Atmospheric correction and retrieval of diffuse attenuation coefficient from GOCI using the 2SeaColor model: A case study in Yangtze estuary (Id. 10555)

Xiaolong Yu\textsuperscript{1}; Suhyb Salama\textsuperscript{1}; Fang Shen\textsuperscript{2}; Wouter Verhoef\textsuperscript{1}; Gaoyan Wu\textsuperscript{1}; Palmer Stephanie\textsuperscript{1}

\textsuperscript{1}Faculty of Geo-Information Science and Earth Observation (ITC), University Twente, the Netherlands
\textsuperscript{2}State Key Laboratory of Estuarine and Coastal Research, East China Normal University, China
Chinese PI(s) & Co-I(s):

Prof. Fang SHEN  State Key Laboratory of Estuarine and Coastal Research (SKLEC), East China Normal University
Prof. Yunxuan ZHOU  State Key Laboratory of Estuarine and Coastal Research (SKLEC), East China Normal University

European PI(s) & Co-I(s):

Dr. David Doxaran  Laboratoire d‘Océanographie de Villefranche (LOV), UPMC/CNRS, France
Dr. Suhyb Salama  Faculty of Geo-Information Science and Earth Observation (ITC), University of Twente, the Netherlands
Xiaolong Yu  Faculty of Geo-Information Science and Earth Observation (ITC), University of Twente, the Netherlands
Background

Methodology
- Atmosphere correction for GOCI images
- The analytical $K_d$ retrieval model: 2SeaColor

Results
- Models validation and inter-comparison
- GOCI-derived $K_d$ maps

Conclusion
The diffuse attenuation coefficient ($K_d$, in m$^{-1}$) describes light penetration and attenuation in the water column. It depends both the optical active water constituents and the distribution of the ambient light field.

Applications of $K_d$: Turbidity in water (Kirk, 1994); Sediment transportation and resuspension (Majozi et al., 2014; Zhang et al., 2006), Heat transfer in the upper water layer (Stramska and Zuzewicz, 2013; Wu et al., 2007), Phytoplankton photosynthesis and photobleaching (Loiselle et al., 2009; McClain et al., 1996); Estimation of primary production in nature waters (Bergamino et al., 2010; Lee et al., 2011).

However, to remotely retrieve $K_d$ in turbid coastal waters, we need develop algorithms for both atmosphere correction (AC) of target sensor and $K_d$ retrieval with satisfied accuracy.
Limitations of some widely used models for AC and $K_d$ retrieval:

- **AC algorithm:**
  NIR-SWIR method (Wang, 2007; Wang et al., 2009e) was validated for turbid waters, but it is not operationally feasible for our target sensor GOCI (Geostationary Ocean Color Imager) since there are no SWIR bands in this sensor.

- **$K_d$ retrieval**
  Lee model (Lee et al., 2005a; Lee et al., 2005d) and Zhang model (Zhang and Fell, 2007) require a fine-tuning on the empirical coefficients and underestimate $K_d$ in extremely turbid waters without tuning parameters (Jamet et al., 2012).
Provide satisfied retrieval results for both clear and turbid water without tuning parameters.
Methodology

✓ 1 Data

✓ 2 AC algorithm

✓ 3 $K_d$ retrieval model
Methodology

1 Data

**In-situ datasets**

- $K_d$ and $R_{rs}$

Fig. a NOMAD dataset
Fig. b COASTLOOC dataset
Fig. c Yangtze dataset

**GOCI images**

7 match-ups between March 4th and 9th (Fig. c)

Location of the field data sets (solid circles in Fig. c) are matchups sites within 1h overpass of GOCI.
Verhoef and Bach (2003) put forward a simplified hypothesis based on the RTE, using four-stream radiation fluxes to model the radiative transfer in the atmosphere. From this theory, the total radiance at the top of the atmosphere \((L_{\text{TOA}})\) can be calculated according to the following equation:

\[
L_{\text{TOA}} = \frac{E_s^0 \cos \theta_s}{\pi} \left[ \rho_{so} + \frac{(\tau_{ss} + \tau_{sd})r(\tau_{do} + \tau_{oo})}{1 - r\rho_{dd}} \right]
\]  

Where, \(\rho\) and \(\tau\) symbols with double subscripts are intrinsic optical properties (reflectances and transmittances) of the entire atmosphere layer. The subscripts are \(s\) for direct solar flux, \(d\) for upward or downward diffuse flux, and \(o\) for radiance in the observer’s direction. Furthermore, \(E_s^0\) is the extraterrestrial solar irradiance (corrected for sun-earth distance), and \(\theta_s\) is the solar zenith angle. \(r\) is surface reflectance.
✓ 2 AC algorithm

Eq. (1) can be simplified to:

\[ L^{\text{TOA}} = L_0 + \frac{G r}{1 - rS} \]  \hspace{1cm} (2)

The parameters \( L_0, S, \) and \( G \) can be simulated by MODTRAN when running the model for surface reflectances of 0.0, 0.5, and 1.0, respectively.

\[ R_{rs} = \frac{r}{\pi} \]  \hspace{1cm} (3)
Methodology

✔ 2 AC algorithm

MODTRAN outputs are spectral variables and dependent on atmosphere conditions.

Look-up-table (LUT) for GOCI images is build:
- various visibilities (5, 10, 20, 30 and 40km)
- aerosol type (Maritime and Urban)
- Atmospheric profile model: mid-latitude winter.
- Atmospheric components: CO₂, water vapor, and ozone gas were set to 392 ppmv, 2.9 g/cm², and 0.31 atm-cm, respectively.
- Relative viewing azimuth, and the zenith angles of sun and viewing were calculated based on location and the times of GOCI image acquisition.

Best-fitted case is found with minimum difference between MODTRAN simulated and field measured Rrs.
Radiative transfer inside the water column can be described by the two-stream radiative transfer equations (Duntley, 1942, 1963) as follows:

\[
\frac{dE_s}{dz} = kE_s
\]

\[
\frac{dE^-}{dz} = -s_E E_s + \alpha E^- - \sigma E^+
\]

\[
\frac{dE^+}{dz} = s'E_s + \sigma E^- - \alpha E^+
\]

$E_s$ is the direct solar flux, $E^-$ and $E^+$ are the diffuse downward and upward fluxes; $z$ is the vertical dimension; $k$ is the extinction coefficient for direct sunlight; $s$ and $s'$ are the backscattering and forward scattering for direct sunlight.
Methodology

3 $K_d$ retrieval model - 2SeaColor forward model

The reflectance result predicted by the 2SeaColor model is $r_{sd}^\infty$, the directional-hemispherical reflectance of the semi-infinite medium, which is linked to IOPs by Salama and Verhoef (2015):

$$r_{sd}^\infty = \frac{\sqrt{1+2x-1}}{\sqrt{1+2x+2\mu_w}}$$

(4)

where $x$ is the ratio of backscattering to absorption coefficients ( $x = b_b/a$), and $\mu_w$ is the cosine of the solar zenith angle beneath the water surface.

The reflectance factor $r_{sd}^\infty$ can be approximated by $Qr_{rs}$, where $Q \approx 3.25$ and $r_{rs}$ is remote sensing reflectance beneath the surface (Morel and Gentili, 1993), which can be converted from above-surface remote sensing reflectance ($R_{rs}$) by Lee et al. (2002):

$$r_{rs} = \frac{R_{rs}}{0.5+1.7R_{rs}}$$

(5)
3 $K_d$ retrieval model – Inverse scheme

We first simulate the $R_{rs}$ spectra using the inherent optical properties (IOPs) parameterizations based on Eqs. (4) and (5).

\[ a_\varphi(\lambda) = (a_0(\lambda) + a_1(\lambda)\ln[a_\varphi(440)])a_\varphi(440) \]

\[ a_{dg}(\lambda) = a_{dg}(\lambda_0)e^{-s(\lambda-\lambda_0)} \]

\[ b_{bp}(\lambda) = b_{bp}(\lambda_0)(\frac{\lambda_0}{\lambda})^Y \]

where the subscript $\varphi$ stands for chlorophyll and $a0$ and $a1$ are given in Lee et al. (1998), the subscript $dg$ stands for the combination of CDOM and detritus and $S$ is the spectral slope for their absorption coefficient, $\lambda_0$ is the reference wavelength and $Y$ is the power law exponent for the particle backscattering coefficient.
3 \( K_d \) retrieval model – Inverse scheme

Initial values of \( a_\varphi(440), a_d(\lambda_0), b_{bp}(\lambda_0) \) and \( Y \) were adopted from Lee et al. (1999) based on band-ratios of \( R_{rs} \), and \( S \) is set equal to 0.015 \( \text{nm}^{-1} \). In this way, no empirical coefficients tuning and field data except the measured \( R_{rs} \) are required in the inversion scheme.

For each set of values for the four unknowns, a spectral optimization program using nonlinear curve fitting to compute the difference between model-simulated \( R_{rs} \) curves and field measured \( R_{rs} \) curves iterates until the minimum is reached. At that point, the absorption and backscattering coefficients are considered to be retrieved.
The downwelling attenuation coefficient of diffuse and direct light is numerically computed as:

\[ K_d = - \frac{1}{E_d} \frac{dE_d}{dz} \]

The total downward planar irradiance of diffuse and direct light is:

\[ E_d(z) = E^-(z) + E_s(0)e^{kz} \]
1 AC for GOCI images

2 $K_d$ models validation and inter-comparison

3 GOCI-derived $K_d$ maps
Results

- 1 AC for GOCI images

Field measured $R_{rs}$ (black line) against best-fitted atmosphere corrected $R_{rs}$ from GOCI images (red circles) for seven matchup sites.
1 AC for GOCI images

Root mean square error (RMSE) between field measured $R_{rs}$ and GOCI atmospherically corrected values for the ten atmosphere conditions (letters refer to aerosol types whereas numbers to visibilities, i.e. U5 refers to Urban aerosol at 5 km visibility).

<table>
<thead>
<tr>
<th>Case</th>
<th>M5</th>
<th>M10</th>
<th>M20</th>
<th>M30</th>
<th>M40</th>
<th>U5</th>
<th>U10</th>
<th>U20</th>
<th>U30</th>
<th>U40</th>
</tr>
</thead>
<tbody>
<tr>
<td>RMSE</td>
<td>0.0311</td>
<td>0.0104</td>
<td>0.0088</td>
<td>0.0106</td>
<td>0.0117</td>
<td>0.0549</td>
<td>0.0313</td>
<td>0.0221</td>
<td>0.0198</td>
<td>0.0188</td>
</tr>
</tbody>
</table>
2 $K_d$ models validation and inter-comparison

(i) Zhang model:

If $\frac{R_{RS}(490)}{R_{RS}(555)} \geq 0.85$, then $K_d(490) = 10^{(-0.843-1.459X-0.101X^2-0.811X^2)} + 0.016$

where $X = \log_{10}\left(\frac{R_{RS}(490)}{R_{RS}(555)}\right)$

If $\frac{R_{RS}(490)}{R_{RS}(555)} < 0.85$, then $K_d(490) = 10^{(0.094-1.302X+0.247X^2-0.021X^2)} + 0.016$

Where $X = \log_{10}\left(\frac{R_{RS}(490)}{R_{RS}(665)}\right)$
Results

2 $K_d$ models validation and inter-comparison

(ii) Lee model:
Total absorption and backscattering coefficient are first derived from remote sensing reflectance by QAA_v5 model (Lee et al., 2002), which is supposed to be more accurate for turbid waters is used in this study (available at http://www.iocccg.org/groups/Software_OCA/QAA_v5.pdf). $K_d$ is then calculated according to (Lee et al. 2005d):

$$K_d(490) = m_0(z, \theta_s)a(490) + m_1(z, \theta_s)(1 - m_2(z, \theta_s)e^{-m_3(z, \theta_s)a(\lambda)})b_b(490)$$

where $\theta_s$ is sun zenith angle, $m_0, m_1, m_2$ and $m_3$ are adopted from Lee et al. (2005d, their Table 2) for corresponding sun zenith angles at the surface layer.
Results

- 2 $K_d$ models validation and inter-comparison

Known versus derived $K_d(490)$ for (a) Zhang model, (b) Lee model and (c) 2SeaColor model, applied to all three data sets. The mini graphs show the results for low $K_d(490)$ observations where $K_d(490) \leq 0.2$ m$^{-1}$. 
### Results

> 2 $K_d$ models validation and inter-comparison

Statistical results of three $K_d$ retrieval models with respect to observations divided into two subsets with a threshold of $K_d(490) = 0.2$ m$^{-1}$.

<table>
<thead>
<tr>
<th></th>
<th>All (N=1764)</th>
<th>$K_d(490) \leq 0.2$ m$^{-1}$ (N=1481)</th>
<th>$K_d(490) &gt; 0.2$ m$^{-1}$ (N=283)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Zhang</td>
<td>Lee</td>
<td>2SeaColor</td>
</tr>
<tr>
<td>Slope</td>
<td>0.769</td>
<td>1.210</td>
<td>0.911</td>
</tr>
<tr>
<td>Intercept</td>
<td>0.014</td>
<td>-0.012</td>
<td>0.008</td>
</tr>
<tr>
<td>$R^2$</td>
<td>0.898</td>
<td>0.782</td>
<td>0.928</td>
</tr>
<tr>
<td>rMAD</td>
<td>21.78</td>
<td>20.16</td>
<td>19.68</td>
</tr>
<tr>
<td>RMSE</td>
<td>0.113</td>
<td>0.174</td>
<td><strong>0.083</strong></td>
</tr>
<tr>
<td>F25</td>
<td>63.27</td>
<td>78.57</td>
<td>73.13</td>
</tr>
<tr>
<td>F100</td>
<td>99.72</td>
<td>97.28</td>
<td>99.32</td>
</tr>
</tbody>
</table>
Results

- $2K_d$ models validation and inter-comparison

In general, **advantages** of the 2SeaColor model are:

(i) no empirical relationships and coefficients are involved;
(ii) no preliminary knowledge on optical property of the interested region is required;
(iii) only the spectrum of remote sensing reflectance is necessary;
(iv) it provides consistently accurate estimations in both clear and turbid waters.

With these advantages, we can state that 2SeaColor has little limitations due to regional variability in optical properties.
Results

3 GOCI-derived $K_d$ maps

Field measured $K_d$ (490) versus GOCI-derived values at matchup sites by (a) the Zhang model, (b) the Lee model and (c) the 2SeaColor model.
Results

3 GOCI-derived $K_d$ maps

$K_d$ derived from atmospherically corrected GOCI image using the 2SeaColor model. Four wavelengths are shown: (a) 440nm, (b) 490nm, (c) 550nm and (d) 660nm.
3 GOCI-derived $K_d$ maps

Field measured spectrums of $R_{rs}(\lambda)$ and $K_d(\lambda)$ in the Yangtze data set.
Results

3 GOCI-derived $K_d$ maps

$K_d$ maps derived from atmospherically corrected GOCI image at wavelength 490 nm using (a) Zhang model, (b) Lee model and (c) 2SeaColor model (‘X’ marks are matchup sites).
Results

3 GOCI-derived $K_d$ maps

Why result from 2SeaColor is satisfied?

The range of $K_d$ (490) maps by 2SeaColor model is in accordance with field measurements from the East China Sea (Qiu et al., 2013).

Concentration of total suspended matter (TSM) in the Hangzhou Bay and the North Branch of Yangtze Estuary is actually 1 ~ 2 orders of magnitudes greater than the one at the matchup sites (Shen et al., 2010), which should results in a relatively large range of $K_d$.

On the other hand, both the Zhang model and Lee model present the trend of underestimation with the increasing of $K_d$ without a fine-tuning on the empirical coefficients as shown in this study and Jamet et al. (2012).
The 2SeaColor model is able to provide stable and accurate estimations of $K_d$ for wide ranges of water turbidity without the need for tuning coefficients.

Reasonable $K_d$ maps are also derived when applying the 2SeaColor to GOCI images reproducing the expected magnitude and range of $K_d$ in the Yangtze Estuary.

The successful application of 2SeaColor to GOCI images enables our further study of variations in the $K_d$ distribution on large spatial and temporal scales.
• ESA-MOST Dragon III
• SKLEC
• LOV

Thank you for your attention!