1. Introduction

1.1. Motivation

There is a strong user interest in monitoring the turbidity of coastal and estuarine waters. Satellite mapping of turbidity is relevant both as an indicator of the optical environment for water quality monitoring purposes (Nechad, Ruddick, & Neukermans, 2009) and as an easily-measurable proxy for suspended particulate matter (SPM) concentration in sediment transport applications (Gippel, 1995). As regards water quality monitoring, turbidity is specifically listed as a mandatory parameter to be measured by EU member states in the Marine Strategy Framework Directive (European Union, 2008). Whereas various turbidity retrieval algorithms have been proposed for specific geographical regions as described in Section 1.2, the advent of large scale automated global processing of satellite data motivates the need for a single standard algorithm, valid everywhere, preferably without the need for region-specific adaptation or calibration. The objective of the present study is to determine whether a single algorithm for retrieval of turbidity can be valid for all coastal and estuarine waters, regardless of geographical region, particle type, size, composition, concentration, etc.

In the present study the term “turbidity” follows the definition of the International Standards Organization ISO 7027 (ISO, 1999), using the 90° side-scattering of light at 860 nm with respect to Formazin, a chemical standard. It is important to note the advantages of this definition as opposed to alternative definitions. For example, the definition of EPA (1993) is based on broadband scattering of light from a tungsten lamp and is thus much more sensitive to particle composition. It is also noted that the ISO definition gives an inherent optical property, in
contrast to related apparent optical properties such as diffuse attenuation or Secchi depth, which are sensitive to the ambient light field (sun zenith angle, etc.). A drawback of the ISO definition is the use of a chemical reference, giving Formazin Nephelometric Unit (FNU), rather than a physical reference, giving units of m$^{-1}$. While a simple and unique conversion factor between FNU and m$^{-1}$ could be defined from measurement of the 90° scattering coefficient of Formazin in m$^{-1}$, such a measurement has not yet been made at 860 nm to our knowledge.

1.2. Algorithms for retrieval of turbidity

Many remote sensing studies have been devoted to retrieve SPM, the parameter of main interest in sediment transport studies, but less to retrieve turbidity. However, this parameter has been used as an effective indicator of water quality and, being an optical property, is more tightly related to the backscattering coefficient, $b_h$, in m$^{-1}$, and thus to reflectance than is SPM. Most of the existing works propose site-specific empirical relationships between turbidity and reflectance at different satellite wavebands by fitting field turbidity measurements with either field- or satellite-derived reflectance. Historically, turbidity measurements have been reported in a generic unit such as the Nephelometric Turbidity Unit (NTU), as specified by the United States Environmental Protection Agency (EPA), but it can be also reported in FTU (Formazin Turbidity Unit) and FNU (Formazin Nephelometric Unit) depending on the method and equipment used. All the aforementioned units are based on calibrations using the same primary standard known as Formazin, but in general the use of NTU is limited to instruments that comply with the EPA Method, FNU pertains to instruments that comply with ISO 7027, and FTU when spectrophotometric equipment is used (absorptometric methods) (Anderson, 2006). Unfortunately, turbidity is often reported indistinctly as NTU, FTU or FNU considering all as comparable or equivalent units regardless of the instrument’s technology used. In general, a good correlation is found between turbidity and reflectance at satellite bands located in the red part of the spectrum for low to moderate turbidity values. Landsat band 3 (630–690 nm) has been used to map turbidity in Guadalquivir River (Spain) for a turbidity range 1.5–8 NTU (Bustamante, Pacios, Diaz-Delgado, & Aragonés, 2009). Choueby (1992) found a good correlation between LISS-I red band (620–680 nm) and turbidity in the range of 15–45 NTU in the Tava reservoir in India. Goodin, Harrington, Druane Nellis, and Rundquist (1996) used SPOT-HRV2 red band (610–680 nm) to map relatively low levels of turbidity, which ranged from 3 to 15 NTU, in the Tuttel Creek reservoir in Kansas, USA. MODIS 250-m resolution band at 645 nm had been successfully used to map turbidity in the Tampa Bay (USA) for turbidities ranging between 0.9–8 NTU (Chen, Muller-Karger, & Hu, 2007). Malteze, Capodici, Ciracolo, and La Loggia (2013) tested and re-calibrated Chen’s algorithm for the Sicilian gulfs (Italy) where measured turbidity ranged 1.5–6 FNU. Petus et al. (2010) developed a regional algorithm for MODIS-Aqua 250-m red band to map turbidity in the Adour River plume (Bay of Biscay, France), where field turbidity values varied between 0.5 and 70 NTU. Ouillon et al. (2008) found that the MERIS 681 nm band showed the best fit for turbidity values ranging between 1–25 FTU in three different tropical coastal waters of New Caledonia, Cuba and Fiji. Potes, Costa, and Salgado (2012) found a good linear relationship between turbidity and the ratio between the green (560 nm) and blue (412.5 nm) MERIS spectral bands for turbidity values ranging from 1 to 60 NTU. Conversely, fewer studies have been performed for very turbid waters. A multiple linear regression analysis using Landsat red (630–690 nm) and near-infrared (750–900 nm) bands was used to predict turbidity in a glacial lake in Alaska where highly scattering rock flower (sediment originated from glacial rock weathering) dominates the particulate fraction and where turbidity varied between 2–957 NTU (Liversedge, 2007).

Semi-analytical models have also been proposed for retrieving turbidity. Gohin (2011) derived satellite turbidity from chlorophyll-a concentration (Chl-a) and non-algal SPM concentrations. First, Chl-a is determined using pre-defined look-up-tables (Gohin, Druon, & Lampert, 2002), and then non-algal SPM is estimated from radiance at 555 nm or 670 nm, depending on the SPM level retrieved, by inverting a semi-analytical model (Gohin et al., 2005). Finally, turbidity is derived from Chl-a (for the phytoplankton part) and non-algal particles using an empirical relationship between SPM and turbidity derived from in situ measurements performed in Boulogne and Dunkerque (France) with values ranging from 0.2 to 100 NTU. Nechad et al. (2009) developed a semi-analytical generic one-band algorithm for turbidity as a function of reflectance for coastal waters. The algorithm, which can be used for any optical sensor measuring somewhere in the spectral range 520–885 nm, was calibrated using field radiometric and turbidity measurements from the Southern North Sea (SNS) and validated using an independent set of seaborne measurements from the same region. The algorithm was calibrated for MERIS bands and the best non-linear least-square regression fit between reflectance and turbidity was obtained at the red 681 nm band for turbidity values in the range 0.6–83 FNU.

1.3. Regionality of algorithms

Standard data products from ocean color satellite missions such as MODIS and MERIS are typically based on algorithms which are calibrated and/or validated with in situ data for a limited number of regions—see Matthews (2011); and Odermatt, Gitelson, Brando, and Schaepman (2012) for reviews of coastal water ocean color algorithms. The performance of ocean color algorithms outside their calibration/validation regions is a key question. If an algorithm has similar performance in all regions, including those not used in the original algorithm calibration, then it can be considered as a “global” algorithm and can be used with confidence everywhere, even where there is no in situ data for validation. If an algorithm performs differently in different regions then it must be considered as a strictly “regional” algorithm and its use outside the calibration regions has greater uncertainty. As regards terminology, the term “region” is not limited here to the geographical sense but could be considered more generally via a range of suitably defined parameters, which could be geographical (e.g. latitude, longitude), optical (e.g. spectral absorption and backscatter coefficients), biogeochemical (e.g. suspended particle composition, size distribution and concentration), hydrographic (influence of river discharge, salinity, depth), or others.

In the present study the potential generality of a turbidity retrieval algorithm using the near infrared band at 859 nm in sediment-dominated waters with moderate to high turbidity is assessed by both analysis of radiative transfer simulations and in situ measurements from various coastal and estuarine waters. The radiative transfer simulations cover a wide range of SPM concentrations, particulate scattering phase functions (SPF) and particulate absorption coefficient. In the context of regionality of algorithms, it is important to note the difference between SPM retrieval algorithms and turbidity (or backscatter) retrieval algorithms. Turbidity (or backscatter) retrieval algorithms will be sensitive to particle size and type only via variability of the SPF. SPM retrieval algorithms will be sensitive to particle size and type via variability of the SPF but will also be sensitive to variability of other factors such as particle density and refractive index. In other words, since turbidity is, like particulate backscatter, an inherent optical property it is not necessary to consider the potential variability of mass-specific optical properties (Neukermans, Loisel, Mériaux, Astoreca, & McKee, 2012), which are, by contrast, an important source of regional variability for retrieval of SPM concentration. For completeness the use of the red 645 nm band in waters with medium to low turbidity is also proposed to ensure that a very wide range of turbidities can be covered.

2. Data and methods

2.1. Remote sensing turbidity algorithm

The semi-empirical single band turbidity ($T$) retrieval algorithm of Nechad et al. (2009) is analyzed in the present study. It relates turbidity
and marine reflectance at wavelength $\lambda$, $\rho_w(\lambda)$, which is defined as $\pi L_w(\lambda) / E^+_{0+}(\lambda)$, where $L_w$ is the water-leaving radiance and $E^+_{0+}$ is the above-water downwelling irradiance, through

$$T = \frac{A^0_{\lambda} \rho_w(\lambda)}{(1 - \rho_w(\lambda) / C^0)} \quad [\text{FNU}]$$  \hspace{1cm} (1)$$

where $A_T$ and $C_T$ are two wavelength-dependent calibration coefficients.

A complete list of symbols is provided in Table 1. The parameter $C$ was calibrated using “standard” inherent optical properties (IOPs) as described in Nechad, Ruddick, and Park (2010), while the $A_T$ coefficient was obtained by a non-linear least-square regression analysis using in situ measurements of $T$ and $\rho_w$ (see Appendix A for a detailed algorithm description). The $A^{645}$ coefficient was derived using in situ data obtained in the Southern North Sea (cruises performed between 2007–2010), while the $A^{859}$ was calibrated using data with $T$ ranging from 0 to 255 FNU from the above mentioned Southern North Sea cruises and two campaigns in the Scheldt estuary (SC) performed in 2010 (Dogliotti et al., 2011). The coefficients at the two wavelengths are given in Table 2.

In the present work we focus on the algorithm on the using the 859 nm band for medium to high turbidity values. At this wavelength the high pure water absorption (~4.4 m$^{-1}$, Kou, Labrie, & Chylek, 1993) avoids saturation of the algorithm expected at lower wavelengths (Bowers, Boujdelas, & Harker, 1998). However, to cover the wider range of turbidities that can be found in coastal waters and to avoid the expected reduced sensitivity of this band at low turbidity, the use of the red band at 645 nm is also proposed. An analysis of the field data used to calibrate the algorithm at the two wavelengths, i.e. from SNS (2007–2010) and SC (2010), showed that for $\rho_w(645)$ values lower than 0.05 (corresponding to modeled $T$ – 15 FNU) there is a good agreement between field and modeled $T$ (Fig. 1). At higher reflectance values, the model sensitivity decreases as one moves away from the linear regime and approaches the saturation region. Thus, an algorithm scheme based on variable bands, as in Shen, Verhoef, Zhou, Salama, and Liu (2010), is suggested based on the reflectance value at 645 nm. In order to assure a smooth transition between the two algorithms, a linear weighting function (between 0 and 1) is applied to the modeled $T$ for $\rho_w(645)$ ranging between 0.05 and 0.07. In summary, we use the 645 nm band when $\rho_w(645) < 0.05$, the 859 nm band when $\rho_w(645) > 0.07$ and blend the two algorithms when $0.05 < \rho_w(645) < 0.07$. In the transition zone, the weight of the algorithm ($w$) changes linearly from 0 at $\rho_w(645) = 0.05$ to 1 at $\rho_w(645) = 0.07$. Thus, blending is done according to the following formula: $T = (1 - w) \cdot T_{645} + w \cdot T_{859}$, where $T_{859}$ is the turbidity calculated using the 645 nm band and $T_{645}$ is the turbidity calculated using the 859 nm band in Eq. (1).

2.2. Field measurements

New in situ measurements were collected from various European and South American coastal and shallow estuarine environments characterized by high concentrations of suspended sediments and a variety of environmental conditions (Fig. 2, Table 3). This dataset was not used to derive or calibrate the algorithm of Section 2.1. Data from the southern North Sea (SNS) were collected during a cruise onboard the Belgica research vessel in 2011, primarily in Belgian coastal waters. This region is a relatively shallow area (~50 m) subject to strong vertical mixing from strong winds and tidal currents (Ruddick & Lacroix, 2006) where sediment resuspension causes relatively high SPm concentrations (~1–200 g m$^{-3}$).

Measurements from the Scheldt (SC) estuary were collected from a fixed pontoon (51°14’ N–4°23’ E) located near the city of Antwerp (Belgium) in June 2012 and October 2013. The Scheldt is a relatively turbid estuary where SPm values can reach up to 400 g m$^{-3}$ and are subject to strong tidal and seasonal variations (Ardnt, Vanderborght, & Regnier, 2007).

Data from the Gironde (GIR) estuary were collected from two fixed pontoons located in Pauillac (45°11’8.3” N–4°44.58” W) and Blaye (45°7’5.1” N–40.040’3” W) in southwest France in June 2012 and August 2013. This macrotidal estuary is characterized by highly turbid waters dominated by suspended sediments. Mean SPm concentrations vary from 150 to approximately 3000 g m$^{-3}$ in surface waters (Allen, Salomon, Bassoullet, Du Penhoat, & Degranpré, 1980).

Measurements from Río de la Plata (RDP) were performed from a fixed pontoon at the Palermo Fishermen Pier (34°0.20.18’ S–58’ 14.36’ W) located in Buenos Aires (Argentina) in November 2012. The city is located in the upper part of the estuary where it is very shallow (~5 m). The river carries a large amount of suspended particulate and dissolved organic matter and high values of SPm have been reported in this region, with mean values ranging from 100 to 300 g m$^{-3}$ and extreme concentrations up to 400 g m$^{-3}$ (C.A.R.P., 1989; Urien, 1967).

A field campaign was performed in the turbid coastal waters of French Guyana (FG) in October 2009. The sampled area is located in relatively shallow waters which are strongly influenced by the Amazon River discharge and characterized by mudflats all along the coast (Froidefond, Lahet, Doxaran, Prost, & Ternon, 2004; Loisel et al., 2009).

Table 1

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\beta$</td>
<td>Volume Scattering Function (VSF)</td>
<td>m$^{-3}$ sr$^{-1}$</td>
</tr>
<tr>
<td>$\phi$</td>
<td>Relative azimuth angle between the sun and sensor</td>
<td>Degrees</td>
</tr>
<tr>
<td>$\gamma_r$</td>
<td>Spectral slope of the particulate beam attenuation spectrum</td>
<td>-</td>
</tr>
<tr>
<td>$\lambda$</td>
<td>Wavelength</td>
<td>nm</td>
</tr>
<tr>
<td>$\rho_w$</td>
<td>Water reflectance</td>
<td>-</td>
</tr>
<tr>
<td>$\theta_w$</td>
<td>Viewing zenith angle</td>
<td>Degrees</td>
</tr>
<tr>
<td>$\chi$</td>
<td>Scattering angle</td>
<td>Degrees</td>
</tr>
<tr>
<td>$\alpha_{DOC}$</td>
<td>Colored dissolved organic matter absorption coefficient</td>
<td>m$^{-1}$</td>
</tr>
<tr>
<td>$\alpha_p$</td>
<td>Non-particle absorption coefficient</td>
<td>m$^{-1}$</td>
</tr>
<tr>
<td>$\alpha_a$</td>
<td>Particle absorption coefficient</td>
<td>m$^{-1}$</td>
</tr>
<tr>
<td>$\sigma_p$</td>
<td>Mass-specific particle absorption coefficient</td>
<td>m$^{-2}$ g$^{-1}$</td>
</tr>
<tr>
<td>$\sigma_{pT}$</td>
<td>Turbidity-specific particle absorption coefficient (at 90°)</td>
<td>m$^{-2}$ FNU$^{-1}$</td>
</tr>
<tr>
<td>$\beta_{SPC}$</td>
<td>Particulate side-scattering coefficient (at 90°)</td>
<td>m$^{-3}$</td>
</tr>
<tr>
<td>$\beta_w$</td>
<td>Total backscattering coefficient</td>
<td>m$^{-3}$</td>
</tr>
<tr>
<td>$\beta_{b}$</td>
<td>Particulate backscattering coefficient</td>
<td>m$^{-3}$</td>
</tr>
<tr>
<td>$\beta_{bT}$</td>
<td>Turbidity-specific particulate backscattering coefficient</td>
<td>m$^{-3}$ FNU$^{-1}$</td>
</tr>
<tr>
<td>$\beta_{p}$</td>
<td>Particulate scattering coefficient</td>
<td>m$^{-3}$</td>
</tr>
<tr>
<td>$\beta_{pT}$</td>
<td>Mass-specific particulate backscattering coefficient</td>
<td>m$^{-3}$ g$^{-1}$</td>
</tr>
<tr>
<td>$\epsilon_w$</td>
<td>Mass-specific particle beam attenuation coefficient</td>
<td>m$^{-2}$ g$^{-1}$</td>
</tr>
<tr>
<td>$	ext{Chl-a}$</td>
<td>Chlorophyll-a concentration</td>
<td>mg m$^{-3}$</td>
</tr>
<tr>
<td>$E^+_{0+}$</td>
<td>Above-water downwelling irradiance</td>
<td>W m$^{-2}$ nm$^{-1}$</td>
</tr>
<tr>
<td>$L_w$</td>
<td>Water-leaving radiance</td>
<td>W m$^{-2}$ sr$^{-1}$ nm$^{-1}$</td>
</tr>
<tr>
<td>$S$</td>
<td>Spectral slope of the particulate absorption spectrum</td>
<td>-</td>
</tr>
<tr>
<td>$SPF$</td>
<td>Scattering phase function</td>
<td>sr$^{-1}$</td>
</tr>
<tr>
<td>$SPM$</td>
<td>Suspended particulate matter</td>
<td>g m$^{-3}$</td>
</tr>
<tr>
<td>$T$</td>
<td>Turbidity</td>
<td>FNU</td>
</tr>
</tbody>
</table>

Table 2

<table>
<thead>
<tr>
<th>$\lambda$ (nm)</th>
<th>$A_T$</th>
<th>$C_T$</th>
</tr>
</thead>
<tbody>
<tr>
<td>645</td>
<td>228.1</td>
<td>0.1641</td>
</tr>
<tr>
<td>859</td>
<td>3078.9</td>
<td>0.2112</td>
</tr>
</tbody>
</table>
transmitted light as compared to the same ratio for a standard suspension of Formazin. This optical measurement technique of turbidity from the side-scattering coefficient, $b_{90}$, is in accordance with ISO 7027 (1999), and determines turbidity in Formazin Nephelometric Unit (FNU). Turbidity was recorded in triplicates that were averaged. Turbidities of the STABLCAL Stabilized Formazin Turbidity 10 or 20, 100 and 800 FNU standards and that of pure water were recorded after each sampling campaign to check the instrument stability.

Above-water marine reflectance measurements were collected with a set of three Trios-RAMSES hyperspectral radiometers with a sampling interval of approximately 3.3 nm and an effective spectral resolution of about 10 nm covering the spectral range 400–900 nm. The protocol described in Ruddick, De Cauwer, Park, and Moore (2006) was followed except that a) for measurements from fixed structures the standard viewing azimuth of 135° relative to sun was modified to 90° when necessary to minimize structure perturbations of the light field, and b) for measurements in estuaries with fetch-limited surface waves the wind speed was set to zero in estimation of the Fresnel reflectance. The hyperspectral reflectance data were integrated over the relative spectral response function of MODIS band 2 to obtain $\rho_w(645)$ and $\rho_w(859)$.
2.3 Radiative transfer simulations

izing 106 measurements considering all sites together (see Table 4). The algorithms was 8 for SNS and SC, 9 for FG, 40 for RdP, and 41 for GIR, totaling 106 measurements considering all sites together (Table 4).

Table 4

<table>
<thead>
<tr>
<th>Location</th>
<th>Platform</th>
<th>Time period</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Temperature [°C]</th>
<th>Salinity [PSU]</th>
<th>SZA [°]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Southern North Sea</td>
<td>Belgica R/V</td>
<td>4-7 Jul 2011</td>
<td>51° 23.14' N</td>
<td>02° 50.00' E</td>
<td>18.5 [17.8-19.2]</td>
<td>33.4 [31.8-34.4]</td>
<td>38.2 [28.7-60.7]</td>
</tr>
<tr>
<td>French Guyana</td>
<td>Papjo R/V</td>
<td>13-17 Oct 2009</td>
<td>04° 44.57' N</td>
<td>52° 10.80' W</td>
<td>NA</td>
<td>NA</td>
<td>36.4 [14.4-57.2]</td>
</tr>
<tr>
<td>Scheldt</td>
<td>St. Anna pontoon</td>
<td>2 &amp; 5 Jun 2012</td>
<td>51° 14.05' N</td>
<td>04° 23.78' E</td>
<td>18.7</td>
<td>5.0 [1.1-6.4]</td>
<td>35.4 [28.6-49.8]</td>
</tr>
<tr>
<td>Gironde</td>
<td>Pauillac pontoon</td>
<td>15-16 Jun 2012</td>
<td>45° 11.83' N</td>
<td>00° 44.58' W</td>
<td>20.7 [20.5-21]</td>
<td>1.9 [1.6-3.0]</td>
<td>43.9 [21.9-74.5]</td>
</tr>
<tr>
<td>Blaye pontoon</td>
<td>12-16 Aug 2013</td>
<td>45° 07.52' N</td>
<td>00° 40.03' W</td>
<td>NA</td>
<td>NA</td>
<td>39.9 [30.9-55.1]</td>
<td></td>
</tr>
<tr>
<td>Rio de la Plata</td>
<td>Fishermen Pier</td>
<td>14-23 Nov 2012</td>
<td>34° 33.65' S</td>
<td>58° 23.93' W</td>
<td>25.4 [22.9-27]</td>
<td>0.16 [0.02-0.18]</td>
<td>31.2 [14.3-74.8]</td>
</tr>
</tbody>
</table>

For the present analysis the quality control criteria retained only marine reflectances that were collected in homogeneous sunny skies (as indicated by the relation $R_{300}/R_{330}^F$ (750) < 0.05 sr$^{-1}$, where $R_{300}$ is the measured sky radiance), and with small deviation from the time-averaged mean reflectance at 859 nm and 645 nm, i.e. a coefficient of variation (standard deviation to mean ratio) CV < 20%. After applying the quality control, the number of field measurements used to evaluate the algorithms was 8 for SNS and SC, 9 for FG, 40 for RdP, and 41 for GIR, totaling 106 measurements considering all sites together (see Table 4).

2.3. Radiative transfer simulations

Radiative transfer simulations were performed using Hydrolight 5.0 (Mobley & Sundman, 2008) at a single wavelength (859 nm) considering only water and mineral particles, i.e. colored-dissolved organic matter (CDOM) absorption and chlorophyll-a concentration were set to zero. Absorption and scattering for water were taken from Kou et al. (1993) and Morel (1974), respectively. Water absorption is so high in the near infrared (~4.4 m$^{-1}$) that light absorption by phytoplankton and CDOM can be assumed negligible in comparison (Babin & Stramski, 2002). Simulations were performed for mineral particle concentrations that varied from 0.1 to 1000 g m$^{-3}$. Fournier-Forand (FF) scattering phase functions (SPFs) with four different backscattering ratios ($bbp/bbp$) were used (0.01, 0.0183, 0.03, and 0.05). These values cover the extreme range of values reported in the literature for coastal waters dominated by non-algal particulate matter which varied between 0.01 and 0.06, with a mean value of approximately 0.02 (e.g. Boss et al., 2004; Chami, McKee, Leymarie, & Khomenko, 2006; Loisel, Mériaux, Berthon, & Poteau, 2007; McKee et al., 2009; Snyder et al., 2008; Tzortziou et al., 2006; Whitmire, Boss, Cowles, & Pegau, 2007). For each SPF, the particle backscatter to absorption ratio ($bbp/bap$) was obtained by varying both $bbp$ and $bap$. The mass-specific particulate absorption, $a_p^s (\lambda)$, was extrapolated from 443 nm using

$$a_p^s (\lambda) = a_p^s (443) \times \exp \left[ -S (\lambda - 443) \right]$$

where $a_p^s (443) = 0.041$ m$^2$ g$^{-1}$ and the spectral slope $S = 0.0123$ nm$^{-1}$ are taken from Babin, Stramski, et al. (2003). The equation was originally developed for the 380–730 nm range and is assumed to be valid beyond 730 nm. Mass-specific particulate scattering, $b_p^s (\lambda)$, was obtained using Eq. (2) and modeling the mass-specific particulate beam attenuation, $C_p^s (\lambda)$, as

$$C_p^s (\lambda) = \left[ a_p^s (555) + b_p^s (555) \right] \times (\lambda/555)^{-\gamma}$$

where $b_p^s (555) = 0.51$ m$^2$ g$^{-1}$ and the spectral slope of the beam attenuation $\gamma = 0.3749$ were taken from Babin, Morel, Fournier-Sicre, Fell, and Stramski (2003). The water was modeled as homogeneous and infinitely deep. Surface downwelling light was modeled for sun zenith angles ($\theta_h$) varying from 0° to 60°, a nominal wind speed of 5 m s$^{-1}$, clear and overcast sky conditions were considered, and the surface irradiance was calculated using the RADTRANX model supplied as part of the Hydrolight code. The particle side-scattering coefficient, $bbp$, is defined here as $2n + \rho (90°)$ where $\rho (\chi)$ is the Volume Scattering Function (VSF) for scattering angle $\chi$ as defined in Mobley, Sundman, and Boss (2002). With this definition a VSF which is independent of angle in the backscattering directions, i.e. with $\rho (\chi) = \rho (90°)$ for $\chi > 90°$, has equal side- and backscattering coefficients. For the four SPFs defined above ($FF = 0.01$, $0.0183$, and $0.05$) the ratio of $bbp/bap$ is 1.47, 1.44, 1.38, and 1.31, respectively.

3. Results

3.1 Sensitivity of the 859 nm algorithm to different factors

3.1.1 Scattering phase function

Natural variability of the SPF is an important potential source of uncertainty for turbidity estimation from Eq. (1), which assumes a single SPF. The radiative transfer simulations are here analyzed to estimate this uncertainty at 859 nm. The results are presented both in terms of side-scattering, directly relevant for turbidity, and in terms of backscattering; because the latter is used in analytical retrievals related uncertainty (ΔSPF) of the SPF is much more significant. FF = 0.01 and FF = 0.0183 are expected to differ due to varying SPFs compared to FF = 0.03, which is used in the retrieval context. The influence of varying SPFs is much more significant for $bbp$, indicating that retrieval of $bbp$, unlike $bbp$ or $bbp$, will have high natural variability associated with the SPF.

This SPF-related uncertainty is further analyzed by quantifying the range of side-, back- and total scattering coefficients that correspond to the same reflectance when different SPFs are considered. Thus, Fig. 4 shows the change in side-, back- and total scattering coefficients that are needed to give the same reflectance as a reference SPF ($bbp/bap = 0.0183$ hereafter referred as FF = 0.0183) when the SPF is varied over a range of SPFs typically found in coastal waters, given by backscattering ratios of 0.01 and 0.03 (referred as FF = 0.01 and
FF = 0.03. This represents directly an important component of retrieval algorithm uncertainty since the SPF is generally not known a priori. The total scattering coefficient, as was previously observed in Fig. 3, shows the highest variability (bold lines in Fig. 4a) and thus greatest sensitivity to SPF variations over the whole reflectance range. This can be explained by the fact that \( b_p \) integrates all scattering angles including forward scattering angles, whereas reflectance is almost insensitive to near-forward scattering. Uncertainty on \( b_p \) varied from 35 and 40% for FF = 0.03 and up to 70 and 85% for FF = 0.01 (bold lines in Fig. 4a), both increasing at lower reflectance values. At high reflectance,
3.1.2. Bidirectional effects

To assess the turbidity algorithm uncertainty due to bidirectional distribution of sea water reflectance at 859 nm, the side-scattering was computed for different viewing geometries and illumination conditions given a fixed SPF (e.g. FF = 0.0183) (Fig. 5). Three viewing geometries, given by viewing zenith angle \( \theta_v \) and relative azimuth angle between the sun and sensor (\( \phi \)), were selected: i) \( \theta_v = 40^\circ \) and \( \phi = 135^\circ \), ii) \( \theta_v = 40^\circ \) and \( \phi = 90^\circ \), and iii) Nadir viewing, i.e. \( \theta_v = \phi = 0^\circ \). The first two are typical above-water radiometry viewing geometries, as used for the calibration dataset. Three sun zenith angles (\( \theta_s \)) were also analyzed: 1) \( \theta_s = 0^\circ \) (at zenith), 2) \( \theta_s = 30^\circ \) and 3) \( \theta_s = 60^\circ \) for clear and overcast skies. The results show that uncertainty is sensitive to varying viewing geometry (up to 7%) and, to a less extent, to illuminating conditions, reaching up to 3.5% (Fig. 5a and b, respectively).

3.1.3. Other factors

In order to quantify the uncertainty in the turbidity retrieval at 859 nm due to natural variability of \( b_{bbp}/a_p \), which is related to the particle type, additional simulations were performed for varying parameters that were previously fixed in Eqs. (2) and (3), i.e. the mass-specific particulate absorption and backscattering coefficients \( a_p^0(443) \) and \( b_p^0(555) \) and absorption and attenuation spectral slopes \( S \) and \( \gamma_v \), as well as different SPF's. Simulations were performed using values taken from previously published studies, i.e. \( a_p^0(443) \) ranging between 0.018 and 0.064 m\(^2\) g\(^{-1}\) and \( S \) varying from 0.011 to 0.0136 nm\(^{-1}\), to cover the variability found in coastal waters reported in Babin, Stramski, et al. (2003), \( b_p^0(555) \) ranged between 0.1 and 0.8 m\(^2\) g\(^{-1}\) (Babin, Morel, et al., 2003), and \( \gamma_v \) from 0.3 to 1.15 to cover the variability found by Boss et al. (2001) in waters dominated by detritus and resuspended sediments. As previously (Figs. 4 and 5), uncertainty in the retrieved \( b_{bbp} \) was calculated by normalizing the values to a reference value for a given combination of input variables. In this case typical mean values found in coastal waters such as \( b_{bbp}/b_p = 0.0183 \), \( a_p^0(443) = 0.041 \) m\(^2\) g\(^{-1}\), \( S = 0.0123 \) nm\(^{-1}\), \( b_p^0(555) = 0.51 \) m\(^2\) g\(^{-1}\), and \( \gamma_v = 0.3749 \) were chosen. Results show that uncertainty can be up to 10% and 15% at high and low reflectance, respectively, when \( b_{bbp}/a_p \) ranges over one order of magnitude (Fig. 6).

In coastal waters where terrestrial inputs are important CDOM absorption in the blue part of the spectrum can be very high (up to 90% of the total absorption). However, it decreases exponentially with increasing wavelength (Bricaud, Morel, & Prieur, 1981), being very low in the NIR compared to water absorption. When considering relatively high CDOM absorption values at 443 nm, e.g. \( a_{CDOM}(443) \) ranging from 0.1 to 0.7 m\(^{-1}\) and mean slope of 0.0176 nm\(^{-1}\) typically found in coastal areas with terrestrial origin (Babin, Stramski, et al., 2003), CDOM absorption in the NIR (859 nm) varies between 6.6 \( 10^{-3} \) and 4.6 \( 10^{-4} \) m\(^{-1}\). Considering that water absorption in the NIR is relatively high (4.4 m\(^{-1}\)), CDOM absorption is only 0.0015 to 0.0105% of the water absorption, reaching up to 0.034% when natural variability in the CDOM absorption spectral slope is considered (Babin, Stramski, et al., 2003). Thus, CDOM absorption, generally high in turbid estuaries, will have negligible influence in the turbidity retrieval algorithm for sediment-dominated waters using NIR bands.

Another source of uncertainty is related to the variability of water absorption with temperature and salinity. Using the Water Optical Properties Processor (WOPP), a computer program to calculate the inherent optical properties (IOPs) of pure water at a specific water temperature and salinity, Röttgers et al. (2012) found that the relative change in water absorption is \(-0.027\%\) per °C and \(-0.009\%\) per PSU (Practical Salinity Unit) for the OLCI NIR band at 866 nm. Thus, in this NIR band the impact on reflectance comes mainly from the uncertainty of the temperature effect. At low reflectance variability of water absorption due to a change from 0 to 20 °C would cause an uncertainty in reflectance of 0.5%, while at high reflectance the impact would be less and could be considered negligible.

3.2. Algorithm performance in different regions

The one-band algorithm (Eq. 1) using 645 nm and 859 nm bands and the switching scheme presented above was applied to reflectance measured at five different locations, covering a variety of suspended particles in terms of turbidity range, composition, size distribution, and refractive index. The dataset used here to evaluate the algorithm is independent of the dataset used for the algorithm development and calibration. To evaluate the algorithm performance, statistics of the linear regression (slope, intercept, root-mean-square error, RMSE) between

![Fig. 5. Particle side-scattering (b\(_{bbp}\)) normalized to b\(_{bbp}\) at a reference condition (b\(_{bbp}\)^0); Nadir viewing, \( \theta_v = 30^\circ \), and FF = 0.0183 for variable: a) observation geometries (\( \theta_s \) and \( \phi \)), and b) illuminating conditions (\( \theta_s \) for clear and overcast skies).](image-url)
modeled ($T_{\text{mod}}$) and field ($T_{\text{field}}$) values were calculated. In order to evaluate the uncertainty, the mean absolute relative percentage error ($\varepsilon$) and bias ($\delta$) were calculated as

$$
\varepsilon = \frac{1}{n} \sum \frac{|T_{\text{mod}} - T_{\text{field}}|}{T_{\text{field}}} \times 100
$$

$$
\delta = \frac{1}{n} \sum \frac{T_{\text{mod}} - T_{\text{field}}}{T_{\text{field}}} \times 100
$$

where $n$ is the number of observations. The correlation between modeled and field measurements ($r$) was also computed (Table 4).

Comparisons showed a general good agreement between field and modeled $T$ for all the sites analyzed. Reflectance increased with increasing $T$ up to ~20 FNU for $\rho_w(645)$ and up to ~1000 FNU for $\rho_w(859)$, and then tended to saturate at higher values, respectively (Fig. 7). A good performance of the algorithm using the 645 nm band for low $T$ can be observed for both the SNS and FG datasets (solid symbols in Fig. 7a). In Fig. 7 solid symbols indicate when the 645 nm and 859 nm band algorithm is used, while empty symbols show when the weighted combination between the two algorithms is used. The portable HACH turbidimeter measures $T$ values up to 1000 FNU, so dilutions were performed with MilliQ water at four stations for which readings exceeded this limit in the Gironde estuary. Even though dilution might change the particle size distribution thus contaminating the measurement, a good agreement with field data was still found for these samples (Fig. 7b). In order to avoid the asymptotic regime where saturation occurs and $T$ retrieval is less reliable, statistical analysis was performed for $T < 1000$ FNU. Equations and statistics of the regressions for each region and considering the entire data base are shown in Table 4. Scatter plot of modeled (using the switching one-band algorithm) vs. field $T$ is shown in Fig. 8. A good correlation between modeled and field values was observed for each site with a correlation coefficient $r > 0.9$ at all sites except for RdP ($r = 0.83$). On average, $T$ was underestimated by ~12% in SNS and SC and overestimated between 7% and 9% in the GIR, RdP and FG. Mean relative errors ranged between 11 and 15% in FG, RdP, SC and GIR, and were highest in SNS ($\varepsilon \sim 22\%$). Slopes of the linear regressions were close to 1, and RMSE ranged between 5 (SNS) and 33 FNU (GIR). When the entire data base was considered ($n = 106$), a general good performance of the algorithm was found with a correlation coefficient $r = 0.97$, relatively small bias (4.8%), and a mean relative error of ~14%. The fitted linear regression explained 97% of the variability with a slope close to 1 (1.07), intercept close to zero ($-1.29$) and RMSE of 20.5 FNU (Table 4).

4. Discussion

4.1. Generality of the 859 nm band algorithm

The single band turbidity algorithm analyzed in the present study (Dogliotti et al., 2011; Nechad et al., 2008) was derived assuming a constant $T$-specific particulate backscattering coefficient ($b_{\text{pp}}/T$). Since $T$ is an optical property (side-scattering), the variability of the side to backscattering ratio is expected to be lower than, for example, the SPM-specific backscattering coefficient ($b_{\text{pp}}/\text{SPM}$) which has been shown to vary by a factor of 3 or 4 and depends on the particle composition i.e. organic- or inorganic-dominated waters (Doxaran et al., 2012; Loisel et al., 2009; Martinez-Vicente, Land, Tilstone, Widdicombe, & Fishwick, 2010; Neukermans et al., 2012). In general, less in situ measurements of concomitant $T$ and $b_{\text{pp}}$ have been performed compared to SPM and $b_{\text{pp}}$ values. Neukermans (2012) estimated the variability of $\beta_y(120°,658\text{ nm})/\beta_y(90°,658\text{ nm})$ to be within a factor of 1.7 with a mean value of 1.2 using in situ VSF measurements carried out in a wide range of water types using the Wetlabs MASCOT instrument.

Fig. 6. Side-scattering for variable $b_{\text{pp}}/\rho_w$ ratio normalized to a reference value obtained using a combination of typical mean values found in coastal waters for SPI, SIOPs, and absorption and attenuation spectral slopes (see text in Section 3.1.3 for values).

Fig. 7. Scatter plot of field $T$ (FNU) versus a) $\rho_w(645)$ and b) $\rho_w(859)$. Measurements performed in SNS, SC, GIR, FG, and RdP are shown in different colors. The black line is the semi-analytical model for each wavelength and solid symbols indicate when the corresponding algorithm is used, while empty symbols indicate when a weighted combination between the two algorithms is used. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
The sensitivity analysis performed using radiative transfer simulations, which covered a wide range of natural conditions, showed that uncertainty in the retrieved turbidity can reach up to 6% when variable SPF and mean specific inherent optical properties (SIOPs) were considered. However, when natural variability of SIOPs was considered, higher values were found (up to 10%). It should be noted that simulated $b_{bbp}/b_{ap}$ ratios ranged over one order of magnitude and this might exaggerate its natural variability. Field measurements of $b_{bbp}(850)$ performed in RdP and GIR campaigns four HACH 2100P and 2100QIS portable turbidimeters (with same technology and design) have been used and inter-instrument comparisons showed a low variability for Formazin standards, with CV varying between 0.2 and 3%, while for natural samples the variability was 0.5–5.9% (RdP) and 0.4–5.9% (GIR).

Fig. 8. a) Scatter plot of modeled versus field T (FNU) measurements performed in SNS, SC, GIR, FG, and RdP. b) Detail of the scatter plot at low T range (0–100 FNU). The dashed line is the 1:1. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Even though the analysis of the general validity of the algorithm for low to medium turbidity, i.e. using the 645 nm band, was beyond the objectives of the present work, possible sources of error due to variable SIOPs found in regions other than the one used to calibrate the algorithm were considered. However, since the 645 nm band is not very sensitive to CDOM absorption, the general conclusion of low sensitivity of the proposed algorithm to CDOM variability is maintained. The sensitivity of one-band algorithms has been shown to depend on both wavelength and turbidity range, with red wavelengths being more sensitive to low T and longer wavelengths to high T (Stramski, Wozniak, & Flatau, 2004; Shen et al., 2010). This is supported by the results found in the present study showing a general good performance of the proposed algorithm with low mean relative errors (<14%) and RMSE (20%) for T ranging from 1 to 1000 FNU. This switching band algorithm makes use of the red 645 nm band for low $T$ (T < 15 FNU or $\rho_{645} < 0.05$) and the NIR 859 nm band for high $T$ values (up to 1000 FNU), while it tends to saturate above 1000 FNU.

4.2. Generality of the 645 nm band algorithm

Even though the analysis of the general validity of the algorithm for low to medium turbidity, i.e. using the 645 nm band, was beyond the objectives of the present work, possible sources of error due to variable SIOPs found in regions other than the one used to calibrate the algorithm (SNS) are discussed in the following paragraphs. Error in the $T$ retrieval due to variability in CDOM and mass-specific particulate (non-algal and phytoplankton particles) absorption coefficient are analyzed. Variability in CDOM absorption results in a proportional error in $T$ retrieval given its linear dependence in Eq. (1) through $A_T$ coefficient (see Eq. A4 in Appendix A). Similar to the error analysis performed in Nechad et al. (2010), the relative error associated with CDOM absorption was calculated and showed to vary between 0.08 and 18% at
645 nm for typical $a_{CDOM}$ values calculated assuming an exponential dependence with wavelength (Bricaud et al., 1981) for two different exponential slope parameters (0.0176 nm$^{-1}$ and 0.0131 nm$^{-1}$) and two levels of CDOM absorption ($a_{CDOM}(443 \ nm) = 0.1 \ m^{-1}$ and $1.0 \ m^{-1}$).

Previous theoretical analysis indicates that the algorithm's sensitivity to variable mass-specific particulate absorption has negligible impact on derived $T$ in the low reflectance regime of Eq. (1) given that the $A_f$ coefficient does not depend on particulate absorption (Eq. A4), but becomes an additional source of error for high reflectance values (Nechad et al., 2010). The authors found that errors associated with non-algal particulate absorption are generally low for $\lambda > 600 \ nm$ and varied between $-1$ and $-14\%$ at 600 nm for SPM $= 10 \ mg \ m^{-3}$ and 100 mg m$^{-3}$, respectively. Errors associated with the phytoplankton absorption may be significant for high Chl-a $> 30 \ mg \ m^{-3}$ at 645 nm, varying between 19% and 57% for Chl-a $= 10 \ mg \ m^{-3}$ and 30 mg m$^{-3}$, respectively following the methodology of Nechad et al. (2010).

5. Conclusions

In the present study we showed that single-band semi-analytical algorithm using 645 nm and 859 nm bands (Dogliotti et al., 2011; Nechad et al., 2009) and a switching scheme can be used to retrieve turbidity from water reflectance in very different regions and is almost insensitive to the sediment type. The impact of the regional variability of the relationship between $T$ and $\rho_w(859)$ caused by natural SPF variability was assessed using radiative transfer simulations. These simulations showed that the corresponding retrieval uncertainty is expected to be low, less than 6%, for very different SPFs typically found in natural coastal waters. The uncertainty on turbidity retrieval associated with bidirectional effects was also analyzed and it was found to be low, typically less than 6%. Considering that technical differences between the instruments used to measure turbidity exists, protocols and specifications of critical components of turbidimeters should be tightened, choosing a set of specifications that are best suited for retrieving the particle side-scattering optical property which has been shown to be less sensitive to varying SPFs and thus useful to retrieve $T$ from remote sensing data. Similarly if the proposed algorithm is used globally, then the turbidity retrieved corresponds strictly to the ISO-based instrument used for algorithm calibration.

A good agreement between modeled and in situ measurements was found for each sampled site where $T$ was modeled with RSME of 5–33 FNU. A satisfactory performance of the algorithm was also found when all the sites were analyzed together, covering a wide range of $T$ (1.8–988 FNU), and showing a high correlation ($r = 0.97$), a mean relative error of 13.7%, and bias of 4.8%.

Results presented in this study suggest that a general algorithm can be found for remote sensing of water turbidity using the red and NIR bands of past, present and future ocean color satellite sensors provided atmospheric correction is possible. The use of the 860 nm band on sensors such as MODIS, MERIS, SeaWIFS, GOCC, OCli, and HICO supposes that a suitable atmospheric correction can be performed there. While this band is generally used precisely for atmospheric correction purposes, it is essential to use an algorithm adapted for turbid waters, giving a non-zero water reflectance either by use of a near infrared marine model (Bailey, Franz, & Wendell, 2010; Doerffer & Schiller, 2007; Moore, Aiken, & Lavender, 1999; Ruddick, Ovidio, & Rijkeboer, 2000; Stumpf, Arnone, Gould, & Ransibrahmanakul, 2003), or by use of short wave infrared (SWIR) bands, when available (MODIS) (Wang, Shi, & Tang, 2011; Wang, Tang, & Shi, 2007). The high resolution bands of the MODIS sensor, like the 645 nm and 859 nm bands, that were originally designed for land and cloud applications have the advantage that they do not saturate over highly turbid waters, but have significantly lower sensitivity (low signal-to-noise ratio) relative to the ocean bands (Franz et al., 2006). For example, Petus et al. (2010) showed that MODIS 859 nm band is not sensitive enough to detect turbidity variations between 0.01 and 10 FNU. Thus, in order to overcome this limitation the use of a shorter wavelength, like the 645 nm band, was here proposed to estimate $T$ in medium to low turbid waters.

The main limitation of this algorithm is related to the range of $T$ rather than the geographic region or particle type. On the data set analyzed, the algorithm is suitable for the 1–1000 FNU turbidity range. However, further analysis of the generality of the algorithm for low $T$, i.e. using the 645 nm band, is required in order to evaluate its sensitivity to variable SPF and bidirectional effects.

Finally, suspended particulate matter concentration, the parameter of main interest in sediment transport studies, could subsequently be retrieved by ocean color remote sensing if a region-specific relation between $T$ and SPM is known, i.e. the global algorithm to retrieve $T$ from reflectance is supplemented with a local algorithm to estimate SPM from $T$. This has the very practical advantage that SPM maps could be obtained from different coastal and estuarine waters using relatively inexpensive and portable nephelometers and SPM analysis to regionally calibrate just the $T$-SPM relation. This would avoid the need of the more expensive radiometric measurements and restrictive illumination conditions (daylight, cloud-free sky) that are needed to calibrate the full reflectance-SPM relationship in any new region. This is similar to the approach suggested by Doerffer (2006) except that the latter study uses total scattering as the retrieved optical parameter whereas our results suggest that uncertainty related to SPF variability is reduced by using $b_{spT}$ instead of $b_r$.

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Appendix A. Turbidity algorithm

The one-band turbidity algorithm used in the present work is briefly described in this appendix, more detailed analysis of the theoretical basis can found in Nechad et al. (2009, 2010) The algorithm relates turbidity ($T$) to water reflectance ($\rho_w$) through

$$T = \frac{A_f \rho_w(\lambda)}{(1-\rho_w(\lambda)/C)} \ [\text{FNU}] \ (A1)$$

where $A_f$ and $C$ are two wavelength-dependent calibration coefficients. For simplicity, wavelength ($\lambda$) will be dropped hereafter. The parameter $C$ is the asymptotic limit of Eq. (1) where $T$ tends to infinity. It is determined only by the type of particles and not by their concentration and is given by

$$C = \gamma \frac{b_{spT}}{a_{pt} + b_{spT}} \ [\text{dimensionless}] \ (A2)$$

where $a_{pt}$ and $b_{spT}$ are the turbidity-specific particulate absorption and backscattering coefficients, and $\gamma$ is a factor relating reflectance to inherent optical properties

$$\gamma = \frac{\pi \gamma f}{Q} \ (A3)$$

where $\gamma$ represents reflection and refraction effects at the surface (Morel & Gentili, 1996), $f$ is a varying dimensionless factor (Morel &
where \( \zeta \) is the non-particulate absorption coefficient, the sum of absorption from pure water and the colored dissolved organic matter absorption coefficient \( \alpha_{CDOM} \). The parameter \( \zeta \) was calibrated using “standard” inherent optical properties (IOPs) since errors in calibration of \( \zeta \) have negligible impact in the linear regime of Eq. (1) where the algorithm is used. Thus, average coefficients and typical IOPs for coastal waters found in the literature were used (more details in Nechad et al., 2010). The \( \zeta \) coefficient was obtained by a non-linear least-square regression analysis using in situ measurements of \( T \) and \( \rho_w \).

References


