day of September 2000, respectively. The former exhibited the highest spatial average of Chl-a in summer 1999 due to noticeable bloom in the Changjiang adjacent waters. The later exhibited the highest spatial average of Chl-a in summer 2000 due to remarkable enhancement of Chl-a induced by typhoon Saomai.

There was a case that even Chl-a enhancement induced by typhoon was noticeable, spatially averaged Chl-a over the entire region showed only a slight increase. The typhoon-induced Chl-a enhancement was obscured by high Chl-a in continental shelf waters. Such case is showed by the image of the first 8-day of October 2004, the 8-day period within which typhoon Meari passed over the ECS (Fig. 4-4C). It was thus a difficulty to investigate the dynamics of phytoplankton biomass and production driven by typhoon alone if the investigation was carried out over the large region of the ECS. We therefore discharged the high Chl-a influenced by Changjiang discharge by using the isopleth of 0.37 mg m\(^{-3}\) derived from the subtraction of standard deviation (0.13 mg m\(^{-3}\)) from the overall mean (0.5 mg m\(^{-3}\)). This Chl-a isopleth from the climatological summer Chl-a image thus becomes the west border of interest region (region-2, Fig.4-5A,B) for elucidating the effects of typhoon on phytoplankton dynamics. Region-2 was implemented only on the long-term variation analysis, whereas for the short-term
variation, analysis emphasized phytoplankton dynamics in the bloom area induced by Meari as will be described in the following subsection.

4.3.2. Enhancements of Chl-a and Primary Production Induced by Typhoon

Daily variation

Meari one of the most intensive typhoons (category 3 on the Saffir-Simpson scale, average pressure < 964 hPa) in 2004 passed over the region-2 from 25 to 29 September. During its recurling from northwestward to northeastward direction, it traveled very slowly, with the typhoon speed ranged from 1 - 5 knot, and stayed for about 2 days (26-27 September) at the region in which the shelf break (200 m isobath) bends from its eastwestward arc to northeastward arc (Fig.4-6).

A post-typhoon remarkable SST cooling was revealed from 3-day (29 September - 1 October 2004) composite MODIS SST image forming prolonged shape of low SST of about 22°C following Meari best track (Fig.4-7G). In contrast, during pre-typhoon condition, this area was occupied by high SST of about 29°C (Fig.4-7A). Collocated with SST cooling area, Chl-a enhancement was remarkably observed with the concentration in the bloom area was > 1 mg m⁻³, and formed more prolonged shape than SST cooling area (Fig.4-7H). During the pre-Meari period, this area only had Chl-a < 0.2 mg m⁻³ (Fig.4-7B). With the pattern resembles to Chl-a bloom, IPP increased from ~ 500 mgC m⁻² d⁻¹ during pre-Meari, to ~ 1400 mgC m⁻² d⁻¹ during post-Meari conditions (Fig.4-7C and Fig.4-7I).

To see the variations in SST, Chl-a and IPP within scale of days, we analyze daily data at a pixel exhibited the highest Chl-a (1.54 mg m⁻³) at 125°E 26.5°N. Prior to Meari (21 September 2001, the last day before Meari when SeaWiFS could see this
location), the point was characterized by high SST (28.97°C), low Chl-a (0.10 mg m$^{-3}$) and IPP (557 mgC m$^{-2}$ d$^{-1}$). Whereas, after typhoon passage, the point showed the lowest SST (22.29°C) on 29 September 2004, the highest Chl-a (1.54 mg m$^{-3}$) and IPP (1654 mgC m$^{-2}$ d$^{-1}$) on 30 September 2004 (Fig.4-7A). Therefore, the water entrainment by Meari dropped SST by as much as 6.68°C, and nutrients pumped into the euphotic zone by entrainment enhanced Chl-a and IPP by as much as 15 and 3 folds, respectively. There were different time lags of SST cooling and intensive biological processes in response to Meari passage. SST responded rapidly just after Meari passed over the bloom area at the first day (26 September). Phytoplankton seemed to need 4 days in utilizing pumped nutrients to reach intensive biological processes (30 September).

To see how large SST, Chl-a and IPP in the bloom area deviated from surrounding areas, we derived their values from the transect T shown in Fig.4-7G-I. The bloom area of transect T was characterized by low SST with the lowest SST 22.6°C, high Chl-a and IPP of 0.89 mg m$^{-3}$ and 1406 mgC m$^{-2}$ d$^{-1}$, respectively. In contrast, surrounding areas were characterized by high SST > 25°C, low Chl-a and IPP of about 0.20 mg m$^{-3}$ and 600 mgC m$^{-2}$ d$^{-1}$, respectively (Fig.4-8B).

**Interannual variations**

Summing pixel by pixel of each 8-day means SeaWiFS Chl-a and IPP over the region-2, yielded the 7-year averages of total Chl-a and IPP of 52.30 ± 14.83 Mg (Megagram = $10^6$ g) and 222 ± 27 GgC (Gigagram = $10^9$ g), respectively. The mean ± standard deviation of total Chl-a and IPP for each year are shown in Table 4-2.
The averages of total Chl-\(a\) (63.65 Mg) and IPP (239 GgC) in 1999 are the highest among 7-year observations. The averages of total Chl-\(a\) and IPP in years 1999 and 2002 were higher than the 7-year mean.

In general, enhancements of Chl-\(a\) and IPP could be observed in 8-day period within or after which 11 typhoons passed over the region-2 (Fig. 4-9). Considering that the extension of high Chl-\(a\) region from Changjiang discharge was masked, the increases of Chl-\(a\) and IPP due to typhoon passages ranged from 2.32 to 63.61 Mg (mean 20.65 ± 19.48 Mg) and from 2.98 to 99.76 GgC (mean 35.64 ± 30.13 GgC), respectively. The highest enhancements of both Chl-\(a\) and IPP were driven by typhoon Saomai that passed over the region-2 from 13 to 15 September 2000.

We related the enhancements of Chl-\(a\) to typhoon wind speed, typhoon transit speed and typhoon pressure. The enhancements of Chl-\(a\) and IPP were positively correlated with typhoon wind speed, but negatively with typhoon speed and typhoon pressure (Fig.4-10).

4.4. Discussion

4.4.1. Interannual Variation in Summer Chl-\(a\) Related to ENSO

SeaWiFS Chl-\(a\) averaged over the region-1 during mid summer 1998 and 1999 were remarkably higher than overall mean value (0.50 mg m\(^{-3}\)). Ishizaka (2000) has reported a remarkable high SeaWiFS Chl-\(a\) during years 1998 and 1999 in the ECS that may be associated with Changjiang plume variation. Summer flood in 1998 was followed by additional floods in June, July and August 1999 (Guo et al., 2004). In fact, Changjiang discharge showed the highest in August 1998 and July 1999 (Fig.4-11).
The reasons for severe floods during summer 1998 and 1999 vary widely as discussed in several publications. Tao et al. (1998) and Huang et al. (1998) noted that the strong El Nino event during 1997-1998 may be the cause for the summer 1998 flood, whereas the summer 1999 flood might be caused by intensification of La Nina event in the winter 1998 (Chen et al., 2001b). In fact, the high summer Changjiang discharges are generally associated with the cold years (La Nina events), whereas, the low summer discharges are associated with the warm years (El Nino event) (Fig.4-11). From these facts, high horizontal nutrient flux from La Nina-affected severe discharges in summer 1998 and 1999 may fuel phytoplankton growth in the ECS.

4.4.2. Enhancements of IPP and INP by Meari

Considering only the bloom patch bordered by the contour of Chl-a = 1.0 mg m\(^{-3}\), Meari increased IPP from 14.1 GgC (prior to Meari, 18-21 September 2004) to 24.2 GgC (post-Meari, 30 September 2004) (Fig.4-12). Meari thus enhanced IPP by as much as 10.1 GgC during single day. This is about a half of the enhancement by Kai-Tak that passed over the northern South China Sea (SCS) reported by Lin et al., (2003). The lower enhancement by Meari than Kai-Tak is due to the facts that, even Kai-Tak was a moderate category 2 (compare to Meari with category 3) on Saffir-Simpson scale, it moved in a near stationary slow speed (0-2.7 knot, compare to Meari with 1-5 knot) and stayed for 3 days in the northern SCS (compare to Meari with 2-day stay).

With pre- and post-Meari IPP of 14.1 and 24.2 GgC, respectively, we assessed pre- and post-Meari INP were about 2.12 and 12.07 GgC, thus about 10 GgC was enhanced by Meari. This result also showed that IPP enhancement by Meari is actually INP potentially available to be exported out of the euphotic zone.
The IPP enhancement by Meari was actually one of the greatest enhancements among 11 investigated typhoons. There are some probable explanations for great enhancement by Meari as follows; 1) Meari was an intensive typhoon formed under the influence of warm year, 2) Meari traveled slowly during its recurving over the shelf break (200 m depth) where the Kuroshio warm current flows which may give heat energy to intensify Meari. The thermodynamics of a tropical cyclone is ideally like heat engine, running between a warm heat reservoir (the sea, at around 26.85°C) and a cold reservoir (15-18 km up in the tropical troposphere, at -73.15°C, Willoughby, 1999) (Fig.4-13). Tropical cyclone intensity is proportional to the difference in absolute temperature between these two reservoirs. Hurricane Opal (October 1995), and the recent Katrina (August 2005) and Rita (September 2005), the two strongest hurricanes during the last 100 years in the Gulf of Mexico were quickly re-intensified (become category 4-5) shortly after moving over the warm waters of the Gulf of Mexico's "Loop Current". The similar phenomenon also happened during Meari that passed over the Kuroshio warm current in the ECS. Even with slight intensifying and still in the same category 3 on the Saffir-Simpson scale, Meari’s wind speed was strengthened from 80 knot (25 September) to 85 knot (26-27 September) just after passed over the shelf break where the Kuroshio warm current flows. The pressure was also depressed from 950 to 945 hPa (Fig.4-14).

4.4.3. Interannual Variation of Typhoon-enhanced Chl-a, IPP and INP Related to ENSO

Considering only the region-2, the less enhancements of Chl-a and IPP were those induced by Todd (September 1998) and Olga (August 1999), respectively (Fig.4-10). It was not surprising because based on their wind speeds (< 83 knot) and pressure
for Todd (> 980 hPa), those two typhoons were in the category 1 on the Saffir-Simpson scale classified as weak typhoons with minimal damage category. And the great enhancements were those induced by Rammasun (July 2002) and Meari (September 2004), the two typhoon with strong wind speed > 88 knot. It is very interesting to find that the less and great enhancements were those induced by typhoons formed during La Nina and El Nino episodes, respectively (Fig. 4-11). It is worth noting that during El Nino years, tropical cyclones have longer lifetimes and, notably, are more intense, with more category 3–5 on Saffir-Simpson scale. During La Nina years, there is a tendency toward more short-lived tropical cyclones, many of which do not reach typhoon intensity (Camargo and Sobel, 2005). Changes in steering flow such as anomalous anticyclones to the east of Philippines and over the ECS during cold years were also likely to suppress the formation and development of typhoon in the western North Pacific (Chan, 2005). In fact, we found a tendency of decreasing typhoon wind speed when the SOI goes to positive value (cold episode) and significant relationship between SOI and typhoon pressure ($r^2=0.36$, n=41, $p<0.05$) in which the more negative SOI, the lower the pressure (the more the intensity) of typhoon is (Fig. 4-15). The weak water column entrainment by La Nina influenced Todd and Olga could only bring less nutrients from deep layer that in turn only fuels less Chl-$\alpha$ and IPP increases. In contrast, Rammasun and Meari that can be classified as category 3 on the Saffir-Simpson scale (pressure between 945 and 964 hPa and wind speed > 88 knot with extensive damage scale) enhanced remarkably Chl-$\alpha$ and IPP (Fig. 4-10).

However, the greatest enhancement was not induced by typhoon formed during El Nino episode but, by Saomai that formed during cold episode, even if its wind speed slightly slower (~84 knot) than Meari and Rammasun. This suggests us that there
may be another factor determining the Chl-a and IPP enhancements in response to typhoon passages. Nitracline depth is one of well known oceanographic parameter determining the Chl-a and IPP enhancements by typhoon (Babin et al., 2004; Lin et al., 2003). For example, here we compare typhoon Saomai and Meari. Even Saomai had slower wind speed (< 84 knot) than that of Meari (> 88 knot), and both typhoons had the same typhoon speed (1-5 knot), Saomai induced the greatest enhancements (~100 GgC, compare to Meari with the enhancement of 74 GgC). From the bottom depth and nitracline depth relationship derived from PN line datasets, we assess that the nitracline depth in the bloom area induced by Saomai was shallower (40~50 m) than that induced by Meari (60~70 m) (not shown). This suggests that besides typhoon parameters (wind speed, pressure etc.), nitracline depth also played an important role in determining IPP enhancement in response to typhoon passages.

Further, we investigated the relationship between IPP enhancement and nitracline depth for all 11 investigated typhoons. The values of nitracline depths were derived along the typhoon track when passed over the region-2. By plotting all 11 typhoons data, it seemed to be no correlation between nitracline depth and IPP enhancement (Fig.4-16). However, if the typhoons were classified into two groups (thus to some degree typhoon speed influence was removed), the typhoons with slow (<10 knot) and fast (>10 knot) moving speed, the significant negative correlations were emerged (Fig.4-16), suggesting indeed nitracline depth also influence IPP enhancement in response to typhoon.

Employing the relationship between typhoon-induced IPP enhancement and typhoon parameters as well as nitracline depth, we construct a multiple regression
equation of IPP enhancement as a function of wind speed, typhoon speed and nitracline depth with the formulation as;

\[ \text{IPPen} = 191.7 + 1.1 (\text{WS}) - 8.3 (\text{TS}) - 1.9 (\text{ND}) \] (4-1),

where IPPen, WS, TS and ND are IPP enhancement, wind speed, typhoon speed and nitracline depth, respectively. This formulation showed a good performance by explaining 78% \((p<0.01)\) of variation in measured IPP enhancement (Fig.4-17A).

By replacing wind speed in Eq.4-1 with typhoon pressure (TP) the formulation became;

\[ \text{IPPen} = 984 - 0.7 (\text{TP}) - 8.7 (\text{TS}) - 2.0 (\text{ND}) \] (4-2).

This equation could also explain well (73%, \(p<0.01\)) the variation in measured IPP enhancement (Fig.4-17B).

Employing Eq.4-1 (because of higher performance than Eq.4-2), we analyze all typhoons passed over the investigated region during the period from 1998 to 2004 to see the interannual variation of typhoon-induced IPP and INP enhancements. We found that, typhoon contribution to annual and summer IPP and INP were in its lowest in 1998 and in its highest in 2004 (Table 4-3). The least typhoon-induced enhancements and contribution revealed in 1998 were due to only three typhoons occurred with low intensity related to La Nina episode. In contrast the greatest enhancements and contributions in 2004 were associated with nine typhoons with some typhoons had the intensity equal to or even more than Meari, which were related to El Nino episode. Depend on typhoon intensity which is modulated by climate changes, typhoon may contribute within the range of 4.0~35.0% to summer INP, the season considered to have minimum seasonal INP in the investigated region (minimum nitrate upward flux associated with strong stratification).
Despite of the accuracies of models used in this study, and considering the lack of information on the role of typhoon on biogeochemical processes in the ECS, this study however does provide a first order approximation of new production driven by typhoon available to be exported out of euphotic zone in the marginal ECS. This study is also the first one exhibiting the link between changes in atmospheric circulation and the functioning of ocean as biological pump through modulation typhoon activities.

4.5. Summary

The significant achievements that can be summarized are as follows;

1. during two-day stay, Meari dropped SST by as much as 7°C, enhanced Chl-a and IPP by as much as 15 and 3 folds, respectively,

2. the high enhancement induced by Meari was caused by some factors as follows; 1) Meari was a warm phase-affected intensive typhoon, 2) slow movement allowing maximum entrainment, 3) intensification of Meari as it traveled over the Kuroshio warm waters which gave more heat energy,

3. considering only the bloom area, Meari enhanced IPP by as much as 10 GgC during a single day, which was actually the INP available to be exported from the euphotic zone,

4. the enhancements of Chl-a and IPP induced by eleven typhoon passages positively correlated with wind speed, but negatively with typhoon speed and pressure,

5. great (less) enhancements were forced by typhoons formed during warm years (cold years) or El Nino event (La Nina event),

6. the least typhoon contribution to summer INP revealed in 1998 (4%) coincided with La Nina episode, and the greatest observed in 2004 (35%) coincided with El Nino episode.
Table 4-1. Mean and standard deviation of 8-day means of Chl-a spatially averaged over the region-1 for each year of 7-year SeaWiFS observation.

<table>
<thead>
<tr>
<th>Year</th>
<th>Mean (mg m$^{-3}$)</th>
<th>Standard deviation (mg m$^{-3}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>0.52</td>
<td>0.07</td>
</tr>
<tr>
<td>1999</td>
<td>0.68</td>
<td>0.14</td>
</tr>
<tr>
<td>2000</td>
<td>0.46</td>
<td>0.12</td>
</tr>
<tr>
<td>2001</td>
<td>0.37</td>
<td>0.07</td>
</tr>
<tr>
<td>2002</td>
<td>0.50</td>
<td>0.13</td>
</tr>
<tr>
<td>2003</td>
<td>0.52</td>
<td>0.12</td>
</tr>
<tr>
<td>2004</td>
<td>0.43</td>
<td>0.05</td>
</tr>
<tr>
<td>Average</td>
<td>0.50</td>
<td>0.13</td>
</tr>
</tbody>
</table>
Table 4-2. Average ± standard deviation of 8-day means of Chl-$\alpha$ and IPP summed over the region-2 for each year of 7-year SeaWiFS observation.

<table>
<thead>
<tr>
<th>Year</th>
<th>Chl-$\alpha$ (Mg)</th>
<th>IPP (GgC)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>44.34 ± 6.72</td>
<td>216 ± 25</td>
</tr>
<tr>
<td>1999</td>
<td>63.65 ± 14.97</td>
<td>239 ± 29</td>
</tr>
<tr>
<td>2000</td>
<td>54.77 ± 19.96</td>
<td>221 ± 30</td>
</tr>
<tr>
<td>2001</td>
<td>49.08 ± 12.64</td>
<td>217 ± 12</td>
</tr>
<tr>
<td>2002</td>
<td>56.69 ± 9.41</td>
<td>237 ± 28</td>
</tr>
<tr>
<td>2003</td>
<td>50.72 ± 11.16</td>
<td>221 ± 25</td>
</tr>
<tr>
<td>2004</td>
<td>49.06 ± 20.05</td>
<td>200 ± 24</td>
</tr>
<tr>
<td>Mean</td>
<td>52.30 ± 14.83</td>
<td>222 ± 27</td>
</tr>
</tbody>
</table>
Table 4-3. Interannual variations of typhoon numbers, typhoon-induced IPP and INP enhancements, typhoon contributions to annual and summer IPP and INP. White and light grey rows indicate La Nina (cold) and El Nino (warm) episodes, respectively.

<table>
<thead>
<tr>
<th>Year</th>
<th>Typhoon numbers</th>
<th>IPP enhancement (GgC)</th>
<th>INP enhancement (GgC)</th>
<th>Contribution to annual IPP (%)</th>
<th>Contribution to annual INP (%)</th>
<th>Contribution to summer IPP (%)</th>
<th>Contribution to summer INP (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>3</td>
<td>219</td>
<td>110</td>
<td>0.3</td>
<td>1.3</td>
<td>1.0</td>
<td>4.0</td>
</tr>
<tr>
<td>1999</td>
<td>6</td>
<td>693</td>
<td>346</td>
<td>1.1</td>
<td>4.1</td>
<td>3.0</td>
<td>12.1</td>
</tr>
<tr>
<td>2000</td>
<td>6</td>
<td>1298</td>
<td>649</td>
<td>1.9</td>
<td>7.7</td>
<td>5.7</td>
<td>22.6</td>
</tr>
<tr>
<td>2001</td>
<td>3</td>
<td>518</td>
<td>259</td>
<td>0.8</td>
<td>3.1</td>
<td>2.3</td>
<td>9.0</td>
</tr>
<tr>
<td>2002</td>
<td>7</td>
<td>1008</td>
<td>504</td>
<td>1.5</td>
<td>6.0</td>
<td>4.4</td>
<td>17.6</td>
</tr>
<tr>
<td>2003</td>
<td>4</td>
<td>235</td>
<td>118</td>
<td>0.4</td>
<td>1.4</td>
<td>1.1</td>
<td>4.1</td>
</tr>
<tr>
<td>2004</td>
<td>9</td>
<td>1962</td>
<td>980</td>
<td>3.0</td>
<td>11.7</td>
<td>8.6</td>
<td>34.9</td>
</tr>
<tr>
<td>Min</td>
<td>3</td>
<td>219</td>
<td>110</td>
<td>0.3</td>
<td>1.3</td>
<td>1.0</td>
<td>4.0</td>
</tr>
<tr>
<td>Max</td>
<td>9</td>
<td>1962</td>
<td>980</td>
<td>3.0</td>
<td>11.7</td>
<td>8.6</td>
<td>35.0</td>
</tr>
<tr>
<td>Ave.</td>
<td>5.1</td>
<td>813</td>
<td>407</td>
<td>1.2</td>
<td>4.8</td>
<td>3.6</td>
<td>14.2</td>
</tr>
</tbody>
</table>
Fig. 4-1. Map of the area selected for the study of interannual variation in Chl-α over the entire region of the ECS. Area in white (region-1) is the region within which weekly means of Chl-α were spatially averaged, black areas are either lands or masked areas discharged from analysis, grey one is the region with the bottom depth less than 50 m. Bottom depth contours of 50, 100 and 200 m are also shown.
Fig. 4-2. Relationship between bottom depth and nitracline depth derived from summer 1995 PN line dataset (ND = 21.72 LN(BD) – 53.26, $r^2=0.95$, n=11, $p<0.01$). ND and BD are nitracline and bottom depth, respectively.
Fig. 4-3. Interannual variation in spatially averaged weekly means of Chl-a over the entire region of the ECS (Fig. 4-1) for the period from 26 June 1997 through 8 November 2004. Bars in grey color indicate the 8-day period within which typhoons passed over. Typhoon's name is also shown for each bar.
Fig. 4-4. Images of 8-day average Chl-α in the region-1. A, B and C are the first 8-day of August 1999, the third 8-day of September 2000 and the first 8-day of October 2004, respectively. The red lines in b and c are the best tracks of typhoon Saomai and Meari, respectively.
Fig. 4-5. A, Chl-a isopleth of 0.362 mg m\(^{-3}\) (black curve) overlaid on climatological summer mean of Chl-a derived from 8-day means of SeaWiFS Chl-a over the period from 26 June 1997 through 8 November 2004. B, the region-2 selected for elucidating phytoplankton dynamics driven by typhoon alone (white region) which is bordered by Chl-a isopleth of 0.362 mg m\(^{-3}\) in the west side (black curve). Isodepths of 50 m (blue curve) and 100 m (red curve) are also shown.
Fig. 4-6. Best track of typhoon Meari (red cross and line) overlaid on the bathymetry map of the ECS. Black contours represent 50, 100 and 200 m isobaths. Numbers indicate the propagation dates of typhoon Meari and the red numbers indicate the dates when typhoon Meari turned the direction with the slowest typhoon speed (1-5 knot).
Fig. 4-7. SST cooling, Chl-α and IPP enhancements revealed from satellite observations. The first row shows 3-day (18-21 September 2004) composite images of MODIS SST (A), SeaWiFS Chl-α (B) and estimated IPP (C) representing pre-typhoon Meari conditions. The second row shows the positions of Meari on 26, 27 and 28 September 2004. The third row is the same as the first row, except the 3-day (29 September - 1 October 2004) composite representing post-Meari conditions. Meari best track is also shown in the third row (red curve). The white line is transect T selected to derive SST, Chl-α and IPP variations across the typhoon best track.
Fig. 4-8. A, temporal variations in SST, Chl-\(a\) and IPP including the days prior to and after typhoon Meari event. The grey bar indicates the period or dates within which typhoon Meari passed over the region-2. B, spatial variations in SST, Chl-\(a\) and IPP at transect T crossing the typhoon-induced phytoplankton bloom patch as shown in Fig. 4-6G-I.
Fig. 4-9. Interannual variations of typhoon-induced Chl-α (open circles) and IPP (closed circles) enhancements spatially integrated over the region-2. J, A, S, O and N in x-axis are July, August, September, October and November, respectively. Grey bars indicate typhoons passed over the region-2.
Fig.4-10. Relationships between typhoon-induced Chl-a (open circle) and IPP (closed circles) enhancements and typhoon parameters. Td98, Ol99, Sa00 and Mr04 are Todd 1998, Olga 1999, Saomai 2000 and Meari 2004, respectively.
Fig. 4-11. Interannual variation in monthly mean Changjiang discharge (source data: Zhu et al., 2001) and Southern Oscillation Index (SOI) (http://www.cpc.ncep.noaa.gov/data/indices/soi). Dashed lines indicate the months during which 11 typhoons passed over the region-2.
Fig. 4-12. Images of SeaWiFS Chl-a and IPP prior to and post Meari (26-28 September 2004). A and B are the 3-day composite images of Chl-a and IPP, respectively, representing the condition prior to Meari. C and D are Chl-a and IPP images on 30 September 2004, respectively, representing the condition after Meari.
Fig.4-13. Tropical cyclones as thermodynamic engines. Air takes up energy, primarily latent heat stored in water vapour, as it spirals into the lower levels of the vortex under the influence of friction. It converges towards the eyewall (a ring of convective clouds that encloses the clear central eye. As the air ascends to the tropopause (the top of the troposphere, where the temperature decreases with height), the vapour condenses, converting the latent heat into sensible heat which is, in turn, converted to mechanical energy that can do work against friction or strengthen the vortex. The energy realized through this cycle is proportional to the difference in temperature between the ocean at roughly 300 K and the upper troposphere at around 200 K. Storm-induced upward mixing of cooler water reduces the ocean-surface temperature by a few degrees, and can have a considerable effect on the fastest wind speed attainable (after Willoughby, 1999).
Fig. 4-14. Temporal (hour scale) variations of Meari pressure (closed circle) and wind speed (open circle) from 20 to 29 September 2004. Grey bar indicates the time within which Meari passed over the bloom area.
Fig. 4-15. Relationships between SOI and typhoon parameters (typhoon wind speed (A) and typhoon pressure (B)). Regression line for SOI and wind speed relationship is not shown because of statistically insignificant ($p>0.05$).
Fig. 4-16. Relationship between nitracline depth and IPP enhancement induced by typhoon. Closed circle and thin solid line (regression line, $r^2=0.97$, $p<0.05$) were associated with typhoons with slow speed (< 10 knot), open circle and dashed line (regression line, $r^2=0.54$, $p<0.05$) were associated those with fast speed (> 10 knot).
Fig. 4-17. Relationships between measured and estimated IPP enhancements calculated using multiple regression 4-1 (A) and 4-2 (B).
Chapter 5

General Discussion

Satellite-based primary production modeling in the ECS is an important prerequisite to understand the role of the ECS in the biogeochemical processes. Efforts to estimate primary production in the regional scale of the ECS needed to consider vertical profile of Chl-a, as its deep Chl-a maximum contributed significant amount to euphotic depth-integrated primary production.

The empirical model originally developed from the western ECS should be applied with caution in the eastern ECS due to differences in the structure of Chl-a vertical profile. The presence of deep Chl-a maximum was also responsible for the inaccuracy of empirical model. For the complex analytical model, inaccuracy in estimating Chl-a vertical profile, was responsible for reducing the performance of the complex analytical model. On the other hand, global model with locally adjusted photosynthetic parameter, even was originally developed from the open ocean, revealed a fair accuracy in predicting primary production in the eastern ECS. However, to reach full applicability for the entire region of the ECS including coastal western and eastern ECS, such a model seemed to be challenged by providing accurate model for its photosynthetic parameter, as well as the problem in retrieving Chl-a from ocean color satellite.

Primary production model and satellite data have been proved as a useful tool to assess primary production enhancement and potential carbon flux induced by typhoon passages. This study showed that primary production enhancements were clearly observed in response to typhoon passages. Besides, typhoon parameters (typhoon wind speed, typhoon speed and pressure), pre-typhoon nitracline depth was
also played an important role in determining primary and new productions enhancements by typhoon. This study is also the first one which demonstrated the link between upper ocean biological productivity in response to typhoon passage and global climate changes (El Nino/La Nina episodes).
Chapter 6

General Conclusions

As a straight line of general conclusions summarized from this study we draw some main points as follows;

1) Attempts to model primary production based on satellite data in the ECS should consider the spatial variabilities in model variable inputs such as Chl-a vertical profile and photosynthetic parameter and the optical property of the region,

2) Primary production model combine with satellite data have been proved as a useful tool to provide a synoptic measurement of primary production allowing us to elucidate its spatial and temporal variation forced by typhoon passages,

3) This study also showed the probable remote forcing by global climate changes (El Nino/La Nina events) in driving biological dynamics in the ECS through typhoon passages.
References


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