Article

Deriving the Vertical Variations in the Diffuse Attenuation Coefficient of Photosynthetically Available Radiation in the North Pacific Ocean from Remote Sensing

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Abstract: Diffuse attenuation coefficient of photosynthetically available radiation (PAR), $K_{PAR}$, is a key product of ocean color remote sensing. Current ocean color algorithms generally detect only the average $K_{PAR}$ within one optical depth, $K_{RS}^{PAR}$. Due to the marked vertical variations of $K_{PAR}$, knowledge of $K_{RS}^{PAR}$ is insufficient to accurately evaluate the submarine light field. By using field in situ observations, a two-step approach, based on the development of an ocean color algorithm for $K_{RS}^{PAR}$ and the relationships between $K_{RS}^{PAR}$ and the average $K_{PAR}$ from the surface down to depth $Z$ ($K_{Z}^{PAR}$), was developed to remotely estimate the vertical variations in $K_{Z}^{PAR}$ in the North Pacific from the MODerate-resolution Imaging Spectrometer (MODIS) imagery. The root mean square difference of log($K_{Z}^{PAR}$) in depths within the euphotic zone was around ±0.059 (in unit of m$^{-1}$ for $K_{Z}^{PAR}$), which corresponded to a deviation of ±15% for the estimated $K_{Z}^{PAR}$ and the penetration depths of PAR. Our study may provide a promising approach to detect the vertical variations of $K_{Z}^{PAR}$ and underwater PAR distributions in the North Pacific Ocean.

Keywords: diffuse attenuation coefficient; photosynthetically available radiation; vertical variation; remote sensing; North Pacific

1. Introduction

Knowledge of the vertical distribution of photosynthetically available radiation (PAR), which is associated with approximately the visible part of the solar radiation spectrum from 400 to 700 nm, is essential to understanding the underwater biogeochemical processes and the response of marine ecosystem to environmental changes [1,2]. Playing a key role in driving water-column photosynthesis, the vertical distribution of PAR, especially in the euphotic layer, is needed for accurately quantifying and modeling the aquatic primary production, which, as a whole for the global oceans, accounts for about half of global carbon fixation and thus alters the global oxygen and carbon balances [3–5]. In addition to being a controlling factor in the water-column photosynthesis, the vertical distribution of PAR is fundamental to predictions of vertical subsurface heating rates, given that more than 98% of total incoming solar irradiance as short wave is converted to heat. Subsurface heating can lead to instabilities of the upper layer, which in turn affects the physical and
biogeochemical processes and photo-oxidation in the water column [6,7]. Moreover, the status of the penetration of PAR is often applied to evaluate the water clarity, which is an important environmental driver in benthic communities such as seagrass and coral reefs [8,9]. Nevertheless, an improvement will be realized in predicting climate changes and monitoring water quality with the improvement in the assessment of the vertical distribution of PAR.

According to the Lambert–Beer Law, PAR decreases approximately exponentially with depth (Z) and can be expressed as follows [1,10–15]:

\[
PAR(Z) = PAR(0^-) \exp\left(-\mathbf{K}_{\mathbf{PAR}}^Z \times Z\right)
\]

where \( PAR(Z) \) and \( PAR(0^-) \) are the PAR values at depth \( Z \) and just below the surface, respectively, while \( \mathbf{K}_{\mathbf{PAR}}^Z \) represents the average diffuse attenuation coefficient of the PAR (\( K_{\mathbf{PAR}} \)) from the surface down to depth \( Z \). Since \( PAR(0^-) \) can be modeled accurately [16], the determination of \( PAR(Z) \) is dependent almost exclusively on the accurate estimation of \( \mathbf{K}_{\mathbf{PAR}}^Z \). As an apparent optical property (AOP), however, \( K_{\mathbf{PAR}} \), as well as \( \mathbf{K}_{\mathbf{PAR}}^Z \), is dependent on not only the water-column inherent optical properties (IOPs) such as absorption and scattering coefficients of water components, but also the angular structure of the submarine light field [11]. As such, \( K_{\mathbf{PAR}} \) is pronouncedly depth-dependent, varying by even more than a factor of two in the euphotic zone [10,12,13,17].

\( K_{\mathbf{PAR}} \), and thus \( \mathbf{K}_{\mathbf{PAR}}^Z \), can be measured in situ relatively simply, but such data are severely limited in their spatial and temporal coverage. On the other hand, satellite remote sensing provides a synoptic, large-scale, and continuous view of the oceans to augment the direct field in situ observations [18]. Efforts were made in past decades to develop and validate ocean color algorithms for \( K_{\mathbf{PAR}} \) [10,12,19–21]. Three types of algorithms are commonly developed for remotely estimating \( K_{\mathbf{PAR}} \): chlorophyll-based approach [12,19,22], IOP-based approach [10,13], and empirical algorithm based on the band ratios of remote sensing reflectance \( R_{\text{rs}} \) [20]. The first two types require remote sensing products, e.g., chlorophyll a (Chl_a) concentration and IOPs of absorption and backscattering coefficients, respectively. An additional uncertainty is likely introduced when propagating these products to estimate \( K_{\mathbf{PAR}} \). Furthermore, the vertical variations in Chl_a and IOPs, especially in stratified waters, may also cause difficulty in evaluating the vertical distributions in \( K_{\mathbf{PAR}} \) [10,22].

The third type, namely, the \( R_{\text{rs}} \)-based empirical algorithm, has been routinely used to estimate the diffuse attenuation coefficients in the global oceans from space [20]. For historic reasons, algorithms for the diffuse attenuation coefficient of downwelling plane irradiance at 490 nm (\( K_{490} \)), rather than \( K_{\mathbf{PAR}} \), are commonly developed [20,21]. For example, the remotely sensed \( K_{490} (K_{\text{RS490}}) \) by the MODerate-resolution Imaging Spectrometer on Aqua (MODIS-Aqua) is estimated as

\[
\log \left( K_{\text{RS490}}^{\text{RS}} - 0.0166 \right) = -0.8515 - 1.8263X + 1.8714X^2 - 2.4414X^3 - 1.0690X^4
\]

where \( X = \log[R_{\text{rs}}(488)/R_{\text{rs}}(555)] \). In order to convert \( K_{\text{RS490}}^{\text{RS}} \) to the remotely sensed \( K_{\text{PAR}}^{\text{RS}} \), an empirical algorithm is commonly used such that [12]

\[
K_{\text{PAR}}^{\text{RS}} = 0.0864 + 0.884K_{\text{RS490}}^{\text{RS}} - 0.00137/K_{\text{RS490}}^{\text{RS}}
\]

Although the combination of Equations (2) and (3) provides an operational approach to estimate \( K_{\text{PAR}}^{\text{RS}} \) from the MODIS-Aqua imagery, two questions are raised when applying it to evaluate the vertical distributions in \( K_{\mathbf{PAR}} \). First, this approach requires the estimation of \( K_{\text{RS490}}^{\text{RS}} \), and an additional uncertainty may be introduced when propagating \( K_{\text{RS490}}^{\text{RS}} \) to estimate \( K_{\text{PAR}}^{\text{RS}} \). Secondly, there exists a gap between \( K_{\text{RS490}}^{\text{RS}} \) and the vertical variations in \( K_{\mathbf{PAR}} \). \( K_{\text{PAR}}^{\text{RS}} \) is approximately equal to \( \mathbf{K}_{\mathbf{PAR}}^{Z_{37}} \), the average \( K_{\mathbf{PAR}} \) from the surface down to one optical
depth at which PAR is reduced to about 37% (e^{-1} ≈ 37%) of its surface value [13], since more than 90% of water-leaving radiance is reflected from the upper one optical depth [23]. \( K_{\text{PAR}} \) is thus valid to evaluate the penetration of PAR in the layer above one optical depth only. Even with vertically uniform profiles of IOPs, \( K_{\text{PAR}} \) can vary by a factor of two; thus, its value below one optical depth may be much lower than \( K_{\text{PAR}}^{RS} \) [10,13]. As such, the penetration depths of PAR in the lower layer may be substantially underestimated if using a constant \( K_{\text{PAR}} \) equal to \( K_{\text{PAR}}^{RS} \) for the whole water column. Therefore, the relationship between \( K_{\text{PAR}}^{RS} \) and \( K_{\text{PAR}}^{Z} \), in addition to the ocean color algorithms for \( K_{\text{PAR}}^{RS} \), has to be elucidated in order to accurately quantify the submarine penetration of PAR from space. Although previous works have reported the estimations of \( K_{\text{PAR}}^{Z} \) on few specific layers, e.g., from the surface down to the base of the euphotic zone [10,14,15,17] or two optical depths [12], an approach to evaluate the vertical profiling in \( K_{\text{PAR}}^{Z} \) continuously has seldom been documented [22].

In this study, we report an attempt to develop an operational approach for remotely estimating the vertical distributions of \( K_{\text{PAR}}^{Z} \) in the oceanic water by using field observations collected in the North Pacific (Figure 1). The approach was designed in a two-step method: (1) to develop an algorithm for estimating \( K_{\text{PAR}}^{RS} \) directly from the \( R_{rs} \) band ratio, and (2) to build the relationships between \( K_{\text{PAR}}^{RS} \) and \( K_{\text{PAR}}^{Z} \). Section 2 describes the data and methods employed in this study. Section 3 shows the results and discusses them in terms of algorithm development and validation. Section 4 shows the application of the algorithms in evaluating the distributional patterns of \( K_{\text{PAR}}^{Z} \) in the North Pacific. Section 5 draws the conclusions.
2. Data and Methods

2.1. Study Area

The general distributional patterns of the biophysical properties, as shown in the climatologically seasonal distributions in spring in the MODIS-Aqua-derived sea surface temperature (SST) and Chl\textsubscript{a}, in the North Pacific are shown in Figure 1. SST generally increases southwards from ~4 °C at 50°N to ~29 °C at the equator, primarily reflecting the latitudinal dependence of heat loss/gain across the air-sea interface. The major circulations, namely, the Kuroshio, the North Pacific Current, the California Current, and the North Equatorial Current, define the approximate boundaries of the North Pacific Subtropical Gyre (NPSG), which is the most extensive gyre of the Earth. The SSTs at the north and south boundaries of the NPSG are about 15 °C and 28 °C, respectively (Figure 1a). Corresponding to the southward-increasing SST, which may indicate the strengthened vertical stratification and thus reduced nutrient supply for phytoplankton growth, Chl\textsubscript{a} generally decreases southwards, varying from ~0.4 mg m\textsuperscript{−3} at 50°N to ~0.1 mg m\textsuperscript{−3} at the equator (Figure 1b). High Chl\textsubscript{a}, e.g., over 1 mg m\textsuperscript{−3}, can be found in the coastal waters, while the most oligotrophic waters are found in the NPSG, with Chl\textsubscript{a} typically around 0.05 mg m\textsuperscript{−3}.

In terms of PAR, as exampled in its climatologically seasonal distribution in spring in the North Pacific (Figure A1 in Appendix A), it generally increases southwards from ~30 E m\textsuperscript{−2} d\textsuperscript{−1} at 50°N to ~45 E m\textsuperscript{−2} d\textsuperscript{−1} at the equator. Latitudinal maximum values of PAR typically occur at about 10–17°N. Seasonally, PAR generally increases from the winter to the summer, then following a decrease to the winter. Seasonal variations are typically more pronounced in the high-latitude region than in the low-latitude region. Such a distributional pattern may primarily reflect the spatial and temporal variability in the incoming total solar irradiance, in addition to modulation of the atmospheric transmission associated with cloud cover and aerosol distributions.

2.2. Field In Situ Observations in the Experiment of Pac\textsubscript{2017}

A total of 65 stations, including 25 optical stations, covering from the equator to 40°N, 143 to 149°E, were occupied between 14 April and 13 May 2017 during the 2017 cruise of the Spring Voyage of the Comprehensive Survey of Water Bodies in the Central and Southern Western Pacific by R/V Kexue (Pac\textsubscript{2017}; Figure 1; Table 1). At each station, vertical distributions of water temperature and salinity were recorded with a SeaBird SBE9/11 conductivity–temperature–depth (CTD) recorder. The mixed layer depth (MLD) was defined as the depth at which the temperature was 0.5 °C lower than the surface temperature [24]. Discrete water samples were collected with depth by using 20 L SBE bottles mounted onto a Rosette sampling assembly. Samples for the determination of the concentrations of Chl\textsubscript{a} were collected through 47 mm GF/F Whatman glass fiber filters by filtering ~2.5 L seawater onboard the ship. The concentrations of Chl\textsubscript{a} were determined by fluorimetry [25], though such a historically widely used technique may produce a relatively high uncertainty in determining Chl\textsubscript{a}, e.g., tending to underestimate Chl\textsubscript{a} when chlorophyll b is unusually distributed in the water. Only the surface concentrations of Chl\textsubscript{a}, whose samples were collected at 10 m depth, are reported here.

At each optical station conducted in the daytime, vertical profiling of the spectral downwelling irradiance [$E\textsubscript{d}(\lambda, Z)$] and upwelling radiance [$L\textsubscript{u}(\lambda, Z)$], along with simultaneous measurements of the sea surface solar irradiance [$E\textsubscript{s}(\lambda)$], between 400 and 700 nm in 2 nm intervals was determined through a Satlantic HyperPro II system by following the manufacturer’s manual (https://www.seabird.com/hyperpro-ii/product-downloads?id=54627923897 (accessed on 1 December 2021)). PAR(Z) (µE m\textsuperscript{−2} s\textsuperscript{−1}) was calculated as follows:

\[
\text{PAR}(Z) = 10^{-5} \frac{N\textsubscript{A}}{hc} \int_{400}^{700} \frac{\lambda}{hc} E\textsubscript{d}(\lambda, Z) d\lambda
\]

(4)

where \(\lambda\) (nm), Z (m), and $E\textsubscript{d}(\lambda, Z)$ (µW cm\textsuperscript{−2} nm\textsuperscript{−1}) represent light wavelength, water depth, and spectral downwelling irradiance at depth Z, while $h$, $c$, and $N\textsubscript{A}$ are the Planck’s constant ($6.625 \times 10^{-34}$ J s), the speed of light ($3 \times 10^8$ m s\textsuperscript{−1}), and the Avogadro constant.
(6.022 × 10^{23} \text{ mol}^{-1}), \text{ respectively}. \text{ The euphotic zonal depth (EZD) was defined as the depth at which PAR(Z) was 1\% of the surface value. According to Equation (1), } K_{Z_{\text{PAR}}} \text{ (in unit of m}^{-1}) \text{ was calculated as } K_{Z_{\text{PAR}}} = \frac{1}{Z} \ln \left[\frac{\text{PAR}(0)}{\text{PAR}(Z)}\right]. \text{ The above-water spectral remote sensing reflectance } [R_{\text{rs}}(\lambda)] \text{ was determined from } E_d(\lambda, Z), L_u(\lambda, Z), \text{ and } E_s(\lambda) \text{ using a data processor in compliance with the Ocean Optics Protocols established for in situ observations in support of ocean color remote sensing measurements [26].}

Table 1. Field experiments used for this study. Detailed sampling dates for experiments can be found in the Appendix A (Table A1).

<table>
<thead>
<tr>
<th>Experiments *</th>
<th>Years</th>
<th>Areas</th>
<th>Optical Stations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pac_2017</td>
<td>2017</td>
<td>143–149°E, 0–40°N</td>
<td>25</td>
</tr>
<tr>
<td>HOT</td>
<td>2009–2019</td>
<td>158°W, 22.75°N</td>
<td>63</td>
</tr>
<tr>
<td>CLIVAR</td>
<td>2006–2007</td>
<td>152–110°W, 0–54°N</td>
<td>20</td>
</tr>
<tr>
<td>JGOFS</td>
<td>1988, 1992</td>
<td>150–140°W, 0–15°N</td>
<td>27</td>
</tr>
<tr>
<td>JGOFS_WOCE</td>
<td>1991</td>
<td>154–151°W, 1–18°N</td>
<td>9</td>
</tr>
<tr>
<td>ZonalFlux</td>
<td>1996</td>
<td>165°E–150°W, 0–2.2°N</td>
<td>9</td>
</tr>
</tbody>
</table>

* Pac_2017: the 2017 cruise of the Spring Voyage of the Comprehensive Survey of Water Bodies in the Central and Southern Western Pacific; HOT: Hawaii Ocean Time-Series; ACE-Asia: Aerosol Characterization Experiment off the coast of Asia; CalCOFI: California Cooperative Oceanic Fisheries Investigation; CCE_LTER: Long-Term Ecological Research Network (LTER) California Current Ecosystem (CCE); CLIVAR: Climate Variability and Predictability; JGOFS: Joint Global Ocean Flux Study; JGOFS_WOCE: JGOFS World Ocean Circulation Experiment.

2.3. Additional Field Datasets

Additional field observations in the North Pacific were downloaded from the Hawaii Ocean Time-Series (HOT) website (https://hahana.soest.hawaii.edu/hot/ (accessed on 1 December 2021)), in which optical observations conducted at the ALOHA (A Long-Term Oligotrophic Habitat Assessment) station (22.75° N, 158.00° W) between 2009 and 2019 were extracted, and the SeaWiFS Biooptical Archive and Storage System (SeaBASS; https://seabass.gsfc.nasa.gov/ (accessed on 1 December 2021)) [27], in which optical observations conducted in multiple cruises were extracted (Figure 1; Table 1). The extracted data included PAR(Z), E_d(\lambda, Z), L_u(\lambda, Z), and E_s(\lambda). Similarly, K_{Z_{\text{PAR}}} \text{ and } R_{\text{rs}}(\lambda) \text{ were determined from these observations by following the methods as used in Section 2.2. To reduce the effect of the variation of the incident solar light on accurately determining } K_{Z_{\text{PAR}}} \text{ and } R_{\text{rs}}(\lambda), \text{ e.g., overpassing clouds might markedly reduce the incoming PAR}(0^°) \text{ and thus PAR}(Z), \text{ casts whose coefficients of variation in } E_u(\lambda) \text{ during sampling were } >1\% \text{ were excluded. Then, 745 sets of observations in } K_{Z_{\text{PAR}}} \text{ and } R_{\text{rs}}(\lambda) \text{ were available.}

The extracted data were pooled into our observations, and, finally, there were 770 sets of observations. About 80\% (617) of stations, which were randomly selected by using Excel® “Rand” function, were used for algorithm development, and the remaining 20\% (153) of stations were used to validate the algorithms.

2.4. Statistics

The performance of the algorithm of the remotely sensed K_{Z_{\text{PAR}}} \text{ and the penetration depth of PAR was evaluated by the root mean square difference (RMSD) such that}

\[
\text{RMSD} = \sqrt{\frac{\sum (D - F)^2}{n}}
\]

where n, F, and D are the number of samples, the observations, and the corresponding derived products, respectively. Unless noted, uncertainty used in this study means RMSD.
3. Results

3.1. Hydrographic Properties Observed In Situ

The in situ observations followed similar patterns to the remotely sensed data (Figure 1), as exampled in the latitudinal variations of SST, surface salinity, MLD, EZD, and surface Chl_a along 143–149°E observed during our cruise (Pac_2017) conducted in April–May 2017 (Figure 2). According to reports in the literature [28–33], this area might be subdivided into three primary hydrographic regimes, the Kuroshio Extension (north of 34°N), the NPSG (15–25°N) and the western Tropical Pacific (south of 10°N), and two transition zones (25–34°N and 10–15°N). The SST, 14.2 ± 3.0, 27.0 ± 1.5, and 29.9 ± 0.4 °C in the Kuroshio Extension, the NPSG, and the western Tropical Pacific, respectively, increased southwards. Surface salinity generally increased from 34.6 ± 0.2 in the Kuroshio Extension to 34.9 ± 0.3 in the NPSG and then decreased to 34.0 ± 0.2 in the western Tropical Pacific, consistent with previous reports [34,35]. The MLD, 138 ± 72 m, was deep and was much deeper than the EZD, 53 ± 21 m, in the Kuroshio Extension, suggesting a replete supply of nutrients by vertical mixing and thus resulting in a large surface Chl_a, 1.12 ± 0.30 mg m⁻³. In the NPSG, as reported in the literature [36], the MLD, 29 ± 10 m, became very shallow and was much shallower than the EZD, 115 ± 17 m. Because the permanent anticyclonic circulation leads to an Ekman downwelling, a deep pycnocline, and a limited supply of nutrients by vertical mixing [36,37], the surface Chl_a, 0.07 ± 0.02 mg m⁻³, was very low in the NPSG. In the western Tropical Pacific, the MLD, 76 ± 20 m, was shallower than the EZD, 103 ± 16 m, and the surface Chl_a, 0.07 ± 0.02 mg m⁻³, was also low. Nevertheless, the MLD, the EZD, and the surface Chl_a were all significantly (p < 0.01) correlated with the SST (r = −0.53, 0.88, and −0.84), suggesting that the temperature-associated status of water stratification is important, if not dominant, in controlling the distributions of the physical, hydrographic, and biological qualities in the North Pacific.

![Figure 2](image-url)

Figure 2. Latitudinal variations in (a) SST (●) and surface salinity (○), (b) mixed layer depth (MLD, ▲) and euphotic zonal depth (EZD, △), and (c) surface Chl_a (■) and calculated remotely sensed diffuse attenuate coefficient of PAR ($K_{PAR}^{RS}$, □) along 143–149°E observed in situ during the Pac_2017 cruise.

3.2. Variations of $K_{PAR}^{RS}$ Observed In Situ

Horizontal variations of $K_{PAR}^{RS}$, as exampled in the latitudinal variations of $K_{PAR}^{RS}$ calculated from the in situ PAR observations in the first optical depth during the Pac_2017 cruise in April–May 2017, are shown in Figure 2c. Ranging between 0.05 and 0.18 m⁻¹, $K_{PAR}^{RS}$ generally decreased southward, following a similar trend to the surface Chl_a. In
actuality, a positive correlation \( r = 0.77 \) was found between \( K_{P_{RS}}^{Z} \) and the surface Chl-a, suggesting that this study area is well classed into the Case-1 waters in which the optical properties are determined primarily by the phytoplankton and their associated living and inanimate materials.

Depth-dependence in \( K_{P_{RS}}^{Z} \) was pronounced in all water conditions. Vertical variations of \( \frac{P_{Z}}{P_{0}} \) and \( K_{P_{RS}}^{Z} \), as exampled in a well-mixed station \((149^\circ E, 38^\circ N)\) and a stratified station in the NPSG \((143^\circ E, 22^\circ N)\) during the Pac_2017 cruise, are shown in Figure 3. More examples can be found in the Appendix A (Figure A2). At both stations, although \( \frac{P_{Z}}{P_{0}} \) followed a generally exponential decay function with depth, it was reduced faster in the upper layer than in the lower layer, resulting in a generally decreasing \( K_{P_{RS}}^{Z} \) with depth, consistent with previous reports \([10,12,13,17]\). In the well-mixed station, \( K_{P_{RS}}^{Z} \), ranging between 0.115 and 0.094 m\(^{-1}\), varied by a factor of 1.2 (Figure 3a). In the stratified station, \( K_{P_{RS}}^{Z} \), ranging between 0.056 and 0.038 m\(^{-1}\), varied by a factor of 1.5 (Figure 3b). Without considering the vertical variations in \( K_{P_{RS}}^{Z} \), i.e., by treating \( K_{P_{RS}}^{Z} \) vertically constant to \( K_{P_{RS}}^{Z} \), the penetration depths of PAR would be markedly underestimated. For example, the estimated EZDs by using constant \( K_{P_{RS}}^{Z} \) equal to \( K_{P_{RS}}^{Z} \) were 42 and 92 m, which were shallower than the observations of 49 and 117 m by 14% and 21% in these two stations (Figure 3). These results indicated that knowing \( K_{P_{RS}}^{Z} \) alone is insufficient to retrieve the submarine penetration of PAR. In order to accurately estimate the penetration depths of PAR at different light levels from space, the relationships between \( K_{P_{RS}}^{Z} \) and \( K_{P_{RS}}^{Z} \) have to be built.

![Figure 3](image-url)

**Figure 3.** Vertical variations of \( \frac{P_{Z}}{P_{0}} \) (black solid lines) and \( K_{P_{RS}}^{Z} \) (pink dash–dotted lines) in (a) a well-mixed station \((149^\circ E, 38^\circ N)\) and (b) a stratified station \((143^\circ E, 22^\circ N)\) during the Pac_2017 cruise. The estimated \( \frac{P_{Z}}{P_{0}} \) by treating \( K_{P_{RS}}^{Z} \) vertically constant to \( K_{P_{RS}}^{Z} \) are shown using the red dashed lines.
3.3. Algorithm Development for Remotely Sensing $K_{\text{PAR}}^Z$ in the North Pacific

In this study, we adapted a two-step method for remotely estimating $K_{\text{PAR}}^Z$: developing an ocean color algorithm for $K_{\text{PAR}}^{RS}$, and building the relationships between $K_{\text{PAR}}^Z$ and $K_{\text{PAR}}^{RS}$. For the first step, $K_{\text{PAR}}^{RS}$ could be related to $R_{\text{rs}}$ band ratio such that (Figure 4)

$$
\log \left( K_{\text{PAR}}^{RS} \right) = (-0.697 \pm 0.006) - (0.951 \pm 0.013)X \quad n = 617, \quad r^2 = 0.896 \quad (6)
$$

![Figure 4. Relationship between log($K_{\text{PAR}}^{RS}$) and log[$R_{\text{rs}}(488)/R_{\text{rs}}(555)$] for data used for algorithm development. $K_{\text{PAR}}^{RS}$ is given in unit of m$^{-1}$. Black solid line—best fit line of Equation (6); dashed lines—one RMSD ($\pm 0.065$) from the best fit line; red solid line—relationship by using the combined Equations (2) and (3).](image)

Again, $X = \log [R_{\text{rs}}(488)/R_{\text{rs}}(555)]$. The RMSD of the derived log($K_{\text{PAR}}^{RS}$) was $\pm 0.065$ (in unit of m$^{-1}$ for $K_{\text{PAR}}^{RS}$), which corresponded to a deviation of about $\pm 16\%$ in the derived $K_{\text{PAR}}^{RS}$. This uncertainty was reduced to $2/3$ of that by applying the commonly used approach, the combined Equations of (2) and (3), which introduced the RMSD of $\pm 0.103$ in the derived log($K_{\text{PAR}}^{RS}$) and the corresponding deviation of about $\pm 27\%$ in $K_{\text{PAR}}^{RS}$. Moreover, the use of the combined Equations (2) and (3) generally introduced an overestimation on $K_{\text{PAR}}^{RS}$.

For the second step, our data indicated that for a given light level ($f$) within the euphotic zone whose corresponding depth is $Z_f$ so that $f = \text{PAR}(Z_f)/\text{PAR}(0^-) \times 100\%$, the average $K_{\text{PAR}}$ from the surface down to $Z_f$, $K_{\text{PAR}}^{Z_f}$, could be linearly related to $K_{\text{PAR}}^{RS}$ ($r^2 > 0.96$; Figure 5), while the derived slope ($A$) was related to log($f$) by a third-order polynomial function such that ($r^2 = 0.96$; Figure 6)

$$
K_{\text{PAR}}^{Z_f} = A \times K_{\text{PAR}}^{RS} \quad (7)
$$

$$
A = (1.250 \pm 0.010) + (0.752 \pm 0.041)\log(f) + (0.510 \pm 0.044)\log(f)^2 + (0.121 \pm 0.013)\log(f)^3 \quad (8)
$$
Zf could thus be determined such that
\[ Z_f = -\log(f) \]

Figure 5. Linear regressions between \( K_{RS}^{Z_f} \) and \( K_{Z_f}^{PAR} \) for selected light levels of \( f = 50\% (\circ, \text{solid black line}), 10\% (\triangle, \text{dashed red line}), \) and \( 1\% (\lozenge, \text{dot–dashed blue line}) \) for data used for algorithm development.

By combining Equations (6) and (7), given the light level of \( f \), \( K_{Z_f}^{PAR} \) could be determined from the Rrs band ratio, and the corresponding \( Z_f \) could then be further determined from Equation (10). The RMSD of \( \log(K_{Z_f}^{PAR}) \) derived from this approach was around ±0.059 (\( K_{Z_f}^{PAR} \) in unit of m\(^{-1}\)), which corresponded to a deviation of ±15\% in both the derived \( K_{Z_f}^{PAR} \) (Figure 7a) and \( Z_f \) (Figure 7b). The above results suggest that, given substantially accurate Rrs spectra by satellites, the remote estimates in \( K_{Z_f}^{PAR} \) and \( Z_f \) at any light levels within the euphotic zone could be accurately estimated from space with a deviation of ±15\% only.

Figure 6. Relationship between the slope of Equation (7) (\( A, \circ \)) and light level (\( f \)). Symbols (\( \circ \)) and error bars represent the regressed \( A \) values and the standard deviations.

The RMSD of \( \log(K_{Z_f}^{PAR}) \) derived from Equation (7) was around ±0.041 (ranging between ±0.015 and ±0.055 for \( f \) varying from 1\% to 70\%; \( K_{Z_f}^{PAR} \) in unit of m\(^{-1}\)), resulting in the corresponding deviation of the derived \( K_{Z_f}^{PAR} \) to be around ±10\%. By definition,

\[
f = \frac{\text{PAR}(Z_i)}{\text{PAR}(0^-)} = \frac{\text{PAR}(0^-)\exp \left( -\frac{Z_i}{\text{PAR}(0^-)} \right)}{\text{PAR}(0^-)}
\]

(9)
$Z_f$ could thus be determined such that

$$Z_f = \frac{\log(f)}{K_{\text{PAR}}^Z}$$

(10)

By combining Equations (6) and (7), given the light level of $f$, $K_{\text{PAR}}^Z$ could be determined from the $R_m$ band ratio, and the corresponding $Z_f$ could then be further determined from Equation (10). The RMSD of $\log(K_{\text{PAR}}^Z)$ derived from this approach was around $\pm 0.059$ ($K_{\text{PAR}}^Z$ in unit of m$^{-1}$), which corresponded to a deviation of $\pm 15\%$ in both the derived $K_{\text{PAR}}^Z$ (Figure 7a) and $Z_f$ (Figure 7b). The above results suggest that, given substantially accurate $R_m$ spectra by satellites, the remote estimates in $K_{\text{PAR}}^Z$ and $Z_f$ at any light levels within the euphotic zone could be accurately estimated from space with a deviation of $\pm 15\%$ only.

**Figure 7.** Comparisons of the derived (a) $K_{\text{PAR}}^Z$ (m$^{-1}$) and (b) the associated penetration depths ($Z_f$, m) by using our approach with the field in situ observations for selected light levels of $f = 50\%$ (○), 10\% (△), and 1\% (●) for data used for algorithm development. Solid lines—1:1 plots; dashed lines—$\pm 15\%$ deviated from the 1:1 plots.

### 3.4. Validation on the Derived $K_{\text{PAR}}^Z$

The performance of the proposed approach was further validated by comparing $K_{\text{PAR}}^Z$ and $Z_f$ derived from $R_m(\lambda)$ to the observations in the dataset used for validation. The resulting RMSD in $\log(K_{\text{PAR}}^Z)$ for light levels within the euphotic zone was around $\pm 0.057$ ($K_{\text{PAR}}^Z$ in unit of m$^{-1}$), which corresponded to a deviation of $\pm 14\%$ in both the derived $K_{\text{PAR}}^Z$ (Figure 8a) and $Z_f$ (Figure 8b). These values were even slightly smaller than those found in the algorithm development, suggesting that there was a general good agreement on the remotely sensed and the observed $K_{\text{PAR}}^Z$ and $Z_f$.

Comparisons of the MODIS-Aqua-derived $K_{\text{PAR}}^Z$ and $Z_f$ to the field in situ observations, by following the protocols [38], are shown in Figure 9. For each station, the main criteria included minimum 5 valid pixels in the $3 \times 3$ pixel array centered on the field station, coefficient of variation of valid satellite pixels lower than 0.15, and time difference within $\pm 3$ h between satellite overpass and field observation. The resulting RMSD in $\log(K_{\text{PAR}}^Z)$ for light levels within the euphotic zone was around $\pm 0.073$ ($K_{\text{PAR}}^Z$ in unit of m$^{-1}$).
m\textsuperscript{-1}), which corresponded to a deviation of ±18% in both the derived $K_{\text{PAR}}^{Z_1}$ (Figure 9a) and $Z_1$ (Figure 9b). These values were slightly larger but still quite comparable to those found in Figures 7 and 8, suggesting that there was a reasonable agreement between the remotely sensed and the observed $K_{\text{PAR}}^{Z_1}$ and $Z_1$. Nevertheless, since only 13 match-up data points were available for this evaluation, further validation when additional field observations became available would be desirable.

Figure 8. The same as Figure 7 but for in situ data used for validation. (a) $K_{\text{PAR}}^{Z_1}$ (m\textsuperscript{-1}) and (b) the associated penetration depths ($Z_1$, m).

Figure 9. The same as Figure 7 but the derivations were made from the MODIS-Aqua match-up data with a satellite overpass time window of ±3 h. Here, dashed lines represent ±18% deviated from the 1:1 plots (solid lines). (a) $K_{\text{PAR}}^{Z_1}$ (m\textsuperscript{-1}) and (b) the associated penetration depths ($Z_1$, m).

4. Application: Distributions of Remotely Sensed $K_{\text{PAR}}^{Z_1}$ and $Z_1$ in the North Pacific

The operational approach developed in this study can be applied to estimate the vertical variations of $K_{\text{PAR}}$ from space. Using MODIS-Aqua monthly Level-3 products
(Reprocessing R2022.0) of R<sub>n</sub>(488) and R<sub>n</sub>(555) extracted from the NASA Ocean Color Web (http://oceancolor.gsfc.nasa.gov (accessed on 1 December 2022)), the climatologically (2002–2022) seasonal distributions in spring in the satellite-derived K<sub>RS</sub><sub>PAR</sub>, K<sub>Z1%</sub><sub>PAR</sub>, and EZD in the North Pacific are shown as examples in Figure 10. K<sub>RS</sub><sub>PAR</sub> is generally >0.1 m<sup>-1</sup> north of 40°N, typically <0.05 m<sup>-1</sup> in the NPSG, and around 0.06 m<sup>-1</sup> south of the NPSG (Figure 10a). Such a distributional pattern generally follows the pattern of the latitudinal variations in SST (Figure 1a) such that high K<sub>RS</sub><sub>PAR</sub> is associated with low SST and vice versa. Enhanced nutrient supply and, hence, primary production by vertical mixing to the surface layer in low SST (Figure 2) may account for high K<sub>RS</sub><sub>PAR</sub> in low-SST conditions. High values of K<sub>RS</sub><sub>PAR</sub> can also be found in the coastal waters, in which land-sourced inputs and/or southward currents are marked, and in the cyclonic circulation gyres, e.g., region surrounded by the Oyashio, the North Pacific Current, and the Alaska Current at about 160°E–135°W, 40–60°N [32]. The upwelling may enhance vertical transport of nutrients to the surface layer to support phytoplankton growth, resulting in the increased turbidity and, thus, high K<sub>RS</sub><sub>PAR</sub>. In contrast, the NPSG is characterized as oligotrophic waters with very low K<sub>RS</sub><sub>PAR</sub> because of the occurrence of a permanent anticyclonic circulation gyre. K<sub>RS</sub><sub>PAR</sub> is slightly enlarged south of the NPSG, possibly due to the slightly enhanced vertical mixing by the blowing of the trade winds [32]. K<sub>Z1%</sub><sub>PAR</sub>, whose value is about 81% of K<sub>RS</sub><sub>PAR</sub>, follows a similar distributional pattern to K<sub>RS</sub><sub>PAR</sub> (Figure 10b). The distributional pattern in the EZD follows a reverse trend to K<sub>Z1%</sub><sub>PAR</sub> (Figure 10c) according to Equation (9). As such, the EZD may vary from <50 m north of 40°N to >110 m in the NPSG.

![Figure 10](image-url). Climatologically (2002–2022) seasonal distributions in spring in the MODIS-Aqua-derived (a) K<sub>RS</sub><sub>PAR</sub>, (b) K<sub>Z1%</sub><sub>PAR</sub>, and (c) EZD in the North Pacific.
The application of the remotely sensed $K_{Z_{PAR}}$ and the associated vertical distributions of the PAR may improve the accuracy in the assessment of the carbon fixation in the North Pacific. For example, the remotely sensed net primary production in the euphotic zone (PP$_{eu}$) is thought to be linearly related to the euphotic zonal depth (EZD) [5]. The EZD is typically underestimated, e.g., by 14 ± 16% using the models integrated in the PP$_{eu}$ model [5] or by 16 ± 11% estimated by treating $K_{Z_{PAR}}$ vertically constant to $K_{RS_{PAR}}$ (Figures 3 and A2). As a result, PP$_{eu}$ may be substantially underestimated in the North Pacific. A reassessment on the contribution of the North Pacific to the global carbon cycle may be needed.

Although an improved approach to remotely estimate $K_{Z_{PAR}}$ in the North Pacific has been proposed in this study, it is necessary to note that it is designed to be applied to the Case-1 waters, given that most of the in situ observations were conducted in the offshore regions. For inshore locations and marginal seas at land–sea boundaries, which are typically characterized as the Case-2 waters, regionally tuned algorithms are generally needed.

5. Conclusions

This study demonstrated that, by adapting a two-step approach, that is, developing an ocean color algorithm for $K_{RS_{PAR}}$ and building the relationships between $K_{Z_{PAR}}$ and $K_{RS_{PAR}}$, the vertical variations in $K_{Z_{PAR}}$ in the North Pacific may be assessed from space. The deviation in the remotely sensed $K_{Z_{PAR}}$ in depths within the euphotic zone, as well as the corresponding penetration depths of PAR, is about ±15%. The significant underestimation in the remotely sensed euphotic zonal depth from methods adapted in current primary production models suggests a substantial underestimation of the primary production in the North Pacific. The variations in $K_{Z_{PAR}}$, together with those in other hydrographic properties (e.g., SST and Chl$_a$), suggest that our dataset covers various water conditions and, thus, may be a good representative to the Case-1 waters in the North Pacific Ocean. A similar approach may be developed for the global oceans when field observations in the global oceans become available.

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Data Availability Statement: Data from this research will be available upon request to the authors.

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Conflicts of Interest: The authors declare no conflict of interest.
Appendix A

Table A1. Sampling dates for the field experiments used for this study.

<table>
<thead>
<tr>
<th>Experiments</th>
<th>Periods</th>
</tr>
</thead>
<tbody>
<tr>
<td>JGOFS</td>
<td>16–28 Sep 1991</td>
</tr>
<tr>
<td>OCEAN_LIDAR</td>
<td>24 Apr–10 May 1996</td>
</tr>
</tbody>
</table>

Figure A1. Climatologically (2002–2022) seasonal distributions in spring in the MODIS-Aqua-derived PAR.
Figure A2. Latitudinal variations in the vertical variations of PAR(Z)/PAR(0−) (black solid lines) and \( K_{\text{PAR}} \) (pink dash–dotted lines) along ~152°W for selected stations conducted in the CLIVAE experiment during 21 February to 27 March 2006. The estimated PAR(Z)/PAR(0−) by treating \( K_{\text{PAR}} \) vertically constant to \( K_{\text{PAR}}^{\text{RS}} \) are shown in the red dashed lines. Latitudes for the stations are (a) 4°N; (b) 14°N; (c) 23°N; (d) 34°N; (e) 45°N; (f) 54°N.
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