Using geostationary satellite ocean color data to map the diurnal dynamics of suspended particulate matter in coastal waters

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Abstract

Total suspended particulate matter (TSM) in coastal waters is often characterized by high concentration and significant diurnal dynamics. Insufficient spatial and temporal resolution limits both cruise sampling and polar-orbiting satellite remote sensing in the mapping of TSM diurnal dynamics in coastal regions. However, the in-orbit operation of the world’s first geostationary satellite ocean color sensor, GOCI, provides hourly observations of the covered area. In this study, we proposed a practical atmospheric correction algorithm for GOCI data in turbid waters. The validation results showed that the GOCI-retrieved normalized water-leaving radiances matched the in situ values well in both quantity and spectral shapes. We also developed a regional empirical TSM algorithm for GOCI data that is applicable in extremely turbid waters. Based on these atmospheric correction and regional TSM algorithms, we generated hourly TSM maps from GOCI Level-1B data. The diurnal variations derived by GOCI were a good match to the buoy data. The hourly GOCI observations revealed that various regions and tidal phases had different diurnal variation magnitudes, with a maximum of up to 5000 mg/l in central Hangzhou Bay. Strong wind events, such as typhoons, can significantly increase TSM in the bay; however, both the GOCI observations and buoy measurements indicated that this increase was episodic, had a short duration, and returned to normal within a day after the passage of a typhoon. Our results suggest that GOCI can successfully map the diurnal dynamics of TSM in turbid coastal waters. Moreover, the significant diurnal dynamics revealed in the hourly GOCI observations implied that caution should be taken in mapping TSM in coastal waters using cruise sampling and conventional polar-orbiting satellite data, as the temporal resolution is insufficient for capturing diurnal variations.

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1. Introduction

Coastal waters are often characterized by high concentrations of total suspended particulate matter (TSM) due to terrestrial inputs and sediment resuspension. Such matter in coastal waters exhibits high diurnal dynamics due to tidal action and high concentrations that can limit the available light underwater, affecting the productivity of upper-layer phytoplankton and the benthic ecosystem (Cloern, 1987; Miller & Cruise, 1995). The dynamics of TSM directly affect the transport of carbon, nutrients, pollutants, and other materials (Ilyina et al., 2006; Mayer et al., 1998), and thus monitoring these dynamics is of great interest and importance. The large spatial and temporal variability of TSM, however, makes it difficult to synoptically map such matter in coastal regions using traditional field sampling methods. Over the past decades, polar-orbiting satellite ocean color data have been used to map the TSM in various coastal regions (Fettweis et al., 2007; Hu et al., 2004; Mertes et al., 1993; Miller & McKee, 2004; Myint & Walker, 2002; Petus et al., 2010; Tassan, 1994; Warrick et al., 2004; Zhang et al., 2010), along with sensors such as the Sea-viewing Wide Field-of-view Sensor (SeaWiFS), the Moderate Resolution Imaging Spectroradiometer (MODIS), and the Medium Resolution Imaging Spectrometer (MERIS). While these polar-orbiting satellites can synchronously map TSM with sufficient accuracy, they can only observe one time per day, per satellite at mid-low latitudes, which is still far below the demand for diurnal variation monitoring. Compared with polar-orbiting satellites, geostationary satellites are uniquely capable of imaging the Earth’s surface from about 70°S to 70°N with a high revisit frequency; thus, they have a significantly greater temporal sampling capacity that greatly enhances our ability to monitor and assess the coastal ocean dynamics (IOC Report 12, 2012). A few recent studies have tried to use geostationary meteorological satellite data (e.g., the...
Spinning Enhanced Visible and Infrared Imager (SEVIRI) radiometer on board the METEOSAT Second Generation (MSG) satellite platform) to map diurnal TSM variations in coastal waters, demonstrating the advantages of geostationary satellites in monitoring highly dynamic coastal waters (Neukermans et al., 2009, 2012; Salama & Shen, 2010). However, geostationary meteorological satellites are designed to observe clouds and other meteorological information, which limits their ability to accurately monitor the low ocean color signals in coastal waters. This is particularly true in small-scale estuaries and bays, due to the satellites’ low radiometric sensitivities, fewer bands, and coarse spatial resolutions.

On June 27, 2010, the world’s first geostationary satellite ocean color sensor, the Geostationary Ocean Color Imager (GOCI), was successfully launched by South Korea. Compared with other polar-orbiting satellite ocean color sensors, GOCI is uniquely capable of monitoring short-term and regional oceanic phenomena (e.g., tide dynamics, red tides, river plumes, and sediment transport) with a high spatial resolution (500 m) and a very high temporal resolution (1 h, eight images per day) (Ryu et al., 2011), which makes it very useful for monitoring the diurnal dynamics of the material in coastal areas. To map the diurnal dynamics of the TSM or other materials in turbid coastal waters using GOCI, two key issues must be solved: the atmospheric correction in turbid waters and the regional algorithm for TSM (or another material) for GOCI.

Exact atmospheric correction in turbid waters remains challenging in ocean color remote sensing. To solve the atmospheric correction problem in turbid waters, several methods have been proposed based on the assumptions of spatial homogeneity of water’s optical and aerosol properties (Hu et al., 2000; Ruddick et al., 2000). A more appropriate approach is to use short-wave infrared (SWIR) bands instead of the near infrared (NIR) bands in highly turbid waters (Wang & Shi, 2005). However, GOCI has no SWIR band. Wang et al. (2012) recently proposed a regional NIR atmospheric correction algorithm for GOCI data in the Western Pacific region, including turbid waters, based on the regional empirical relationship between spectral normalized water-leaving radiiances and the diffuse attenuation coefficient at 490 nm. These turbid water atmospheric correction methods are strongly dependent on local water and aerosol optical properties. However, in many coastal waters with tidal dynamics, terrestrial inputs, and anthropogenic aerosol effects, the water and aerosol optical properties are often highly heterogeneous.

In general, the inversion of TSM can be analytic, semi-analytic and empirical. However, due to the high variation in TSM components and water optical properties in coastal waters (Binding et al., 2005), it remains difficult to implement applicable analytic and semi-analytic models (Zhang et al., 2010). Numerous empirical algorithms have been developed for various estuaries and coasts based on the relationship between TSM and the band ratio for the remote sensing reflectance (or even one band) of different satellites (Curran et al., 1987; Doxaran et al., 2002a, 2003; Larouche & Boyer-Villemaire, 2010; Tassan, 1994; Zhang et al., 2010). Such empirical algorithms are highly reliant on in-situ datasets, and thus are only regionally applicable.

In this study, taking Hangzhou Bay (HZB) as a sample region, we aim to develop and validate the atmospheric correction and regional TSM algorithms to map TSM diurnal dynamics using GOCI data in turbid coastal waters. Driven by strong tides and huge terrestrial input, the HZB waters are extremely turbid and exhibit significant diurnal dynamics (Xie et al., 2009), which makes them ideal for testing the mapping ability of geostationary satellite as applied to TSM diurnal dynamics in coastal waters.

2. Data and methods
2.1. Study site
HZB is the largest bay along the southeastern coast of China, located immediately south of the mouth of the Changjiang River, as shown in Fig. 1(a). HZB is a wide, shallow, funnel-shaped estuary covering an area of approximately 8500 km², and is about 100 km wide at the mouth and 86 km long (Guo et al., 2009). The average water depth in the bay is 8–10 m during low tides. HZB is one of the world’s most strongly tidal bays, with an amplitude of 3–4 m at the mouth,
4–6 m further upstream and a maximum of 9 m at the top (Lin et al., 2005; Xie et al., 2009). Two large rivers, the Changjiang and the Qiantang, drain into the bay. The Changjiang is the world’s fifth largest river in terms of volume discharge, and the fourth largest in terms of sediment load. Its average water discharge is 925 km$^3$ yr$^{-1}$, with a sediment load of 4.8 × 10$^8$ t yr$^{-1}$. A portion of the water and sediment from the Changjiang Estuary enters the HZB, profoundly influencing both its hydrodynamics and its sedimentation (Xie et al., 2009). The Qiantang is the major river that directly inputs the HZB, with an average water discharge of 42 km$^3$ yr$^{-1}$ and sediment load of 7.9 × 10$^6$ t yr$^{-1}$ when considered together with the smaller rivers surrounding the HZB (Xie et al., 2009).

2.2. Field samples

Two cruises with almost identical sampling stations were conducted in the East China Sea during the “908 Project” initiated by the China’s State Oceanic Administrator (SOA). The two cruises covered 120–128°E and 29–33°N, including the HZB, the Changjiang River Estuary and the East China Sea shelf, as shown in Fig. 1(b). The summer cruise occurred from July 18 to August 23, 2006, and the winter cruise occurred from December 23, 2006 to February 4, 2007. The TSM, the remote sensing reflectance and other environmental parameters were synchronously measured at the sampling stations on both cruises.

2.2.1. Measurement of TSM

TSM was measured gravimetrically on pre-weighed cellulose acetate membrane filters (47 mm diameter, 0.45 μm pore size). Sufficient samples of surface-layer water (fixed at a 2 m depth) were collected by a rosette Niskin sampler and filtered at each station to determine the weight of the particulate matter. The samples were rinsed three times using 50 ml distilled water to remove dissolved salts. The filters were frozen and stored below −20 °C until laboratory processing. In the laboratory, the filters were dried in an oven at 40 °C for 6–8 h, then placed in the silica gel drier for 6–8 h, and finally re-weighed to obtain the TSM concentration. There was at least one blank sample at each station to make the blank correction. A total of 377 TSM samples were taken, not only during the day, but also at night: 190 from the summer cruise and 187 from the winter cruise.

2.2.2. Measurement of remote sensing reflectance

The remote sensing reflectance was measured aboard ship using an ASD FieldSpec HandHeld Spectroradiometer (FSHH, 325–1075 nm). At the daytime stations, under suitable solar illumination (generally from 9:00 AM to 3:00 PM), we measured the upward radiance from the sea’s surface ($L_t$) and the standard reflecting plate ($L_p$), and the downward sky radiance ($L_{sky}$). To avoid sun-glint contamination, the zenith and azimuth angles used to measure $L_t$ were about 40° and 135° (referring to solar orientation), respectively. In addition, we selected optimal place to minimize the effects of ship shading and foam. We set the integrating time according to the incident radiance intensity, and corrected the dark current for each change in integrating time. Before the cruise, an absolute calibration was performed in the laboratory using an NIST traceable lamp. According to the radiances measured by ASD FSHH with units of mW/(cm$^2$·μm·sr), the remote sensing reflectance ($R_{rs}$ with unit of sr$^{-1}$) can be calculated as

$$R_{rs}(λ) = \frac{β_{p}(λ) [L_t(λ) − β_s(λ)L_{sky}(λ)]}{\pi L_p(λ)}$$

where $β_{p}$ and $β_{s}$ are the standard plate and air–sea interface reflectance, respectively. $β_{s}$ is influenced by sky conditions, sea-surface roughness, and viewing and illumination geometries, and may also be wavelength dependent due to the different angle distributions of the downward sky radiance.
radiance just above the sea’s surface at different wavelengths. However, for the above-water \( R_w \) measurement with zenith and azimuth angles of about 40° and 135° under clear or overcast sky conditions, the wavelength dependence of \( \beta_s \) is small (Mobley, 1999). In general, \( \beta_s \) varies from 0.022 to 0.05 (Lee et al., 1996). In this study, we estimated \( \beta_s \) by assuming a black ocean at the referenced NIR wavelengths and wave-length independence (Doxaran et al., 2002b). It is worth noting that the referenced NIR wavelengths changed with water turbidity; that is, longer NIR wavelengths were used for higher turbidity. In addition, in extremely turbid waters, \( \beta_s \) is fixed at 0.05 when the estimated value is larger than 0.05. Because remote sensing reflectance can only be measured during the day and under clear skies, the effective number was much less than that of the TSM measurement, with 35 measurements taken during the summer cruise and 31 during the winter cruise.

In addition to the two aforementioned cruises, a buoy designed to monitor the ocean environment was moored at the mouth of the HZB (at 122.3625°E and 30.5509°N, as shown in Fig. 1(c)) by the Shanghai Oceanic Central Meteorological Observatory. At a frequency of 3 h, the buoy instantaneously measured the surface water salinity and temperature (Alex Compact-CTW with accuracies of 0.05 °C and 0.05 mS/cm conductivity), turbidity (Alec CLW using the infrared back-scattering method with an accuracy of 2%), current (Alec AEM with accuracies of 1 cm/s and 2° for velocity and direction, respectively), wind (R. M. Young, with accuracies of 0.3 m/s and 3° for speed and direction, respectively) and other ocean parameters (e.g., chlorophyll, aquatic pCO2). The topography around the buoy was very complex due to the presence of islands to the north and south sides, and isobaths to the west and east that varied quickly as Fig. 1(c) illustrates.

We took measurements near the buoy from June 26 to 27, 2011 to calibrate the buoy data (primarily for the aquatic pCO2). We arrived at station A (at 122.3610°E and 30.5503°N, as shown in Fig. 2(a)) near the buoy at about 9:00 PM on June 26, and measured the TSM (during the day and night) and water-leaving radiance (only during the day) until 12:00 PM on June 27 at a frequency of 3 h to match the buoy measurements. Thus, there were six TSM samples and two water-leaving radiance samples at station A. At about 12:30 PM, we completed the measurements at station A and began moving westward to station B (at 122.1200°E and 30.5463°N, as shown in Fig. 2(a)). We arrived at station B at about 3:00 PM, and took one TSM sample and one water-leaving radiance sample. It should be noted that just before our measurement, Typhoon Meari passed through the mouth of HZB on the evening of June 25, which might have changed the environmental conditions within the bay.

### 2.3. GOCI data

GOCI has eight visible to near-infrared bands (412 nm, 443 nm, 490 nm, 555 nm, 660 nm, 680 nm, 745 nm and 865 nm) with high signal-to-noise ratios, enabling more accurate retrieval of TSM and other ocean color information. The GOCI observing area is about 2500 km × 2500 km (116°E–143.92°E, 24.75°N–47.25°N for central directions) centered on 130°E and 36°N, covering the coasts of Eastern China, the Korean peninsula, and Japan, along with the corresponding shelves and open oceans (Ryu et al., 2011).

We applied four days (April 5 and June 24, 26 and 27, 2011) of GOCI Level-1B data from the Korea Ocean Satellite Center (KOSC). The GOCI data from 5 April 2011 were applied due to low cloud coverage, and the other 3 days of GOCI data corresponded to our cruise date in the HZB. There were eight observations from 8:28 AM to 3:28 PM local time, at a frequency of 1 h, for each day of GOCI Level-1B data. The GOCI Level-1B data contain the total radiances at the top of the atmosphere received by GOCI over eight bands for each time observation. Solar and sensor geometries were obtained from the Level-2 files, which were generated from the Level-1B files using the GOCI Data Process Software (GDPS) distributed by KOSC.

### 3. Algorithm development and validation

#### 3.1. Atmospheric correction of GOCI data in turbid waters

In 2004, we proposed a practical atmospheric correction algorithm for turbid waters using ultraviolet wavelengths (UV-AC) (He et al., 2004) and based it on the assumption that the water-leaving radiance at ultraviolet wavelengths could be neglected compared with longer wavelengths. In turbid coastal waters, water-leaving radiances increase largely at the visible light (VIS) and NIR due to the strong particulate scattering caused by high TSM concentrations. However, strong absorptions by high concentrations of detritus and CDOM cause a rapid decrease in the blue part of the VIS, and still more in the UV. In some extremely turbid waters, the water-leaving radiance at the UV may be much less than at the NIR. In such cases, the UV band is better than the NIR band at obtaining a more accurate atmospheric correction. As a result, the water-leaving radiance at the UV band can be neglected compared with that at the VIS and NIR, and thus the UV band can be used to estimate the aerosol scattering reflectance. In this study, we apply the UV-AC algorithm to the GOCI Level-1B data (see the Appendix A for details), but because GOCI has no UV band, we use the 412 nm band as the referenced band to retrieve the remote sensing reflectance for the further inversion of TSM.

The UV-AC algorithm’s performance has been validated in various estuary waters (the Changjiang River Estuary, the Mississippi River Estuary, and the Orinoco River Estuary) with strong spatial and temporal variability of both CDOM and detritus absorptions (Tzortziou et al.,...
The results suggest that the UV-AC algorithm is quite promising for the retrieval of water-leaving reflectance at longer VIS and NIR wavelengths, given its adept retrieval of TSM in turbid coastal waters (He et al., 2012). In highly turbid waters with a water-leaving reflectance at 865 nm (including the atmosphere diffuse transmission coefficient) larger than 0.01, the average absolute relative errors of the water-leaving reflectance retrieved by the UV-AC algorithm were 30.8% (412 nm), 22.9% (443 nm), 15.9% (490 nm), 13.1% (510 nm), 8.9% (555 nm), 6.0% (670 nm), 11.5% (765 nm) and 13.7% (865 nm). The validation results also reveal that using the 412 nm band when working with polar-orbiting satellite ocean color sensors that lack the UV band, such as SeaWiFS, MODIS, and MERIS, proved promising in turbid waters (He et al., 2012). The UV-AC algorithm’s performance was also stable across a day’s changing illuminations (He et al., 2012).

Here, we further validate the GOCI-retrieved \( L_{\text{wn}} \) using the UV-AC algorithm by in situ values. There are three match-up samples of \( L_{\text{wn}} \), with two samples at station A and one sample at station B (Fig. 2(a)). Taking advantage of the hourly GOCI observations, we match the GOCI retrieval with the in-situ measurements according to the closest time. For station A, we compare the GOCI-retrieved \( L_{\text{wn}} \) at 9:28 AM and 12:28 PM with the in-situ values measured at 9:16 AM and 12:08 PM on June 27. For station B, we compare the GOCI-retrieved \( L_{\text{wn}} \) at 3:28 PM with the in situ value measured at 3:21 PM on June 27. Moreover, because the 865 nm is too close to the referenced NIR wavelengths to estimate the air–sea interface reflection coefficient, introducing relatively larger noise for the ASD measurement in the moderately turbid waters, we only compare the \( L_{\text{wn}} \) at the first seven bands of GOCI, as illustrated in Fig. 2(b)–(d). In general, the GOCI-retrieved \( L_{\text{wn}} \) matches the in situ values well in both quantity and spectral shape, with average absolute relative errors of 25.0% (412 nm), 11.8% (443 nm), 9.9% (490 nm), 6.6% (555 nm), 13.9% (660 nm), 6.8% (680 nm) and 29.1% (745 nm) indicating that the UV-AC algorithm is capable of retrieving the water-leaving radiance for turbid waters in the HZB. The results for station A are similar to the simulative validation, with only slight overestimation at the shorter wavelengths and underestimation at 865 nm (He et al., 2012). However, for station B, the GOCI-retrieved values underestimate the \( L_{\text{wn}} \) for all wavelengths. This may be due to the small-scale patch of water turbidity around this station, which was very close to the “Xiaoyangshan” international harbor (as shown in Fig. 1(c)) in Shanghai, China’s biggest city. Cui et al. (2010) used the 13 strict match-ups in the coastal waters of the Bohai Sea to determine that
the average relative errors of MERIS-retrieved \( L_{\text{wn}} \) are 20% (412 nm), 18% (443 nm), 15% (490 nm), 15% (510 nm), 16% (560 nm), 17% (620 nm), and 18% (665 nm). Therefore, the accuracy of the GOCI-retrieved \( L_{\text{wn}} \) using the UV-AC algorithm in the HZB is better than that of the MERIS-retrieved \( L_{\text{wn}} \) in the Bohai Sea. Overall, the consistencies of the \( L_{\text{wn}} \) between GOCI retrieval and in situ measurement are quite encouraging, despite the existence of uncertainties.

### 3.2. Regional TSM algorithm for GOCI in the HZB

Several empirical TSM algorithms have been developed in the Changjiang River Estuary. Using a cruise section in the mouth of the Changjiang River, Han et al. (2006) found a linear relationship between the TSM on the logarithmic scale and the combined index of the 550 nm and 670 nm bands of the Chinese Moderate Resolution Imaging Spectroradiometer (CMODIS). Zhang et al. (2010) also developed an empirical TSM algorithm for MODIS using three bands (488 nm, 555 nm and 645 nm) and based on an in situ dataset covering the outer mouth and shelf of the Changjiang River. It is worth noting that the in situ TSM concentrations used by Han et al. (2006) and Zhang et al. (2010) (Han and Zhang, hereafter) to develop their empirical models were less than 1000 mg/l and did not cover the HZB, which has the highest water turbidity with a maximal TSM concentration of up to 5000 mg/l. Bai et al. (2011) recently found that Han and Zhang’s TSM models had limited application in extremely turbid waters such as those of the Changjiang River Estuary and HZB. Moreover, Bai et al. (2011) noted that it would be better to use the near-infrared wavelength to retrieve the TSM in the HZB, and developed an empirical TSM model for MERIS using the ratio of normalized water-leaving radiiances between 779 nm and 560 nm.

The in situ TSM and \( R_{\text{rs}} \) at the GOCI bands measured during the summer and winter cruises of the “908 Project” indicate a good linear relationship between the TSM on the logarithmic scale and the band ratio of \( R_{\text{rs}} \) at 745 nm and 490 nm from moderately to extremely turbid waters (TSM from 8 mg/l to 5275 mg/l) in the Changjiang River Estuary and HZB (Fig. 3):

\[
\text{TSM} = 10^{1.0758 - 1.1230 \cdot \text{Ratio}}, \quad \text{(R = 0.96, SD = 0.22, N = 41, P < 0.0001),}
\]

\[
\text{Ratio} = \frac{R_{\text{rs}}(745 \text{ nm})}{R_{\text{rs}}(490 \text{ nm})}
\]

where \( R \) and SD are the correlation coefficient and standard deviation, respectively. The deviations of the regressed offset and slope in Eq. (2) are 0.0676 and 0.0536, respectively.
As Fig. 3(b) reveals, the prediction error from Eq. (2) is small for a TSM concentration ranging from 8 mg/l to 300 mg/l, and relatively large for extremely turbid waters with a TSM concentration larger than 300 mg/l. The average relative bias and absolute relative errors are 13.3% and 25.2% for TSM ≤ 300 mg/l, and 14.2% and 56.0% for TSM > 300 mg/l, respectively. We expect that there are two explanations for the relatively large difference between the predicted and in situ TSM in highly turbid waters. First, the composition, size, and shape of the particles are natural complexity in highly turbid waters. It is very necessary and interesting to examine the particles’ composition and size, and this is beyond the scope of this study due to data limitations. Second, it is very difficult to exactly match the TSM and reflectance in terms of the measurement time. Although we tried to measure the reflectance and take the TSM sample simultaneously at each station, there were unavoidable small time differences. Even a minor time difference (several minutes) may cause a significant difference in the TSM concentration of highly turbid waters due to small spatial–temporal scales such as sediment flocculation. In addition, TSM measurement errors (including water volume measurement, weighting and sedimentation in the sampling bottles) might contribute to the prediction errors in highly turbid waters. The effect of weighting errors should be less important for high TSM concentrations due to the larger weight of sediment on the filters. During the filtering process on board, we stirred the collected water samples fully to avoid sedimentation, and thus its effect is expected to be small enough to be ignored. However, an error caused by water volume measurement might be significant because less water sample volume was used in highly turbid waters.

Although extremely turbid waters present uncertainties, the TSM algorithm in Eq. (2) is encouraging because it covers large TSM concentration ranges from 8 mg/l to 5275 mg/l. Aside from the uncertainty of the TSM algorithm itself, the accuracy of the TSM retrieval is also affected by the accuracy of the ratio \( R_{rs}(745 \text{nm})/R_{rs}(490 \text{nm}) \), which is retrieved through atmospheric correction. If the satellite-retrieved \( R_{rs}(745 \text{nm})/R_{rs}(490 \text{nm}) \) ranging from 0.0 to 2.0 (TSM from 11.9 to 2098 mg/l) has ±20% errors, then the average relative errors of the TSM retrieval by Eq. (2) are 32.1% for TSM ≤ 300 mg/l, and 96.7% for TSM > 300 mg/l.

![Fig. 6. Same as Fig. 4, but for 26 June 2011, with the black areas covered by clouds.](image-url)
4. Results and discussion

4.1. TSM diurnal variations in HZB mapped by GOCI data

Using the $R_{rs}$ retrieved by the atmospheric correction algorithm and the regional TSM model following Eq. (2), we obtained hourly maps of the TSM concentration in the HZB, courtesy of the GOCI data on April 5 and June 24, 26 and 27, 2011 (Figs. 4–7, respectively). Under hourly GOCI observations, the TSM diurnal variation in the bay is evident. To further quantify this TSM diurnal variation, we selected five representative points from the top to the mouth of HZB (Fig. 8). The different locations throughout HZB clearly exhibited different diurnal variation magnitudes. As curve C in Fig. 8(b) reveals the maximum magnitude was up to 5000 mg/l in the central part of the bay. The minimum variation occurred in the northwestern part of the bay (corresponding to curve B in Fig. 8(b)) due to the relatively deeper water depth. Along the northern bank, there is a deep tidal channel with a water depth of up to 13 m, and Point B is located in this deep channel (Fig. 1(c)).

Interestingly, both the TSM concentration and the magnitude of the diurnal variation on April 5 were an order larger than they were on June 24, 26 and 27. On April 5, the TSM concentration throughout the bay was greater than 100 mg/l, and greater than 1000 mg/l in most areas. However, on June 24 and 27, the TSM was generally less than 100 mg/l. On June 27, the TSM was generally less than 50 mg/l, and correspondingly the diurnal variation magnitudes on June 24 and 27 were much smaller compared with those on April 5. It is notable that the TSM increased slightly on June 26 compared to June 24 and 27, due to the passage of Typhoon Meari on June 25. We discuss the reasons for these TSM differences and the effects of the typhoon in Section 4.3.

4.2. Comparison of TSM diurnal variations mapped by GOCI and buoy measurement

To examine how well GOCI captures the TSM dynamics in HZB, we compared the GOCI-retrieved TSM with the water turbidity measured by the buoy (Fig. 9). The diurnal variations of the TSM retrieved by GOCI were a good match with the buoy-measured water turbidity. To quantitatively validate the former, we must convert the latter to a TSM concentration.

Water turbidity is often used as a proxy for TSM concentration, but different water turbidity sensors calibrated with the same turbidity standard in the lab may not give the same results in the field, and...
the correlation between turbidity and TSM is most often sensor specific (Boss et al., 2009; Neukermans et al., 2012). Using the matched shipboard-measured TSM at station A and the buoy-measured water turbidity from June 26 to 27 (as shown in Fig. 10(a)), we reveal a good linear relationship between turbidity and TSM (Fig. 10(b)), with a linear regression line described by

\[
\text{TSM} = -0.72 + 1.45 \times \text{Turbidity} \quad (R = 0.94, SD = 2.13, N = 5, P = 0.0187).
\]

(3)

The deviations for the regressed offset and slope in Eq. (3) are 3.02 and 0.31, respectively. It is notable that our June 26–27, 2011 HZB cruise occurred at neap tide, when the water was relatively clear with a maximum TSM of only 21.07 mg/l near the buoy station. Therefore, the extrapolation to high TSM using Eq. (3) may be limited, suggesting that more synchronous measurements of the TSM and buoy turbidity should be conducted to ensure the confident calibration of buoy turbidity in the future. Here, we expect the linear relationship between buoy turbidity and TSM concentration to be steady and applicable across the TSM range.

Using Eq. (3), we convert the buoy-measured water turbidity to TSM concentration, and compare it with the GOCI-retrieved TSM (Fig. 11). Fig. 11(a) and (b) shows that all of the diurnal variations in the TSM retrieved by GOCI match the buoy measurements quite well, except at 9:28 AM on June 24, indicating that GOCI can reliably map the diurnal dynamics of TSM in the bay. The anomaly of the GOCI-retrieved TSM at 9:28 AM on June 24 was caused by thin cloud contamination. As Fig. 11(c) illustrates, a strip of cloud passed through the buoy station, although it was relatively thin and not masked by the atmospheric correction. Fig. 11(d) presents the scatter plot of all of the matched TSM between GOCI retrieval and buoy measurement. Compared to the buoy measurement, the average relative error of the GOCI-retrieved TSM is 10.9% on the logarithmic scale. Moreover, the buoy-derived TSM on April 5 was higher by more than an order of magnitude, compared to that on June 24, 26 and 27. The buoy-derived TSM was also slightly higher on June 26 than that on June 24 and 27, supporting the results derived from the GOCI-retrieved TSM maps.

4.3. Factors controlling the TSM diurnal dynamics in the HZB

Numerous factors can affect temporal variations of the TSM in HZB including the tide, winds, and river runoff (Su & Wang, 1989). In general, the TSM concentrations inputted by the Changjiang and Qiantang Rivers are much higher in the wet season with large discharge from May through October, than in the dry season with low discharge from November through April (Chen et al., 2006; Mao et al., 2010). Fig. 12(a) and (b) shows the daily average discharges of the Changjiang and Qiantang Rivers in April and June 2011. The daily Changjiang River discharge from June 24 to 27 was about three times larger than that on April 5. The Qiantang River discharge in June 2011 was similarly much higher than that in April 2011. However, as both the GOCI retrieval and

![Fig. 8. The TSM diurnal variation in the HZB retrieved by GOCI. (a) The locations of selected representative points in the HZB, and the diurnal variations of TSM at representative points on (b) 5 April 2011, (c) 24 June 2011, (d) 26 June 2011 and (e) 27 June 2011.](image-url)
buoy measurements reveal, although river discharges were much higher in June than they were in April, the TSM concentration from June 24 to 27 was an order lower than that on April 5, indicating that the river discharge is not a dominant factor controlling the magnitude of TSM diurnal variation in the bay. The lack of covariance between the water turbidity and wind speed measured by the buoy (Fig. 12(c)) also indicates that normal wind is not a main factor controlling TSM diurnal variation in the bay.

As Fig. 12(d) shows, water turbidity covaries closely with salinity and tidal elevation, suggesting that the tide is the main factor controlling TSM diurnal variation in the bay. In fact, the surface water salinity covaries with the tidal status in the bay. During the flood tide, the salinity increases along with the tidal elevation due to the intrusion of shelf water with high salinity; however, the salinity decreases during the ebb tide. In addition, previous studies using the model simulation and in situ measurement have also revealed that the diurnal variations in salinity are controlled by the tides in the Changjiang River Estuary and HZB (Rong & Li, 2012; Wang et al., 2007; Wu et al., 2010), with high and low salinity corresponding to the flood and ebb tides, respectively. Aside from the tidal-induced TSM diurnal variation, episodic strong wind events such as typhoons may enhance vertical mixing and sediment resuspension, and thus can systematically increase TSM.

4.3.1. The effect of spring-neap tides on TSM diurnal variation

Fig. 13(a) and (b) shows the hourly tidal elevations from the tidal tables of the “Tanhu” tide gauge station (Fig. 1(c)) in April and June 2011, respectively. A spring tide occurred on April 5 and a neap tide occurred from June 24 to 27. A large tidal current during spring tide can have a maximum velocity of up to 2.2 m/s (Fig. 13(c)) and can induce strong bottom shear stress and enhance sediment resuspension, which is likely what caused the high turbidity in the bay on April 5. In contrast, a tidal current during neap tide is much smaller, with weak bottom shear stress and sediment resuspension, which explains the lower turbidity on June 24, 26 and 27. In addition, Rong and Li (2012) used a three-dimensional hydrodynamic model to determine that the water column was strongly stratified during neap tides but became weakly stratified during spring tides in the mouth of the Changjiang River, whereas for some phases of the spring tide the stratification was completely destroyed, favoring sediment resuspension. Therefore,

**Fig. 9.** Comparison between GOCI-retrieved TSM and the buoy-measured water turbidity, (a) on 5 April 2011 and (b) from 24 to 27 June 2011.

**Fig. 10.** Calibration of the buoy-measured water turbidity by the shipboard-measured TSM. (a) Comparison between the buoy measured turbidity and cruise-measured TSM. (b) Linear regression between the buoy-measured turbidity and cruise-measured TSM.
TSM diurnal variation typically varies with the spring–neap cycle, with greater amplitudes during spring tides and smaller amplitudes during neap tides.

4.3.2. Episodic TSM variations caused by a typhoon

Typhoon Meari passed northward through the mouth of the HZB between June 24 and 27 2011. Meari was a relatively weak typhoon with a maximum wind speed of about 30 m/s. On June 23, the typhoon was in the southeastern East China Sea. It moved swiftly north-by-northwest and arrived at the mouth of HZB at 9:00 PM local time on June 25 before quickly moving northward at speeds of more than 30 km/h.

During the passage of Typhoon Meari, the TSM in HZB increased rapidly and significantly. The buoy-derived TSM shown in Fig. 14 reveals that TSM increased from 30 mg/l to 180 mg/l when the typhoon passed through the mouth of the bay. It is worth noting that the maximum wind speed recorded by the buoy was significantly lower than that recorded in the central part of the bay, due to the islands along the north and south sides of the buoy station, which interrupted air flow and reduced wind speeds. Compared with the buoy measurements at one station, the GOCI-retrieved TSM had the advantage of synchronously mapping the dynamics covering the whole bay. Taking advantage of the hourly observations, GOCI clearly captured the increase of TSM in the southern part of the bay on June 26 (Fig. 6) compared to that on June 24 (Fig. 5). Moreover, the turbid waters in the southern part of the bay caused by the typhoon decreased significantly from morning to afternoon, and had almost returned to normal by mid-afternoon on June 26 (referring to the distribution on 27 June) once the typhoon had passed (Fig. 6). The quick recovery of the increased TSM caused by the typhoon is also evidenced by the buoy-derived TSM data revealed in Fig. 14. Therefore, typhoon-induced TSM increases in the bay are expected to be episodic and of short duration, possibly caused by the rapid deposition of resuspended sediment.

5. Conclusions

In this study, we took HZB as a sampling region and quantitatively applied and validated GOCI data to map the diurnal dynamics of TSM in turbid coastal waters. We validated the UV-AC algorithm for HZB and developed a regional TSM algorithm that was used to study the diurnal dynamics in HZB using GOCI. Tide is the main driver of the diurnal spatial and temporal variability, and episodic events have a strong influence, but temporally limited influence.

The geostationary ocean color satellite is ideal for environmental monitoring (e.g., water quality, pollution, oil spills, harmful algal blooms, sediment and carbon transport, air–sea carbon flux, and

Fig. 11. Comparison between GOCI-retrieved and buoy-derived TSM. (a) Comparison on 5 April 2011; (b) comparison on 24, 26, and 27 June 2011; (c) pseudo color image of the GOCI Level-1B data at 9:28 AM on 24 June 2011 (R: 660 nm, G: 555 nm: B: 412 nm); (d) scatter plot comparison of the matching TSM between the GOCI retrievals and buoy measurements on 5 April 2011 and from 24 to 27 June 2011.
oceanic engineering) in highly dynamic coastal waters, thanks to a much higher sampling frequency than polar-orbiting ocean color satellites and higher spectral, spatial, and radiative resolutions than geostationary meteorological satellites. In addition, the hourly observations may improve the satellite’s estimation of ocean primary production and the consequences of carbon fluxes. Moreover, these high frequency observations may greatly enhance our ability to monitor and assess the oceanic dynamics by providing insight into their processes.

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Fig. 12. (a) The daily averaged discharge of the Changjiang River measured at the Datong hydrological station during April and June 2011 (from http://xxfb.hydroinfo.gov.cn). (b) The daily averaged discharge of the Qiantang River measured at the Lanxi hydrological station during April and June 2011 (from http://www.qgj.cn). (c) The relationship between water turbidity and wind speed measured by the buoy from 4 to 6 April 2011. (d) The relationship between water turbidity and salinity measured by the buoy, and the tidal elevation from the tidal tables of the “Tanhu” tide gauge station from 4 to 6 April 2011.
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Appendix A. UV-AC algorithm for GOCI data

The details of the UV-AC algorithm for GOCI data are as follows:

1. Calculating the Rayleigh-scattering corrected reflectance ($\rho_{rc}$) for all bands as

$$\rho_{rc}(\lambda) \equiv \rho_t(\lambda) - \rho_r(\lambda) = \rho_a(\lambda) + t_v(\lambda) \rho_w(\lambda), \quad (A1)$$

where $\rho_t$ is the total reflectance at the top of the atmosphere as measured by GOCI after correcting for sea-surface whitecaps and sun glint; $\rho_r$ is the Rayleigh-scattering reflectance by air molecules in the absence of aerosols; $\rho_a$ is the aerosol scattering reflectance including the interactive scattering between air molecules and aerosol; $\rho_w$ is the desired water-leaving reflectance; $t_v$ is the diffuse transmission coefficient from the sea surface to the satellite sensor; and $\lambda$ is the wavelength.

The definition of the reflectance in Eq. (A1) is

$$\rho(\lambda) = \pi L(\lambda)/F_0(\lambda) \cos \theta_0, \quad (A2)$$

where $L$ is the upward radiance, $F_0$ is the extraterrestrial solar irradiance and $\theta_0$ is the solar zenith angle. The purpose of the atmospheric correction is to retrieve $\rho_w$ from $\rho_t$. The Rayleigh-scattering component $\rho_r$ can be accurately calculated using a
vector radiative transfer model along with knowledge of the atmospheric pressure and wind speed at the sea’s surface (Gordon & Wang, 1992; Gordon et al., 1988). Here, we use the general Rayleigh-scattering look-up table generated by He et al. (2006) to estimate \( \rho_s \) for all GOCI bands. The exact calculation of \( \tau_s \) must consider the Rayleigh, aerosol, and absorption gas transmittances. However, for the GOCI observations in HZB, the viewing zenith angle is generally less than 60°, and the aerosol transmittance can be neglectable (Gordon et al., 1983). Moreover, for the GOCI bands, gas absorption is largely contributed by ozone, and the absorptions by other gases can be neglected. Therefore, \( \tau_s \) can be estimated by Rayleigh optical thickness and ozone concentration, and the only unknown in Eq. (A1) is the aerosol scattering reflectance \( \rho_a \).

(2) Assuming that \( \rho_a \) at the referenced 412 nm band can be neglected, we obtain the aerosol scattering reflectance at 412 nm by \( \rho_a^{(412nm)} = \rho_a^{(412nm)} \).

(3) Estimating the aerosol scattering reflectance at 865 nm uses the moderately accurate extrapolation model (Wang & Gordon, 1994) as

\[
\rho_a^{(865nm)} = \rho_a^{(412nm)} \left[ \frac{\epsilon_{RL}^{(745nm, 865nm)}}{\epsilon_{RL}^{(412nm, 865nm)}} \right]^{-\left(865 - 412\right)/\left(865 - 745\right)},
\]

(A3)

where \( \epsilon_{RL}^{(745nm, 865nm)} = \rho_{rc}^{(745nm)}/\rho_{rc}^{(865nm)} \). To avoid overestimating the aerosol scattering reflectance in the open ocean and continental shelf, we limit \( \rho_a^{(865nm)} \) to \( \rho_{rc}^{(865nm)} \).

(4) Assuming that the spectrum of aerosol scattering reflectance is “white” (or flat), then \( \rho_a \) at all GOCI bands are equal to \( \rho_a^{(865nm)} \).

(5) Substituting \( \rho_a \) into Eq. (A1), we obtain the water-leaving radiance \( L_w \) according to Eq. (A2). Furthermore, we obtain the normalized water-leaving radiance \( L_{wn} \) and \( R_{wn} \) as

\[
\begin{align*}
L_{wn}(\lambda) &= L_w(\lambda)/\tau_s(\lambda) \\
R_{wn}(\lambda) &= L_{wn}(\lambda)/F_0(\lambda)
\end{align*}
\]

(A4)

where \( \tau_s \) is the diffuse transmission coefficient from the sun to the sea surface. Similar to \( \tau_s \), \( \tau_s \) can be estimated using the Rayleigh optical thickness and ozone concentration.

The UV-AC algorithm assumes that \( L_w^{(412nm)} \) can be neglected. Thus, the non-zero \( L_w^{(412nm)} \) will initially overestimate the aerosol scattering reflectance. However, this overestimation will be reduced when it is extrapolated to the longer VIS and NIR. The error reduction property of the UV-AC algorithm is more advantageous than the NIR- or SWIR-based aerosol scattering extrapolation methods, in which the error is magnified when it is extrapolated to the short wavelengths. Specifically, in Eq. (A3), although \( \rho_a^{(412nm)} \) overestimates the actual aerosol scattering reflectance, as previously mentioned, \( \epsilon_{RL}^{(745nm, 865nm)} \) will reduce such an overestimation because \( \epsilon_{RL}^{(745nm, 865nm)} \) is generally larger than the actual value (usually larger than 1.0) (He et al., 2012). We theoretically deduced the error of the \( \epsilon_{RL}^{(745nm, 865nm)} \) caused by neglecting \( L_w^{(412nm)} \), and the results show that it is generally larger than \( \epsilon_{RL}^{(412nm, 865nm)} \) (He et al., 2012). In addition, the final estimated aerosol scattering reflectance is generally less than \( \rho_a^{(412nm)} \), as \( \epsilon_{RL}^{(745nm, 865nm)} \) is generally larger than 1.0, and thus the retrieved \( L_{wn}^{(412nm)} \) is positive as \( \rho_a^{(412nm)} > \rho_a^{(412nm)} \).

Furthermore, the UV-AC algorithm assumes a flat spectrum for aerosol scattering reflectance. In most cases involving coastal and maritime aerosols, such an assumption is reasonable (Gordon & Wang, 1994). However, for the tropospheric aerosol, the “white” spectrum approximation may present large extrapolation errors at the short wavelengths (Gordon & Wang, 1994). In this study, the effect of the tropospheric aerosol is expected to be small because we focus on the atmospheric correction in HZB. In addition, two other concerns prompt us to use the “white” spectrum approximation. One is the extremely complex aerosol in the coastal regions, and the other is the fact that it is the challenge of retrieving the exact epsilon value in the high turbid waters using ocean color satellite data, even in the open ocean (He et al., 2011; Mélin et al., 2010; Wang et al., 2005). In fact, the “white” aerosol assumption has been applied in the vicarious calibration of automated global data processing for the CZCS (Coastal Zone Color Scanner) data (Evans & Gordon, 1994).

Appendix B. Supplementary data

Supplementary data associated with this article can be found in the online version, at http://dx.doi.org/10.1016/j.rse.2013.01.023. These data include Google maps of the most important areas described in this article.

References


