Remote sensing of the absorption coefficients and chlorophyll a concentration in the U.S. southern Middle Atlantic Bight from SeaWiFS and MODIS-Aqua

Xiaoju Pan1, Antonio Mannino1, Mary E. Russ1,2, Stanford B. Hooker1

1NASA Goddard Space Flight Center, Greenbelt, Maryland, USA
2University of Maryland, Baltimore County, Goddard Earth Sciences and Technology Center

Accepted by the Journal of Geophysical Research – Oceans September 2008

Index terms:
0480 Remote sensing
0422 Bio-optics
4219 Continental shelf and slope processes
4805 Biogeochemical cycles, processes, and modeling

Key Words: absorption, chlorophyll, remote sensing, SMAB, SeaWiFS, MODIS

Corresponding Author:
Xiaoju Pan
NASA Goddard Space Flight Center, Mail Code 614.7, Build 22 Room 248
Greenbelt, Maryland 20771, USA
Phone: +1 301 286 8966
FAX: +1 301 286 5337
E-mail: xpan@neptune-web.gsfc.nasa.gov
Abstract

At present, satellite remote sensing of coastal water quality and constituent concentration is subject to large errors as compared to the capability of satellite sensors in oceanic waters. In this study, field measurements collected on a series of cruises within U.S. southern Middle Atlantic Bight (SMAB) were applied to improve retrievals of satellite ocean color products in order to examine the factors that regulate the bio-optical properties within the continental shelf waters of the SMAB. The first objective was to develop improvements in satellite retrievals of absorption coefficients of phytoplankton ($a_{ph}$), colored dissolved organic matter (CDOM) ($a_{g}$), non-pigmented particles ($a_{d}$), and non-pigmented particles plus CDOM ($a_{dg}$), and chlorophyll $a$ concentration ($[\text{Chl}_a]$).

Several algorithms were compared to derive constituent absorption coefficients from remote sensing reflectance ($R_{rs}$) ratios. The validation match-ups showed that the mean absolute percent differences (MAPD) were typically <35%, although higher errors were found for $a_{d}$ retrievals. Seasonal and spatial variability of satellite-derived absorption coefficients and $[\text{Chl}_a]$ was apparent and consistent with field data. CDOM is a major contributor to the bio-optical properties of the SMAB, accounting for 35-70% of total light absorption by particles plus CDOM at 443 nm, as compared to 30-45% for phytoplankton and 0-20% for non-pigmented particles. The overestimation of $[\text{Chl}_a]$ from the operational satellite algorithms may be attributed to the strong CDOM absorption in this region. River discharge is important in controlling the bio-optical environment, but cannot explain all of the regional and seasonal variability of biogeochemical constituents in the SMAB.
1. Introduction

Satellites such as the Sea-viewing Wide Field-of-view Sensor (SeaWiFS) and the MODerate-resolution Imaging Spectrometer (MODIS) have been widely applied to the study of biogeochemical processes [IOCCG, 1999; McClain et al., 2004]. Based on bio-optical theory, the satellite measurement, here remote sensing reflectance ($R_s$), is often related to inherent optical properties (IOPs) such as the absorption coefficient ($a$) and scattering coefficient ($b$) [Garver and Siegel, 1997; Gordon et al., 1988; Kirk, 1994; Maritorena et al., 2002; Mobley, 1994]. IOPs are often related to relevant biogeochemical constituents such as chlorophyll $a$ concentration ([Chl$_a$]), dissolved organic carbon (DOC), particulate organic carbon (POC), and suspended sediment [Ferrari et al., 2003; Mannino et al., 2008; Rochelle-Newall and Fisher, 2002; Siegel et al., 2002].

Although chlorophyll $a$ pigment plays a critical role in understanding the bio-optical properties in oceanic waters, it is insufficient to fully characterize the biogeochemical properties, especially in coastal waters where colored dissolved organic matter (CDOM) and sedimentary resuspended matter often overwhelm phytoplankton in the contribution to bio-optical properties [Gordon and Morel, 1983; IOCCG, 2006; Kirk, 1994; Mobley, 1994]. In general, IOPs are composed of four components: pure water, phytoplankton, CDOM, and non-pigmented particles [Kirk, 1994; Mobley, 1994]. Absorption from components other than pure water is often considered to be strongly correlated to [Chl$_a$] in oceanic Case 1 waters, while such an assumption often breaks down in Case 2 waters (e.g. coastal and inland waters) [Gordon and Morel, 1983; IOCCG, 2000; Kirk, 1994; Mobley, 1994].
In theory semi-analytic (SA) models, which apply spectral deconvolution, are applicable to retrieve constituent IOPs from $R_{rs}$ [IOCCG, 2006]. For instance, the GSM01 model [Garver and Siegel, 1997; Maritorena et al., 2002] produces [Chl$_a$], absorption coefficient of CDOM and non-pigmented particles ($a_{dg}$), and particulate backscattering coefficient ($b_{bp}$). Unfortunately there are at least two problems with SA models applied to coastal waters. First, SA models require detailed knowledge of IOP relationships, which vary regionally or seasonally in coastal waters and in fact are regionally specific or empirically derived [Babin et al., 2003a, 2003b; Magnuson et al., 2004]. Second, SA models are equally sensitive to signals at all wavelengths and require them to be equally accurate. In coastal waters the satellite-derived water-leaving radiances ($L_w$) at shorter wavelengths (e.g. 412 and 443 nm) often contain some error. Incorrect atmospheric correction due to inadequate information on aerosol absorption and the selection of inappropriate aerosol model, along with the weak signal-to-noise ratio due to strong CDOM absorption, often causes errors on $L_w$ derivation in coastal waters [Bailey and Werdell, 2006; IOCCG, 2000; Siegel et al., 2000, 2005]. Consequently, the application of SA models in coastal waters faces a significant challenge because of the requirement of highly accurate $L_w$ at all wavelengths.

Empirical algorithms do not require a full understanding of fundamental bio-optical theory. For example, they provide a direct link between satellite-sensed radiance and relevant bio-optical parameters such as [Chl$_a$] and diffuse attenuation coefficient ($K_d$) on global and regional scales [Harding et al., 2005; Mueller, 2000; O’Reilly et al., 1998, 2000; Signorini et al., 2005]. The creation of empirical algorithms, however, requires a sufficient size of highly accurate field measurements spanning all seasons and
adequate spatial coverage for the regions of interest. Thus, empirical algorithms are subject to updates as the dataset increases in size. In the work presented here, a set of self-consistent field $R_{ss}$ data is applied to derive absorption coefficients of oceanic constituents in the U.S. southern Middle Atlantic Bight (SMAB). Absorption coefficients are very important bio-optical properties in the study of radiative transfer modeling and heat budget [Mobley, 1994], carbon flux (e.g. [Chl_a], primary production, DOC, and POC) [Arrigo and Brown, 1996; Behrenfeld et al., 2005; IOCCG, 2006; Mannino et al., 2008; Marra et al., 2007; Rochelle-Newall and Fisher, 2002], water quality (e.g. diffuse attenuation coefficient) [Mueller, 2000], and oceanic physical processes (e.g. salinity distribution) [Rochelle-Newall and Fisher, 2002]. The objectives of this work were: 1) to develop and validate satellite algorithms in deriving constituent absorption coefficients for phytoplankton, non-pigmented particles, and CDOM and [Chl_a] near the ocean surface within the SMAB to within ±35% uncertainty, 2) to determine the relative importance of phytoplankton, CDOM, and detritus in sunlight absorption, and 3) to evaluate the seasonal and regional impacts of river discharge on biogeochemical constituents in the SMAB.

2. Methods

2.1. Study region and field experiments

This study focuses on the SMAB from the Delaware Bay (DB) mouth to the region south of the Chesapeake Bay (CB) mouth (Figure 1). This region is well recognized for the significant impacts by riverine discharge from the Delaware and Chesapeake Bays, which account for most of the salinity variability of the SMAB [Acker
Multiple cruises were conducted in this region during 2004-2006, including the Bio-physical Interactions in Ocean Margin Ecosystems cruises (BIOME) during 30 March to 1 April 2005 (BIOME1), 26 to 30 July 2005 (BIOME2), 9 to 12 May 2006 (BIOME3), and 2 to 6 July 2006 (BIOME4), and the Chesapeake Bay Plume cruises (CBP) during 27 May and 3 November 2005, and 6 September and 28 November 2006, and the Chesapeake Bay Hydrological survey (CBH) during 5 May, 5 July, 1 September, 15 October, and 15 November 2004, and 10 January, 26 May, 21 June, 19 August, and 23 September 2005 (Figure 1). The collected bio-optical data included, but not limited to, phytoplankton pigments, IOPs (absorption), and $R_s$ (only on BIOME and CBP cruises in 2005). Water samples were collected at multiple depths with Niskin bottles.

2.2. Pigments and absorption coefficient

Pigment samples were collected on 25 mm GF/F filters under a gentle vacuum (<5 in Hg) and stored in liquid nitrogen in the field before transfer to a -80°C freezer in the laboratory. Pigments were analyzed at Horn Point Laboratory by reverse-phase high-performance liquid chromatography (HPLC) with a C8 column on the HPLC system equipped with photodiode array detector [Van Heukelem and Thomas, 2001].
was calculated as the sum of concentration from monovinyl Chl_a, divinyl Chl_a, and chlorophyllide a.

Absorption coefficients of particles ($a_p$) and non-pigmented particles ($a_d$) were determined by the quantitative “filter pad” method following the recommendations of Mitchell et al. [2002]. Particulate samples were collected on 25 mm GF/F filters under a gentle vacuum (<5 in Hg) and stored in liquid nitrogen in the field before transfer to a -80°C freezer in the laboratory. Non-pigmented particulate samples were defined as the detritus component of particulate samples after two cold methanol extractions (first 5 ml for 10 minutes, then 10 ml for 1 hour) [Kishino et al., 1985]. Artificial seawater pre-filtered through 0.2 µm Whatman Nuclepore filters was used to rinse off methanol and to hydrate the GF/F filters of both blanks and samples. CDOM samples were collected by filtering seawater through pre-combusted (6 hours at 450°C) GF/F filters and stored under refrigeration (4 to 8 °C). In the laboratory, CDOM samples were warmed to room temperature and filtered through 0.2 µm Whatman Nuclepore (polycarbonate) or Gelman Supor (polyethersulfone) filters prior to analysis [Mannino et al., 2008].

Absorbance spectra were measured using a double-beam Cary 100 Bio Ultraviolet-Visible scanning spectrophotometer through 250-800 nm (CDOM in Suprasil quartz 10 cm pathlength cells) or 300-800 nm (particles) in 1 nm intervals. Blank GF/F filters hydrated with 0.2 µm pre-filtered artificial seawater and ultraviolet (UV) oxidized Milli-Q water were used as the blank and reference for particulate absorbance and CDOM absorbance, respectively. Null correction was made by subtracting the mean of absorbance at 790 to 800 nm for particulate samples for each spectrum. No null correction for CDOM absorption was made since the raw absorbance of seawater samples
at 690-700 nm was within the noise level of the instrument [Mannino et al., 2008]. The multiple-scattering effect for calculating particulate absorption coefficient \( a_p \) and non-pigmented particulate absorption coefficient \( a_d \) was corrected following the method of Mitchell [1990], from which non-pigmented particulate samples were assumed to have the same multiple-scattering amplification factor to total particulate samples [Mitchell et al., 2002]. Phytoplankton absorption coefficient \( a_{ph} \) was calculated as \( a_{ph} = a_p - a_d \).

The absorption coefficient by CDOM and non-pigmented particles \( a_{dg} \) was calculated as the sum of \( a_d \) and CDOM absorption coefficient \( a_g \). Total absorption coefficient \( a \) was calculated as \( a = a_w + a_p + a_g \), where pure water absorption coefficient \( a_w \) was adopted from Pope and Fry [1997].

The absorption coefficient from non-pigmented particles, CDOM, or their sum \( a_x \) was fitted to an exponential function as:

\[
a_x(\lambda) = a_x(\lambda_0) \exp[-S_x(\lambda - \lambda_0)]
\]

Here, \( S_x \) represents the exponential slope for absorption coefficient from non-pigmented particles \( S_d \), CDOM \( S_g \), or their sum \( S_{dg} \). We selected the reference wavelength, \( \lambda_0 \), equal to 380 nm. In Equation (1), the wavelengths analyzed were 350 to 600 nm for \( a_g(\lambda) \), and 380 to 730 nm for \( a_d(\lambda) \) but excluding 400 to 480 nm and 620 to 710 nm to avoid the chlorophyll pigment peaks due to methanol’s incapability to extract some pigments as discussed by Jeffrey et al. [1997], and 380 to 600 nm (but excluding 400 to 480 nm) for \( a_{dg}(\lambda) \) [Babin et al., 2003b].
The phytoplankton absorption coefficient ($a_{ph}$) is typically related to [Chl$_a$] as a power function [Bricaud et al., 1995, 1998; Prieur and Sathyendranath, 1981]:

$$a_{ph}(\lambda) = A_0(\lambda)[Chl \_a]^{A(\lambda)} \quad (2)$$

The modification of Equation (2) provides an expression of $a_{ph}$ from its value at a reference wavelength (here 670 nm):

$$a_{ph}(\lambda) = B_0(\lambda)[a_{ph}(670)]^{B(\lambda)} \quad (3)$$

$A_0(\lambda)$ and $B_0(\lambda)$ are derived coefficients. [Chl$_a$] can also be determined from $a_{ph}(670)$ by a power function similar to Equation (3).

2.3. Apparent optical properties from in situ measurements

The remote sensing reflectance ($R_{rs}$) spectra (bands centered at 320, 340, 380, 395, 412, 443, 465, 490, 510, 532, 555, 560, 625, 665, 670, 683, 710, 780, and 860 nm, and each band is 10 nm wide at full-width half max) were determined with a BioPro in-water profiling spectroradiometer (Biospherical Instruments, Inc., San Diego, CA), as described in detail by Mannino et al. [2008]. The instrument was deployed multiple times for each station, and the absolute uncertainty was less than 5%. The $R_{rs}$ at 551 nm was calculated from a linear interpolation of values at 532, 555, and 560 nm [Mannino et al., 2008]. The $R_{rs}$ at 488 nm was assumed to be equivalent to the value at 490 nm. $R_{rs}$ measured at 6 stations during BIOME1 cruise, 19 stations during BIOME2 cruise, and 3
and 6 stations during CBP cruises on 27 May and 3 November 2005 were included for analysis in this paper.

2.4. Satellite ocean color validation

The method to process satellite images was described in Mannino et al. [2008] following Bailey and Werdell [2006] protocols. SeaWiFS and MODIS-Aqua observations were processed from Level 1 to Level 2 using the SeaWiFS Data Analysis System software (SeaDAS version 5.1.1 and msi12 version 5.6.3). The pixels were masked after atmospheric correction by any of the following flags: land, cloud or ice, high top-of-atmosphere radiance, low normalized water-leaving radiance at 551 or 555 nm, stray light, sun glint, or atmospheric correction failure) [Bailey and Werdell, 2006]. Pixels with $L_w(412) < 0.2 \text{ mW cm}^{-2} \text{ km}^{-1} \text{ sr}^{-1}$ were excluded to minimize the impacts from atmospheric over-correction in causing negative or significantly reduced water-leaving radiance [Siegel et al., 2002]. The 3x3pixel arrays centered on the field stations, each with ~1 km resolution (sensor native), were analyzed. The satellite observations which occurred within ±8 hours and ±32 hours of in situ measurements were considered for match-up analysis to allow for inclusion of sufficient data points.

2.5. Analysis and validation methods

Three curve-fitting functions, linear function (Model II regression) after log-transformation (log_linear_model; Equation 4), fourth-order polynomial function after log-transformation similar to OC4V4 [O’Reilly et al., 1998, 2000] (log_polynomial_model; Equation 5), and one-phase exponential decay function
(exponential_model; Equation 6), were developed to correlate $R_{ns}$ band ratio to the relevant absorption products:

\[
\log[a_i(\lambda)] = C_0(\lambda) + C_1(\lambda)R
\]

\[\text{(4)}\]

\[
\log[a_i(\lambda)] = D_0(\lambda) + D_1(\lambda)R + D_2(\lambda)R^2 + D_3(\lambda)R^3 + D_4(\lambda)R^4
\]

\[\text{(5)}\]

\[
a_i(\lambda) = G_0(\lambda) + G_1(\lambda) \exp[-G_2(\lambda) \frac{R_{ns}(\lambda_1)}{R_{ns}(\lambda_2)}]
\]

\[\text{(6)}\]

Here, $R = \log[R_{ns}(\lambda_1)/R_{ns}(\lambda_2)]$, and $\lambda_1$ and $\lambda_2$ represent the various bands evaluated, and $C_i, D_i$ and $G_i$ are wavelength-specific derived coefficients, and $a_i$ is the analyzed absorption coefficient as $a_{ph}, a_d, a_g,$ or $a_{dg}$. In addition, a one-phase exponential function to determine $R_{ns}$ from $a_i$ similar to Mannino et al. [2008] was also developed, and $a_i$ was then calculated by its reverse function (reverse_exponential_model; Equation 7):

\[
\frac{R_{ns}(\lambda_1)}{R_{ns}(\lambda_2)} = H_0(\lambda) + H_1(\lambda) \exp[-H_2(\lambda)a_i(\lambda)]
\]

\[\text{(7)}\]

Four products, $a_{ph}(670), a_d(380), a_g(380),$ and $a_{dg}(380)$, whose surface measurements were represented as the site values, were analyzed with the above equations. The mean absolute percent difference (MAPD) and root mean square error (RMSE) between the modeled products ($C_{alg}$) and field measurements ($C_{field}$) were calculated.
2.6. Monthly time series analysis

Monthly Level 3 mapped MODIS-Aqua images (4 km resolution) from July 2002 to December 2006 were downloaded from the NASA ocean color website (http://oceancolor.gsfc.nasa.gov) on 8 January 2008. Empirical algorithms developed from this paper were applied to calculate products such as $[\text{Chl}_a]$, $a_{ph}$, $a_{ds}$, and $a_g$. Three stations—Location A (75.90W, 36.93N), B (75.30W, 36.93N), and C (74.77W, 36.93N) representing a transect from the Chesapeake Bay mouth to an outer shelf location—were selected and plotted to demonstrate a monthly time series.

3. Results and Discussion

3.1. Absorption spectra

We observed seasonal transitions in phytoplankton absorption coefficients in April-May and October-November periods. Therefore, at least two seasonal algorithms (May-October and November-April) are required to describe phytoplankton absorption relationships (Figure 2a and Table 1). In general, the phytoplankton absorption ratio $[a_{ph}(\lambda)/a_{ph}(670)]$ in May-October is higher than that in November-April (Figure 2a). This coincides with historical observations that the dominant phytoplankton taxa are diatoms.
in winter and spring, but the phytoplankton assemblage transitions to a greater proportion of dinoflagellates, cryptophytes, and cyanobacteria in summer and fall [Adolf et al., 2006; Marshall and Alden, 1993]. Although phytoplankton absorption coefficients are subject to seasonal variation, the relationship between $a_{ph}(670)$ and [Chl _a] ($r^2=0.964$, N=230; Figure 2b) is relatively constant seasonally due to the dominant contribution from Chl _a to $a_{ph}(670)$ [Jeffrey et al., 1997]:

$$[Chl \_a] = 70.632 \times [a_{ph}(670)]^{1.184} \tag{10}$$

Seasonal variability of riverine discharge, along with other factors such as wind forcing and direction, may cause the seasonal variability of phytoplankton taxonomic composition, pigment package effect, and therefore normalized phytoplankton absorption spectra (Figure 2a) [Babin et al., 2003b; Bricaud et al., 1995, 1998; Trees et al., 2000]. The pigment package effect refers to a consequence of the fact that in the natural waters pigment molecules are not uniformly distributed but are contained within discrete packages such as chloroplasts, cells, and cell colonies, which causes a flattening of the phytoplankton absorption peak due to self-shading wherever pigments are localized within cell membranes [Duysens, 1956].

Equation (1) described the exponential decay characteristics of $a_d$, $a_g$, and $a_{dg}$ rather well with coefficients of determination ($r^2$) of >0.95 for $a_d$, and >0.99 for $a_g$ and $a_{dg}$.

The exponential slopes ($S$) covered a wide range with mean ± 1 standard deviation of 0.0122±0.0023 (N=247, ranging from 0.0084 to 0.0260) for $a_d$, 0.0170±0.0011 (N=300, ranging from 0.0137 to 0.0221) for $a_g$, and 0.0148±0.0014 (N=222, ranging from 0.0122
to 0.0205) for $a_{dg}$. However, the general exponential decay relationships of $S$ (ordinate) versus absorption coefficients (abscissa) for coastal regions that are significantly impacted by freshwater discharge as suggested by Carder et al. [1989] was not apparent for the SMAB dataset. Part of the reason was that the dataset presented in this paper (Figure 1) did not extend far into the estuaries to cover the full range of water types from freshwater to oceanic waters. The observations above suggest that using mean $S$ values may cause significant errors in predicting $a_x$ spectra. Multiple algorithms to retrieve $a_d$, $a_g$, and $a_{dg}$ at multiple wavelengths may be required, as discussed in the next section.

3.2. Algorithm development and validation

Three $R_{rs}$ band ratios (412/555, 443/555, and 490/555 for SeaWiFS, and 412/551, 443/551, and 488/551 for MODIS-Aqua) were compared to determine the best algorithm performance. No matter which of the curve-fitting models from Equations (4)-(7) was selected, algorithms based on $R_{rs}$ band ratio at 490/555 or 488/551 performed similar to, or better than, the other two band ratios. Figure 3 shows examples of model performance on predicting $a_{ph}(670)$, $a_d(380)$, $a_g(380)$, and $a_{dg}(380)$ from Equations (4)-(7), respectively. Performance from other models, i.e. Equation (4) on predicting $a_d(380)$, $a_g(380)$, and $a_{dg}(380)$, showed similar results to Figure 3 (data not shown). Due to the possibly poor quality of satellite water-leaving radiance at shorter wavelengths (e.g. 412 and 443 nm) in coastal waters [Bailey and Wer'dell, 2006; Siegel et al., 2002], we selected $R_{rs}$ band ratios at 490/555 or 488/551 for further analysis. Although the selection of only two bands will cause interdependence of satellite-derived products, it still provides
valuable information on bio-optical properties of the SMAB in the absence of appropriate
semi-analytic algorithms based on more bands.

The log_linear_model (Equation 4) proved suitable to validate satellite-derived
$a_{ph}$, $a_d$, and $a_{dg}$ with relatively high $r^2$ and relatively low MAPD and RMSE (Figures 4-5
and Table 2). The $r^2$ for the log_linear_model ranged from 0.92 to 0.93 for all three
products of $a_{ph}(670)$, $a_d(380)$, and $a_{dg}(380)$, while 0.84 to 0.95 for the other three models.
The MAPD for the log_linear_model was typically similar to, or lower than, the other
three models, regardless of the satellite sensor (SeaWiFS or MODIS-Aqua) and overpass
satellite/in situ time window selected ($\pm$8 hours or $\pm$32 hours) (Figures 4a-b and 5a-b).
This was also supported by the validation results from the RMSE comparisons (Figures
4c-d and 5c-d). The exponential_model may provide lower MAPD and RMSE in some
cases (Figures 4-5), but was not selected due to its relatively low $r^2$ (0.84 to 0.87). In
contrast, the reverse_exponential_model yielded the best validation results for $a_g(380)$
with the highest $r^2$ (0.90 versus 0.77-0.86) and typically lowest MAPD and RMSE
(Figures 4-5 and Table 4) as compared to the other models. Mannino et al. [2008]
showed the same model for $a_g$ but did not include the CBH stations in their validation
analysis. The selected $R_{rs}$ band ratio models can also be applied to derive $a_d$, $a_g$, and $a_{dg}$ at
multiple wavelengths (Tables 2-3). Except for the higher MAPD for $a_d$ (34.8-57.5% for
SeaWiFS and 41.9-65.3% for MODIS-Aqua), the selected regression methods typically
limited MAPD for $a_g$ and $a_{dg}$ to within 30% for wavelengths between 350 and 555 nm
(Table 4). The exponential decay slopes $(S)$ for $a_d$, $a_g$, and $a_{dg}$ can be derived from non-
linear regression methods with $R_{rs}$ band ratio models at multiple wavelengths (e.g. 355,
380, 400, 412, 443, 490, 510, 531, and 555), and agree reasonably well with field

15
derivations (Table 4). The GSM01 [Garver and Siegel, 1997; Maritorena et al., 2002] and its regional version (GSM01-CB) [Magnuson et al., 2004] resulted in relatively high MAPD and RMSE when compared with the field measurements (Figures 4-5). The GSM01 model was developed for global ocean application, and thus is not optimized for the variability of in-water constituents observed in near-shore coastal regions, such as variable or region-specific $S$ values. However, even the optimized GSM01 model for this coastal region (GSM01-CB; primarily Chesapeake Bay and near-shore coastal ocean) developed by Magnuson et al. [2004] did not perform significantly better than GSM01, e.g. MAPD = 42.6% and 46.0% for $a_{ph}(670)$ and $a_{dg}(380)$ from GSM01-CB, versus 43.3% and 40.7% from GSM01 for SeaWiFS ±8 hour overpass window (Figures 4-5). The lack of adequate knowledge of backscattering coefficients, as well as the higher uncertainty of $R_{rs}$ at shorter wavelengths (e.g. 412 and 443 nm) from satellite measurements, may account for the performance of GSM01 and GSM01-CB for this region [Bailey and Werdell, 2006; IOCCG, 2000; Magnuson et al., 2004; Siegel et al., 2000, 2005].

By applying the regression results shown in Table 1, phytoplankton absorption coefficients at other visible wavelengths can also be derived from $a_{ph}(670)$, which can be derived from satellite radiance observations as shown in Table 2. The validation match-ups based on this approach yielded similar MAPD accuracy levels for $a_{ph}(\lambda)$ at 412, 443, 488, 490, 510, 667, and 678 nm to $a_{ph}(670)$ (21.5-26.1% versus 25.7% for SeaWiFS and 19.0-28.1% versus 21.2% for MODIS-Aqua), but relatively higher MAPD at 531, 551, and 555 nm (27.2-30.9% for SeaWiFS and 33.1-43.1% for MODIS-Aqua) due to the relatively higher measurement errors from weaker absorption at these wavelengths (Table 4). If we exclude those stations with extremely low $a_{ph}$ (e.g. $<0.003$ m$^{-1}$ at 555 nm), the
MAPD for $a_{ph}(\lambda)$ was at the same accuracy level for all visible wavelengths evaluated.

The satellite derivation of [Chl_a] from Equation (10) had slightly lower but still reasonable accuracy level compared to that for $a_{ph}(670)$ with MAPD of 32.3±28.2% (N=29) for SeaWiFS and 28.8±20.6% (N=14) for MODIS-Aqua (Table 4).

The validation match-ups between field measurements of absorption constituents and satellite derivations within ±8 hours demonstrated reasonable agreement (Figures 6-7). The SeaWiFS and MODIS-Aqua match-ups have similar $r^2$ varying from 0.80-0.97, and show slope values from 0.52 to 0.73 and from 0.46 to 0.80, respectively. Including the stations applied to develop the algorithms, which increases dataset size by 10-12 points for SeaWiFS and 5-6 points for MODIS-Aqua, yielded improvements in all the slopes (e.g. 0.73-0.87 for SeaWiFS and 0.57-0.64 for MODIS-Aqua). When extending the match-up dataset from ±8 hours to ±32 hours of the satellite overpass window, similar improvements were also found due to significant increase of dataset size by 50% to 140% (data not shown). The validation match-ups between field measurements of [Chl_a] and satellite derivations after log-transformation agreed well with $r^2=0.86$ to 0.95, slope=0.87 to 0.96, and RMSE=0.20 to 0.24 (Figure 7). Similar statistical results were found for $a_{ph}(670)$ when including those stations used to develop the algorithms (data not shown).

It implies that these empirical algorithms are relatively successful and should improve as the size of the dataset increases.

Above all, the satellite-derived absorption coefficients from selected functions (log-linear model for $a_{ph}$, $a_d$, and $a_{dg}$, and reverse-exponential model for $a_g$) yielded relatively good results for the SMAB. Since the dataset used for algorithm development did not include stations from the CBH cruises (because AOP data was not collected at
those stations), the validation analyses should improve after excluding those CBH stations. For example, MAPD improved from 25.7% to 21.9% for $a_{ph}(670)$, from 35.3% to 29.5% for $a_d(380)$, from 25.2% to 20.9% for $a_g(380)$, from 26.5% to 23.8% for $a_{dg}(380)$, and from 32.3% to 24.2% for [Chl_a] for SeaWiFS ±8 hours overpass time window. Therefore, in the future the addition of complete datasets from the lower bay locations will expand the dynamic range of the algorithms and may significantly improve the model capability in more turbid areas of the SMAB.

3.3. Seasonal variability

These empirical algorithms can be applied to study the spatial and seasonal variability of coastal ocean constituents. Figure 8 shows typical examples of the spatial distribution of the absorption coefficients within the SMAB during four seasons: summer (June-August), fall (September-November), winter (December-February), and spring (March-May). The satellite images clearly show the gradients from high to low constituent concentration between the coast and the open ocean as well as the riverine/esturine outflow impact along the coast (Figure 8). The seasonal variability of phytoplankton absorption [e.g. $a_{ph}(443)$] may be due primarily to the river discharge rate from the bay mouths [Acker et al., 2005; Adolf et al., 2006; Marshall and Alden, 1993; Marshall et al., 2006]. The monthly mean flow rates out of the Chesapeake Bay for these selected images were 510, 1648, 2983, and 1463 m$^3$ s$^{-1}$ for August 2005, November 2005, February 2006, and May 2006, respectively (Data sources: http://waterdata.usgs.gov/nwis/; written communication from Gary Fisher, U.S. Geological Survey, 17 July 2007). Consequently, lower phytoplankton abundance
occurred during the dry season in summer 2005 compared to the other three seasons. The
seasonal variability of detritus absorption [e.g. $a_d(443)$] is complex. At least two primary
sources of detritus from riverine/esturine outflow and sedimentary resuspension
controlled $a_d$ in the SMAB. The significant contribution from storm-driven sedimentary
resuspension in autumn may explain the higher detritus content in November 2005. The
seasonal variability of CDOM absorption [e.g. $a_g(443)$] may be controlled primarily by
the degree of riverine inputs of degraded terrestrial vegetation to the SMAB [Del Vecchio
and Blough, 2004; Mannino et al., 2008].

CDOM plays a critical role in contributing to sunlight absorption and thus impacts
primary production in the SMAB by reducing the amount of photosynthetically active
radiation (PAR) available for phytoplankton growth [Arrigo and Brown, 1996]. At 443
nm, CDOM accounted for 35-70% of total light absorption excluding water absorption
($a_{sw}$), as compared to 0-20% for non-pigmented particles, and 30-45% for phytoplankton
(Figures 9-10). Pure seawater absorption [$a_0(443) = 0.007 \text{ m}^{-1}$; Pope and Fry, 1997]
typically accounts for a negligible fraction (~3.0%) of $a(443)$ in the SMAB. The
relatively low contribution of detritus absorption within coastal ocean regions was also
reported by Siegel et al. [2002] and may explain the performance of the satellite-derived
$a_d(\lambda)$ in the validation analysis (Table 4, and Figures 4-6). The satellite-derived $a_d(\lambda)$
underestimates $a_d(\lambda)$ with respect to field samples collected in the near-shore ocean
region (<20 m bottom depth) and overestimates $a_d(\lambda)$ in water with very low $a_d(\lambda)$
(Figures 6, 8, and 10). The gradients of high to low percentages of $a_{ph}$ and $a_d$, and low to
high percentage of $a_g$ from the coast to the open ocean were consistent with field
measurements (Figure 10). During the dry season (e.g. August 2005), CDOM accounts
for a higher percentage of total absorption than during the wet season (e.g. February 2006) (Figure 9). Such a phenomenon may be explained by the impact of river discharge in contributing nutrients to support phytoplankton growth as well as the export of terrestrial CDOM. During the wet season, CDOM and phytoplankton abundance are both elevated, but phytoplankton blooms increase the relative percentage of phytoplankton absorption compared to CDOM absorption. During the dry season, CDOM and phytoplankton are both low, but the reported higher primary productivity and mature grazer community may result in a higher percentage of phytoplankton to be grazed and degraded which in turn reduces the relative percentage of phytoplankton absorption [Adolf et al., 2006; Marshall and Nesius, 1996; Marshall et al., 2006] and increase CDOM through grazer and microbial processing of organic matter [Nelson et al., 2004; Steinberg et al., 2004]. Since phytoplankton pigments have a much weaker relationship with CDOM absorption than with phytoplankton absorption, the significant contribution of CDOM absorption may pose complications for applying global operational algorithms (e.g. OC4V4 and OC3M) [O’Reilly et al., 1998, 2000] to coastal regions. For CDOM-rich Case 2 waters such as the Chesapeake Bay, OC4V4 has been found to significantly overestimate [Chl_a], especially for offshore regions of the SMAB [Harding et al., 2005; Magnuson et al., 2004]. Our results also support this conclusion. For example, match-ups within ±8 hours showed that OC4V4 performed better for the lower CB region (e.g. CBH stations) with MAPD of 33.8% as compared to 79.6% for whole SMAB region (data not shown). It implies that the relative difference between our approach and operational algorithms would be relatively small in near-shore regions but high in offshore regions. The spatial distribution and the seasonal variability of [Chl_a] based on our approach displayed
similar trends as those from OC4V4 and OC3M algorithms, but significantly reduced the
overestimation by operational Chl_a algorithms in the offshore region of the SMAB
(Figure 11). In general, the ratios of [Chl_a] based on operational Chl_a algorithms to our
approach increase with the increase of CDOM contribution to light absorption (Figures 9
and 11). OC4V4 and OC3M [Chl_a] were higher by 0-0.5 times for the inner-shelf
region, 0.4-1.2 times for the middle shelf region, and 1-2 times for the outer shelf region
(Figure 11). This higher ratio trend toward offshore demonstrates the impact of CDOM
on ocean color products in the SMAB.

The satellite derivations of absorption coefficients provide tools to study
biogeochemical processes and radiative transfer. For examples, DOC and salinity can be
strongly correlated to CDOM absorption [Del Vecchio and Blough, 2004; Mannino et al.,
2008; Rochelle-Newall and Fisher, 2002], and primary productivity is correlated to
phytoplankton absorption [Behrenfeld et al., 2005; Marra et al., 2007]. The knowledge of
absorption also provides methods to study other IOPs from space. For example, the
expression of $R_s$ from absorption and backscattering [Garver and Siegel, 1997; Gordon
et al., 1988; Maritorena et al., 2002] and the empirical expression of absorption from $R_s$
ratio make it possible to express backscattering into $R_s$. The knowledge of backscattering
might significantly improve the capability of semi-analytical models in deriving ocean
color products from space [Magnuson et al., 2004].

Although we have shown the significant impact of river discharge on
biogeochemical constituents in the SMAB, the direct link between them should be
interpreted with caution. First, the impact of river discharge on the coastal region of
Chesapeake Bay is different from Delaware Bay. The lower Chesapeake Bay is subject to
nutrient limitation for phytoplankton growth, in contrast to light availability in the
Delaware Bay [Harding et al., 1986; Marshall and Alden, 1993]. Therefore, an increase
in river discharge is more likely to cause a phytoplankton bloom in the lower Chesapeake
Bay by driving more nutrients downstream, while an increase in turbidity from higher
river discharge may decrease primary production in the lower Delaware Bay. Second, the
impact of river discharge is subject to seasonal variability and distance from the bay
mouths, as shown in the following for the coastal region of Chesapeake Bay. In the inner-
shelf region, the correlation coefficients ($r$) between river discharge rate and
biogeochemical products are low (e.g. $r=0.05-0.12$ for [Chl$_a$] and $-0.01-0.14$ for $a_g$) for
all seasons except for summer (Figure 12). The poor correlation may be due to averaging
out the higher frequency responses (less than one week) for the export of nutrients and
CDOM, respectively, from the bays. During summer, the vertical stratification is well
developed [Verity et al., 2002], and the strength of river discharge represents the flux of
nutrients for phytoplankton growth. Thus, the correlation between biogeochemical
products and river discharge improves for summer (e.g. $r=0.48$ for [Chl$_a$] and $a_g$)
(Figure 12). In the middle shelf region, however, river discharge is significantly
correlated to biogeochemical products in winter but poorly correlated during other
seasons (e.g. $r=0.79$ for [Chl$_a$] and $a_g$ in winter and $r=0.09-0.36$ in other seasons)
(Figure 12). During winter, low water temperature and a less mature grazer community
may cause the phytoplankton biomass to be linked directly to nutrient availability, which
is driven primarily by river discharge and by wind-induced vertical mixing of nutrients
from depth [Adolf et al., 2006; Marshall and Alden, 1993]. As the zooplankton and
bacterial communities develop into spring and summer, lower phytoplankton biomass
and higher primary productivity are expected [Adolf et al., 2006] and the direct response of the biological system to river discharge dissipates. The outer shelf region shows a similar pattern but a lower correlation coefficient (e.g. $r=0.49$ for $[\text{Chl}_a]$ and $a_g$ in winter, and $-0.13-0.31$ in other seasons) with river discharge than the middle shelf region (data not shown).

Other physical factors than river discharge, such as water temperature and wind forcing, anthropogenic activities, and even climate change, can also impact phytoplankton abundance, productivity, and carbon distributions in the SMAB. For example, the direction and distribution of the Chesapeake Bay plume is highly dependent on the wind stress direction. During winter and early spring northerly winds (downwelling favorable) and the along-shore southward current force the Chesapeake Bay [Rennie et al., 1999; Verity et al., 2002] and Delaware Bay [Sanders and Garvine, 2001] plumes to flow southward along the coast. As winds reverse later in spring the southerly along-shore flow weakens, and the Chesapeake Bay plume broadens and flows offshore, primarily to the south and east. Upwelling-favorable conditions can initiate local phytoplankton blooms and contribute additional particles to surface waters [Johnson et al., 2001]. The Chesapeake Bay estuarine ecosystem has experienced a large increase in anthropogenic nutrient loading and reductions in the past half century which have affected the floral composition and biomass [Harding, 1994; Paerl et al., 2006]. Furthermore, climate forcing (e.g. hurricanes, drought, etc.) significantly influences phytoplankton dynamics (e.g. by reducing vertical stratification, increasing sedimentary resuspension, and redistributing particles from hurricane forcing) [Miller and Harding, 2005; Paerl et al., 2006].
Above all, the impacts from physical factors (e.g. river discharge, wind forcing, and bathymetry) on bio-optical constituents (e.g. [Chl\textsubscript{a}] and \(a\textsubscript{g}\)) are complicated and cannot be explained by a single factor [Harding, 1994]. Nevertheless, we found that the variability of an optical property, the diffuse attenuation coefficient at 490 nm (K\textsubscript{490}), represents the variability of multiple bio-optical constituents. In the Chesapeake Bay inner-shelf site, the correlation coefficient (r) of [Chl\textsubscript{a}], \(a\textsubscript{ph}, a\textsubscript{d}, a\textsubscript{g},\) and \(a\textsubscript{dg}\) to K\textsubscript{490} was 0.66-0.74, while 0.94-0.99 in the middle shelf location, and 0.93-0.98 in the outer shelf locations. These results also imply that absorption is the dominant contributor to the diffuse attenuation coefficient at offshore locations but scattering contributes significantly at near-shore locations.

4. Conclusions

Several important conclusions can be made from the present analyses of absorption coefficients and [Chl\textsubscript{a}] derived from ocean color remote sensing. The empirical algorithms demonstrate successful retrieval of absorption coefficients and [Chl\textsubscript{a}] within a reasonable uncertainty (e.g. ±35%), and demonstrate significant improvements from the standard semi-analytic model (e.g. GSM01 and GSM01-CB) and operational algorithms (e.g. OC4V4 and OC3M). Field observations and satellite derivations both demonstrate that CDOM is the major contributor to water-column light absorption at shorter wavelengths (e.g. <500 nm), especially during the dry seasons and on the outer shelf where it can account for 35-70% of absorption by particles plus CDOM at 443 nm. River discharge plays a principal role in controlling the distribution of biogeochemical constituents, but is subject to seasonal and regional variability.
Acknowledgements

This research was supported by an appointment to the NASA Postdoctoral Program at the Goddard Space Flight Center, administered by Oak Ridge Associated Universities through a contract with NASA. This work was funded with support from the NASA New Investigator Program, Interdisciplinary Science, and Earth Observing System programs and by NOAA through a grant in support of the Coastal Observatories program (NA03NOS4730220). We thank the captains and crews of R/V Cape Henlopen, Hugh R. Sharp, and Fay Slover. H. Throckmorton, P. Bernhardt, K. C. Filippino, C. Makinen, and M. Linksweiler provided assistance in collecting particulate samples. We are grateful to J. Morrow, J. Brown, D. D’Alimonte, and J.-N. Druon for deploying the profiling radiometer, and L. Van Heukelem and C. Thomas for analyzing HPLC pigments, and the Ocean Biology Processing Group at GSFC. J. O’Reilly kindly provided the high-resolution bathymetry data.
References


Mueller, J.L. (2000), SeaWiFS algorithm for the diffuse attenuation coefficient, K(490), using water-leaving radiances at 490 and 555 nm, in O'Reilly, J.E. and co-authors: 


Figure captions

Figure 1. Map of the study area within the southern Middle Atlantic Bight (SMAB).
Symbols representing the sampling stations from the following cruises are: △ – BIOME1 (30 March to 1 April 2005), ⊿ – BIOME2 (26 to 30 July 2005), ○ – BIOME3 (5 to 9 May 2006), × – BIOME4 (2 to 6 July 2006), □ – CBP (four daily cruises), and ■ – CBH (ten daily cruises).

Figure 2. Log-transformed linear regression of phytoplankton absorption coefficient at 670 nm [$a_{ph}(670)$] to a) $a_{ph}(443)$, and b) chlorophyll-a concentration [Chl_a]. The solid and dashed lines in the upper figure represent the regression for summer-fall season (May to October) and winter-spring season (November to April), respectively.

Figure 3. Absorption algorithms derived from field observations of remote sensing reflectance ($R_{rs}$) from a) log_linear_model for $a_{ph}(670)$: $\log[a_{ph}(\lambda)] = C_0(\lambda) + C_1(\lambda) R$, where $R = \log[R_{rs}(\lambda_1)/R_{rs}(\lambda_2)]$; and b) log_polynomial_model for non-pigmented particulate absorption coefficient at 380 nm [$a_d(380)$]: $\log[a_d(\lambda)] = D_0(\lambda) + D_1(\lambda) R + D_2(\lambda) R^2 + D_3(\lambda) R^3 + D_4(\lambda) R^4$; and c) exponential_model for CDOM absorption coefficient at 380 nm [$a_g(380)$]: $a_g(\lambda) = G_0(\lambda) + G_1(\lambda) \exp[-G_2(\lambda) \frac{R_{rs}(\lambda_1)}{R_{rs}(\lambda_2)}]$; and d) reverse_exponential_model for absorption coefficient by non-pigmented particles plus CDOM at 380 nm [$a_{dg}(380)$]
\[
\frac{R_{rs}(\lambda_1)}{R_{rs}(\lambda_2)} = H_0(\lambda) + H_1(\lambda) \exp[-H_2(\lambda)a_{dg}(\lambda)].
\]
Regression lines from \(R_{rs}\) band ratios of 412/555, 443/555, and 490/555 are represented as solid, dotted, and dashed lines, respectively.

Figure 4. Validation results comparing SeaWiFS observations with field measurements of \(a_{ph}(670)\), \(a_d(380)\), \(a_g(380)\), and \(a_{dg}(380)\) from multiple models (log_linear_model, log_polynomial_model, exponential_model, reverse_exponential_model). Figures a) and b) show the mean absolute percent difference (MAPD), while c) and d) show the root mean square error (RMSE) of the validation results within 8 hours and 32 hours of the satellite overpass, respectively. The data from stations applied to develop the algorithms were not included in this analysis. The satellite derived \(a_{ph}(670)\) and \(a_{dg}(380)\) from GSM01 model [Maritorena et al., 2002] and GSM01-CB model [Magnuson et al., 2004] are also shown for comparison.

Figure 5. Validation results comparing MODIS-Aqua observations with field measurements of \(a_{ph}(670)\), \(a_d(380)\), \(a_g(380)\), and \(a_{dg}(380)\) from multiple models. See Figure 4 for details.

Figure 6. Comparisons of SeaWiFS and MODIS-Aqua and field observations of the absorption coefficients of a) \(a_{ph}(670)\), b) \(a_d(380)\), c) \(a_g(380)\), and d) \(a_{dg}(380)\). The values are plotted on log scale. The satellite derivations of \(a_{ph}(670)\), \(a_d(380)\), and \(a_{dg}(380)\) were from the log_linear_model, while \(a_g(380)\) from the reverse_exponential_model. The match-ups procedure is limited to within ±8 hours, and the data from stations used to
develop algorithms are excluded for validation analyses. The statistical results are based on log-transformation of the data and shown on upper left for SeaWiFS and lower right for MODIS-Aqua. The solid lines represent the 1:1 lines, while dashed lines and dotted lines represent the regression for SeaWiFS and MODIS-Aqua respectively.

Figure 7. Comparisons of SeaWiFS and MODIS-Aqua and field observations of [Chl_a] for satellite overpass window of a) ±8 hours and b) ±32 hours. The data from stations used to develop algorithms of \( a_{ph} \) are excluded for this analysis. The solid lines represent the 1:1 lines, while dashed lines and dotted lines represent the regression for SeaWiFS and MODIS-Aqua respectively. See Figure 6 for detail.

Figure 8. The distribution of \( a_{ph}, a_d, \) and \( a_g \) at 443 nm within the SMAB for 5 August and 3 November 2005, and 15 February and 12 May 2006 representing four seasons. The derived images for 5 August 2005 and 15 February 2006 were from MODIS-Aqua, while the other two were from SeaWiFS.

Figure 9. The distribution of the relative percentage of \( a_{ph}, a_d, \) and \( a_g \) to their sum at 443 nm within the SMAB. See Figure 8 for detail.

Figure 10. The relative percentage of \( a_{ph}, a_d, \) and \( a_g \) to their sum at 443 nm from field measurements grouped into two regions (near shore region with bottom depth <20 m, and offshore region with bottom depth ≥20 m) within the SMAB.
Figure 11. The distribution of [Chl_a] calculated from operational ocean color algorithms (OC4V4 for SeaWiFS and OC3M for MODIS-Aqua) and from the empirical method described in this paper (OC_SMAB; \( [Chl_a] = 70.632 \times [a_{ph}(670)]^{1.184} \)), and their ratio \([\text{OC4V4 or OC3M}/\text{OC_SMAB}]\) within the SMAB. See Figure 8 for detail. The responding scales of the color bar are in log units for [Chl_a] and in linear units for the ratio.

Figure 12. Monthly time series of a) [Chl_a], b) \(a_g(443)\), and c) diffuse attenuation coefficient at 490 nm (K490) from MODIS-Aqua Level-3 images (4x4 km resolution) for a near shore location (75.90W, 36.93N; solid circle) and a middle shelf location (75.30W, 36.93N; open circle). [Chl_a] and \(a_g(443)\) are calculated from algorithms developed in this paper, while K490 is a direct product from the Level-3 images. Monthly river discharge rates at the mouth of Chesapeake Bay (open triangle; Data sources: http://waterdata.usgs.gov/nwis/; written communication from Gary Fisher, U.S. Geological Survey, 17 July 2007) are also shown for comparison.
Table 1. Regression results of phytoplankton absorption coefficient \( (a_{ph}) \) to \( a_{ph}(670) \) from Equation (3): 
\[
a_{ph}(\lambda) = B_0(\lambda)[a_{ph}(670)]^{B_1(\lambda)}.
\]
Log-transformation was applied to the data and Model II linear regression was adopted to calculate \( \log[B_0(\lambda)] \) and \( B_1(\lambda) \). The selected wavelengths for \( a_{ph} \) analysis were the visible bands for SeaWiFS and MODIS-Aqua.

<table>
<thead>
<tr>
<th>( \lambda ) (nm)</th>
<th>May-October (N=196)</th>
<th>November-April (N=51)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>( B_0 )</td>
<td>( B_1 )</td>
</tr>
<tr>
<td>412</td>
<td>1.296</td>
<td>0.835</td>
</tr>
<tr>
<td>443</td>
<td>1.525</td>
<td>0.843</td>
</tr>
<tr>
<td>488</td>
<td>1.023</td>
<td>0.846</td>
</tr>
<tr>
<td>490</td>
<td>1.015</td>
<td>0.851</td>
</tr>
<tr>
<td>510</td>
<td>0.842</td>
<td>0.911</td>
</tr>
<tr>
<td>531</td>
<td>0.694</td>
<td>0.983</td>
</tr>
<tr>
<td>551</td>
<td>0.603</td>
<td>1.047</td>
</tr>
<tr>
<td>555</td>
<td>0.587</td>
<td>1.067</td>
</tr>
<tr>
<td>667</td>
<td>0.899</td>
<td>1.005</td>
</tr>
<tr>
<td>678</td>
<td>1.039</td>
<td>1.002</td>
</tr>
</tbody>
</table>
Table 2. Statistical results for absorption coefficients of phytoplankton \((a_{ph})\), non-pigmented particles \((a_d)\), and CDOM plus non-pigmented particles \((a_{dg})\) at selected wavelengths from log-linear model: \(\log[a_x(\lambda)] = C_0(\lambda) + C_1(\lambda)R\), where \(R = \log[R_x(\lambda_1)/R_x(\lambda_2)]\). The size of the dataset is \(N=25\).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>(R_x(490)/R_x(555))</th>
<th>(R_x(490)/R_x(551))</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(C_1)</td>
<td>(C_0)</td>
</tr>
<tr>
<td>(a_{ph}(670))</td>
<td>-2.602</td>
<td>-1.467</td>
</tr>
<tr>
<td>(a_d(380))</td>
<td>-2.797</td>
<td>-1.319</td>
</tr>
<tr>
<td>(a_d(400))</td>
<td>-2.812</td>
<td>-1.387</td>
</tr>
<tr>
<td>(a_d(412))</td>
<td>-2.849</td>
<td>-1.427</td>
</tr>
<tr>
<td>(a_d(443))</td>
<td>-3.048</td>
<td>-1.633</td>
</tr>
<tr>
<td>(a_d(490))</td>
<td>-3.260</td>
<td>-1.950</td>
</tr>
<tr>
<td>(a_d(510))</td>
<td>-3.584</td>
<td>-2.105</td>
</tr>
<tr>
<td>(a_d(531))</td>
<td>-3.676</td>
<td>-2.223</td>
</tr>
<tr>
<td>(a_d(555))</td>
<td>-3.315</td>
<td>-2.297</td>
</tr>
<tr>
<td>(a_{dg}(380))</td>
<td>-1.394</td>
<td>-0.434</td>
</tr>
<tr>
<td>(a_{dg}(400))</td>
<td>-1.489</td>
<td>-0.576</td>
</tr>
<tr>
<td>(a_{dg}(412))</td>
<td>-1.535</td>
<td>-0.651</td>
</tr>
<tr>
<td>(a_{dg}(443))</td>
<td>-1.593</td>
<td>-0.879</td>
</tr>
<tr>
<td>(a_{dg}(490))</td>
<td>-1.649</td>
<td>-1.190</td>
</tr>
<tr>
<td>(a_{dg}(510))</td>
<td>-1.725</td>
<td>-1.308</td>
</tr>
<tr>
<td>(a_{dg}(531))</td>
<td>-1.633</td>
<td>-1.406</td>
</tr>
<tr>
<td>(a_{dg}(555))</td>
<td>-1.983</td>
<td>-1.564</td>
</tr>
</tbody>
</table>
Table 3. Statistical results for CDOM absorption coefficient \( (a_g) \) at selected wavelengths from the reverse exponential model:

\[
\frac{R_s(\lambda_1)}{R_s(\lambda_2)} = H_0(\lambda) + H_1(\lambda) \exp[-H_2(\lambda)a_g(\lambda)].
\]
The size of the dataset is \( N=34 \).

<table>
<thead>
<tr>
<th>Parameter</th>
<th>( R_s(490)/R_s(555) )</th>
<th>( R_s(490)/R_s(551) )</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>( H_0 )</td>
<td>( H_1 )</td>
</tr>
<tr>
<td>( a_g(355) )</td>
<td>0.538</td>
<td>3.149</td>
</tr>
<tr>
<td>( a_g(380) )</td>
<td>0.534</td>
<td>3.015</td>
</tr>
<tr>
<td>( a_g(400) )</td>
<td>0.540</td>
<td>2.940</td>
</tr>
<tr>
<td>( a_g(412) )</td>
<td>0.523</td>
<td>2.849</td>
</tr>
<tr>
<td>( a_g(443) )</td>
<td>0.531</td>
<td>2.857</td>
</tr>
<tr>
<td>( a_g(490) )</td>
<td>0.547</td>
<td>3.138</td>
</tr>
<tr>
<td>( a_g(510) )</td>
<td>0.493</td>
<td>2.352</td>
</tr>
<tr>
<td>( a_g(531) )</td>
<td>0.494</td>
<td>2.271</td>
</tr>
<tr>
<td>( a_g(555) )</td>
<td>0.335</td>
<td>1.798</td>
</tr>
</tbody>
</table>
Table 4. The mean absolute percent difference (MAPD) and root mean square error (RMSE) from validation match-ups for $a_{ph}$, $a_d$, $a_g$, and $a_{dg}$ at selected wavelengths. The derived exponential decay slope ($S$) for $a_d$ ($S_d$), $a_g$ ($S_g$), and $a_{dg}$ ($S_{dg}$) from non-linear regression and chlorophyll $a$ concentration ([Chl $a$]) are also compared. Data used for algorithm development are not included in this analysis. The size of the datasets are $N=22$, 36, 8, and 19 for $a_{ph}$, $a_d$, or $a_{dg}$, and $N=29$, 45, 14, and 25 for [Chl $a$], and $N=31$, 47, 14, and 25 for $a_{g}$ for SeaWiFS ±8 and ±32 hours and MODIS ±8 and ±32 hours overpass windows, respectively.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>SeaWiFS (±8h)</th>
<th>SeaWiFS (±32h)</th>
<th>MODIS (±8h)</th>
<th>MODIS (±32h)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>MAPD</td>
<td>RMSE</td>
<td>MAPD</td>
<td>RMSE</td>
</tr>
<tr>
<td>$a_{ph}$ (443)</td>
<td>23.5</td>
<td>0.0501</td>
<td>23.4</td>
<td>0.0470</td>
</tr>
<tr>
<td>$a_{ph}$ (670)</td>
<td>25.7</td>
<td>0.0371</td>
<td>30.1</td>
<td>0.0329</td>
</tr>
<tr>
<td>[Chl $a$]</td>
<td>32.3</td>
<td>3.6808</td>
<td>32.5</td>
<td>3.0339</td>
</tr>
<tr>
<td>$a_d$ (380)</td>
<td>35.3</td>
<td>0.1447</td>
<td>38.2</td>
<td>0.1350</td>
</tr>
<tr>
<td>$a_d$ (443)</td>
<td>41.8</td>
<td>0.0787</td>
<td>42.1</td>
<td>0.0726</td>
</tr>
<tr>
<td>$S_d$</td>
<td>13.4</td>
<td>0.0018</td>
<td>11.7</td>
<td>0.0016</td>
</tr>
<tr>
<td>$a_g$ (380)</td>
<td>25.2</td>
<td>0.1636</td>
<td>20.7</td>
<td>0.1359</td>
</tr>
<tr>
<td>$a_g$ (443)</td>
<td>22.8</td>
<td>0.0516</td>
<td>20.1</td>
<td>0.0432</td>
</tr>
<tr>
<td>$S_g$</td>
<td>5.8</td>
<td>0.0011</td>
<td>5.5</td>
<td>0.0011</td>
</tr>
<tr>
<td>$a_{dg}$ (380)</td>
<td>26.5</td>
<td>0.2104</td>
<td>25.2</td>
<td>0.1965</td>
</tr>
<tr>
<td>$a_{dg}$ (443)</td>
<td>24.4</td>
<td>0.0923</td>
<td>24.4</td>
<td>0.0870</td>
</tr>
<tr>
<td>$S_{dg}$</td>
<td>11.9</td>
<td>0.0020</td>
<td>12.7</td>
<td>0.0021</td>
</tr>
</tbody>
</table>
Figure 1. Map of the study area within the southern Middle Atlantic Bight (SMAB).

Symbols representing the sampling stations from the following cruises are: △ – BIOME1 (30 March to 1 April 2005), ▽ – BIOME2 (26 to 30 July 2005), ○ – BIOME3 (5 to 9 May 2006), × – BIOME4 (2 to 6 July 2006), □ – CBP (four daily cruises), and ■ – CBH (ten daily cruises).
Figure 2. Log-transformed linear regression of phytoplankton absorption coefficient at 670 nm [$a_{ph}(670)$] to a) $a_{ph}(443)$, and b) chlorophyll $a$ concentration [Chl$_a$]. The solid and dashed lines in the upper figure represent the regression for summer-fall season (May to October) and winter-spring season (November to April), respectively.
Figure 3. Absorption algorithms derived from field observations of remote sensing reflectance ($R_{rs}$) from a) log_linear_model for $a_{ph}(670)$: $\log[a_{ph}(\lambda)] = C_0(\lambda) + C_1(\lambda) R$,
where $R = \log[R_{rs}(\lambda_1)/R_{rs}(\lambda_2)]$; and b) log_polynomial_model for non-pigmented particulate absorption coefficient at 380 nm [$a_d(380)$]:

$$\log[a_d(\lambda)] = D_0(\lambda) + D_1(\lambda) R + D_2(\lambda) R^2 + D_3(\lambda) R^3 + D_4(\lambda) R^4;$$

and c) exponential_model for CDOM absorption coefficient at 380 nm [$a_g(380)$]:

$$a_g(\lambda) = G_0(\lambda) + G_1(\lambda) \exp[-G_2(\lambda) \frac{R_{rs}(\lambda_1)}{R_{rs}(\lambda_2)}];$$

and d) reverse_exponential_model for absorption coefficient by non-pigmented particles plus CDOM at 380 nm [$a_{dg}(380)$]

$$\frac{R_{rs}(\lambda_1)}{R_{rs}(\lambda_2)} = H_0(\lambda) + H_1(\lambda) \exp[-H_2(\lambda) a_{dg}(\lambda)].$$

Regression lines from $R_{rs}$ band ratios of
412/555, 443/555, and 490/555 are represented as solid, dotted, and dashed lines, respectively.
Figure 4. Validation results comparing SeaWiFS observations with field measurements of $a_{ph}(670)$, $a_{d}(380)$, $a_{g}(380)$, and $a_{dg}(380)$ from multiple models (log_linear_model, log_polynomial_model, exponential_model, reverse_exponential_model). Figures a) and b) show the mean absolute percent difference (MAPD), while c) and d) show the root mean square error (RMSE) of the validation results within 8 hours and 32 hours of the satellite overpass, respectively. The data from stations applied to develop the algorithms were not included in this analysis. The satellite derived $a_{ph}(670)$ and $a_{dg}(380)$ from GSM01 model [Maritorena et al., 2002] and GSM01-CB model [Magnuson et al., 2004] are also shown for comparison.
Figure 5. Validation results comparing MODIS-Aqua observations with field measurements of $a_{ph}(670)$, $a_d(380)$, $a_g(380)$, and $a_{dg}(380)$ from multiple models. See Figure 4 for details.
Figure 6. Comparisons of SeaWiFS and MODIS-Aqua and field observations of the absorption coefficients of a) $a_{ph}(670)$, b) $a_d(380)$, c) $a_g(380)$, and d) $a_{dg}(380)$. The values are plotted on log scale. The satellite derivations of $a_{ph}(670)$, $a_d(380)$, and $a_{dg}(380)$ were from the log_linear_model, while $a_g(380)$ from the reverse_exponential_model. The match-ups procedure is limited to within ±8 hours, and the data from stations used to develop algorithms are excluded for validation analyses. The statistical results are based on log-transformation of the data and shown on upper left for SeaWiFS and lower right for MODIS-Aqua. The solid lines represent the 1:1 lines, while dashed lines and dotted lines represent the regression for SeaWiFS and MODIS-Aqua respectively.
Figure 7. Comparisons of SeaWiFS and MODIS-Aqua and field observations of [Chl$_a$] for satellite overpass window of a) ±8 hours and b) ±32 hours. The data from stations used to develop algorithms of $a_{ph}$ are excluded for this analysis. The solid lines represent the 1:1 lines, while dashed lines and dotted lines represent the regression for SeaWiFS and MODIS-Aqua respectively. See Figure 6 for detail.
Figure 8. The distribution of $a_{ph}$, $a_d$, and $a_g$ at 443 nm within the SMAB for 5 August and 3 November 2005, and 15 February and 12 May 2006 representing four seasons. The derived images for 5 August 2005 and 15 February 2006 were from MODIS-Aqua, while the other two were from SeaWiFS.
Figure 9. The distribution of the relative percentage of $a_{ph}$, $a_{al}$, and $a_{s}$ to their sum at 443 nm within the SMAB. See Figure 8 for detail.
Figure 10. The relative percentage of $a_{ph}$, $a_d$, and $a_g$ to their sum at 443 nm from field measurements grouped into two regions (near shore region with bottom depth $<20$ m, and offshore region with bottom depth $\geq 20$ m) within the SMAB.
Figure 11. The distribution of [Chl_a] calculated from operational ocean color algorithms (OC4V4 for SeaWiFS and OC3M for MODIS-Aqua) and from the empirical method described in this paper (OC_SMAB; $[Chl\_a] = 70.632 \times [a_{ph}(670)]^{1.184}$), and their ratio $[(OC4V4 \ or \ OC3M)/OC\_SMAB]$ within the SMAB. See Figure 8 for detail. The responding scales of the color bar are in log units for [Chl_a] and in linear units for the ratio.
Figure 12. Monthly time series of a) [Chl_a], b) $a_g(443)$, and c) diffuse attenuation coefficient at 490 nm (K490) from MODIS-Aqua Level-3 images (4x4 km resolution) for a near shore location (75.90W, 36.93N; solid circle) and a middle shelf location (75.30W, 36.93N; open circle). [Chl_a] and $a_g(443)$ are calculated from algorithms developed in this paper, while K490 is a direct product from the Level-3 images. Monthly river discharge rates at the mouth of Chesapeake Bay (open triangle; Data sources: http://waterdata.usgs.gov/nwis/; written communication from Gary Fisher, U.S. Geological Survey, 17 July 2007) are also shown for comparison.