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Satellite Observations of Ocean Colour [and Discussion]

I. S. Robinson and E. G. Mitchelson

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Satellite observations of ocean colour

By I. S. Robinson
Department of Oceanography, The University, Southampton SO9 5NH, U.K.

Ocean colour is unique among the properties of the sea that can be measured from satellites because visible wavelength radiation penetrates below the surface skin and contains direct information about the bulk water quality in the upper few metres of the sea. With reference to Nimbus CZCS and Landsat MSS data, this review examines three aspects of ocean-colour monitoring: the removal of atmospheric effects, which make up a large proportion of the visible signal reaching the satellite; the calibration of the ocean colour signal in terms of more useful ocean parameters such as sediment load, chlorophyll concentration, water depth or the depth of the euphotic zone; and the potential applications of colour-monitoring satellites to oceanography. Algorithms for atmospheric correction are now well developed, but calibration for chlorophyll or sediment is less certain, particularly where inorganic sediments or land-derived yellow substance, as well as the local phytoplankton population, are present in the sea. Oceanographic applications include the estimate of total productivity, the identification of blooms, and the location of productive areas to assist ship surveys. The greatest potential for satellite observations of ocean colour may lie in the synoptic spatial information contained in the images, but this awaits serious exploitation by dynamical oceanographers.

1. Introduction

With the launch of the first manned orbiting satellites in the early 1960s it became clear that the colour of the sea not only could be viewed clearly from space, in the absence of cloud cover, but also contained information of potentially great significance to oceanography. What could be viewed with the human eye could also be photographed relatively easily, so that the earliest colour images of the oceans available for scientific analysis were those from visible wavelength cameras on board manned satellites such as Gemini, Apollo and Skylab. Such images gave at best a qualitative impression of patterns of ocean colour variability, suffering from geometric distortions. The development of scanning radiometers made possible measurements capable of geometric registration and radiometric calibration. By using several sensors sensitive to different portions of the visible spectrum, an approximate measure could be made of the spectral signature, or 'colour', of the upwelling radiatian reaching the satellite.

This paper is concerned with the oceanographic use that can be made of both single and multiple waveband radiance measurements in the visible region. Most of the quantitative analyses for marine applications have been performed on visible-wavelength data from the multispectral scanners (MSS) carried on the Landsat series of Earth-monitoring satellites, and the Coastal Zone Colour Scanner (CZCS) flown on the Nimbus 7 satellite. The sensor and orbital characteristics are listed in table 1. The Landsat MSS, designed for land applications, has coarser spectral and radiometric resolution than the CZCS, which was designed particularly for ocean-colour observations (Hovis *et al.* 1980). Landsat, with a smaller field of view, has much higher spatial resolution, which permits its use in coastal and estuarine situations, but this is won at the expense of a very long sampling interval. The new thematic mapper (TM) recently launched

Table 1. Satellite and sensor parameters

	CZCS		MSS	TM
platform	Nimbus 7		Landsat series	Landsat 4
approx. altitude/km	955		920 (Landsats 1–3)	705
orbit	near-polar, Sun-synchronous, northbound pass at local noon		near-polar, Sun-synchronous, 14 southbound passes per day at approx. 10h00 local solar time	near-polar, Sun-synchronous, $14\frac{1}{2}$ southbound passes per day equator crossing at 09h45 local solar time
swath width/km	2300 at 20° forward tilt; 1600 at no tilt		185	185
daytime overpass frequency	every day		every 18 days (2 adjacent days out of 18 in overlap zone)	every 16 days (7 and 9 days in overlap zone)
pixel size/m	825 × 825		56 × 79 (nominally 80 m resolution)	30 × 30
visible and near visible wavebands/nm	443 ± 10		500–600	450–520
	520 ± 10		600–700	520–600
	550 ± 10		700–800	630–690
	670 ± 10		800–1100	760–900
	700–800			1550–1750
saturation radiance in the above channels/ (mW cm^{-2} μm^{-1} sr^{-1})	min. gain	max. gain	24.8	—
	11.46	5.41	24.4	
	7.64	3.50	20.0	
	6.21	2.86	17.6	
	2.88	1.34		
	23.90	23.90		
radiometric digitizing resolution (no. of digital levels available)	255		63	255

on Landsat 4 promises even more spatial detail, at improved radiometric and spectral resolutions.

Three broad research areas can be identified as needing attention before the satellite colour scanner can become a credible and calibrated oceanographic instrument. The first is the problem of removing from the signal received at the sensor all radiation that does not emerge from below the sea surface, i.e. light backscattered by the atmosphere and reflected at the sea surface, and also of correcting for the absorption by the atmosphere of part of the radiation leaving the water. Early visible images of the ocean raised considerable interest in the potential for studying the sea surface by interpreting the sun glint patterns (see, for example, Cox 1974), but more powerful radar techniques have now been developed to study surface-roughness signatures, and the viewing directions of the present scanning visible radiometers are chosen to minimize specular reflexion. I shall therefore treat it here as noise rather than signal.

The second broad study area is the interpretation of the optical signal received from the sea in terms of an oceanographically more meaningful parameter, such as the suspended sediment load, or the concentration of chlorophyll pigmentation. This is strictly the field of optical oceanography, hitherto rather neglected in the U.K., but the coming of the satellite sensor has given it new importance and value, and raised new scientific problems requiring solution.

Finally we need to consider the contributions that satellite observation of ocean colour can make to oceanographic science as a whole. Remote-sensing scientists have been active in the first two areas, but the third has received less attention because the oceanographer has tended to be sceptical of the value of visible remote sensing owing to limitations by cloud cover. It is, however, the only satellite surveillance technique that can penetrate a few metres below the water surface. This paper aims to present to ocean scientists the present capabilities and limitations of the method, and to point towards those areas of oceanography where it promises to make exciting new advances possible.

2. Atmospheric corrections

(a) *Physical principles*

The photons reaching the sensor of a satellite outside the Earth's atmosphere arrive from the incident solar radiation by a variety of routes, including scattering in the atmosphere, scattering within the sea, reflexion at the sea surface and multiple combinations of these possibilities (Sturm 1981*a*; Sorensen 1981). The satellite sensor measures the radiant energy in a given spectral window, incident on the aperture area, arriving from a very narrow cone of directions whose projection on the Earth is the instantaneous field of view, defining the pixel area. The observed variable is therefore the radiance as a function of wavelength and viewing direction, $L_t\ (\lambda, \theta, \phi)$.

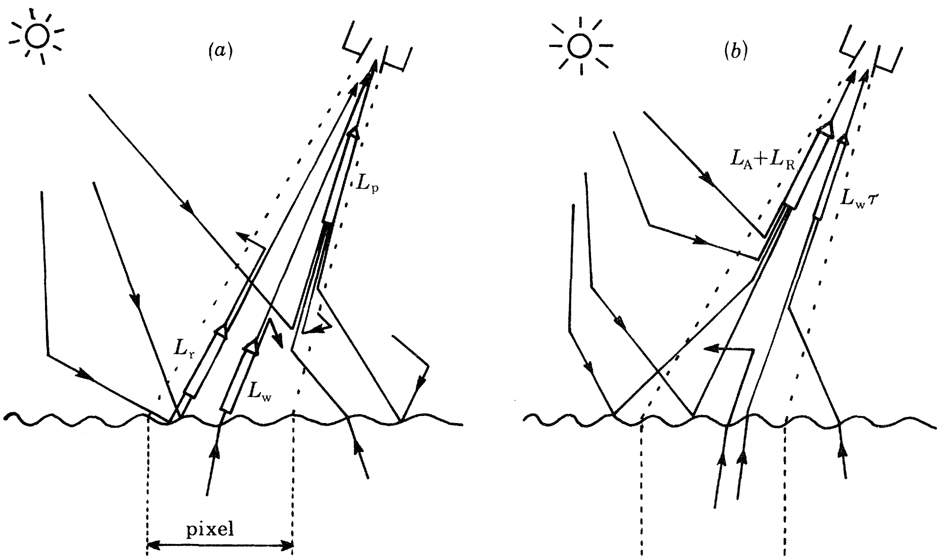

Figure 1. Optical pathways.

The optical quantity related to L_t that contains oceanographic information is $L_w\ (\lambda, \theta, \phi)$, the water-leaving radiance in the direction of the sensor. Some of L_w is scattered or absorbed by the atmosphere before reaching the sensor (figure 1*a*). An additional radiance L_r leaves the sea surface owing to specular reflexion of sunlight (sun glint) or direct reflexion of atmospherically scattered light (sky glint). There is also a path radiance L_p contributed by atmospheric scattering into the sensor field of view of photons, some of which may previously have reflected beneath or at the surface outside the pixel area. Thus

$$L_t = L_w \tau_1 + L_r \tau_1 + L_p, \tag{1}$$

where τ_1 is the direct transmittance through the atmosphere. The task of atmospheric correction is to extract L_w, given L_t.

Given the variety of available paths, the likelihood of multiple scattering, the three-dimensional nature of scattering and the number of unknown quantities such as the optical properties of the atmospheric scatterers and the sea-surface roughness in a given image, it is impossible to determine L_r and L_p from first principles. Progress has been made by simulating the detailed photon-by-photon optical response of typical atmospheres by using Monte-Carlo statistical techniques. These have been used to test the validity of various simplifying assumptions and have led to an approximate model of atmospheric scattering, developed by Gordon (1978, 1981). First, the radiance received at the sensor is divided into those photons that have not penetrated below the sea surface and those that have (figure 1*b*). Given that satellite colour sensors are now designed to avoid direct sun glint, the former are all dependent on random atmospheric scattering processes. The latter will not all have upwelled from the pixel in the instantaneous field of view because of scattering, but provided that the sea properties do not vary strongly with position it is acceptable to relate this part of L_t to L_w from the given pixel. Thus

$$L_t = L_A + L_R + \tau L_w. \tag{2}$$

Here τ is now the diffuse transmittance of the atmosphere, which can be calculated by using the expression given by Sturm (1983), which includes the effect of ozone absorption. L_A is the radiance that would be due to photons scattered by suspended particles in the atmosphere (aerosol scattering) if the atmosphere itself did not scatter, and L_R is due to the Rayleigh scattering that would occur in the absence of aerosols. Equation (2) therefore contains the simplifying approximation justified by Gordon (1978) that, despite multiple scattering, L_A and L_R are independent. Now L_R can be calculated (see Viollier *et al.* (1980) and Sturm (1981*a*)) for a given viewing angle and solar illumination conditions, whereas L_A cannot be estimated *a priori* for a satellite image because the aerosol content varies with time and space. Since $L_A + L_R$ may contribute up to 80% of L_t in the blue part of the spectrum, it is essential that L_A be estimated as accurately as possible if small variations of L_w from scene to scene, and within a scene, are to be reliably detected.

The multispectral nature of the information from the sensor is therefore used to achieve an atmospheric correction, with the assumption that

$$L_{A,\lambda_2} = S(\lambda_2, \lambda_1)\, L_{A,\lambda_1}. \tag{3}$$

This appears to be a valid approximation provided that the aerosol phase function (i.e. the function controlling the scattering direction) and the ratio of the aerosol optical thicknesses, t_A, at the two wavelengths, does not vary within the scene (i.e. on a length scale of several hundred pixels), even though the aerosol scattering may vary significantly. Furthermore, if the aerosol phase function is independent of wavelength too, then S may be represented by

$$S(\lambda_2, \lambda_1) = \epsilon(\lambda_2, \lambda_1)\, F_{0,\lambda_2}/F_{0,\lambda_1}, \tag{4}$$

where $\epsilon(\lambda_2, \lambda_1) = \omega_{A,\lambda_2} t_{A,\lambda_2}/\omega_{A,\lambda_1} t_{A,\lambda_1}$, ω is the probability that a photon will backscatter on interaction with the aerosol and F_0 is the solar irradiance at the top of the atmosphere.

If the sensor records a waveband centred at λ' for which the sea can be assumed to absorb all incident radiation (i.e. $L_{w,\lambda'} = 0$) then $L_{A,\lambda'}$ can be obtained from (2) after calculating $L_{R,\lambda'}$. The significance of (3) is that $L_{A,\lambda}$ can then be obtained from $L_{A,\lambda'}$. Since $L_{R,\lambda}$ can be calculated, the atmospheric correction is determined at wavelength λ provided that $S(\lambda, \lambda')$ is known.

(b) *Algorithms for* CZCS

For the CZCS, channel 4 (wavelength 670 ± 10 nm) is taken as λ', for which L_w is assumed to be zero. Therefore it is required to know $S(\lambda_i, 670)$, with $i = 1, 2, 3$ corresponding to the wavelengths of channels 1 to 3 of the sensor. Several approaches can be used, leading to different algorithms. In each case S is assumed not to vary with position within a scene.

If $t_{A,\lambda_i}/t_{A,670}$ can be measured in one single location simultaneously with the satellite overpass, $S(\lambda_i, 670)$ can be determined from (4) if it is assumed that $\omega_{A,\lambda_i}/\omega_{A,670} = 1$. Alternatively if L_{w,λ_i} were measured at one place synoptic with the overpass, $S(\lambda_i, 670)$ could be determined for the whole scene by using (2) and (3), provided that τ could be accurately estimated. Because τL_w is typically only 20% of L_t, small errors in τ would be relatively unimportant, as shown by Gordon (1981). Because in fact most CZCS scenes appear to indicate that the aerosol correction varies significantly over the extent of a scene, some of the assumptions concerning the form of S may lead to error and it is preferable to make several independent estimates of S from synoptic observations in different parts of the scene. In this way the validity of assuming S to be constant can be tested, and a mean value of S obtained for the whole scene. Results of such tests remain to be published, although Gordon (1981) has raised questions of the validity of using ground measurements of $t_{A,\lambda}$ to estimate $S(\lambda, 670)$.

A further problem in the adoption of Gordon's approach occurs in that $L_{w,670}$ is not always zero in turbid coastal water (Bukata *et al.* 1980). Sturm (1981*b*) has avoided this by calculating $L_{A,670}$ for the darker sea pixels only, for which $L_{w,670}$ is expected to be zero. This, however, prohibits the pixel-by-pixel calculation of $L_{A,670}$, which has to be estimated for the rest of the scene by taking only the viewing and solar geometry into account, and not the aerosol variability. Smith & Wilson (1981) alternatively adopt $L_{w,670} = 0$ in a preliminary atmospheric correction, after which the chlorophyll concentration is calculated. This can then be used to estimate a likely value of $L_{w,670}$ for each pixel enabling an iterative correction procedure to be followed. This is limited by the accuracy of the chlorophyll algorithms (see §3).

Synoptic measurements of optical quantities at ground level are difficult and costly to obtain at sea, and the goal of a universal atmospheric correction algorithm is that it should operate by using only satellite data. If the form of S could be universally established, the problem would be solved. Gordon & Clark (1980) have used $\epsilon(\lambda, 670) = (670/\lambda)^n$, where n is known as the Ångström exponent, but there appears to be no clear consensus in the literature as to whether the exponent should be positive (as implied by Ångström 1964), negative (Quenzel 1970) or zero (De Luisi *et al.* 1972).

The most comprehensive self-contained algorithm yet to appear is presented by Sorensen (1981) and Sturm (1983). In this the scene is initially corrected by assuming that $L_{w,670} = 0$ and $\epsilon = 1$. The darkest pixels, remote from land and cloud, and preferably near the centre of the scene, are identified in the resulting $L_{w,\lambda}$ images. Provided that it is oceanographically reasonable, it is assumed that these pixels correspond to clear water with a chlorophyll concentration less than 0.3 μg l^{-1}. In this case $L_{w,520}$ and $L_{w,550}$ can be calculated by the method of Gordon & Clark (1981), enabling an accurate estimate of $S(520, 670)$ or $S(550, 670)$ to be made for those dark-water pixels. $S(443, 670)$ is extrapolated from these. These S are then applied to the whole scene for a pixel-by-pixel aerosol correction assuming $L_{w,670} = 0$, and if necessary the Smith & Wilson iteration can be included for the cases where $L_{w,670} \neq 0$. Although it has not yet been exhaustively tested, such an approach promises to work well for open ocean waters where patches of low chlorophyll occur and the water turbidity is dominated by chlorophyll with a reasonably well understood optical signature. Such an approach is not so promising in turbid coastal waters, and research waits to be done on testing the accuracy of these algorithms in a variety of different shallow sea and ocean locations.

(c) *Application to Landsat* MSS

The same approach to an atmospheric correction should be valid whatever sensor is used. However, attempts to apply the above method to Landsat MSS data over the sea areas reveal that the sensor response, designed for land use, is incompatible with accurate sea observations. Indirectly this serves to illustrate the important design requirements of an ocean colour sensor.

The particular problems are the poor spectral and radiometric resolutions of the MSS. The sensor bands are nominally 100 nm wide, and in practice the radiation entering the sensor is filtered through a spectral window that is not uniformly flat. The equations presented in §2*a* are appropriate for narrow bandwidths, but because S is probably nonlinearly frequency dependent it is much less precise to apply (3) over a broad band, selecting the central wavelength at which to define S. A more accurate approach would require the inclusion of an integral convolving S with the frequency window of the sensor. Landsat has the advantage of a longer-wavelength sensor centred at 950 nm (channel 7), for which L_{w} must almost always be zero. However, the problem of identifying an appropriate $L_{\mathrm{A},950}$ for use as $L_{\mathrm{A},\lambda}$ is compounded with a 300 nm bandwidth and irregular spectral window in channel 7. Moreover it is not clear that it is appropriate to extend the physical assumptions underlying (2), (3) and (4) so far into the infrared region. MacFarlane (1981) has made an attempt along these lines to develop a haze-removal technique for Landsat sea scenes, but has been restricted by the poor radiometric sensitivity of the MSS, particularly channel 7. Not only is $L_{\mathrm{w,c7}}$ zero, but $L_{\mathrm{A,c7}}$ is also small and is not resolved by the sensor with sufficient accuracy to determine L_{A} confidently for the other channels in each pixel.

Attempts to provide atmospheric corrections for Landsat MSS have therefore tended to be much cruder, at best subtracting from a whole scene the radiance level of the darkest pixel (Johnson 1975; Robinson & Srisaengthong 1981) on the assumption that the water-leaving radiance is zero and the signal is entirely due to path radiance. No attempt is usually made to apply a differential atmospheric correction across a scene, and most attempts at quantitative interpretation of Landsat data over sea have incorporated the atmospheric effects in empirically derived regression constants (see, for example, Klemas *et al.* 1974).

It is clear that to facilitate atmospheric corrections to ocean colour data, future sensors must be designed with narrow wavebands well spaced throughout the visible and into the near infrared. The sensitivity must be sufficient at the higher wavelengths to resolve small aerosol-path radiance differences above the instrument noise or digitizing level. Such criteria have influenced the design of the successor to the CZCS (Hovis 1981) and the European OCS (see, for example, Giannini 1981).

3. Water-quality parameters from ocean colour

(a) *Optical principles*

The removal of atmospheric effects is only the first step in making use of satellite radiance measurements, because it still remains to interpret the observations of ocean colour in terms of oceanographically interesting parameters. Techniques in the science of optical oceanography

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have tended during most of this century to concentrate on measuring absorption and scattering under water (Jerlov 1976) in terms of their effect on productivity or as a measure of suspended material. Only in the last decade has the prospect of space observation provoked much more study of what we can learn by viewing ocean colour from above, as reviewed recently by Sathyendranath & Morel (1983).

The water-leaving radiance L_w is influenced as much by the incident radiation as by the water's optical properties. In fact the parameter of most use is the subsurface reflectance ratio, $R = E_u/E_d$, where E_u is the upwelling irradiance and E_d the downwelling irradiance just below the surface. Now for a flat surface,

$$L_w(\theta, \phi) = RE_d\{1-\rho(\theta', \phi)\}/Qn^2, \tag{5}$$

where n is the refractive index of water and ρ is the Fresnel reflectivity of the water–air interface ($n \sin\theta' = \sin\theta$ relates the subsurface ray inclination θ' to the viewing direction θ of the satellite). Equation (5) is valid for a rough surface provided that θ is not close to the critical angle of 48° (Austin 1974). Assuming that the ocean is a perfect Lambertian reflector, $Q = \pi$ and is direction-independent. However, observations at sea (Austin 1980) suggest that Q should be approximately 5.0 for near-nadir observations. Given L_w from the satellite sensor, R can then be estimated from (5), by calculating E_d given the known Sun illumination and atmospheric properties based on the atmospheric correction already discussed. To improve accuracy the dependence of n on salinity and temperature should be incorporated (Singh *et al.* 1983). Since Q and E_d are only weakly wavelength-dependent, it is often more satisfactory to use

$$\frac{R(\lambda_1)}{R(\lambda_2)} = \frac{L_w(\lambda_1)}{L_w(\lambda_2)}\,\frac{n(\lambda_1)^2}{n(\lambda_2)^2}\,\frac{[1-\rho(\lambda_2)]}{[1-\rho(\lambda_1)]},$$

i.e. the reflectance spectral ratio is obtained from the remotely observed radiance ratio (i.e. colour) of the water.

R itself depends on the balance of absorption and backscattering of light by the water and its suspended contents. Since particle sizes are somewhat larger than the wavelengths of light, Mie scattering theory applies. Preisendorfer (1961) has shown that the total absorption coefficient, a, of the water and the total backscattering coefficient, b_b, are the linear sums of the individual coefficients due to the different absorbers and scatterers. It is therefore useful to express R in terms of a and b_b. Morel & Prieur (1977) concluded from theoretical modelling studies that

$$R = 0.33\, b_b/a \tag{6}$$

is a satisfactory approximation for most marine situations where b_b/a is small (i.e. less than 0.3). Gordon *et al.* (1975) had proposed $R = f\{b_b/(a+b_b)\}$, but (6) lends itself better to the study of spectral ratios of reflectance, absorption and backscattering coefficients. Given a knowledge of the spectral absorption and backscattering characteristics of the different water contents, as discussed below, it is therefore possible to predict the water colour that would be viewed by a satellite. The inversion of this process must underlie any physically based algorithm for water quality analysis using satellite colour observations.

(*b*) *Spectral characteristics of different water types*

Pure sea water provides the baseline against which to compare other types. Morel & Prieur (1977) and Smith & Baker (1981) present typical observations of the spectral dependence of

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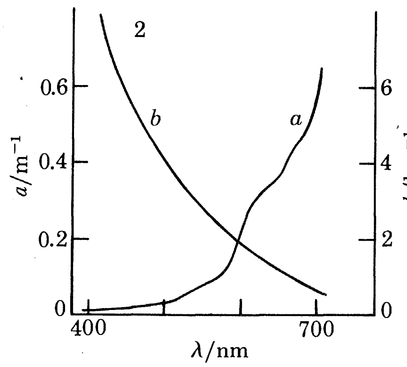
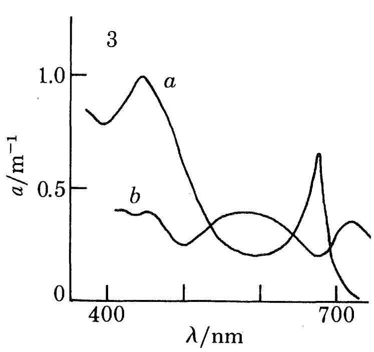
a and b_b, summarized in figure 2. Absorption decreases and backscatter increases with decreasing wavelength, leading to the characteristic blue colour of pure seawater.

Water containing phytoplankton has much more complex spectral characteristics. This is because along with the living cells, which contain chlorophyll, are the associated detrital products, biogenic particulate material and dissolved organic compounds containing phaeophytin *a*. Each of these has its own absorption and scattering properties, and each is found in different proportions according to the species or even the age of the population. Thus observations of phytoplankton-rich sea water *in situ* (see, for example, Prieur & Sathyendranath 1981; Smith &





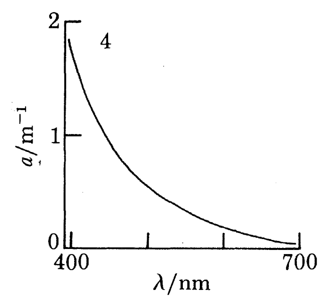


FIGURE 2. Absorption (a) and scattering (b) of pure sea water.

FIGURE 3. Typical spectral variation of: (a) absorption due to chlorophyll, normalized at 440 nm (typical values of specific absorption for chlorophyll at 440 nm vary between 0.01 and 0.1 m^{-1} $(mg\ m^{-3})^{-1}$, depending on age and species) (after Prieur & Sathyendranath 1981); (b) specific backscattering coefficient (units are typically 10^{-3} m^{-1} $(mg\ m^{-3})^{-1}$).

FIGURE 4. Absorption spectrum of yellow substance.

Baker 1981; Kiefer *et al.* 1979) show a variety of spectra, in agreement with the theoretical predictions of Morel & Bricaud (1981). Certain general characteristics emerge, however, as sketched in figure 3. There is a strong absorption peak at 440 nm and a lesser one at around 675 nm; over the rest of the spectrum in general the absorption decreases with wavelength and the backscatter is fairly uniform apart from decreases corresponding to the absorption peaks.

In some locations such as the Baltic Sea, strong concentrations of *yellow substance* (dissolved organic matter) are found in the absence of phytoplankton. This is believed to be strongly related to land run-off (Hojerslev 1980*a*). The absorption spectrum is exponential in form (figure 4) but is the reverse of sea water.

The other, and most diverse, contribution to ocean colour is the presence of *suspended particulate material* not related to phytoplankton. This may be resuspended bottom sediment, river-borne sediment, or eroded coastal and beach material, or it may be due to dumping of waste or dredge spoil. Its spectral characteristics are as diverse as its composition in terms of natural material colour and size distribution, and there appears to be no consensus in the literature about a standard spectrum shape.

Given so many different contributions to the upwelling radiance spectrum, it is apparent that there is little possibility of being able to invert the problem and obtain an analytical relation between chlorophyll content, or suspended solids concentration and the colour radiance or reflectance signal. None the less, the knowledge that has been gained about the optical properties of different substances in sea water does enable a general framework to be constructed from

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which empirical calibration algorithms may be formulated. A fundamental step is to divide all waters into two categories for optical purposes. Case 1 waters are seas whose optical properties are dominated by phytoplankton and their degradation products only, and case 2 waters have non-chlorophyll-related sediments or yellow substance instead of or in addition to phytoplankton. Figure 5 shows typical reflectance spectra for (*a*) case 1 water, and for (*b*) sediment-dominated and (*c*) yellow-substance-dominated case 2 waters. The combination of absorption and backscatter in case 1 water tends to decrease the reflectance below the clear water spectrum at wavelengths below around 540 nm and slightly increases it at higher λ. Increasing chlorophyll content enhances this effect with a definite minimum appearing at 440 nm due to the

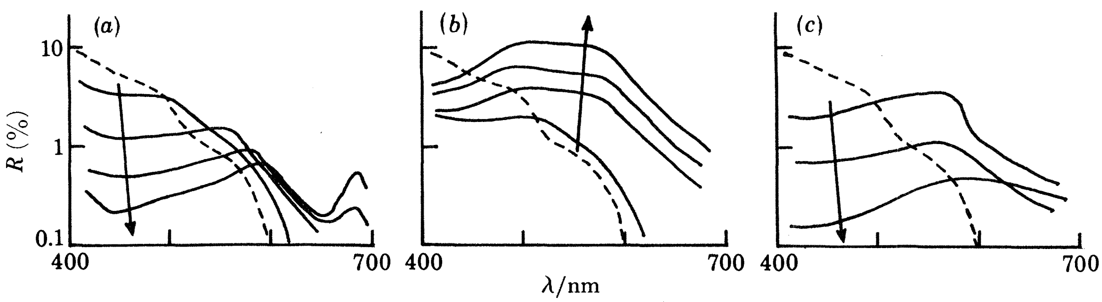

FIGURE 5. Typical reflectance spectra: (*a*) case 1 water; (*b*) case 2 water dominated by suspended sediment; (*c*) case 2 water dominated by yellow substance. – – –, Spectrum for clear blue water; ——→, increasing concentration. (After Sathyendranath & Morel (1983).)

chlorophyll absorption, and at 660 nm, although the red absorption maximum is masked by a reflectance peak at 685 nm due to fluorescence. Although the absolute reflectance values differ with different chlorophyll species, the spectral shape and its variation with chlorophyll concentration remain similar, giving encouragement to the formulation of universal chlorophyll algorithms based on reflectance spectral ratios.

In figure 5*b*, the backscattering by the sediment produces higher reflectance ratios at the larger wavelengths, although there is some reduction at small wavelengths due to small amounts of chlorophyll absorption. Because the spectral shape does not change much with sediment load, the water colour does not change and a sediment algorithm based on spectral ratios appears less hopeful, except in strongly coloured sediments where figure 5*b* would not apply. Instead, an algorithm relating suspended load to the absolute reflectance in a single waveband might be more appropriate. The yellow-substance-dominated waters (figure 5*c*) show a distinct colour change with increasing concentration as the blue reflectance becomes less and less.

From these generalized spectra, it appears that there is sufficient change in optical properties correlated with the concentration of the dissolved or suspended material to permit calibration of colour images in terms of water quality, but only if the water has a single component. This fundamental limitation provides the proper perspective for the development of interpretative algorithms. Optimistic attempts at producing universal algorithms appear to be ill-conceived, but undue pessimism is also out of place. Theoretical models of reflectance such as those by Bukata *et al.* (1981), which confirm that there can be no unique relation between optical and water quality parameters, must be weighed against the fact that we very often have information from sources other than the satellite, which may resolve ambiguities in the remotely sensed data.

(c) Calibration algorithms for CZCS data

Most of the proposed algorithms have been of the form

$$C \text{ or } S \text{ or } K(\lambda) = A(L_{443}/L_{550})^B \quad \text{or} \quad A(L_{520}/L_{550})^B, \tag{7}$$

where C is the total pigment concentration (chlorophyll *a* plus phaeophytin *a*) (micrograms per litre), S is the dry mass concentration of total suspended solids (milligrams per litre), and $K(\lambda)$ is the diffuse attenuation coefficient of the near-surface water for a given λ. Although $K(\lambda)$ is itself an optical property rather than an oceanographic parameter, its value lies in that it has been extensively measured at sea and used for the bio-optical classification of sea water (Smith & Baker 1978*a*, *b*; Baker & Smith 1982).

Equation (7) is simply the result of regression analyses of the logarithms of the satellite radiance ratios and the water properties measured synoptically at sea. The only physical understanding entering (7) is the choice of a spectral ratio as a variable. It should be clear from the above section that the best fit to such a form is likely to be found in case 1 water. Sathyendranath & Morel (1983) and Morel & Gordon (1980) have reviewed the available algorithms (see, for example, Clark 1981; Clark *et al.* 1980; Gordon & Clark 1980; Gordon *et al.* 1982; Morel 1980; Smith & Baker 1982; Smith & Wilson 1981; Sturm 1983) and show this to be so. Case 1 waters give a fairly large negative exponent for the phytoplankton algorithm, and achieve a logarithmic accuracy of $\pm\frac{1}{4}$. By definition, the variables C, S and $K(\lambda)$ are not independent in case 1 water, all being controlled by the phytoplankton, and therefore correspondingly good algorithms are achieved for S and K, a logarithmic accuracy of one-sixth being claimed for the seston algorithm. In general the use of the L_{443}/L_{550} ratio is preferred, particularly at low chlorophyll concentrations. At high concentrations the absorption reduces L_w (443) to such low levels that the atmospheric correction becomes inaccurate, and the L_{520}/L_{550} ratio has to be used instead, with a further loss of accuracy because the wavelengths are so close to one another.

The success achieved with case 1 waters can be misleading if taken out of context. The only reason that coherence exists between the spectral reflectance ratios and sediment is that the particulate load is controlled by the biomass, which has a chlorophyll colour signature. In case 2 waters, no such correlation can be found between sediment load and reflectance ratios (see, for example, Hojerslev 1974). Indeed, in case 2 waters, where there may be at least three independent components influencing the subsurface optics, it is theoretically impossible to resolve any of them from two spectral ratio measurements, which are not completely independent because they are so close to each other within the spectrum. The only useful algorithms in these circumstances are likely to be limited to a particular location and seasonal condition, and heavily dependent on synoptic sea data. In those case 2 waters where yellow substance is not dominant, calibration for pigment still appears to be possible (Gordon & Morel 1981) and recent observations in the Irish Sea suggest that sediment may be related to the wideband reflectance magnitude whereas the spectral ratio continues to indicate the chlorophyll. None the less it may well be more promising to calibrate the imagery in terms of the attenuation coefficient $K(\lambda)$ (Austin & Petzold 1981), which can then be compared with C and S by ground measurements without the need for data collection at the same time as a clear-weather satellite overpass.

Other types of calibration algorithm, involving sums and differences of different bands, or the ratio of differences, have been tried. Most show some limited success in accounting for a

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synoptic data set, but their value in predicting calibrations of other scenes than the one they were constructed from is small, because they appear to lack any sound physical basis for their form.

A different and potentially more useful approach, applicable to case 1 and 2 waters, is that promoted by Hojerslev (1980*b*, 1981*a*, *b*). Concerned not only that the colour index (i.e. the spectral reflectance ratio) has no unique relations with C or S, but also that it gives no indication of the vertical variability of C and S that often occurs, he draws attention instead to a strong correlation between colour index and the depth of the euphotic zone (i.e. the depth at which the downwelling energy flux is 1% of the surface value). This observed result is supported by theoretical models, and although it may not be as useful a measure as C or S, it is capable of useful exploitation by physical and biological oceanographers.

(*d*) *Calibration of Landsat data*

Given the broader bandwidths and the poorer atmospheric corrections, the attempts to calibrate Landsat data in terms of water quality have been crude compared with the CZCS work. It is much harder to pick out the spectral peaks and troughs with a 100 nm band and chlorophyll has therefore been very hard to detect, except in case 1 waters where it is related to particulate scatterers (Le Fèvre *et al.* 1982). Because of their high spatial resolution, Landsat data have been applied mostly in coastal and estuarine regions where type 2 water predominates. Calibration algorithms are therefore only locally applicable. Because of the infrequent repeat cycle, synoptic sea data have also been very hard to obtain, and it is much harder to test the accuracy of calibrations.

Most attempts have been made to relate the satellite radiances to the load of suspended solids, and algorithms fall into two broad categories, those that use colour and those that use intensity. Since in general non-chlorophyll-related suspended solids do not have a strong colour signature, it is more useful to seek simple relations between S and the reflectance in a single band. Often band 5 has been the most useful because it is less contaminated by atmospheric effects. For low concentrations (less than 60 mg l^{-1}) a linear relation is satisfactory (see, for example, Munday & Alfoldi 1979; Robinson & Srisaengthong 1981) whereas over a wider range of loads the reflectance is correlated with $\log_{10} S$ (Klemas *et al.* 1974).

The most successful use of colour correlation with S has come from the use of chromaticity analysis (Alfoldi & Munday 1978). Radiance values are mapped onto a two-colour chromaticity transform plane and S is correlated with the position along the resulting locus. Atmospheric contamination is treated by shifting the whole locus to a normalized position. The success of this technique in studying the Bay of Fundy is probably due to the highly coloured red sediments which are brought down by local rivers, resulting in a strong colour signature to the concentration.

Because of their high resolution, Landsat MSS data have also been used in hydrographic applications to measure shallow water depths. This requires a spectral calibration because in clear water it is the lower wavelengths that penetrate further and hence give a higher reflectance if the bottom is less than about 20 m deep. Cracknell *et al.* (1982) attempted a quantitative calibration, but the influence of suspended sediment and bottom albedo tend to confine this to a qualitative use.

4. Applications to oceanography

The value to science of a new measurement technique is only confirmed when advances in scientific understanding accrue from the new data. When the new technique simply improves the accuracy or precision of a previous method, the new data can usually be applied to modify an existing theoretical framework. In the case of ocean colour, however, the satellite sensor is producing data different in kind from oceanographic information obtained from traditional sources, and a full appreciation of the value of remote-sensing measurements calls for the construction of new theoretical models in which to exploit the data. The unique feature of satellite scanners is their capacity to provide a synoptic view of a large area of sea, with fine spatial resolution, and their ability to repeat that view over months or years. It is in this areal coverage that we should seek to find the principal scientific applications in oceanography, rather than looking to the satellite measurements to duplicate the time resolution and depth penetration achieved by observations based on ships and buoys.

Because the colour scanner is not an all-weather instrument and cannot penetrate cloud it is difficult to use it in the operational role envisaged for some radar sensors, and unreasonable to look to it for observing temporal variability on timescales less than a few days. Of course, where cloud cover is infrequent the daily overpasses of Nimbus 7 can be used semi-operationally. This is true off the California coast where charts showing the position and boundaries of water masses of different ocean colour as detected by CZCS are issued to fishing vessels every few days by N.A.S.A./J.P.L. as part of a wider experimental programme to use satellites in support of commercial fisheries.

The obvious application, for which the CZCS was designed, is the study of the spatial and temporal variability of primary production in the world's ocean. Already studies such as those by Smith & Wilson (1981), Yentsch & Garfield (1981), Anderson *et al.* (1981) and Gordon *et al.* (1980) have demonstrated the feasibility of mapping synoptic chlorophyll concentrations, in case 1 waters at least, to a good degree of accuracy over areas the size of the Southern California Bight or the Gulf of Maine – Georges Bank region. This would have been impossible to achieve with research vessels because covering such a wide area with even a coarse sampling interval would have taken several days, during which time the plankton populations could be expected to have grown or decayed, and been advected through the region. The accuracy of the satellite calibration of individual scenes also appears to be comparable with that achieved with continuous sampling fluorometers. To the designers of the colour scanner such calibrated images appear to be the successful end-product of the project, whereas to the ocean scientist they represent the beginning of a new type of synoptic oceanography. We find ourselves at the early descriptive phase, discovering new aspects of the spatial distribution of chlorophyll and sediment with every new image, but only slowly beginning to formulate the questions that will lead to a new theoretical understanding of the factors controlling the observed distributions. Although little has yet been published, it is now possible to estimate the total primary productivity of a sea area, and its seasonal variations (Smith & Baker 1982; Smith *et al.* 1982). Because the colour data are most strongly influenced by the near-surface chlorophyll, a model must be used to assess the contribution of the rest of the productive depth (Platt & Herman 1982). With a knowledge of the typical spatial variability it will be possible to evaluate how representative are the ship-based observations that have been used in the past for global productivity estimates. Moreover, the observed spatial distribution of chlorophyll, and its evolution in time, should enable much more to be learned about the way in which physical processes – advection by residual flows and upwelling, mixing across fronts, stirring by tides or storms, etc. – control the biological processes, as for example in a Gulf Stream ring (Gordon *et al.* 1982).

The particular study area that has already benefited significantly from the CZCS imagery has been the identification of productive regions in the vicinity of oceanic or shallow sea fronts (see, for example, Yentsch & Garfield 1981; Yentsch 1983; Pingree & Mardell 1981; Pingree *et al.* 1982). Comparison with infrared imagery reveals the relative positions of thermal and colour fronts, or where productive patches occur in the absence of a surface thermal signature. It is worth noting that in the work so far reported, satellite observations have been used to extend the usefulness of, but not to replace, shipborne measurements. Indeed, in process-oriented studies designed to supply data for ecosystem modellers (Esaias 1981) the satellite image can identify the location of the most active areas of productivity where physical and biological parameters require detailed ship-borne study for a period of days to weeks (e.g. during the time of an algal bloom) whereas during the exercise the wider area in which the particular process is set can be regularly surveyed by the satellite. Holligan *et al.* (1983) demonstrate the effectiveness of combined ship and satellite surveying of a dinoflagellate bloom in the English Channel.

Besides the biological science that is developing, satellite ocean-colour data also promise advances in dynamical oceanography. Treating the plankton or suspended sediment as a passive tracer, the satellite images reveal patterns related to residual flows, eddies, frontal instabilities or mixing processes. Many such features may have a colour signature even when they do not appear on thermal infrared imagery. Attempts to obtain dynamical information from colour images have mostly been qualitative, e.g. using CZCS images to display sediment movement in headland eddies (Mardell & Pingree 1981) or describing sediment movement from Landsat images of the Louisiana Bight (Rouse & Coleman 1976), the Bay of Fundy (Amos & Alfoldi 1979), the Solent (Srisaengthong 1982) or the Baltic (Horstmann & Hardtke 1981). Others have attempted to map small-scale features in estuaries by using Landsat data, with a view to pollution-control applications (Klemas 1980). However, Gower *et al.* (1980) sought quantitative information from a Landsat image of the chlorophyll–sediment signature of mesoscale eddies in the North Atlantic. By spectral analysis of the image they obtained an estimate of the wave-number of spatial variability.

This last example represents the innovative approach required if the potential of the spatial coverage of satellite data is to be fully realized. In general, dynamical theories have been constructed with time as the principal independent variable. We tend to study temporal evolution and variability because oceanographers have become good at measuring time series cf dynamical variables by using conventional ship-and-buoy sampling. Now that we can sample equally well in the spatial domain, using satellites, there is every reason to develop more theoretical approaches to understanding spatial variations. Hitherto the testing of models of spatial variability has largely been limited to advective systems in which the length scales could be inferred from a point-measured time series, but the satellite scanner has removed that limitation. The further attraction of obtaining the length scales of dynamical processes from images is that the radiometric–ocean-parameter calibration is no longer so important. Provided that spatial variability of atmospheric effects is removed, accurate length scales can be obtained even when there is only a crude calibration algorithm available, in just the same way as frequency peaks can be readily obtained from a time series whose amplitude calibration is

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suspect. This field of study awaits development, but modern image-processing computers and software are already available for this type of data analysis.

One other application of satellite colour, or visible wavelength, data to dynamical oceanography is the estimation of the radiant shortwave heat flux into the upper ocean. Gautier (1981) has calculated this by using the visible channel on the geostationary GOES meteorological satellite. A development of this may be to use spectral information from the CZCS to estimate the depth of energy penetration into the sea, and thence to model more accurately the thermal structure of the upper ocean and its horizontal variability (Simpson & Dickey 1981; Dickey & Simpson 1983).

5. Conclusions

There is no doubt that we now have sufficient understanding of atmospheric scattering and absorption of visible light to be able to determine an accurate ocean-colour signature from space, given a sufficiently sensitive radiometer measuring in narrow bandwidths covering the visible spectrum, and preferably into the near infrared. The CZCS has demonstrated the validity of the technique, and its successors should show further improvements in accuracy. Experience with the CZCS also suggests that chlorophyll concentration can be measured by satellite to within a factor of two in case 1 waters, although there is a need to test this further with data from a greater variety of locations, whereas waters containing sediment and yellow substance not related to the local plankton population require much more investigation before a universal calibration algorithm can be considered. Looking to the future, it may be possible to adapt aircraft-borne solid-state narrow-band spectrometers for satellite use (Gower & Borstad 1981). By measuring the fluorescence lines these detect the chlorophyll directly, irrespective of the presence of other material.

The application of ocean-colour data is much less developed than the sensor and calibration technologies, which appear to be pushing forward independently of the scientific demand. There is a need for oceanographers to develop a better theoretical framework in which to make use of the spatial data from satellites. The most immediate requirement is for the colour imagery of the oceans to be surveyed and the features revealed on them to be catalogued to stimulate the creation of new hypotheses concerning such spatial problems as the patchiness of productivity and its relation to physical conditions, or the use of chlorophyll and sediment as tracers to reveal the length scales of a variety of dynamical processes. Ocean-colour measurements from space are not to be regarded as a substitute for ship observations, but should be seen as a very different type of data, requiring a different framework for analysis, and with the potential to promote fresh scientific insights in oceanography.

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References

Alfoldi, T. T. & Munday, J. C. 1978 *Can. J. remote Sensing* **4**, 108–126.
Amos, C. L. & Alfoldi, T. T. 1979 *J. Sedim. Petr.* **49**, 159–174.
Anderson, F. P., Shannon, L. V., Mostert, S. A., Walters, N. M. & Malan, O. G. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 381–386. New York and London: Plenum Press.
Ångström, A. 1964 *Tellus* **16**, 64–75.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Austin, R. W. 1974 In *Optical aspects of oceanography* (ed. N. G. Jerlov & E. Steeman Nielson), pp. 317–334. London: Academic Press.
Austin, R. W. 1980 *Bound.-Layer Met.* **18**, 269–285.
Austin, R. W. & Petzold, T. J. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 239–256. New York and London: Plenum Press.
Caker, K. S. & Smith, R. C. 1982 *Limnol. Oceanogr.* **27**, 500–509.
Bukata, R. P., Brinton, J. E., Jerome, J. H., Jain, S. C. & Zurick, H. H. 1981 *Appl. Opt.* **20**, 1704–1714.
Bukata, R. P., Jerome, J. H., Brinton, J. E. & Jain, S. C. 1980 *Appl. Opt.* **19**, 2487–2488.
Blark, D. K. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 227–237. New York and London: Plenum Press.
Clark, D. K., Baker, E. T. & Strong, A. E. 1980 *Bound.-Layer Met.* **18**, 287–298.
Cox, C. S. 1974 In *Optical aspects of oceanography* (ed. N. G. Jerlov & E. Steeman Nielson), pp. 51–75. London: Academic Press.
Cracknell, A. P., MacFarlane, N., McMillan, K., Charlton, J. A., McManus, J. & Ulbricht, K. A. 1982 *Int. J. remote Sensing* **3**, 113–137.
De Luisi, J. J., Blifford, I. H. & Takamine, J. A. 1972 *J. geophys. Res.* **77**, 4529–4538.
Dickey, T. D. & Simpson, J. J. 1983 *Tellus.* (In the press.)
Esaias, W. E. 1981 *Oceanus* **24**, (3), 32–38.
Gautier, C. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 201–206. New York and London: Plenum Press.
Giannini, L. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 395–402. New York and London: Plenum Press.
Gordon, H. R. 1978 *Appl. Opt.* **17**, 1631–1636.
Gordon, H. R. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 257–265. New York and London: Plenum Press.
Gordon, H. R., Brown, O. B. & Jacobs, M. M. 1975 *Appl. Opt.* **14**, 417–427.
Gordon, H. R. & Clark, D. K. 1980 *Bound.-Layer Met.* **18**, 299–313.
Gordon, H. R. & Clark, D. K. 1981 *Appl. Opt.* **20**, 4175–4180.
Gordon, H. R., Clark, D. K., Brown, J. W., Brown, O. B. & Evans, R. H. 1982 *J. mar. Res.* **40**, 491–502.
Gordon, H. R., Clark, D. K., Mueller, J. L. & Hovis, W. A. 1980 *Science, Wash.* **210**, 63–66.
Gordon, H. R. & Morel, A. Y. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 207–212. New York and London: Plenum Press.
Gower, J. F. R. & Borstad, G. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 329–338. New York and London: Plenum Press.
Gower, J. F. R., Denman, K. L. & Holyer, R. J. 1980 *Nature, Lond.* **288**, 157–159.
Hojerslev, N. K. 1974 *Inherent and apparent optical properties of the Baltic* (*Rep. Inst. Phys. Oceanogr. Univ. Copenhagen*, no. 23). (70 pages.)
Hojerslev, N. K. 1980*a* *On the origin of yellow substance in the marine environment* (*Rep. Inst. Phys. Oceanogr. Univ. Copenhagen*, no. 42). (16 pages.)
Hojerslev, N. K. 1980*b* *Bound.-Layer Met.* **18**, 203–220.
Hojerslev, N. K. 1981*a* In *Application of remote sensing data on the continental shelf* (ed. N. Longdon & G. Levy), pp. 73–76. Paris: European Space Agency, SP-167.
Hojerslev, N. K. 1981*b* In *Oceanography from space* (ed. J. F. R. Gower), pp. 347–353. New York and London: Plenum Press.
Holligan, P. M., Viollier, M., Dupouy, C. & Aiken, J. 1983 *Continental Shelf Res.* (In the press.)
Horstmann, U. & Hardtke, P. G. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 429–438. New York and London: Plenum Press.
Hovis, W. A. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 213–225. New York and London: Plenum Press.
Hovis, W. A. *et al.* 1980 *Science, Wash.* **210**, 60–63.
Jerlov, N. G. 1976 *Marine optics*. Amsterdam: Elsevier.
Johnson, R. W. 1975 In *Remote sensing of Earth resources*, vol. 4 (ed. F. Shahrokhi), pp. 565–576. Tullahoma, Tennessee: University of Tennessee.
Kiefer, D. A., Olson, R. J. & Wilson, W. H. 1979 *Limnol. Oceanogr.* **24**, 664–672.
Klemas, V. 1980 *Int. J. remote Sensing* **1**, 11–28.
Klemas, V., Bartlett, D., Philpot, W. & Rogers, R. 1974 *Remote Sensing Envir.* **3**, 153–174.
Le Fèvre, J., Viollier, M., Le Corre, P., Dupouy, C. & Grall, J. R. 1982 *Estuar. coast. Shelf Sci.* **16**, 37–50.
MacFarlane, N. 1981 In *Matching remote sensing technologies and their applications*, pp. 499–508. Reading: Remote Sensing Society.
Mardell, G. T. & Pingree, R. D. 1981 *Oceanologica Acta* **4**, 63–68.
Morel, A. Y. 1980 *Bound.-Layer Met.* **18**, 177–201.
Morel, A. Y. & Bricaud, A. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 313–327. New York and London: Plenum Press.

Morel, A. Y. & Gordon, H. R. 1980 *Bound.-Layer Met.* **18**, 343–355.
Morel, A. Y. & Prieur, L. 1977 *Limnol. Oceanogr.* **22**, 709–722.
Munday, J. C. & Alfoldi, T. T. 1979 *Remote Sensing Envir.* **8**, 169–183.
Pingree, R. D. & Mardell, G. 1981 *Phil. Trans. R. Soc. Lond.* A **302**, 663–678.
Pingree, R. D., Mardell, G. T., Holligan, P. M., Griffiths, D. K. & Smithers, J. 1982 *Continental Shelf Res.* **1**, 99–116.
Platt, T. & Herman, A. W. 1982 *Int. J. remote Sensing* **4** (2). (In the press.)
Preisendorfer, R. W. 1961 In *Radiant energy in the sea* (*Int. Union Geophys. Geod. Monogr.*, no. 10), pp. 11–29.
Prieur, L. & Sathyendranath, S. 1981 *Limnol. Oceanogr.* **26**, 671–689.
Quenzel, H. 1970 *J. geophys. Res.* **75**, 2915–2921.
Robinson, I. S. & Srisaengthong, D. 1981 In *Application of remote sensing data on the continental shelf* (ed. N. Longdon & G. Levy), pp. 221–232. Paris: European Space Agency, SP-167.
Rouse, L. J. & Coleman, J. M. 1976 *Remote Sensing Envir.* **5**, 55–66.
Sathyendranath, S. & Morel, A. Y. 1983 In *Remote sensing applications in marine science and technology* (NATO ASI series) (ed. A. P. Cracknell), pp. 323–357. Dordrecht: D. Reidel.
Simpson, J. J. & Dickey, T. D. 1981 *J. phys. Oceanogr.* **11**, 309–323.
Singh, S. M., Cracknell, A. P. & Charlton, J. A. 1983 *Int. J. remote Sensing* **4** (In the press.)
Smith, R. C. & Baker, K. S. 1978*a* *Limnol. Oceanogr.* **23**, 247–259.
Smith, R. C. & Baker, K. S. 1978*b* *Limnol. Oceanogr.* **23**, 260–267.
Smith, R. C. & Baker, K. S. 1981 *Appl. Opt.* **20**, 177–184.
Smith, R. C. & Baker, K. S. 1982 *Mar. Biol.* **66**, 269–279.
Smith, R. C., Eppley, R. W. & Baker, K. S. 1982 *Mar. Biol.* **66**, 281–288.
Smith, R. C. & Wilson, W. H. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 281–294. New York and London: Plenum Press.
Sorensen, B. M. (ed.) 1981 *Recommendations of the 2nd International Workshop on atmospheric correction of satellite observation of sea water colour.* (49 pages.) Ispra, Italy: Commission of the European Communities Joint Research Centre.
Srisaengthong, D. 1982 Ph.D. thesis, University of Southampton.
Sturm, B. 1981*a* In *Remote sensing in meteorology, oceanography and hydrology* (ed. A. P. Cracknell), pp. 163–197. Chichester: Ellis Horwood.
Sturm, B. 1981*b* In *Oceanography from space* (ed. J. F. R. Gower), pp. 267–279. New York and London: Plenum Press.
Sturm, B. 1983 In *Remote sensing applications in marine science and technology* (NATO ASI series) (ed. A. P. Cracknell), pp. 137–167. Dordrecht: D. Reidel.
Viollier, M., Tanre, D. & Deschamps, P. Y. 1980 *Bound.-Layer Met.* **18**, 247–267.
Yentsch, C. S. 1983 In *Remote sensing applications in marine science and technology* (NATO ASI series) (ed. A. P. Cracknell), pp. 263–297. Dordrecht: D. Reidel.
Yentsch, C. S. & Garfield, N. 1981 In *Oceanography from space* (ed. J. F. R. Gower), pp. 303–312. New York and London: Plenum Press.

Discussion

E. G. Mitchelson (*Department of Physical Oceanography, Marine Science Laboratories, U.C.N.W., Menai Bridge, U.K.*). Dr Robinson has emphasized the severe difficulties of making a satisfactory atmospheric correction in the presence of high levels of suspended solids, which can invalidate the assumption that the water-leaving radiance at 670 nm is zero. In the circumstances, it seems prudent to try to establish whether or not there is a useful colour signature leaving the sea surface when seston concentrations are high.

In collaboration with N. J. Jacob and J. H. Simpson, I have attempted to do this by making a series of observations in the case 2 waters of the northern Irish Sea, with a multichannel irradiance meter. This instrument has the same channels as the CZCS, with a bandwidth of ± 10 nm and sufficient sensitivity to permit the measurement of upwelling irradiance even under conditions of low surface illumination. Parallel measurements of total pigments, total seston (organic and inorganic fractions), and yellow substance were made by using standard methods (Strickland & Parsons 1968; Kalle 1963).

Figure D1 shows a comparison of the blue:green ratio of upwelling light just below the sea surface with the values of total pigment concentration, which extend over one decade. These

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observations were made in the presence of inorganic seston concentrations of up to 8 mg l^{-1} (see figure D2). While the regression analysis shows that the blue:green ratio is strongly dependent on total pigments C ($r = 0.92$, 37 d.f.), including total seston as an additional variable has no significant effect. The negative slope of the regression line, -3.9, is somewhat greater than those determined by other workers in case 1 waters (Smith & Baker 1982; More 1980; Gordon & Clark 1980).

Although the colour of the upwelling light is not markedly dependent on the seston concentration, the intensity is. Figure D2 shows a plot of the reflectance at 550 nm against the inorganic fraction of the total seston. There is considerable scatter, but a significant relation between the variables can be detected ($r = 0.77$, 37 d.f.).

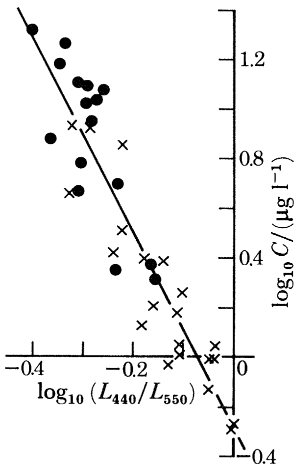

FIGURE D1

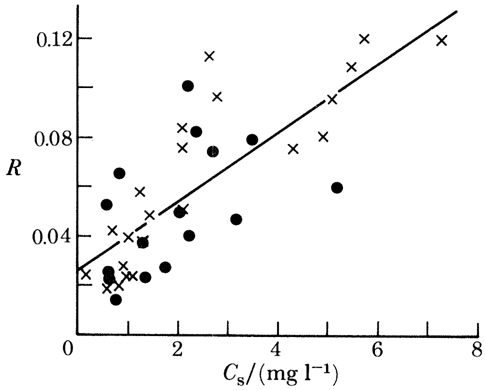

FIGURE D2

FIGURE D1. Total pigments, C, plotted against the blue:green ratio, L_{440}/L_{550}. The line represents the result of a regression analysis, which yields $\log_{10}C = 3.9\log_{10}(L_{440}/L_{550}) - 0.29$. The data points are from two cruises in the Irish Sea. ×, 7–9 July 1982; ●, 12–14 July 1982.

FIGURE D2. Relative reflectance, R, at 550 nm plotted against concentration of inorganic seston, C_s. R is the ratio of upwelling radiation reaching the surface normalized by the above-surface downwelling irradiance measured by an uncalibrated broadband sensor. The regression line is $R = 0.014\,C_s + 0.026$. Data points from two different cruises are identified as in figure 1.

These results, admittedly based on a small sample, show that there is useful information leaving the sea surface in the upwelling irradiance signal, even in case 2 waters dominated by suspended sediment. As well as confirming the above relations we would hope that further radiometric measurements at the sea surface will provide a basis for the testing of the atmospheric algorithm, which now appears as a critical problem in the development of this aspect of remote sensing.

References

Gordon, H. R. & Clark, D. K. 1980 *Bound.-Layer Met.* **18**, 299–315.
Kalle, K. 1963 *Dt. hydrogr. Z.* **16**, 153–166. [In German.]
Morel, A. 1980 *Bound.-Layer Met.* **18**, 177–201.
Smith, R. C. & Baker, K. S. 1982 *Mar. Biol.* **66**, 269–279.
Strickland, J. D. H. & Parson, T. R. 1968 *Bull. Fish. Res. Bd Can.*, no. 167. (31 pages.)

I. S. ROBINSON. Because most of the reflectance measurements reported in the literature are for case 1 waters, these results reported by Miss Mitchelson for case 2 waters are of particular

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interest. They lend support to the hypothesis that in case 2 waters a chlorophyll algorithm can be constructed on the blue: green spectral ratio of reflectance while inorganic suspended solids can be derived from the broad band magnitude of reflectance.

However, these observations must be treated as location-specific and not necessarily applicable to other sea areas. It will be interesting to see if these results can be repeated in other locations, particularly coastal zones where the inorganic suspended sediment concentration is much greater than the 8 mg l^{-1} encountered here, or in areas where more highly coloured sediment is found. It is observations at sea such as those reported here that will open up the archived CZCS data for the British continental shelf seas to quantitative interpretation.

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