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# Removal of surface-reflected light for the measurement of remote-sensing reflectance from an above-surface platform

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**Abstract:** Using hyperspectral measurements made in the field, we show that the effective sea-surface reflectance  $\rho$  (defined as the ratio of the surface-reflected radiance at the specular direction corresponding to the downwelling sky radiance from one direction) varies not only for different measurement scans, but also can differ by a factor of 8 between 400 nm and 800 nm for the same scan. This means that the derived water-leaving radiance (or remote-sensing reflectance) can be highly inaccurate if a spectrally constant  $\rho$  value is applied (although errors can be reduced by carefully filtering measured raw data). To remove surface-reflected light in field measurements of remote sensing reflectance, a spectral optimization approach was applied, with results compared with those from remote-sensing models and from direct measurements. The agreement from different determinations suggests that reasonable results for remote sensing reflectance of clear blue water to turbid brown water are obtainable from above-surface measurements, even under conditions of high waves.

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28. A Microsoft Excel template of this processing scheme is available for interested practitioners.

## 1. Introduction

The remote-sensing reflectance of water ( $R_{rs}$ ,  $\text{sr}^{-1}$ ) is defined as the ratio of the water-leaving spectral radiance ( $L_w$ ,  $\text{W m}^{-2} \text{nm}^{-1} \text{sr}^{-1}$ ) to downwelling spectral irradiance just above the surface ( $E_d(0^+)$ ,  $\text{W m}^{-2} \text{nm}^{-1}$ ).  $R_{rs}$  (or  $L_w$ ) is the basis for development of remote-sensing algorithms as well as for satellite sensor vicarious calibration [1–3]. Because of various technique limitations and the random motion of the water surface, accurate determination of  $R_{rs}$  remains a challenge [2–5]. The measurement of  $R_{rs}$  in marine environments usually involves one of these approaches: 1) measure the vertical distributions of upwelling radiance ( $L_u(z)$ ) and downwelling irradiance ( $E_d(z)$ ) within the water, and then propagate these measurements upward across the sea surface to calculate  $R_{rs}$  [6]; 2) use one sensor to measure  $L_u$  a few centimeters below the surface and use another sensor to measure  $E_d(0^+)$  above the surface, and then propagate  $L_u$  across the surface to calculate  $R_{rs}$  [7]; 3) measure all relevant quantities from an above-surface platform [1–5,8–12], and then calculate  $L_w$  (or  $R_{rs}$ ) by removing surface-reflected light ( $L_{SR}$ ). This third approach is widely used in the field and for continuous measurements [3,4,11,12], although each approach has its own advantages and disadvantages [2,10].

When measurements are made from above the sea surface [see Fig. 1(a)], the measured signal is the total upwelling radiance ( $L_T$ ), which is the sum of the water-leaving radiance ( $L_W$ ) and the surface-reflected radiance ( $L_{SR}$ ). It is necessary to avoid viewing surface foam, the shadow of the platform structure, and obvious solar glint spots. Some surface-reflected light (mostly from downwelling sky radiance, but possibly including some sun glint) is inevitable, however. A correction is therefore required to remove the surface-reflected light from  $L_T$  in order to compute  $L_W$  and  $R_{rs}$  [2,4,8]. One approach to the removal of surface-reflected radiance was proposed by Mobley [13] (a similar description can be found in Morel [1]). In this technique, all  $L_{SR}$  is expressed as the product of  $\rho$  – an effective surface reflectance – and sky radiance ( $L_{sky}$ ) measured for an angle reciprocal to the measurement of  $L_T$  (see Fig. 1 in Ref. [13]). The value of  $\rho$  depends on sea state, sky conditions, and viewing geometry [13,14]. Two approaches have then been proposed for the determination of  $\rho$ : One is to derive the value of  $\rho$  from measured  $L_T$  and  $L_{sky}$  by assuming  $L_W$  approaches 0 at near-infrared wavelengths (e.g. at 780 nm) [1]; the other is to use a table of  $\rho$  values derived from numerical simulations with various wind speeds and viewing geometries [13]. Both approaches [1,13], however, assume that the  $\rho$  value is spectrally constant. To minimize the impact of sun glint on the derivation of  $L_W$  (or  $R_{rs}$ ), Hooker et al. [2] and Zibordi et al. [4] suggested filtering out the higher measured total radiance ( $L_T$ ) values, and reasonably good results were achieved for  $L_W$  in the 412–555 nm range (larger uncertainties were found at 670 nm [4]). Here, after describing the general dependence of  $\rho$ , we show with hyperspectral measurements that  $\rho$  in general varies with wavelength, and that the spectral variation can be significant. We further compare two physical-mathematical approaches and a direct measurement scheme for the removal of  $L_{SR}$  in deriving  $R_{rs}$ .

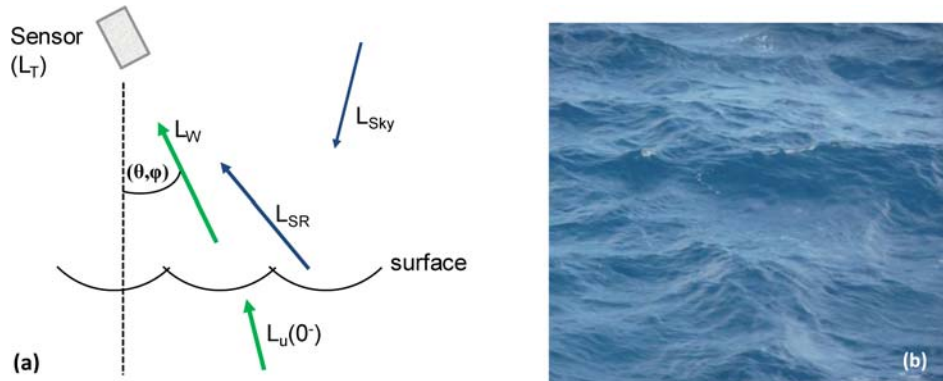


Fig. 1. (a) Schematic illustration of above-surface measurement of  $L_T$ . (b) Example of roughened sea surface when looking down from an above-surface platform. The different shades of blue result from light reflected from different parts of the sky.

## 2. Theoretical background

When a radiance instrument takes measurements of spectral upwelling radiance ( $L_T(\lambda)$ ) from an above-surface platform, it collects not only the radiance emerging from below the water surface (the so-called water leaving radiance,  $L_W(\lambda)$ ), but also surface-reflected light ( $L_{SR}(\lambda)$ ). If the measurement angle is  $\theta$  from nadir and  $\varphi$  (azimuth) from the solar plane [see Fig. 1(a)], then for a level sea surface,  $L_{SR}(\lambda)$  comes from the zenith angle  $\theta' = \theta$  and the same azimuthal angle ( $\varphi$ ). For the more common situation of a constantly moving, roughened, surface (see Fig. 1(b) for an example), and for typical instrument integration times of order of one second or longer (integration time is much shorter for multiband sensors [3]),  $L_{SR}(\lambda)$  actually comes from a large portion of the sky (see Fig. 1 and Fig. 2 of Mobley [13]) and may include solar radiance (sun glint). So, in general, the spectral upwelling radiance measured from an angular geometry  $(\theta, \varphi)$  is

$$L_T(\lambda, \theta, \varphi) = L_w(\lambda, \theta, \varphi) + \sum_i w_i F(\theta_i', \varphi_i', \theta, \varphi) L_{sky}(\lambda, \theta_i', \varphi_i'). \quad (1)$$

Here subscript “ $i$ ” represents the  $i^{\text{th}}$  small wave facet viewed by the sensor;  $w_i$  is the relative weighting of solid angle of the  $i^{\text{th}}$  facet to the sensor’s field-of-view solid angle;  $F$  is the Fresnel reflectance of the  $i^{\text{th}}$  facet; and  $L_{sky}$  is the downwelling radiance incident onto the  $i^{\text{th}}$  facet, which is reflected into sensor’s viewing angle.

Because  $L_{SR}(\lambda)$  is assembled in an unknown manner according to Eq. (1), removal of  $L_{SR}(\lambda)$  becomes a challenge in the field when measurements are taken from above the sea surface (or sea-surface remote sensing in analogy to satellite remote sensing). For this removal, to a first order approximation, Eq. (1) is simplified as [1,13]

$$L_T(\lambda, \theta, \varphi) = L_w(\lambda, \theta, \varphi) + \rho(\theta, \varphi) L_{sky}(\lambda, \theta', \varphi). \quad (2)$$

Here  $L_{sky}(\theta', \varphi)$  is the sky radiance in the same plane as that of  $L_T$ , but with  $\theta'$  the reciprocal (specular) angle of  $\theta$  [2,4,8,9].  $\rho(\theta, \varphi)$  is the effective surface reflectance that accounts for reflected sky light from all directions for the given sensor direction, and is assumed to be independent of wavelength.  $\rho(\theta, \varphi)$  equals the Fresnel reflectance of the sea surface only if the surface is flat (without waves). Values of  $\rho(\theta, \varphi)$  for various viewing directions, sun zenith angles, and wind speeds were evaluated with numerical simulations in [13]. Based on these simulations, it was suggested to use  $\theta = 40^\circ$  from the nadir and  $\varphi = 135^\circ$  from the sun to minimize  $L_{SR}$  when measuring  $R_{rs}$  in the field.

Comparing Eqs. (1) and (2) gives

$$\rho(\theta, \varphi) = \frac{\sum_i w_i F(\theta_i', \varphi_i', \theta, \varphi) L_{sky}(\lambda, \theta_i', \varphi_i')}{L_{sky}(\lambda, \theta', \varphi)}, \quad (3a)$$

or,

$$\rho(\theta, \varphi) = \frac{L_T(\lambda, \theta, \varphi) - L_w(\lambda, \theta, \varphi)}{L_{sky}(\lambda, \theta', \varphi)}. \quad (3b)$$

Because  $L_{sky}$  in general has different spectral shapes for different directions [1] (e.g., for a clear sky at noon,  $L_{sky}$  from the horizon appears whiter than that from the zenith),  $\rho$  in Eq. (2) or Eq. (3a) will in general be spectrally dependent, especially when solar light is inevitably reflected into sensor’s viewing angle by roughened surface, unless the sky is completely overcast.

### 3. Data and methods

To demonstrate the spectral variation of  $\rho$ , hyperspectral measurements over clear oceanic waters were utilized, where the contribution of water to  $L_T$  is negligible in the longer wavelengths. The measurements were made on Feb. 23, 1997 around 12:50 pm (local time), for waters near Hawaii at 21.33 N, 158.16 W. The sky was clear with no clouds, the water appeared blue, the wind speed was around  $8 \text{ m s}^{-1}$ , and the surface wave amplitude was  $\sim 2$  to 3 feet ( $\sim 1 \text{ m}$ ).

Upwelling total radiance ( $L_T$ , 9 scans), downwelling sky radiance ( $L_{sky}$ , 5 scans), and “gray-card” radiance ( $L_G$ , 3 scans) reflected from a standard diffuse reflector (Spectralon®) were measured with a handheld spectroradiometer (SPECTRIX [15]), which covers a spectral range  $\sim 360 - 900 \text{ nm}$  with a spectral resolution  $\sim 2 \text{ nm}$  and has an integration time about 1.5 seconds for the collection of  $L_T$ . The orientation to measure  $L_T$  was  $30^\circ$  from nadir and  $90^\circ$  from the solar plane.  $L_{sky}$  was measured in the same plane as  $L_T$ , but at an angle  $30^\circ$  from zenith. Downwelling irradiance was determined by assuming that the Spectralon is a lambertian reflector, so that  $E_d = \pi L_G / R_G$ , with  $L_G$  the average of the three scans and  $R_G$  the

reflectance of the diffuse reflector (~10%). The measurement was taken at the bow of a large ship with a sensor to water-surface distance about 5 meters. The SPECTRIX has a 10° field of view, which then observed a surface area of ~1 m<sup>2</sup> for this setup.

To evaluate the value and variations of the effective surface reflectance ( $\rho$ ), Eq. (2) is converted to reflectances, where the total remote-sensing reflectance ( $T_{rs}$ , ratio of  $L_T$  to  $E_d$ ) and sky remote-sensing reflectance ( $S_{rs}$ , ratio of  $L_{Sky}$  to  $E_d$ ) were calculated for each  $L_T$  and  $L_{Sky}$  scan, respectively. From Eq. (2), these  $T_{rs}$ ,  $R_{rs}$ , and  $S_{rs}$  are related as

$$T_{rs}(\lambda, \theta, \varphi) = R_{rs}(\lambda, \theta, \varphi) + \rho(\theta, \varphi) S_{rs}(\lambda, \theta', \varphi), \quad (4a)$$

or

$$R_{rs}(\lambda, \theta, \varphi) = T_{rs}(\lambda, \theta, \varphi) - \rho(\theta, \varphi) S_{rs}(\lambda, \theta', \varphi). \quad (4b)$$

Further,  $\rho$  is calculated as

$$\rho(\theta, \varphi) = \frac{T_{rs}(\lambda, \theta, \varphi) - R_{rs}(\lambda, \theta, \varphi)}{S_{rs}(\lambda, \theta', \varphi)}. \quad (5)$$

For the calculation of  $\rho$ ,  $R_{rs}(\lambda, 30^\circ, 90^\circ)$  (assumed equal to  $R_{rs}(\lambda, 0^\circ, 0^\circ)$ ), and for  $\lambda$  in a range of 400 – 800 nm) was estimated with the bio-optical model of Morel and Maritorena [16] and using a chlorophyll-a concentration of  $\text{Chl} = 0.05 \text{ mg m}^{-3}$ , which is an estimate based on observations of MODIS for these waters in February. The model coefficients of Morel and Maritorena [16] cover wavelengths up to 700 nm. For the study here, the model coefficients of wavelengths greater than 700 nm are considered the same as that at 700 nm, except for the attenuation coefficient of pure water, which was replaced with the absorption coefficient of clear natural water [17].

## 4. Results

### 4.1 Variation of $\rho$

For this station,  $T_{rs}$  and  $S_{rs}$  are presented in Figs. 2(a) and 2(b), respectively. Because the sea surface is roughened by waves, as commonly encountered in the field, we did not get identical  $T_{rs}$  for the 9 independent measurements of  $L_T$ . This is because each  $L_T$  measurement observed a different sea surface, hence a different sky coverage, and thus a different  $L_{SR}$ . Nor did we get identical measurements of the 5  $S_{rs}$  because the boat was also constantly moving, and thus the sensor could not maintain the exactly same angular geometry for the different sky-viewing measurements.

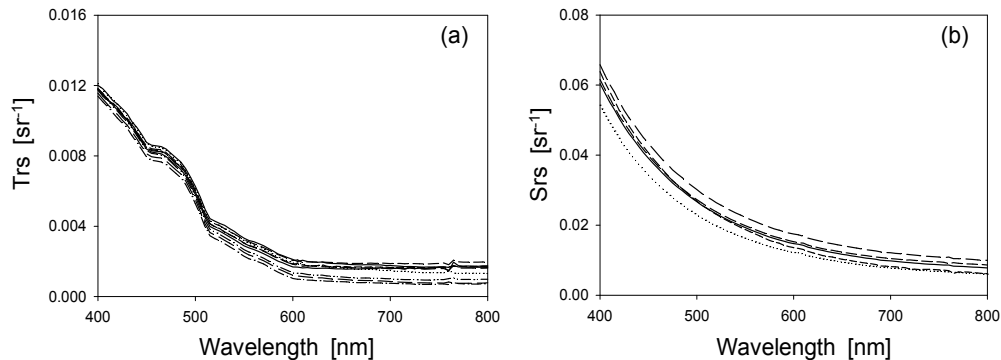


Fig. 2. Measured  $T_{rs}$  (a) and  $S_{rs}$  (b).

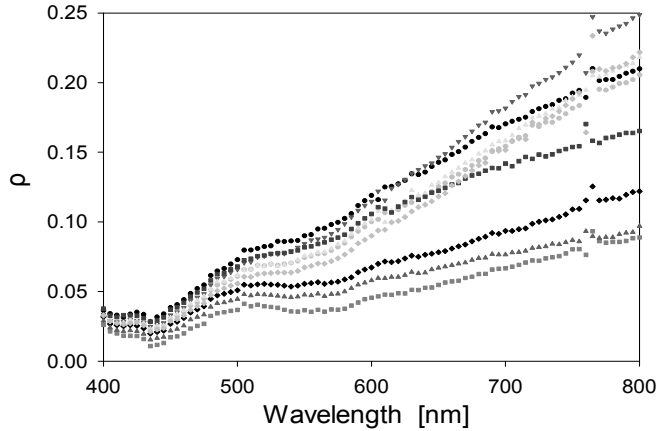


Fig. 3.  $\rho$  values calculated from measured  $T_{rs}$  and  $S_{rs}$ .  $R_{rs}$  was modeled with  $\text{Chl} = 0.05 \text{ mg m}^{-3}$  based on the bio-optical model of Morel and Maritorena [16].

For illustration purposes, Fig. 3 shows values of  $\rho$  calculated for the 9  $T_{rs}$  scans and with  $S_{rs}$  from the first measurement used as the denominator in Eq. (5). It is seen, not surprisingly, that the  $\rho$  values differ among the different  $L_T$  measurements. More importantly, the  $\rho$  values differ spectrally, and this difference can be as high as a factor of eight between 400 nm and 800 nm (a factor of five between 400 nm and 700 nm). The increase of  $\rho$  with wavelength occurs mainly because (1)  $T_{rs}$  collects  $L_{SR}$  from all directions, including the sun and near-horizon directions [recall the whitish patches in Fig. 1(b)]. Compared to sky light from zenith, radiances from these directions are richer in the longer wavelengths. (2)  $S_{rs}$  is measured from one fixed angular geometry, and this  $S_{rs}$  is usually blue rich (dominated by contributions from Rayleigh scattering).

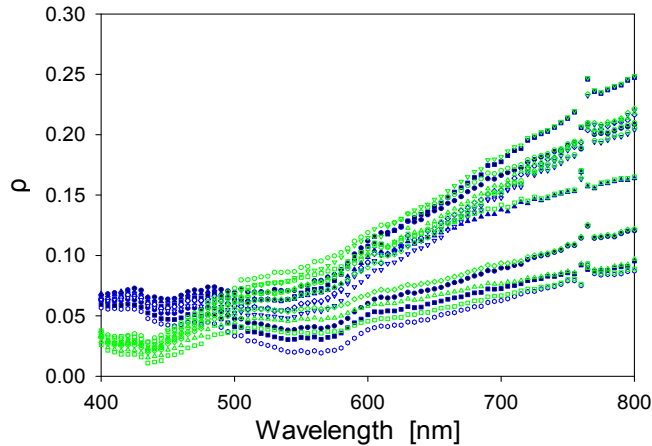


Fig. 4. Similar as Fig. 3, but with two different Chl values. Green:  $\text{Chl} = 0.05 \text{ mg m}^{-3}$ ; blue,  $\text{Chl} = 0.1 \text{ mg m}^{-3}$ .

To evaluate the impacts of incorrect  $R_{rs}$ , which were estimated from a spectral model with roughly estimated chlorophyll concentration, on the calculated  $\rho$  values, Fig. 4 compares the  $\rho$  values calculated from the 9  $T_{rs}$  measurements and the first  $S_{rs}$ , but with  $\text{Chl} = 0.05$  and  $0.1 \text{ mg m}^{-3}$ , respectively. For wavelengths in the range of  $\sim 400 - 500 \text{ nm}$ , because  $R_{rs}$  makes strong contributions to  $T_{rs}$ , wide variations of  $\rho$  values were found, which highlights the limitation of calculating the effective  $\rho$  from field measurements when the water contribution

is high. For wavelengths longer than  $\sim 550$  nm, however, it is found that the impact of different Chl values (then different  $R_{rs}$ ) on  $\rho$  is nearly negligible. This is because for such clear waters phytoplankton contribution to  $R_{rs}$  is nearly negligible at the longer wavelengths. This is further illustrated in Fig. 5 via scatter plots between  $\rho(\text{Chl} = 0.05)$  and  $\rho(\text{Chl} = 0.025)$ , and between  $\rho(\text{Chl} = 0.05)$  and  $\rho(\text{Chl} = 0.1)$ . This figure shows that Chl has very little impact on  $\rho$  values of  $\rho > 0.07$  (corresponding to  $\sim 550$  nm for the measurements in this study). The same results were found when the first  $S_{rs}$  was replaced by any of the other measurements of  $L_{sky}$  (results not shown here).

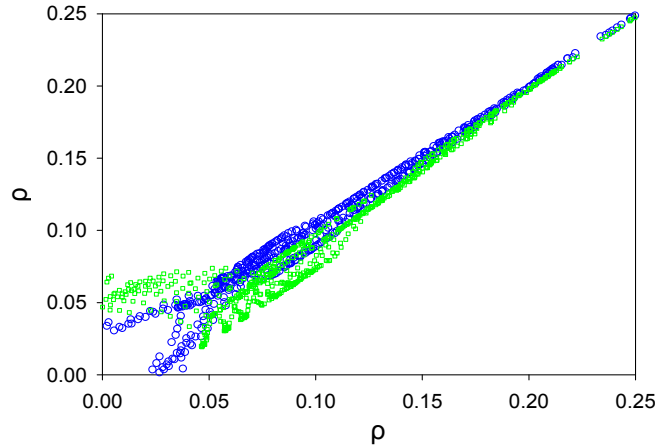


Fig. 5. Scatter plot between  $\rho(\text{Chl} = 0.05)$  and  $\rho(\text{Chl} = 0.025)$ , blue symbol; and between  $\rho(\text{Chl} = 0.05)$  and  $\rho(\text{Chl} = 0.1)$ , green symbol.

If there are clouds in the sky (assuming the sun itself is not blocked by clouds), this  $\rho$  value could vary widely with wavelength, because  $S_{rs}$  could be measured from a small portion of the clear sky (very blue) or aimed at a cloud, while  $L_{SR}$  will include radiance from clouds (nearly white) and the background blue sky. These results indicate that applying a  $\rho$  value calculated in the near infrared (e.g. 780 nm) to the shorter wavelengths will cause large uncertainties in  $R_{rs}$  in the blue bands [2], unless the measurements are made under nearly ideal conditions (no clouds, low wind, no foam on surface, and very short integration time).

#### 4.2 Removal of $L_{SR}$

The above analysis indicates that when the sea surface is not flat, 1)  $\rho$  is not a constant among measurement scans; and 2)  $\rho$  values change with wavelength, at least for the longer wavelengths ( $> \sim 550$  nm) in this study. With such an observation, even if wind speed and angular geometry are all known exactly (note that the effective  $\rho$  also depends on the orientation of waves), it will still be a daunting challenge to accurately remove  $L_{SR}$  via Eq. (2) or Eq. (4). Earlier, Hooker et al. [2] and Zibordi et al. [4] proposed to filter out the higher  $L_T$  measurements before applying Eq. (4b) for the removal of  $L_{SR}$ . This technique is generally supported by the results shown in Fig. 3, where higher spectral contrast of  $\rho$  is found for the high  $\rho(800)$  value (high  $L_T$ ). However, because it can never be known exactly which  $L_{sky}$  is reflected into the view of an  $L_T$  measurement, it is unclear how to select a proper  $\rho$  value that is relevant for the smaller  $L_T$  measurements, as using the smallest  $L_T$  to derive  $L_W$  via Eq. (4b) may result in underestimation of  $L_W$  [18].

For this station, the wind speed was about  $8 \text{ m s}^{-1}$ , so a  $\rho$  value of 0.05 was assumed for the angular geometry (based on Fig. 8 in Mobley [13]) and applied for the calculation of  $R_{rs}$  via Eq. (4). Since there were 9 measurements of  $T_{rs}$  and 5 measurements of  $S_{rs}$ , 45  $R_{rs}$  were derived. Figure 6 shows the average  $R_{rs}$  with  $\pm 1$  standard deviation as computed from the 45 spectra. As a qualitative check, the modeled  $R_{rs}$  for  $\text{Chl} = 0.05 \text{ mg m}^{-3}$  is also included in



Fig. 6. It is found that the average  $R_{rs}$  from measurements match modeled  $R_{rs}$  reasonably well for the ~400-550 nm range, but there are significant differences for the longer wavelengths. Since there are large uncertainties in the modeled  $R_{rs}$  (resulted from, likely, both inaccurate Chl value and imprecise  $R_{rs}$  model), we are not expecting the two  $R_{rs}$  matching each other exactly. However, because the water-leaving radiance (or  $R_{rs}$ ) of such waters is nearly negligible at longer wavelengths, it can be safely argued that the  $R_{rs}$  derived from Eq. (4) is overestimated for those wavelengths. This observation is consistent with Fig. 13 (left) of Mobley [13].

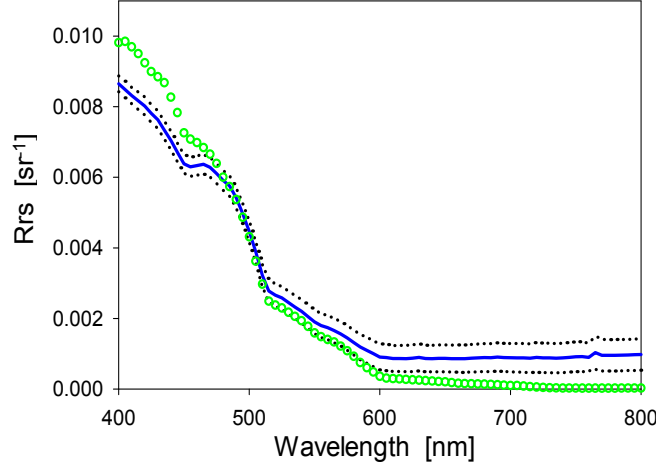


Fig. 6. Average  $R_{rs}$  (solid blue) and the one standard deviation (dotted lines), calculated using Eq. (4b). Green line is modeled  $R_{rs}$  with Chl = 0.05 mg m<sup>-3</sup> using the Morel-Maritorena bio-optical model [16].

The commonly measured properties (except grey card) are  $L_T$  at angle  $(\theta, \varphi)$  and  $L_{Sky}$  at  $(\theta', \varphi')$ . Also, because the actual total  $L_{SR}$  is not measured directly, we re-write Eq. (1) as

$$L_T(\lambda, \theta, \varphi) = L_w(\lambda, \theta, \varphi) + w_0 F(\theta, \varphi) L_{sky}(\lambda, \theta', \varphi') + \sum_{i=1} w_i F(\theta_i', \varphi_i', \theta, \varphi) L_{sky}(\lambda, \theta_i', \varphi_i'). \quad (6)$$

Here  $w_0$  represents the weighting of sky light coming from the specular direction  $(\theta', \varphi')$ , and the sum now includes sky light from all other directions. Further, since sky light from the specular direction  $(\theta', \varphi')$  dominates  $L_{SR}$  from the field-of-view centered at  $(\theta, \varphi)$  [13], Eq. (6) is approximated (by setting  $w_0 = 1$ ) as,

$$L_T(\lambda, \theta, \varphi) \approx L_w(\lambda, \theta, \varphi) + F(\theta, \varphi) L_{sky}(\lambda, \theta', \varphi') + \sum_{i=1} w_i F(\theta_i', \varphi_i', \theta, \varphi) L_{sky}(\lambda, \theta_i', \varphi_i'), \quad (7)$$

or, in terms of reflectance,

$$T_{rs}(\lambda, \theta, \varphi) \approx R_{rs}(\lambda, \theta, \varphi) + F(\theta, \varphi) S_{rs}(\lambda, \theta', \varphi') + \sum_{i=1} w_i F(\theta_i', \varphi_i', \theta, \varphi) S_{rs}(\lambda, \theta_i', \varphi_i'). \quad (8)$$

Now in Eq. (8) both  $T_{rs}$  and  $S_{rs}$  for the specular direction are directly determined from measurements.  $F(\theta, \varphi)$  is the Fresnel reflectance of water surface for  $(\theta, \varphi)$ , which is known for a given angular geometry. For the calculation of  $R_{rs}$  from Eq. (8) for any measurement of  $L_T$  and  $L_{sky}$ , it is thus necessary to determine the last term on the right-hand side of Eq. (8).

Because it is not known yet how this last term varies spectrally, this term is assumed for expediency to be spectrally independent [9]. Thus Eq. (8) becomes

$$T_{rs}(\lambda, \theta, \varphi) \approx R_{rs}(\lambda, \theta, \varphi) + F(\theta, \varphi) S_{rs}(\lambda, \theta', \varphi) + \Delta(\theta, \varphi), \quad (9a)$$

or [9],

$$R_{rs}(\lambda, \theta, \varphi) \approx T_{rs}(\lambda, \theta, \varphi) - F(\theta, \varphi) S_{rs}(\lambda, \theta', \varphi) - \Delta(\theta, \varphi). \quad (9b)$$

Thus, for each set of spectral  $T_{rs}$  and  $S_{rs}$ , there is a spectrally constant value ( $\Delta$ , a bias) that must be determined before  $R_{rs}$  can be derived. For oceanic waters where  $R_{rs}$  is negligible in the red and near infrared,  $\Delta$  can be estimated by assuming  $R_{rs}$  near 750 nm is 0 [19]. For coastal turbid waters, however, this assumption is no longer valid. For such environments, one approach [19,20] is to model the spectral  $R_{rs}$  as a function of spectral inherent optical properties (IOPs), and then solve for  $\Delta$  by comparing modeled spectral  $R_{rs}$  with spectral  $R_{rs}$  derived from Eq. (9b) using all measured spectral information (so-called spectral optimization) [21–24].

Basically, for optically deep waters, the spectral  $R_{rs}$  can be conceptually summarized as

$$R_{rs}(\lambda, \theta, \varphi) \approx \text{Fun}(a(\lambda), b_b(\lambda), \theta, \varphi), \quad (10)$$

where  $a(\lambda)$  is the absorption coefficient, and  $b_b(\lambda)$  is the backscattering coefficient. The inherent optical properties  $a(\lambda)$  and  $b_b(\lambda)$  can be modeled with bio-optical models of optically active components [22,24,25], so that Eq. (10) becomes explicit functions as

$$\begin{cases} R_{rs}(\lambda_1, \theta, \varphi) \approx \text{Fun}(a_w(\lambda_1), b_{bw}(\lambda_1), P, G, X, \theta, \varphi), \\ R_{rs}(\lambda_2, \theta, \varphi) \approx \text{Fun}(a_w(\lambda_2), b_{bw}(\lambda_2), P, G, X, \theta, \varphi), \\ \dots \\ R_{rs}(\lambda_n, \theta, \varphi) \approx \text{Fun}(a_w(\lambda_n), b_{bw}(\lambda_n), P, G, X, \theta, \varphi). \end{cases} \quad (11)$$

Here  $\lambda_1$  to  $\lambda_n$  are the sensor's wavelengths,  $a_w$  and  $b_{bw}$  are the known absorption and backscattering coefficients of pure seawater, and  $P$ ,  $G$ , and  $X$  represent the magnitude of the absorption coefficient of phytoplankton, gelbstoff, and the backscattering coefficient of particles, respectively. To derive the value of  $\Delta$  in Eq. (9b), an objective function is defined as

$$\text{Err} = \frac{\left[ \int_{400}^{675} (R_{rs} - \tilde{R}_{rs})^2 + \int_{750}^{800} (R_{rs} - \tilde{R}_{rs})^2 \right]^{0.5}}{\int_{400}^{675} R_{rs} + \int_{750}^{800} R_{rs}}, \quad (12)$$

with  $\tilde{R}_{rs}$  from Eq. (11) while  $R_{rs}$  from Eq. (9b).  $\int_{400}^{675}$  represents the average of an array between 400 nm and 675 nm. The upper bound of wavelength (800 nm) can be extended to a longer wavelength for turbid lake or river waters when sensor has measurements in those wavelength ranges.  $\text{Err}$  is then a function of 4 variables ( $P$ ,  $G$ ,  $X$ , and  $\Delta$ ) for optically deep waters, and they are derived numerically when  $\text{Err}$  reaches a minimum – spectral optimization [22,26].  $R_{rs}$  is therefore computed by applying this numerically derived  $\Delta$  to Eq. (9b). Note that in the correction of  $L_{SR}$  the focus is the estimation of  $\Delta$ , although values of  $P$ ,  $G$ , and  $X$  are also determined.

For the measurements at this station, again, 45 spectral  $R_{rs}$  were determined with this spectral optimization method, and their average and standard deviation are presented in Fig. 7. The overestimations of  $R_{rs}$  in the longer wavelengths ( $> \sim 550$  nm) are generally removed, as compared to Fig. 6. At the same time, the average  $R_{rs}$  matches the modeled  $R_{rs}$  (with Chl =  $0.05 \text{ mg m}^{-3}$ ) very well in the  $\sim 400$ -550 nm range, although it is not our intention (the measured and Chl-modeled  $R_{rs}$  do not necessarily represent the same water environments).

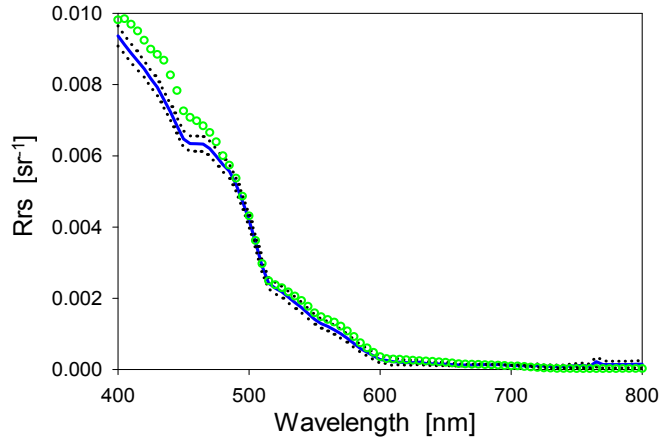


Fig. 7. Similar as Fig. 6, but  $R_{rs}$  was calculated based on Eq. (9) and with a spectral optimization scheme. Green line is modeled  $R_{rs}$  with  $\text{Chl} = 0.05 \text{ mg m}^{-3}$  using the Morel-Maritorena [16] bio-optical model.

To further test the above evaluation and the optimization approach of removing  $L_{SR}$ , new measurements (September 13, 2010; ~11 am local time) were carried out (with SPECTRIX) over turbid river water (Pearl River, Mississippi, USA. Figure 8 shows color photos of the water and sky when measurements were taken). This shallow (~0.5 m) and very turbid water makes it nearly impossible to obtain  $R_{rs}$  from measurements of vertical profiles of  $L_u$  and  $E_d$  [6]. During the experiment, the surface was calm [see Fig. 8(a)] and the sky was blue [Fig. 8(b)] with some thin cirrus clouds. Two different measurement schemes were carried out. One followed the traditional scheme [10] that measures  $L_G$ ,  $L_T$  and  $L_{Sky}$  (see Section 3), with  $\theta = 30^\circ$  from nadir and  $\phi = 90^\circ$  from the solar plane, and the sensor to water-surface distance was ~1 meter (the sensor then covered a surface area ~0.05 m<sup>2</sup>).  $R_{rs}$  were derived, separately, from these measurements following the simple approach [Eq. (4b),  $\rho = 0.022$  is used for calm surface.  $R_{rs-simp}$  in the following] and following the optimization approach ( $R_{rs-opt}$  in the following) mentioned above.

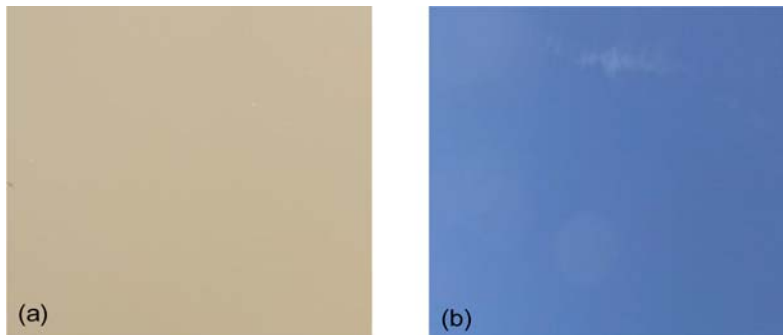


Fig. 8. Color photos of the river water (a) and sky (b) measured on September 13, 2010, ~11 am local time.

Another measurement followed a novel scheme proposed by Ahn et al [27], where a small black tube (~4 cm in diameter) was placed in front of the sensor to block  $L_{SR}$  (see Fig. 9 for a schematic illustration). When  $L_T$  was measured the tube was dipped just below the sea surface (~5 cm) while the sensor itself was kept above the surface. Therefore there will be no  $L_{SR}$  into the sensor in this setup and the instrument records a direct measurement of  $L_W$ .  $R_{rs}$  ( $R_{rs-direct}$  in the following) was then derived as the ratio of measured  $L_W$  to  $E_d$  (from measurement of  $L_G$ ).

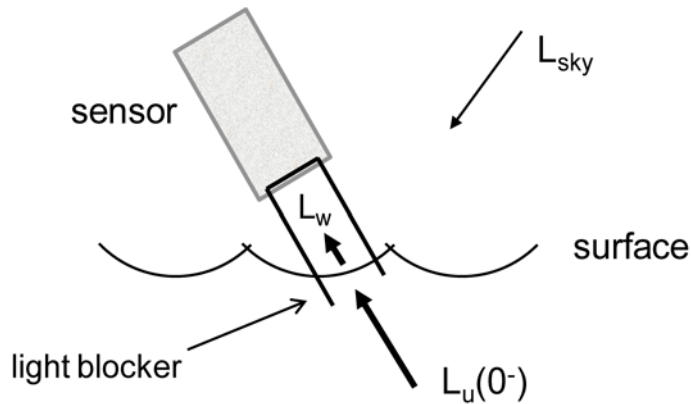


Fig. 9. Scheme to measure  $L_w$  directly (re-drawn from Ahn et al [27]). The open box is a black tube to block surface reflected light, which is inserted just below ( $\sim 5$  cm) the surface when measuring  $L_w$ .

Figure 10 shows the derived  $R_{rs}$  from the three measurement-determination schemes; blue is  $R_{rs-direct}$ , green is  $R_{rs-opt}$ , and cyan is  $R_{rs-simp}$ . The three  $R_{rs}$  curves show similar spectral shapes, which are typical of turbid, high-CDOM river waters (note the yellow-brown color in Fig. 8).  $R_{rs-simp}$  is considerably higher than both  $R_{rs-direct}$  and  $R_{rs-opt}$ , suggesting incomplete removal of  $L_{SR}$  even for this quite calm situation (it may be that some sun glitter could not be completely avoided for the  $(30^\circ, 90^\circ)$  viewing geometry and integration times of  $\sim 1$ -2 seconds). On the other hand,  $R_{rs-direct}$  and  $R_{rs-opt}$  are very consistent across the  $\sim 400$ -850 nm range, with a coefficient of variation about  $\sim 11\%$  (which is about 46% between  $R_{rs-simp}$  and  $R_{rs-direct}$ ). The slight negative  $R_{rs}$  (both  $R_{rs-direct}$  and  $R_{rs-opt}$ ) for wavelengths shorter than 400 nm may result from a combination of 1) SPECTRIX has lower signal-to-noise ratio for wavelengths shorter than  $\sim 400$  nm [15], and 2) the extremely low upwelling signal in the blue-to-UV wavelengths of this CDOM-rich water. Nevertheless, the deduced  $R_{rs}$  of this turbid water (along with the result of blue oceanic waters) strongly indicates that Eq. (9b) with an optimization scheme to determine the value of  $\Delta$  is adequate in obtaining  $R_{rs}$  in the field when measurements are made above the sea surface under un-ideal conditions and that  $L_{SR}$  is not blocked during measurements. However, neither  $R_{rs-direct}$  nor  $R_{rs-opt}$  are error free, because  $R_{rs-direct}$  encounters some self-shading and/or contributions from reflectance inside the tube, while  $R_{rs-opt}$  suffers from the approximation from Eq. (6) to Eqs. (7) and (9a).

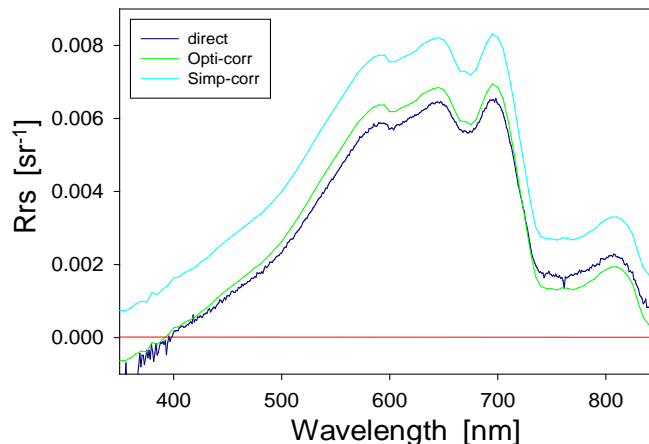


Fig. 10. Comparison between directly measured  $R_{rs}$  (blue line) of water showing in Fig. 8(a) and  $R_{rs}$  obtained after correcting surface-reflected light.

## 5. Conclusions

Using measurements from a clear-water station, we demonstrated that the effective surface reflectance ( $\rho$ ) varies not only with each measurement scan but also with wavelength. Consequently, application of a spectrally constant  $\rho$  value for the removal of  $L_{SR}$  from above-surface measurements is a crude approximation, especially if the sea surface is significantly roughened by waves and the sensor has a long integration time (as do most high-spectral resolution sensors). Earlier studies [2,4,18] have shown that it is wise to filter out the higher  $L_T$  measurements before the derivation of  $L_W$  when the simple formula [e.g., Eq. (4b)] is used for the derivation. Here we show that for clear to turbid waters, a spectral optimization scheme [28] is also adequate to remove  $L_{SR}$  in  $L_T$  measurements and derive reasonable  $R_{rs}$ . Further, the scheme to block  $L_{SR}$  by equipping a black tube in front of the sensor and dipping it just below the surface shows promise to obtain reliable measurement of  $L_W$  without the difficult post-processing. Further effort is required by the remote-sensing community to evaluate these approaches for a wide range of environments and measurement conditions (e.g. Hooker et al. [2] and Toole et al [7]) and then to establish a consensus for the optimum way to determine  $R_{rs}$  in the field when measurements are made from an above-surface platform, especially for situations such as turbid waters and partly cloudy skies.

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