

DEVELOPMENT OF CHLOROPHYLL ALGORITHM FOR GEOSTATIONARY OCEAN COLOR IMAGER (GOCI)

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ABSTRACT ... Chlorophyll concentration is an important factor for physical oceanography as well as biological oceanography. For these necessity many oceanographic researchers have been investigated it for a long time. But investigation using vessel is very inefficient, on the other hands, ocean color remote sensing is a powerful means to get fine-scale (spatial and temporal scale) measurements of chlorophyll concentration. Geostationary Ocean Color Imager (GOCI), for ocean color sensor, loaded on COMS (Communication, Ocean and Meteorological Satellite), will be launched on late 2008 in Korea. According to the necessity of algorithm for GOCI, we developed chlorophyll algorithm for GOCI in this study. There are two types of chlorophyll algorithms. One is an empirical algorithm using band ratio, and the other one is a fluorescence-based algorithms. To develop GOCI chlorophyll algorithm empirically we used bands centered at 412 nm, 443 nm and 555 nm for the DOM absorption, chlorophyll maximum absorption and for absorption of suspended solid material respectively. For the fluorescence-based algorithm we analyzed in-situ remote sensing reflectance (R_{rs}) data using baseline method. Fluorescence Line Height (ΔFlu) calculated from R_{rs} at bands centered on 681 nm and 688 nm, and $\Delta Flu_{(area)}$ are used for development of algorithm. As a result $\Delta Flu_{(area)}$ method leads the best fitting for squared correlation coefficient (R^2).

KEY WORDS: Chlorophyll Concentration Algorithm, Ratio algorithm, Fluorescence-based algorithm, Geostationary Ocean Color Imager (GOCI)

1. INTRODUCTION

The influence of phytoplankton on the color of seawater has been studied for several decades. It is well understood that chlorophyll-a, the primary photosynthetic pigment in phytoplankton, absorbs relatively more blue and red light than green, and the spectrum of backscattered sunlight or color of ocean water progressively shifts from deep blue to green as the concentration of phytoplankton increases.

Light energy absorbed by the antennae systems of photosystem I (PS I) and photosystem II (PS II) can undergo any one of the three fates: it can be used for photosynthesis or excess energy can be dissipated as heat or it can be re-emitted as chlorophyll-a (Chl-a) fluorescence. The re-emitted Chl-a fluorescence originates mostly from PS II (about 95%), the contribution from PS I being negligible. The amount of Chl-a fluorescence is very small (2%) and is dependent on the absorption of light by chlorophyll molecules and the rate of transfer of electrons between PS II and PS I. In fact, the ratio of the amount of photons emitted as fluorescence to the total number of absorbed photons defines the yield of fluorescence, which provides information concerning changes in the efficiency of photochemistry and heat dissipation (Maritorena et al., 2000).

Although the total amount of Chl-a fluorescence is small, measurement of this signal is rather easy because the spectrum of fluorescence is different from that of the

absorbed light, with the peak of fluorescence emission being of longer wavelength than that of absorption. For the first time, Tyler and Smith in 1967 noticed a distinct, Gaussian-shaped red peak in the upwelling irradiance measured in the San Vicente reservoir near San Diego, California (Tyler and Smith, 1970). Several other researchers in the later years repeatedly detected this feature around 683-685 nm in their radiance or reflectance measurements in oceanic waters (Neville and Gower, 1977; Gower and Borstad, 1987; Gitelson, 1992; Abbott and Letelier, 1999; Ahn and Shanmugam, 2007). The peak around this wavelength was indeed found closely correlated with the in-situ Chl-a concentrations, representing an important complement to remotely estimate phytoplankton pigment concentrations in natural water bodies.

2. METHOD

2.1 In-situ bio-optical datasets

In-situ bio-optical measurements were performed in Korean and neighboring sea waters through the years 1999-2007 onboard the research vessels EARDO, Olympic and so on. During each cruise, the water samples were collected with buckets/Niskin bottles and sub samples were filtered onboard onto Whatman GF/F glass microfiber filters for measurement of chlorophyll (Chl) and suspended sediments (SS) concentrations.

These filters were frozen by liquid nitrogen at dark conditions until analysis took place at the laboratory with the standard spectrophotometric method for determining Chl and oven-drying method for SS (Jeffrey and Humphrey, 1975; Ahn et al., 1999). The dissolved organic matter (DOM) was analyzed spectrophotometrically based on the procedure given in Ahn et al. (1999). Simultaneously, radiometric measurements such as down-ward spectral irradiance ($E_d(\lambda)$) and total water leaving radiance ($L_{wT}(\lambda)$) and sky radiance ($L_{sky}(\lambda)$) were performed at various sample sites using an ASD FieldSpec Pro Dual VNIR Spectroradiometer with spectral range from 350-1050 nm and spectral sample interval of 1.4 nm. This instrument was calibrated several times to confirm its stability. Most of these measurements were generally made in near-solar noon and under nearly cloudless conditions. The data recorded in units of $mW\ cm^{-2}\ \mu m^{-1}\ sr^{-1}$ needed to be corrected for the contribution of skylight reflection and air-sea interface effects. Thus, the measured $L_{wT}(\lambda)$ was corrected for the sky light reflection and the air-sea interface effects using the following formula,

$$L_w(\lambda) = L_{wT}(\lambda) - F_r(\lambda) \times L_{sky}(\lambda) \quad (1)$$

The values of $L_{sky}(\lambda)$ were obtained from the sky radiometer and F_r value was assumed to be constant 0.025 (Austin, 1974). The remote sensing reflectance ($R_{rs}(\lambda)$) was obtained by normalizing the corrected water-leaving radiance ($L_w(\lambda)$) to spectral down-welling irradiance (E_d) as follows,

$$R_{rs}(0^+ \lambda) = \frac{L_w(\lambda)}{E_d(0^+ \lambda)} \quad (2)$$

The $R_{rs}(\lambda)$ can also be obtained by using $L_w = 0.54 L_u(0^+)$, and $E_d(0^+) = 0.96 E_d(0^+)$, where 0.54 is a mean coefficient summarizing the effect of internal reflection of the up-welling radiance flux during transmission through the interface, and 0.96 accounts for the loss of the down-welling flux by reflection at the air-sea interface (Austin, 1974). From remote sensing reflectance the normalized water-leaving radiance (L_{wN}) may be obtained (as the water-leaving radiance can be changed with light field conditions of surface layer of the ocean) using the following relation

$$L_{wN}(\lambda) = L_w(\lambda) \frac{F_0(\lambda)}{E_d(\lambda)} \quad (3)$$

where F_0 is the mean extraterrestrial solar irradiance at a given spectral band.

2.2 Theoretical background of chlorophyll algorithm

2.2.1 Band ratio algorithm: For the open ocean, NASA-SeaWiFS operational algorithms (OC2v4 and OC4v4; O'Reilly et al., 1998, 2000) are generally used by lots of ocean color researchers. The functional forms of these algorithms are:

$$\langle chl \rangle = 10^{(a_0 + a_1 R + a_2 R^2 + a_3 R^3)} + a_4 \quad (4)$$

for the OC2v4,

$$\langle chl \rangle = 10^{(a_0 + a_1 R + a_2 R^2 + a_3 R^3 + a_4 R^4)} \quad (5)$$

for the OC4v4.

In Eq. (4), R is the \log_{10} of the ratio between remote sensing reflectances, R_{rs} , measured at 490 and 555 nm. For the OC4v4 (Eq. (5)), R is the \log_{10} of the ratio of R_{rs} measured at 443 and 555, 490 and 555, or 510 and 555 nm, depending on its value (the maximum is chosen). Coefficient values for OC2v4 are 0.319, -2.336, 0.879, -0.135 and -0.071 respectively. And in Eq. (5), coefficient values are 0.366, -3.067, 1.930, 0.649 and -1.532 respectively.

2.2.2 Fluorescence algorithm: Fluorescence algorithms are simplistic in their nature by neglecting the complex algal physiological and irradiance effects (but less when compared to uncertainties associated with the blue-green ratio and atmospheric correction algorithms) on fluorescence and are based on the following parameters: (1) the height of the fluorescence peak (also referred to as the Fluorescence Line Height - FLH) at 681, and (2) the area delimited by the ΔFlu curve ($\Delta Flu_{(area)}$). These parameters can be achieved by constructing a linear baseline between troughs at 660 and 730 nm (regions of significant absorption by Chl-a and seawater), on either side of the fluorescence peak in the R_{rs} spectrum. At these wavelengths the reflectances are less sensitive to Chl-a concentrations, but largely depend on the concentrations of inorganic sediments and other detritus matters from which the scattered photons might induce a change in the baseline with minimal influence on the area and height of the 688 nm peak above the baseline. Thus, the height of the peak and the area above the baseline between 660 nm and 730 nm largely depend on phytoplankton concentration and can be used as their quantitative measure of Chl-a in ocean waters.

Fluorescence Line Height (FLH) represents a simple mathematical form of spectral bands that provide an apparent peak above the linear baseline between 660 and 730 nm, which is a relative measure of the amount of radiance leaving the sea surface in the Chl-a fluorescence emission band (presumably a result of chlorophyll fluorescence). The FLH is measured by defining the baseline underneath the fluorescence curve and subtracting it from the peak values as follows,

$$\Delta Flu = XY = CX - YC \quad (6)$$

which can be expanded to give

$$\Delta Flu = XY = CX - \left[\left(\frac{R_L - R_S}{\lambda_{(L)} - \lambda_{(S)}} (\lambda_{(F)} - \lambda_{(S)}) \right) + R_S \right] \quad (7)$$

where ΔFlu is the FLH denoted by XY , CX is the fluorescence peak signal, and YC is the fluorescence base (Fig. 1). R_S and R_L refer to the remote sensing reflectance at shorter (660) and longer (730) wavelengths respectively, while λ_S and λ_L are the shorter and longer wavelengths respectively, and λ_F is the fluorescence wavelength.

ΔFlu curve above the baseline represents a proportion of the fluorescence energy emitted weighed by the angular acceptance of the radiometer and the attenuation coefficients of the water for excitation and emission wavelengths, while the background below this line originates from Raman plus elastically backscattered radiation. The integral of the fluorescence emission spectrum above the baseline leads to the $\Delta Flu_{(area)}$ in Eq. (5),

$$\Delta Flu_{(Area)} = \int_{660}^{730} \Delta Flu(\lambda) d\lambda \quad (8)$$

Normalizing this by the integral of absorption within the excitation (blue and green) spectral domain leads to the quantum yield of Chl-a fluorescence. The $\Delta Flu_{(area)}$ parameter is important to overcome some problems with the former two that become either saturated at high Chl-a concentrations or weak in phytoplankton-less oceanic waters. This would result in an observed non-linear relationship between the FLH and Chl-a concentrations (Huot et al., 2005). The most likely mechanism for saturation of FLH at very high Chl-a is reabsorption of fluoresced photons, which depends on the absorption efficiency of the cell and, therefore, on its size and internal pigment content (Morel and Bricaud, 1981).

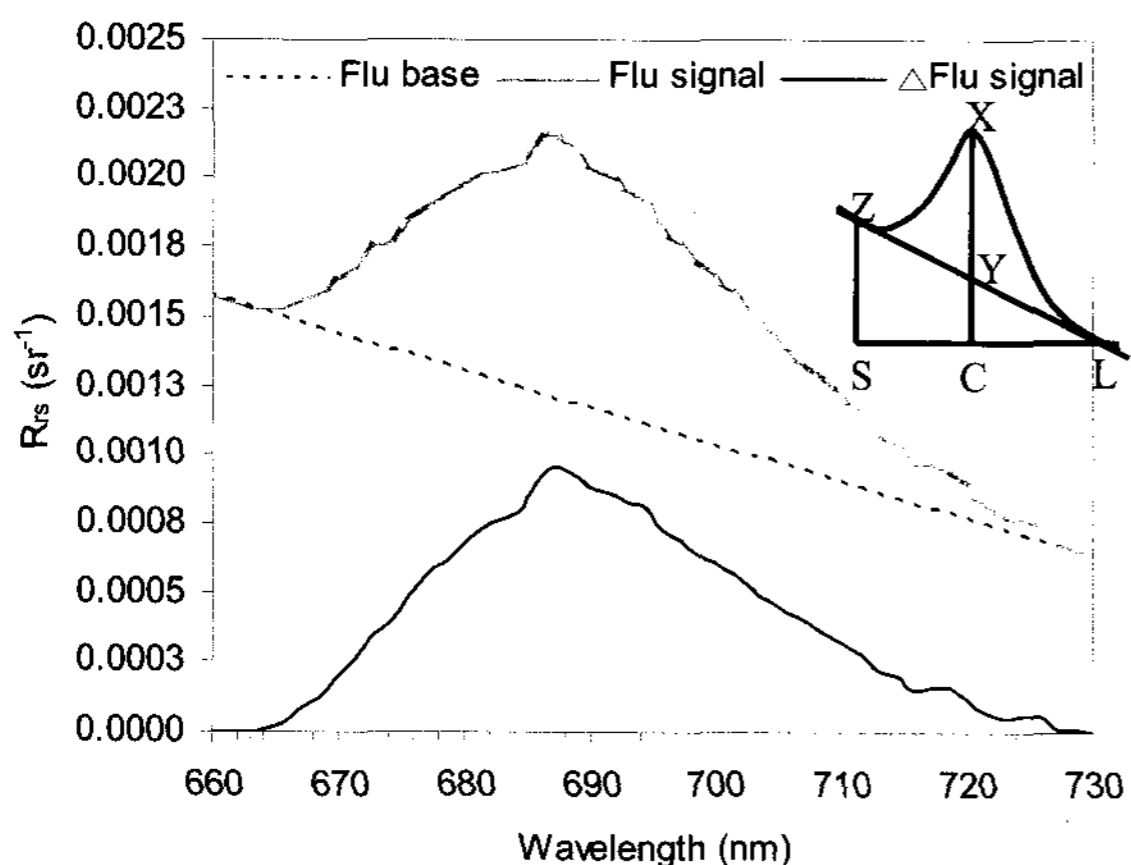


Figure 1. Schematic representation of the fluorescence line height (ΔFlu) estimation using the R_{rs} spectrum

3. RESULTS

3.1 Simple ratio algorithm

The basic form of GOCI band ratio chlorophyll algorithm is like blow.

$$\langle chl \rangle = A * \left(\frac{(R_{rs}(443) + R_{rs}(490)) - R_{rs}(412)}{R_{rs}(555)} \right)^B \quad (8)$$

where, $R_{rs}(412)$, $R_{rs}(443)$, $R_{rs}(490)$ and $R_{rs}(555)$ are R_{rs} values measure at 412, 443, 490 and 555 nm respectively. And A and B are coefficients.

To develop for GOCI chlorophyll band ratio algorithm we used in-situ chl-a concentration and R_{rs} data obtained on the Korean and neighboring sea waters from 1999 to 2007. Fig. 2 shows the statistical relationship between in-situ Chl-a concentrations ($mg\ m^{-3}$) and combination of R_{rs} values ($R_{rs}(412)$, $R_{rs}(443)$, $R_{rs}(490)$ and $R_{rs}(555)$) ($N=203$). The regressions against in-situ Chl-a concentrations yielded the squared correlation coefficients (r^2) 0.74.

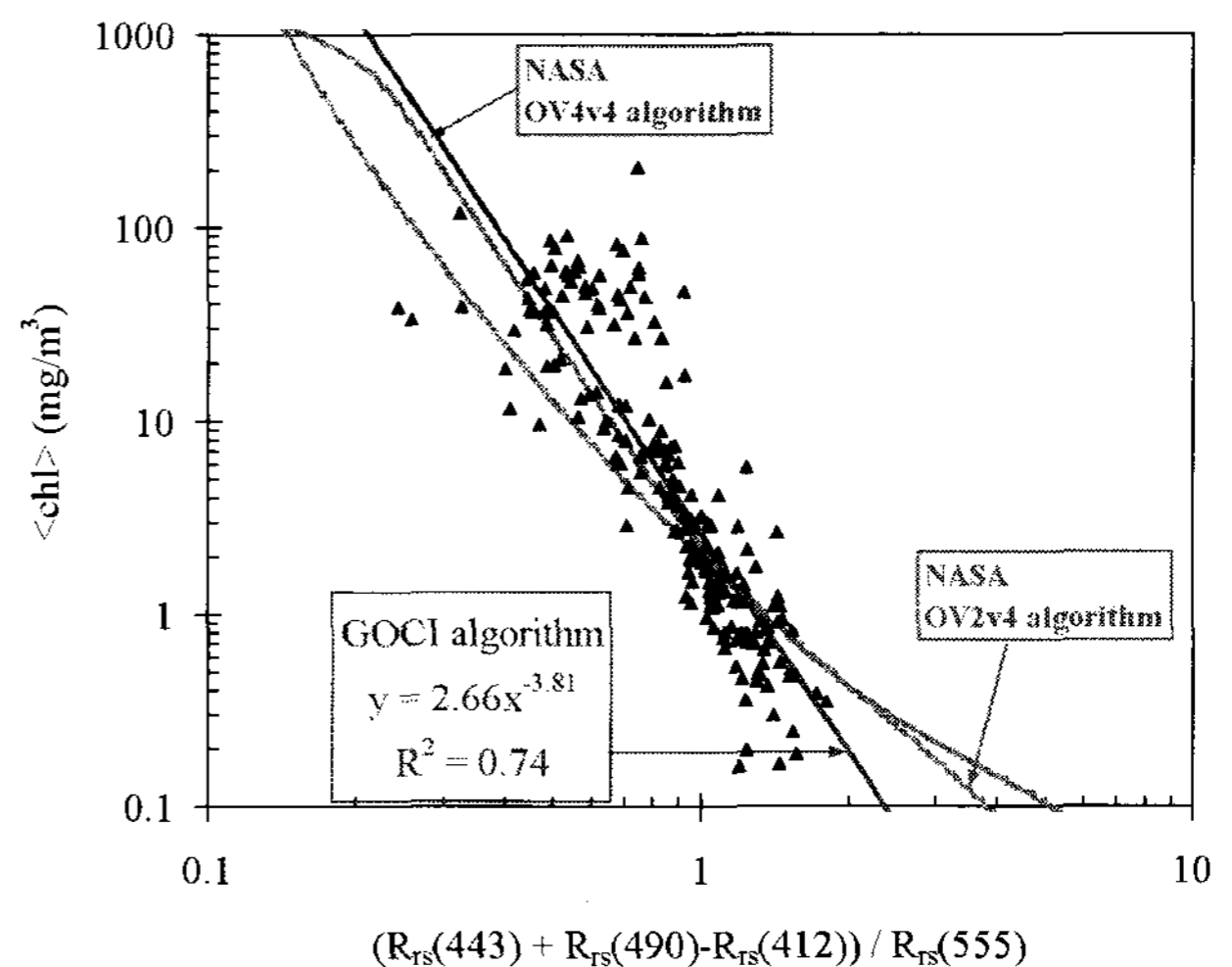


Figure 2. Relationship between in-situ Chl-a concentrations ($mg\ m^{-3}$) and combination of R_{rs} values ($R_{rs}(412)$, $R_{rs}(443)$, $R_{rs}(490)$ and $R_{rs}(555)$). The solid line is the best-fit regression to our bio-optical dataset ($N=203$).

3.2 Fluorescence algorithm

Fig. 3a and b show the statistical relationships of the log-transformed $\Delta Flu(680)$ and $\Delta Flu_{(area)}$ versus log-transformed in-situ Chl-a concentrations ($N=203$). The regressions against in-situ Chl-a concentrations yielded the squared correlation coefficients (r^2) 0.88 for the $\Delta Flu(680)$ and 0.66 for the $\Delta Flu_{(area)}$. It should be noted that the area-based fluorescence algorithm remains effective even if there are changes in the physiological aspects of phytoplankton that can cause the variability in the magnitude and spectral-position of the fluorescence peak. Since the GOCI avoids inclusion of bands that are susceptible to the pronounced O_2 absorption feature at

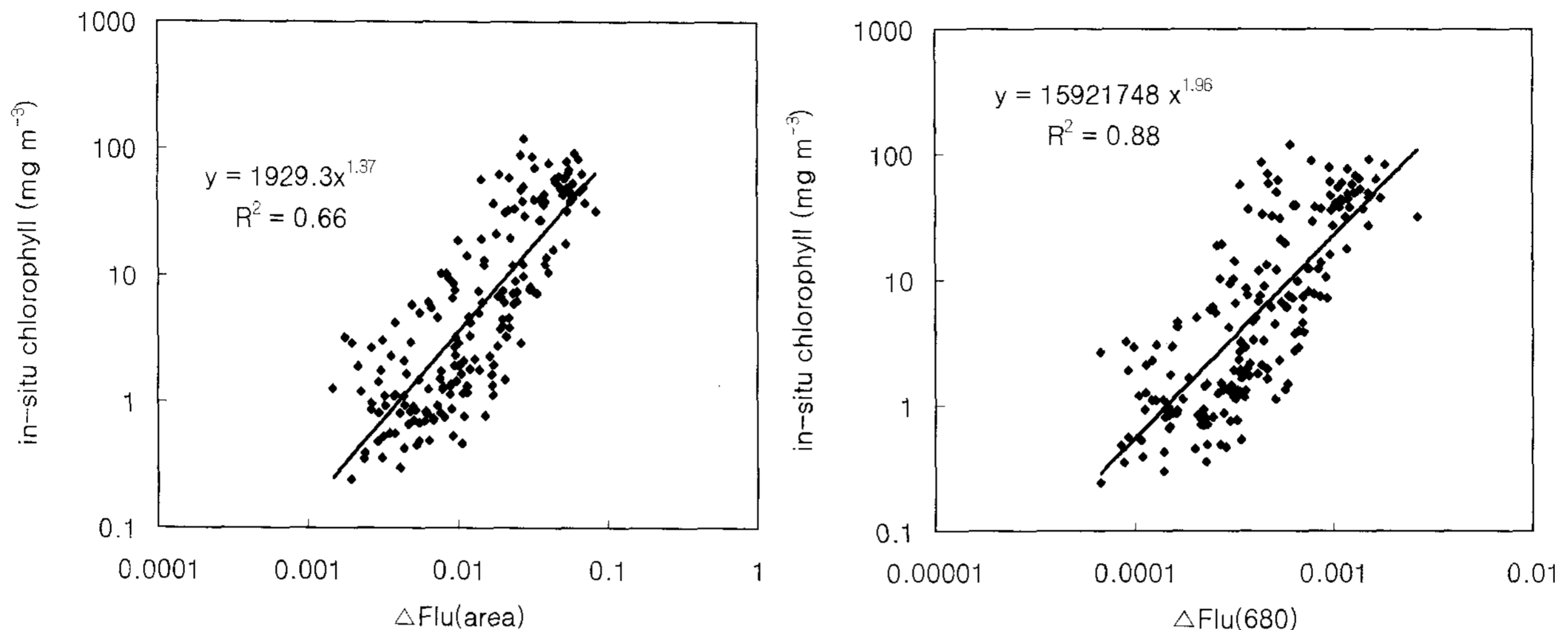


Figure 3a and b. Relationships between in-situ Chl-a concentrations (mg m^{-3}) and $\Delta\text{Flu}_{(\text{area})}$ and $\Delta\text{Flu}(680)$. The solid line is the best-fit regression to our bio-optical dataset ($N=203$).

687 nm and H_2O absorption feature at 730 nm (which erode the long wavelength portion of the fluorescence peak), the above parameters can be accurately estimated and used to quantify phytoplankton pigment concentrations in Case II waters.

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REFERENCE

Abbott, M.R., P.J. Richerson, and T.M. Powell, 1982. In situ response of phytoplankton fluorescence to rapid variations in light, *Limnology and Oceanography*, 27, pp. 218-225.

Ahn, Y.H et al., 1999. Development of redtide and water turbidity algorithms using ocean color satellite. KORDI Report, BSPE 98721-00-1224-01.

Ahn, Y.H., and P. Shanmugam, 2007. Derivation and analysis of the fluorescence algorithms to estimate chlorophyll a concentrations in ocean waters, *Journal of Optics A: Pure and Applied Optics*, 9, pp 352-362.

Austin, R.W., 1974. Inherent spectral radiance signatures of the ocean surface, In: Ocean color analysis. Scripps Institute of Oceanography. La Jolla. CA.

Gitelson, A., 1992. The peak near 700nm on reflectance spectra of algae and water: relationships of its magnitude and position with Chl-a concentration, *International Journal of Remote Sensing*, 13, pp 3,367-3,373.

Gower, J.F.R., and G.A. Borstad, 1987. On the use of solar-stimulated fluorescence signal from chlorophyll a for airborne and satellite mapping of phytoplankton, *Advanced Space Research*, 7, pp 101-106.

Huot, Y., C.A. Brown, and J.J. Cullen, 2005. New algorithms for MODIS sun-induced chlorophyll fluorescence and a comparison with present data products, *Limnology and Oceanography*, 3, pp 108-130.

Jeffrey, S.W., and G.F. Humphrey, 1975. New spectrophotometric equations for determining chlorophyll a, b and c in higher plants, algae and natural phytoplankton. *Biochimie Physiologie Pflanzen*, 167, pp. 374-384.

Maritorena, S., A. Morel, and B. Gentili, 2000. Determination of the fluorescence quantum yield by oceanic phytoplankton in their natural habitat. *Applied Optics*, 39, pp. 6,725-6,737.

Morel, A., and A. Bricaud, 1981. Theoretical results concerning light absorption in a diverse medium, and application to specific absorption of phytoplankton. *Deep Sea Research*, 28, pp. 1,375-1,393.

Neville, R.A., and J.F.R. Gower, 1977. Passive remote sensing of phytoplankton via chlorophyll a fluorescence, *Journal of Geophysical Research*, 82, pp 3,487-3,493.

O' Reilly, J. E., Maritorena, S., Mitchell, B. G., Siegel, D. A., Carder, K. L., Garver, S. A., Kahru, M. and McClain, C., 1998. Ocean color chlorophyll algorithms for SeaWiFS, *Journal of Geophysical Research*, 103, pp. 24,937-24,953.

Tyler, J.E., and R.C. Smith, 1970. Measurements of spectral irradiance under water, *Ocean series (Gordon and Breach, New York)*, 1, pp 103.