Lecture Notes on Coastal and Estuarine Studies

Managing Editors:
Richard T. Barber Christopher N.K. Mooers
Malcolm J. Bowman Bernt Zeitzschel

4

Howard R. Gordon
André Y. Morel

Remote Assessment of Ocean Color for Interpretation of Satellite Visible Imagery

A Review

Springer-Verlag
New York Berlin Heidelberg Tokyo 1983

Authors

Howard R. Gordon
Department of Physics
University of Miami
Coral Gables, FL 33124, USA

André Y. Morel
Laboratoire de Physique et Chimie
Université Pierre et Marie Curie
06230 Villefranche-sur-Mer FR

Library of Congress Cataloging in Publication Data
Gordon, Howard R.
Remote assessment of ocean color for interpretation of satellite visible imagery.
Bibliography: p
1. Optical oceanography—Remote sensing. 2. Colors—Analysis. I. Morel, André Y. II. Title.
GC180.G67 1983 551.46′01 83-16984

9 8 7 6 5 4 3 2 1

ISBN-13: 978-0-387-90923-3 e-ISBN-13: 978-1-4684-6280-7
DOI: 10.1007/978-1-4684-6280-7

CONTENTS

I. INTRODUCTION 1

II. THE PHYSICS OF OCEAN COLOR REMOTE SENSING .. 3

A. Irradiance Ratio and Upwelling Subsurface Radiance....................... 6
B. Atmospheric Effects 19

III. IN - WATER ALGORITHMS 24

A. The Phytoplankton Pigment Algorithms 26
B. The 'K' Algorithms 36
C. The Seston Algorithms 41
D. The Analytic Algorithm 42
E. Relationship Between the Algorithms 47
1. Relationships Between the Radiance Ratios r_{ij} 48
2. Seston-Pigment Relationships 49
3. Pigment-'K' Relationships 55
4. Preliminary Conclusions 62

IV. ATMOSPHERIC CORRECTION 68

V. APPLICATION OF THE ALGORITHMS TO CZCS IMAGERY 72

VI. SUMMARY AND CONCLUSIONS 80

VII. APPENDIX I: THE COASTAL ZONE COLOR SCANNER (CZCS) 84

VIII. APPENDIX II: RECENT DEVELOPMENTS 87

A. Clear Water Radiance Concept 87
B. Accuracy of Pigment Estimates 91
C. Applications 96

IX. REFERENCES 103

Acknowledgments

This work is a contribution to the research encouraged by the IAPSO Working Group on Optical Oceanography, in particular its third term of reference ('Examination of ocean optical properties with their application to other aspects of oceanography, including physical oceanography, ocean dynamics, heat absorption and climatology, marine biology, and sedimentology.'). Due to the important amount of oceanographic data used in this study which have been acquired over a long period, all of the support received cannot be acknowledged. By restricting ourselves to the latest, we wish to acknowledge the Centre National de la Recherche Scientifique (under contracts ERA 278 and GRECO 034), the Centre National d'Etudes Spatiales (under contracts 80-81/250), the Centre National d'Exploitation des Oceans (under contracts 79/2084, 79/2010, and 81/203), the National Aeronautics and Space Administration (under contract NAS 5-22963 and grant NAGW-273), and the National Atmospheric and Oceanic Administration (under contract NA-79-SAC-00741).

We also wish to acknowledge the support and effort of the following individuals: L. Prieur for providing the unpublished data used in Figure 5b; D.K. Clark for providing Figure 13; J.W. Brown, O.B. Brown, and R.H. Evans for processing the imagery presented in Appendix II; and D.K. Clark and D.L. Ball for additional processing of Figures AII-3 through AII-5.

I. INTRODUCTION

Since the pioneering work of Clarke et al. (1970) it has been known that chlorophyll a (or, more generally, pigments) contained in phytoplankton in near-surface waters produced systematic variations in the color of the ocean which could be observed from aircraft. As a direct result of this work, NASA developed the Coastal Zone Color Scanner (CZCS), which was launched on Nimbus-G (now Nimbus-7) in October 1978. (A short description of the CZCS is provided in Appendix I.)

Shortly before launch, at the IUCRM Colloquium on Passive Radiometry of the Ocean (June 1978), a working group on water color measurements was formed to assess water color remote sensing at that time. A report (Morel and Gordon, 1980) was prepared which summarized the state-of-the-art of the algorithms for atmospheric correction, and phytoplankton pigment and seston retrieval, and which included recommendations concerning the design of next generation sensors.

The water color session of the COSPAR/SCOR/IUCRM Symposium 'Oceanography from Space' held in Venice (May 1980, i.e., in the post-launch period) provided the opportunity for a reassessment of the state-of-the-art after having gained some experience in the analysis of the initial CZCS imagery. Such an assessment is the purpose of this review paper, which will begin with an outline of the basic physics of water color remote sensing and the fundamentals of atmospheric corrections. The present state of the constituent retrieval and atmospheric correction algorithms will then be critically assessed, considering primarily the papers presented at the Symposium, and also some new material. Samples of imagery are presented and the initial comparisons with experiments are discussed. Finally, an appendix (Appendix II) is provided which summarizes the major developments in water color remote sensing which have taken place during the preparation of the report, and prior to the IAMAP Assembly in Hamburg (August 1981).

II. PHYSICS OF OCEAN COLOR REMOTE SENSING

A typical water color remote sensing situation is depicted in Fig. 1, which shows the satellite's radiometer aiming toward a spot on the sea surface (marked PIXEL), hence measuring the radiance* $L_T(\theta,\phi)$ emerging from the Earth's atmosphere. The goal of this measurement is to determine the concentration of the various constituents of the water (e.g., phytoplankton, total suspended material, yellow substances, etc.). The radiance $L_T(\theta,\phi)$ consists of photons which have been multiply scattered from the atmosphere, ocean, and sea surface, and thus depends in a complex manner on the optical properties of the atmosphere (and their distribution with altitude) and

---

* A radiometer consists of a flat detector of surface area A, capable of measuring the radiant power Φ falling on it in a spectral band $\Delta\lambda$ centered on the wavelength λ. When the detector is located at a position specified by the vector $\underset{\sim}{r}$, is aimed in a direction specified by the unit vector $-\underset{\sim}{\xi}$, and is viewing the universe with a solid angle of acceptance $\Delta\Omega$, it measures the radiance traveling in the direction $\underset{\sim}{\xi}$ defined by

$$L(\underset{\sim}{r},\underset{\sim}{\xi},\lambda) \equiv \Phi \;/\; A\ \Delta\lambda\ \Delta\Omega.$$

In the text the direction of $\underset{\sim}{\xi}$ is specified by the angles θ and ϕ. When λ is omitted from the argument of L it is to be understood that the statement or relationship is true for all wavelengths.

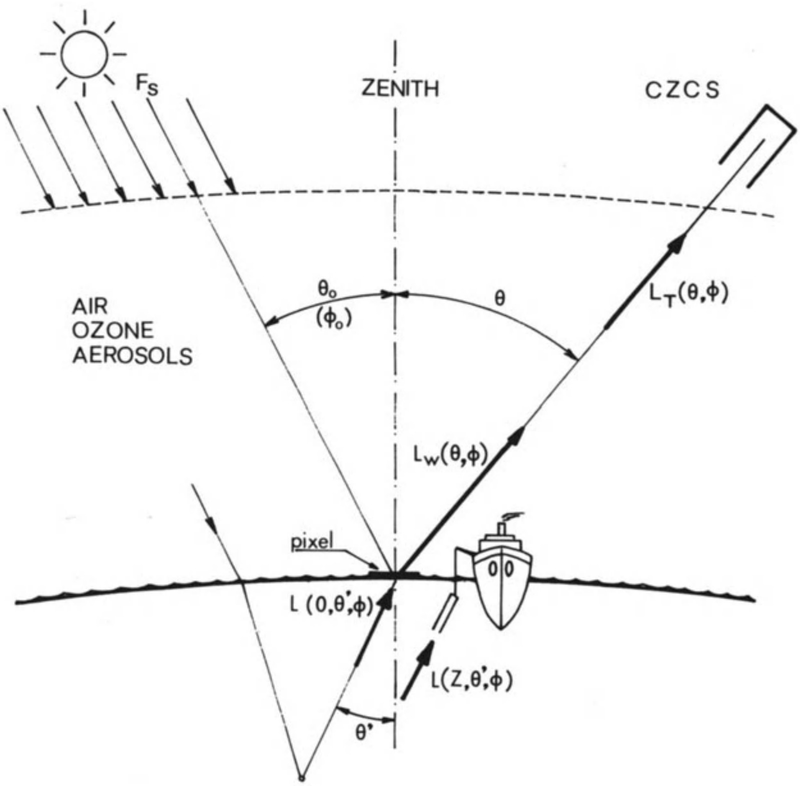

Figure 1. Schematic geometry and nomenclature for upwelling radiances.

on the optical properties of the ocean (and their distribution with depth). Through a series of Monte Carlo simulations (Gordon, 1976) of radiative transfer in an ocean-atmosphere system characterized by realistic and vertical distributions of all pertinent optical properties in the atmosphere, and a representative set of optical properties for the ocean (assumed homogeneous), three simplifying facts emerge. First, the vertical structure of the atmosphere has only minor influence on $L_T(\theta,\phi)$. Next, the ocean and atmosphere may be decoupled for the purpose of computing $L_T(\theta,\phi)$. Finally, the effect of ocean constituents on $L_T(\theta,\phi)$ can be adequately simulated by placing a hypothetical lambertian reflector of albedo R just beneath the sea surface (i.e., at depth z=0). The required subsurface albedo R is the irradiance ratio or irradiance reflectance defined by

$$R \equiv E_u(0)/E_d(0),$$

where the upward and downward irradiances, E_u and E_d respectively, are defined by

$$E_u(0) = \int_0^{2\pi} \int_0^{\pi/2} |\cos\theta'| L(0,\theta',\phi') \sin\theta' d\theta' d\phi',$$

$$E_d(0) = \int_0^{2\pi} \int_{\pi/2}^{\pi} |\cos\theta'| L(0,\theta',\phi') \sin\theta' d\theta' d\phi',$$

and $L(0,\theta',\phi')$ is the radiance a submerged observer would measure (in the ocean to be simulated) with a radiometer aimed at an angle θ' from the nadir and an azimuth angle ϕ' relative to the sun (see Fig. 1). Mathematically,

$$(1) \qquad L_T(\theta,\phi) = L_1(\theta,\phi) + R\, L_2(\theta,\phi)/(1 - rR),$$

in which $L_1(\theta,\phi)$ is the contribution to $L_T(\theta,\phi)$ from all photons which never penetrated the ocean (but may have reflected from the surface), $L_2(\theta,\phi)$ is the radiance which results from photons which penetrated the surface and interacted once with the lambertian surface for the case R=1, and r is a constant equal to about 0.5 (see, e.g., Austin, 1974 and Gordon, 1976). The factor 1/(1−rR) in the second term accounts, to all orders, for photons which are internally reflected in the ocean; since R very rarely exceeds 0.1, this factor is of minor importance.

The interpretation of $L_T(\theta,\phi)$ in terms of ocean constituents thus requires understanding the relationship between R and the constituents, while the retrieval of the constituents from $L_T(\theta,\phi)$ requires an accurate estimate of $L_1(\theta,\phi)$, which typically accounts for more than 90% of $L_T(\theta,\phi)$.

A. Irradiance Ratio and Upwelling Subsurface Radiance

The physics of 'ocean color' is well understood. The ocean color is physically described by the spectral values of reflectance $R(\lambda)$. The first theoretical approaches were made by using the two-flow (Schuster's) method, as an approximate solution of the radiative transfer problem (Gamburtsev, 1924, cited in Kozlyaninov, 1972; Duntley, 1942; Joseph, 1950; and Kozlyaninov and Pelevin, 1965). R was found to be related to the absorption coefficient, a, and backscattering coefficient, b_b, through (Joseph, 1950)

$$R = \frac{(1 + 2b_b/a)^{1/2} - 1}{(1 + 2b_b/a)^{1/2} + 1},$$

where b_b is linked to the volume scattering function $\beta(\gamma)$ according to

$$b_b = 2\pi \int_{\pi/2}^{\pi} \beta(\gamma) \sin\gamma \, d\gamma,$$

and to the scattering coefficient, b, through $b_b = \bar{b}_b b$, where $\bar{b}_b$ is the dimensionless backscattering ratio, and b is given by

$$b = 2\pi \int_{0}^{\pi} \beta(\gamma) \sin\gamma \, d\gamma.$$

Symbols and definitions used here follow the terminology adopted by IAPSO (International Association for Physical Sciences of the Ocean, see Morel and Smith, 1982). The Gamburtsev-Duntley expression for R is

$$R = \frac{b_b^*/a^*}{1 + (b_b^*/a^*) + (1 + 2b_b^*/a^*)^{1/2}},$$

where a^* and b_b^* are the absorption and backscattering coefficients for the diffuse light stream, called 'hybrid' optical properties by Preisendorfer (1961). They are related to the 'true' or 'inherent' properties of the ocean by

$$a^* = a/\bar{\mu}_d \quad \text{and} \quad b_b^* \cong b_b/\bar{\mu}_d.$$

$\bar{\mu}_d$ is the average cosine for the upper hemisphere which characterizes the downward radiance distribution and is obtained by forming the ratio

$$\bar{\mu}_d = E_d/\mathring{E}_d,$$

where $\mathring{E}_d$ is the downward scalar irradiance

$$\mathring{E}_d = \int_0^{2\pi} \int_{\pi/2}^{\pi} L(z,\theta',\phi')\sin\theta' d\theta' d\phi'.$$

Note that the average cosine for the entire radiance field, $\bar{\mu}$, which will be used later (III.B), is defined as $\bar{\mu} = E/\mathring{E}$, where E is the magnitude of the (downward) vector irradiance defined to be $(E_d - E_u)\underset{\sim}{z}$, with $\underset{\sim}{z}$ a unit vector in the nadir direction, and $\mathring{E}$ is the scalar irradiance

$$\mathring{E} = \mathring{E}_d + \mathring{E}_u = \int_0^{2\pi} \int_0^{\pi} L(z,\theta',\phi')\sin\theta' d\theta' d\phi'.$$

In the limit $b_b << a$, which is valid for nearly all oceanic and coastal waters and for the spectral domain considered, and also with the realistic approximation $b_b/a = b_b^*/a^*$, the above formulae simply reduce to

$$R \cong \frac{1}{2} b_b/a.$$

The main value of this formulation is the demonstration that, at least to first order, R is governed by the ratio of backscattering to absorption.

Full computations of radiative transfer in the ocean have now been carried out for various inherent properties of the medium and various illumination conditions. Gordon et al. (1975) used a Monte Carlo technique and their computations of R were fitted to polynomial expansion

$$R = \sum_{1}^{3} r_n X^n \tag{2}$$

with

$$X = \frac{b_b/a}{1 + b_b/a}.$$

In this expansion the first term is predominant, with r_1 having a value of 0.32 for solar illumination with the sun near the zenith and 0.37 for totally diffuse illumination. Note that unlike a^* and b_b^*, a and b_b are rigorously summable over the constituents of the ocean, i.e.,

$$a = a_w + \sum (a)_i \tag{3}$$

and

$$b_b = (b_b)_w + \sum (b_b)_i, \tag{4}$$

where w refers to water, and i to the i-th constituent. The ocean constituents influence R through their effect on (a) and (b_b). In general (Morel and Prieur, 1977),

$$(a)_i = (a)_i^o C_i$$

(5)

$$(b_b)_i = (b_b)_i^o C_i,$$

where $(a)_i^o$ and and $(b_b)_i^o$ are the absorption and back scattering coefficients per unit concentration of the i-th constituent (i.e., the specific absorption and backscattering coefficients for the i-th constituent).

By the use of the successive scattering order method (Prieur and Morel, 1975; Prieur, 1976) an alternative expression was obtained,

$$(6) \qquad R = 0.33\ (b_b/a)\ (1 + \Delta).$$

The second order term Δ was studied as a function of phase function and of the radiance distribution within the submarine light field. In any case, Δ remains weak (typically less than ±5%) and for practical purpose may be neglected. The usefulness of this very simple relationship has been demonstrated in many papers (Morel and Prieur, 1977; Smith and Baker, 1978b), and recently confirmed by Kirk (1981). At least for remote sensing studies, the use of more complicated relationships or attempts to refine them through new calculations of radiation transfer are not needed until a better knowledge of the absorbing and scattering properties of the materials present in the sea is reached.

All of the computations described above were carried out for a homogeneous ocean (i.e., a and $\beta(\gamma)$ independent of depth), whereas in practice the ocean is usually highly stratified. Based on the Monte Carlo simulations of radiative transfer on strongly stratified media, Gordon and Clark (1980a) have concluded that R is still approximately given by Eq. 2 if X is replaced by

$$\bar{X} = \int_0^{\tau_{90}} g(\tau)X(\tau)d\tau \Big/ \int_0^{\tau_{90}} g(\tau)d\tau,$$

with

$$g(\tau) = \exp\left[-2\int_0^{\tau} K(\tau')d\tau'/c(\tau')\right],$$

where K_d is the attenuation coefficient for downwelling irradiance, defined as

$$K_d(z,\lambda) = d(Ln[E_d(z,\lambda)])/dz,$$

c is the beam attenuation coefficient (with $c = a + b$), and τ is the optical depth which is related to the true depth z through

$$\tau = \int_o^z c(z)\, dz.$$

The upper limit on the integrals above is the optical depth at which the downwelling irradiance falls to 1/e of its value just beneath the surface. For a homogeneous ocean, the depth z_{90} corresponding to τ_{90} is simply $1/K_d$ and is called the 'penetration depth.' Only 10% of the contributions to R results from photons which have reached depths greater than z_{90} (Gordon and McCluney, 1975).

When the optical properties of the ocean are largely governed by the concentration (C) of a single constituent, then Gordon and Clark (1980a) have shown that the 'effective' concentration (the concentration which would yield the same R in a homogeneous ocean) is

$$(7) \qquad C_f = \int_o^{z_{90}} C(z) f(z,\lambda) dz \Big/ \int_o^{z_{90}} f(z,\lambda) dz,$$

where

$$f(z,\lambda) = \exp[-2 \int_o^z K_d(z',\lambda) dz'].$$

Thus a constituent at a depth z_{90} would have to be about an order of magnitude more concentrated than at the surface to have the same effect on R.

Interpretation of Remotely Sensed Ocean Color

In the case of a remote sensor, the quantity actually measured is a radiance L. After atmospheric correction the quantity to be interpreted is $L_w(\theta,\phi)$, the water-leaving radiance. (Austin (1974) calls L_w the inherent sea surface radiance; however, we shall adopt the term 'water-leaving radiance' to avoid the confusion which may arise because radiance is an apparent optical property rather than an inherent optical property, according to standard terminology.) This is the radiance just above the surface which emerges from the ocean (see Figure 1). For a flat surface it is related to the upwelling radiance just beneath the surface $L(\theta,\phi)$ by

$$L_w(\theta,\phi) = L(\theta,\phi)[1 - \rho(\theta',\theta)]/m^2, \tag{8}$$

where m is the refractive index of water, $\rho(\theta',\theta)$ the Fresnel reflectivity (subsurface) for an incident angle θ', and $\sin\theta' = (1/m)\sin\theta$. Since m varies only slightly with wavelength, $L_w(\theta,\phi)$ and $L(\theta,\phi)$ have essentially the same spectral composition.

To the approximation that the ocean is a diffuse Lambertian reflector (beneath the surface) the radiance $L(\theta',\phi')$ has a constant value L_u (for downward looking paths), and

$$E_u = \pi L_u,$$

or

$$L_u = (E_d(0)/\pi)R.$$

From radiance distribution measurements (Tyler, 1960; Smith, 1974) and also from theory (Prieur, 1976), it appears that the above assumption is an oversimplification. According to Austin (1974), if

L_u is the radiance in the nadir direction, π must be replaced a factor closer to 5. This is principally due to the fact that $L(\Theta',\emptyset')$ increases rapidly with Θ' for $\Theta > \Theta_c'$ the critical angle ($\sim 48^o$). For $\Theta' < \Theta_c'$ the assumption that $L(\Theta',\emptyset')$ is constant in a given situation appears to be well satisfied, and hence

$$L_u = (E_u(0)/Q)R.$$

According to theoretical predictions, Q depends slightly on wavelength λ. If L_u is used in absolute units, this factor must be taken into consideration. If radiances are used through ratios as $L_u(\lambda_1)/L_u(\lambda_2)$ for the two wavelengths λ_1 and λ_2, the spectral change of the above mentioned factor should be considered, but has not been, due to the lack of experimental data. At present, the influence of this quantity on the algorithms discussed below is felt to be small.

When the sea surface is roughened by the wind one expects the relationship between $L_w(\Theta,\emptyset)$ and $L(\Theta',\emptyset')$ to be more complex than Eq. 9. However, Austin (1974) has shown that when $L(\Theta',\emptyset')$ is completely diffuse the dependence of the surface transmittance $(L_w(\Theta,\emptyset)/L(\Theta',\emptyset'))$ on wind speed is weak, and Eq. 9 can be applied with an error of less than 10% for wind speeds less than 10 m/s and view angles less than about 50 degrees. Even though $L(\Theta',\emptyset')$ is typically not totally diffuse, the fact that it is essentially constant for $\Theta' < \Theta'c$ implies that this conclusion is still valid. Again, if the radiances are used through ratios, the ratio of water-leaving radiances will be the same as the ratio of subsurface radiances.

Along with being well understood theoretically, ocean color has also been well documented experimentally. Since 1970, considerable data (see Table 1) have been acquired concerning spectral values $E_u(\lambda)$, $R(\lambda)$ and more recently, $L_u(\lambda)$. Some of them remain unpublished. The waters studied vary from the 'desert' blue waters (e.g., Sargasso Sea) to those rich in phytoplankton, in suspended sediments, in dissolved yellow substance, or influenced by diverse combinations of these various components. As a matter of fact, algorithms now in use in CZCS data interpretation, or others more advanced in use for airborne measurements, originate from analytical or statistical studies of this 'radiometric sea-truth' data bank generated by ship-bound oceanographers.

Table 1

Cruise	Zone	Number of Stations	Optical Measurement
Vislab/SIO	Crater Lake Gulf Stream Gulf of Calif.	9	E_d, E_u[1]
SCOR/UNESCO Discoverer (1970)	Sargasso Sea Central-East Pacific	18	E_d, few E_u[2]
,,	,,	23	E_d, E_u[3]
Cineca-Charcot II (1971)	Upwelling off Mauritania	10	E_d, E_u[4]
Harmattan (1971)	Central-East Atlantic	16	E_d, E_u[4]
Cineca-Charcot V (1974)	Upwelling off Mauritania	30	E_d, E_u[5]
Guidom (1976)	South of Cape Verde Islands	11	E_d, few E_u[6]
Mares-Beagle II (1972)	Galapagos Islands	5	E_d[7]
Antiprod (1977)	South Indian and Antarctic	13	E_d, E_u[8]
4 Cruises (1974-76)	Japan Sea and Coastal Zones	20	E_d, E_u[9]
2 Cruises (1978)	Tokyo Bay Hisatsuru	13	E_d, E_u[10]

Table 1 (Cont.)

Cruise	Zone	Number of Stations	Optical Measurement
Researcher (1977)	Gulf of Mexico	10	E_d, E_u, L_u[11]
Crockett (1978)	Lake Erie	2	E_d, L_u[11]
Gyre (1978)	Gulf of Mexico	10	E_d, E_u, L_u[11]
New Horizon (1979)	Southern Calif. Bight	10	E_d, E_u, L_u[11]
Oceanus (1979)	Western North Atlantic	12	E_d, E_u, L_u[11]
David Starr Jordan (1979)	Eastern Pacific	13	E_d, L_u[11]
Oceanographer (1980)	North Pacific (Japan-Seattle)	13	E_d, L_u[11]
Desteiguer (1981)	San Diego to Hawaii	2	E_d, L_u[11]
Fos (1979)	Fos Gulf, Etang Berre	8	E_d, E_u[12]
Emicort (1977)	off Marseille Etang Berre	5	E_d, E_u[12]
C-Fox (1979)	Vancouver Island	10	E_d, E_u[12]