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OCEAN COLOR DETERMINATION THROUGH A SCATTERING ATMOSPHERE

ROBERT J. CURRAN

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GSFC — GODDARD SPACE FLIGHT CENTER
GREENBELT, MARYLAND

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ABSTRACT

Measurements made of the surface level albedo for ocean water containing various concentrations of phytoplankton indicate a strong correlation between wavelength dependent albedo ratios and phytoplankton chlorophyll concentration. To sense surface level albedo ratios from space platforms it is necessary to correct for the scattering and absorption properties of the atmosphere for the wavelengths in question. Atmospheric scattering models were constructed to calculate corrections at two wavelengths, 0.46 and 0.54 μm. The relationship between albedo ratios at the top of the atmosphere and at surface level were made for several different aerosol optical depths. Assuming a natural background uncertainty in the aerosol optical depth of 0.1 it is found that the chlorophyll concentration may be determined to within one standard deviation of from 0.5 to 2.5 milligrams per cubic meter, depending upon the solar zenith angle at the time of measurement. By remotely sensing the aerosol optical depth to a greater accuracy it appears feasible to detect chlorophyll concentrations to an uncertainty approaching 0.1 milligram per cubic meter.

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INTRODUCTION

Ocean color is the spectral response of the ocean to the input of both incident solar radiation, which is attenuated by the earth's atmosphere, and the contribution of diffuse sky radiation. This spectral response may be quantitatively defined in terms of the surface albedo $A_S(\lambda)$, which is a function of the wavelength λ. The magnitude of the surface albedo is a result of the scattering and absorption properties of particulate constituents and dissolved materials present in the ocean water. Since two types of albedos will be used a detailed definition will be presented in the following section.

The angular distribution of radiation leaving the top surface of the ocean may be considered to be the result of two different physical processes. The first is the component of the incident radiation which is specularly reflected from the surface of the ocean. The second component consists of the upwelling radiation coming from below the air-sea interface. The angular distribution of the specularly reflected solar radiation can be modeled to some precision and has been discussed by several authors. (Cox and Munk, 1965).

The radiance returned to space from below the water surface is several orders of magnitude smaller than that of the specular reflection of direct sunlight. Because the desired spectral information comes from the upwelling radiation emanating from beneath the water surface, it is necessary to restrict viewing of the ocean surface to geometries which do not include the specularly reflected radiation. The calculations made to determine the effect of a scattering atmosphere in observing ocean color assume that the sun glint pattern resulting from specular reflection has been avoided. Therefore, the ocean surface will be assumed to act simply as a Lambert reflector of radiation. The albedo $A_S(\lambda)$ of such a surface is defined as:

$$A_S(\lambda) = \frac{\pi I(\lambda)}{F_S(\lambda)} \tag{1}$$

where the wavelength dependent radiance $I(\lambda)$ is independent of the viewing direction for the reflecting surface as described above. Here, $F_S(\lambda)$ is the irradiance incident on the ocean surface from above.

The irradiance, $F_s(\lambda)$, consists of the directly transmitted solar irradiance $u_o F_t(\lambda)$ (where u_o is the cosine of the solar zenith angle and $F_t(\lambda)$ is the wavelength dependent transmitted solar radiation) and a diffuse incident component as denoted by $F_D(\lambda)$. Thus, the incident irradiance is described by the following relationship:

$$F_s(\lambda) = u_o F_t(\lambda) + F_D(\lambda) \tag{2}$$

Measurements have been reported of the spectral dependence of the ocean surface albedo by Ramsey (1968), by Clark, Ewing, and Lorenzen (1970) and White (1969). The measured albedo as reported by these authors is directly equivalent to the above definition of surface albedo if I (λ) in equation (2) is the nadir radiance at the surface. The nadir radiance is the upwelling radiation as measured by a photometer viewing the nadir direction and will be referred to as such in the following. Some of the measured data with corresponding values of chlorophyll concentration are shown in Figure 1.

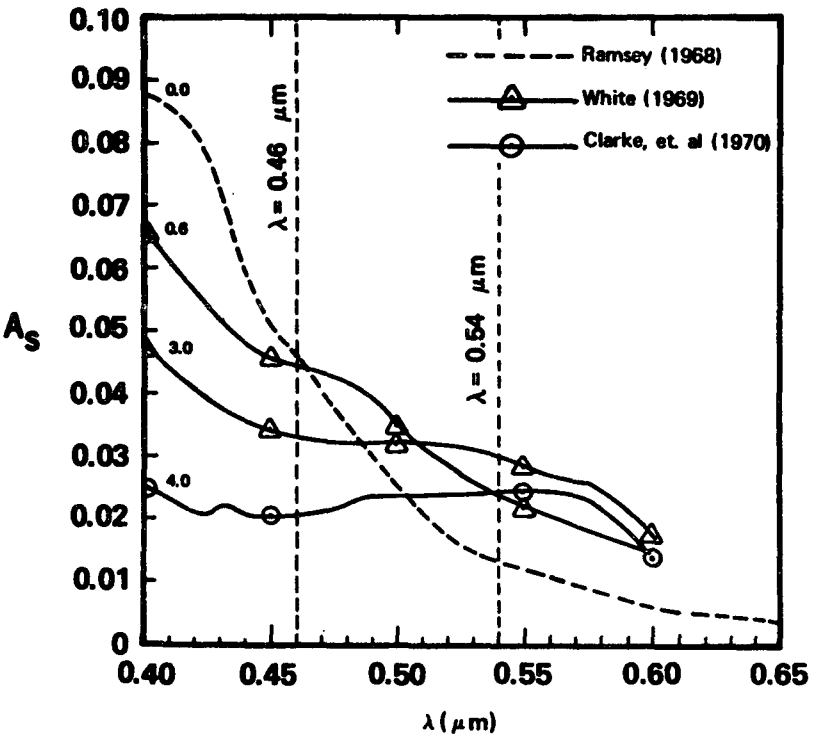

Figure 1. Measured wavelength dependence of ocean surface level albedo. The values adjoining the curves on the left hand side of the figure indicate the measured chlorophyll concentration in mg/m^3

It has been noted by several authors, as for example, Clarke, et al. (1969), Clarke, et al. (1970) that the surface albedo of the ocean may be used as an indicator for the amount of chlorophyll present in the water. The measurements presented by Clarke, et al. (1970) suggest a correlation between the concentration of chlorophyll in the water and the wavelength dependent surface albedo.

The experimental information given by Ewing, et al. and that presented by Ramsey (1968) and White (1969) will be used to determine the wavelength dependent surface albedo as a function of chlorophyll concentration. This data consists of color ratios for chlorophyll concentrations divided into two widely spaced groups. Because of the lack of data the model used was a linear relationship between chlorophyll concentration and surface albedo. The experimental approach of Clarke, et al (1970) to the determination of chlorophyll concentration is the formation of ratios of albedo between two wavelengths such that the effects of chlorophyll concentration are maximized.

The wavelengths chosen by Clarke, et al. (1970) were 0.46 and 0.54 μm. In considering the relationship between chlorophyll concentration and ocean color as expressed in Figure 1, albedo ratios between these two wavelengths appear to optimize the effects of changing chlorophyll concentration. Calculations made using the measured spectral albedos give the relationship between the color ratio, A_s (0.54)/A_s (0.46), and chlorophyll concentration as is shown in Figure 2. Several measured values are shown on this diagram. A linear fit was applied to the values with high chlorophyll concentrations. Because of the large variability of the ocean color ratio for low chlorophyll concentrations, no attempt has been made to alter the linear form of this region of the graph.

The relationship between the surface color ratio and the chlorophyll concentration may be differentiated in order to determine the corresponding uncertainties between chlorophyll concentration and surface ratio. The relationship between the uncertainties is graphically displayed in Figure 3 for the chlorophyll concentration range of interest. In evaluating the errors inherent in measuring the surface color ratio through a turbid atmosphere we will use the relationship displayed in Figure 3 to determine the correspondence between the uncertainty in measuring ocean color and the uncertainty in estimating chlorophyll concentration. As measured data becomes available concerning the relationship between the sea surface color ratio and chlorophyll concentration, the relationships of Figures 2 and 3 will most certainly be altered. The analysis to follow will be made in terms of the color ratio and only when the analysis is complete will the corresponding errors in chlorophyll concentration be included. The procedure will allow the analysis to be easily

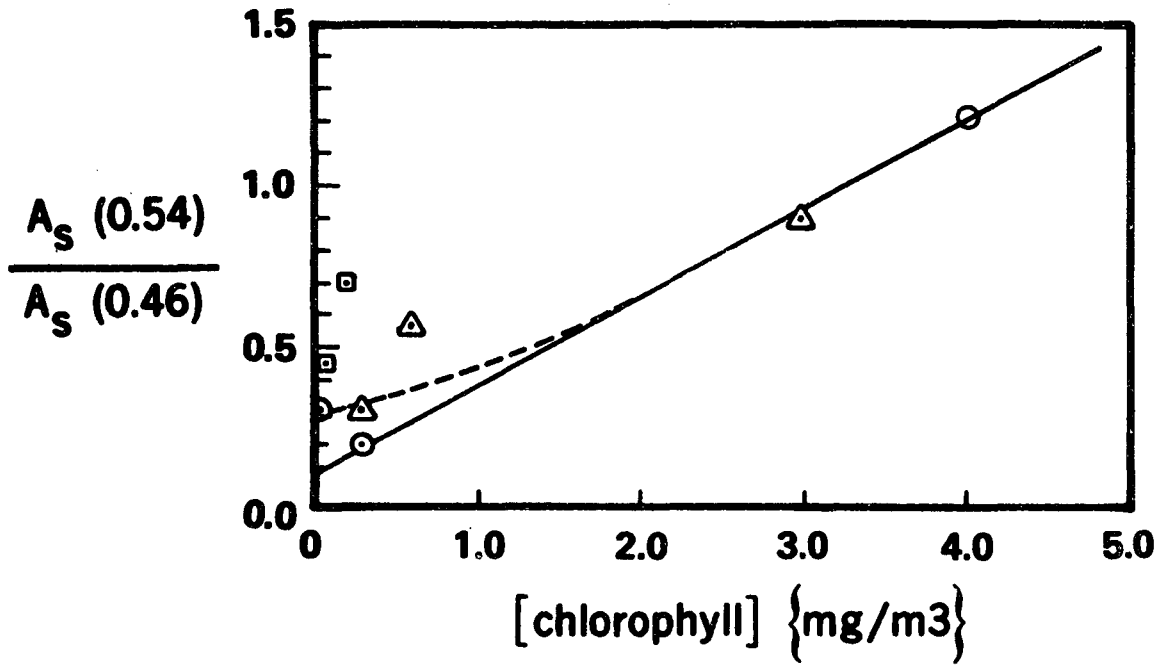

Figure 2. Measured surface level color ratios for several values of chlorophyll concentration. The source of the data points correspond to the legend of figure 1 with square enclosed data points due to Ramsey.

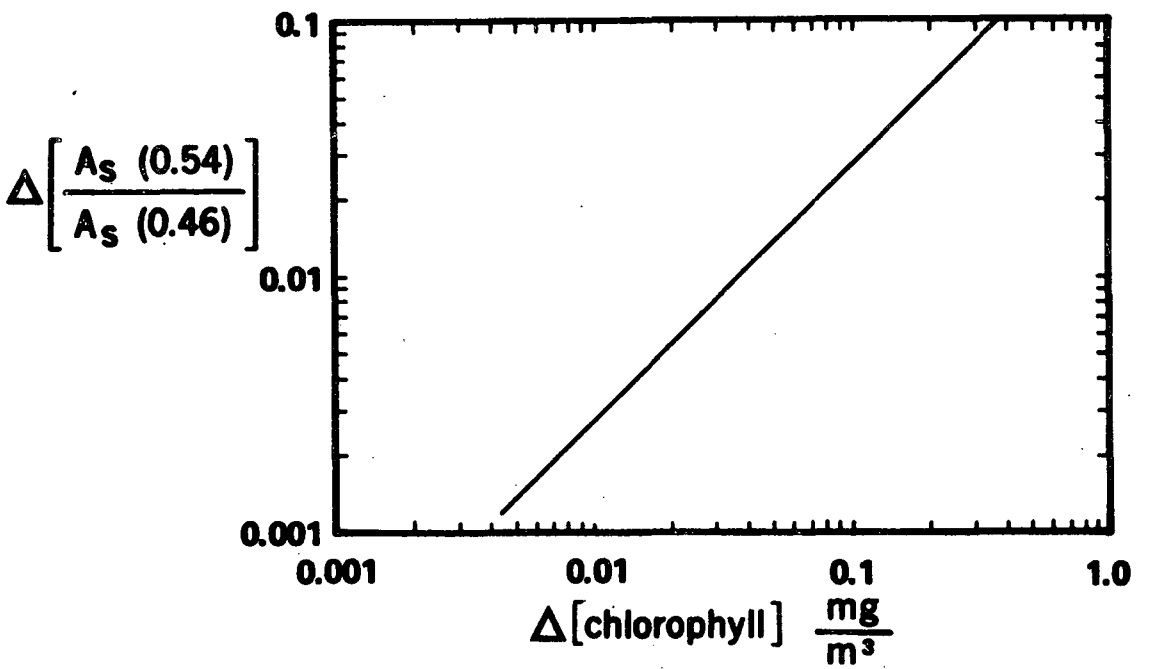

Figure 3. Relationship between the uncertainty in color ratio and the uncertainty in chlorophyll concentration with the assumption of a linear fit to the color ratio data. The uncertainty in both cases refers to one standard deviation from the mean value.

modified as our knowledge of the relationship between color ratio and chlorophyll concentration becomes more complete.

ATMOSPHERIC TRANSFER MODEL

In the determination of the obscuring effects of the atmosphere to observing ocean color from satellite altitudes a mixed molecular-aerosol scattering model was used. The molecular constituent was distributed with altitude according to the climatological mean as defined by the U.S. Standard Atmosphere. The aerosol constituent was distributed in altitude according to the Elterman 1964 clear air model (Elterman (1964)). However, the aerosol distribution was renormalized using as an input parameter the aerosol optical depth, τ_A. This optical depth is defined in terms of the particle number density n(z) and the volume scattering coefficient β_A (λ, z) for aerosols by the following:

$$\tau_A(\lambda) = \int_0^\infty n(z)\, \beta_A(\lambda, z)\, dz \tag{3}$$

where the integration is over altitude z.

The aerosol particulates were assumed to consist of a size distribution of spherical paricles with an index of refraction of 1.54. The imaginary part of the index of refraction was assumed to be zero thus precluding any absorption effects. This choice of refractive index approximates the optical parameters of both sodium chloride and silicate materials and thus seemed a reasonable assumption for the atmospheric particulates in maritime regions.

The particle size distribution assumed was the power law distribution with size parameter ν^* defined by:

$$dn(z, r) = c(z)\, r^{-(\nu^* + 1)}\, dr \tag{4}$$

where r is particle radius in micrometers and z is the altitude in kilometers. This distribution expresses the number density, dn (z, r) of particulates in the interval with the normalizing factor c(z) used to produce the proper total number of particulates in an atmospheric column. As has been noted by Herman, et al. (1971), for observation angles excluding the near forward or near backscatter directions, the effect of the size parameter, ν^*, on the angular distribution of scattered radiance is negligible.

The importance of the size parameter to these calculations is in determining the wavelength dependence of the aerosol optical depth. Because we are comparing scattered radiances at two different wavelengths this

wavelength dependence is of consequence. Calculations were made, assuming a particle size range of 0.01 to 10.0 μm, to determine the wavelength dependent scattering properties for several values of the size parameter ν^*. In Figure 4 are displayed the optical depths normalized to the aerosol optical depth at $\lambda = 0.50\ \mu$m, for three values of ν^* believed to be in the range typically encountered in nature. The curves displayed in Figure 4 assume that the size parameter is a constant with altitude and show strong dependence of aerosol optical depth with wavelength.

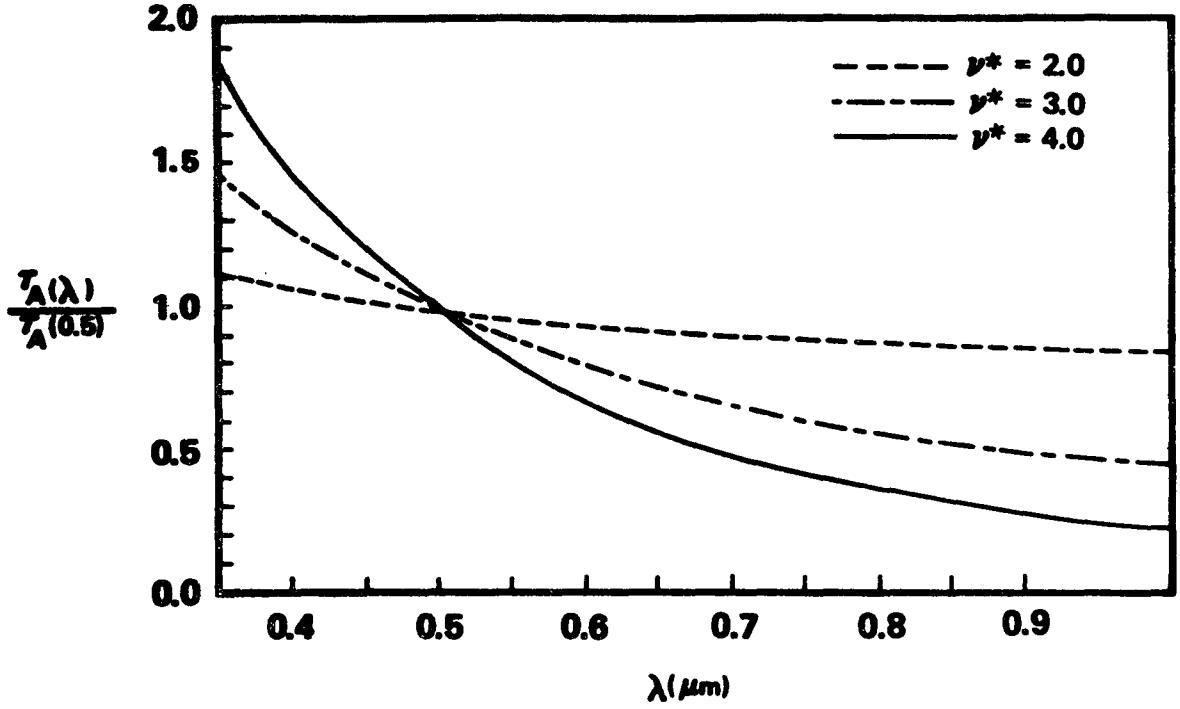

Figure 4. Wavelength dependence of the aerosol optical depth relative to the aerosol optical depth at $\lambda = 0.50\ \mu$m for three size parameters in the Junge distribution.

Because of the interest in forming ratios of radiances between the wavelength 0.54 and 0.46 μm, values for the ratio of optical depths at these same wavelengths were formed using the same distribution. The ratios of optical depth as a function of the size parameter ν^* are displayed in Figure 5. For naturally occurring aerosol distributions the ratio $\tau_A\ (0.54)/\tau_A\ (0.46)$ may be assumed to vary between the limits of 1.0 and 0.4. This range of values agrees with those of Quenzel (1970) who uses both a power law distribution and logarithmic Gaussian distribution. This latter distribution was specifically developed to be used in the interpretation of measurements of aerosol optical depth made in the Atlantic Ocean and thus presumably is representative of maritime aerosol distributions.

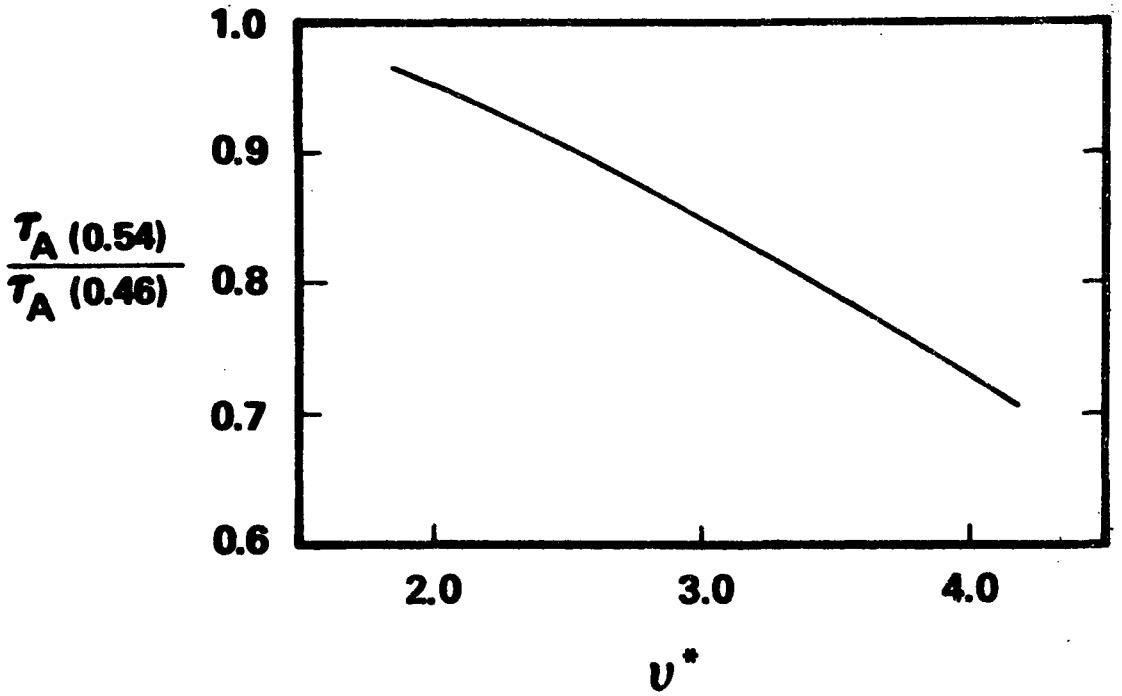

Figure 5. Radio of the aerosol optical depths at $\lambda = 0.54\ \mu m$ and $\lambda = 0.46\ \mu m$ for different values of the size parameters, ν^*. The lower boundary of the shaded areas are for $\nu^* = 4.0$ and the upper boundary for $\nu^* = 2.0$. The 3.0 value of ν^* is indicated by dashed lines.

NATURALLY OCCURRING AEROSOL OPTICAL DEPTHS

The mean aerosol optical depth at 0.50 μm wavelength, as suggested by Elterman (1964) for clear conditions, is approximately 0.24. Generally the particulates in the atmosphere over the oceans are considered to be some fraction of this value, in the range of 0.05 to 0.15. This lower value is reasonable if the continents are considered the primary source of the particulates and thus the particulates must be transported from continental to maritime regions. However, the oceans are known sources of salt and because NaCl is very hygroscopic and the water vapor is readily available, these particulates may be assumed to swell hygroscopically in the moisture laden air above the ocean waters.

In light of a process of this sort, a possible indication of the maritime optical depth is the world wide frequency distribution of visibility over the oceans. Analysis indicates that Elterman clear air estimate may be a rough mean for the 0.50 μm wavelength maritime optical depth. The visibility may easily range to values one tenth of the clear air value but seldom to ten times this value. Therefore, the frequency distribution of aerosol optical depth is

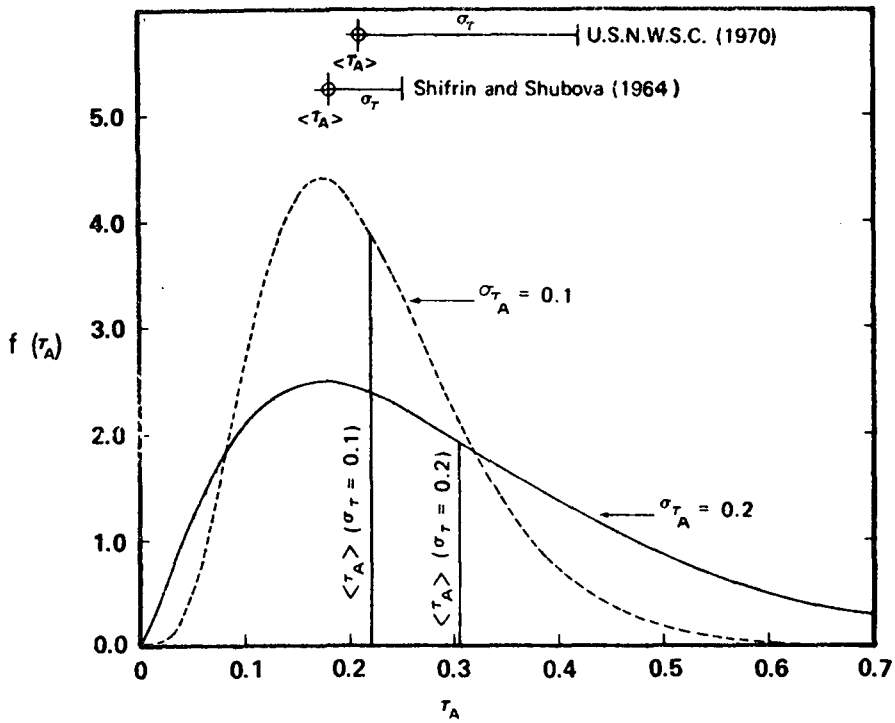

Figure 6. Assumed frequency distributions of aerosol optical depth (at $\lambda = 0.50\ \mu m$). Also indicated are mean values of aerosol optical depth and one standard deviation from the mean value. Visibility data as supplied by the U.S. Naval Weather Service Command were used to determine yearly mean values of the aerosol optical depth and its variability for the coastal area of the New England states. The scale height of the aerosol height distribution was assumed to be one kilometer.

assumed to be a skewed distribution toward increasing optical depth. The frequency distributions displayed in Figure 6 are gamma distributions with mode at 0.20 optical depth and standard deviations of 0.1 and 0.2 in optical depth. These distributions display the general features expected of the maritime optical depth. Also displayed in figure 6 are measured mean values of aerosol optical depth and standard deviations from these mean values.

Many of the measurements of ocean color will be made in regions of the ocean which are situated near the continents. These regions are known to be areas of upwelling and should be good locations for which the phytoplankton may grow and thus are prime candidates for the consideration of ocean color. However, the aerosol populations in these coastal regions are presumably a mixture of maritime and continental particle types and thus this mixture may follow the frequency distribution curve as displayed in figure 6 more closely than the particulate concentrations over more central regions of the oceans.

CALCULATIONS

The input information necessary for the calculations was: the molecular optical depth corresponding to the wavelengths of interest; the aerosol optical depth, τ_A; the solar zenith angle, θ_o; and the surface level albedo. Calculations of the transfer of radiation through the distribution of scatterers as defined by the model were made using a modified version of the computer subroutines for radiative transfer of Herman and Browning as described by Herman, et al. (1971). These routines numerically integrate the equation of transfer for the distribution of scatterers given. The results of the calculations were diffusely scattered radiances for a grid of zenith angles and azimuth angles with unit input solar irradiance. These radiances are easily renormalized to the measured solar irradiance incident at the top of the earth's atmosphere for the particular wavelengths of concern. The present study is concerned only with the radiances measured by a satellite looking in the nadir direction.

The surface level albedo was previously defined in terms of the measured radiance from a surface level and the total irradiance incident on that surface. As was further mentioned this albedo was measured experimentally by Ramsey (1968). The relative radiance measurements made by the aircraft borne spectrophotometer are normalized at ground level to the total irradiance. The incident irradiance is as described in equation (2). This albedo is quite different from the albedo as measured from satellite altitudes. The irradiance incident at the top of the atmosphere is the solar irradiance which is a plane parallel beam of radiation, having essentially one direction of propagation, and thus no diffuse component. Thus, the satellite borne radiometer must be continually calibrated with respect to the incident solar irradiance. The distinction between these two types of albedo is very important because it is noted that the surface level albedo contains information as to the transmissivity and scattering power of the atmosphere at the wavelength in question. This added information, which is a result of the dependence of the total irradiance at surface level on optical depth, is not available to the satellite borne instrument. In order to distinguish between the two types of albedo measures we will term the satellite measured albedo the geometric (this term is often used in astronomical contexts for exactly the same definition). Mathematically, the geometric albedo is defined relative to the nadir radiance as:

$$A_g(\lambda) = \frac{\pi I(\lambda)}{u_o F_o(\lambda)} \tag{5}$$

where μ_o is the cosine of the solar zenith angle and $F_o(\lambda)$ is the wavelength dependent solar irradiance measured at the top of the earth's atmosphere. The discussion of the results which are to follow will be in terms of the geometric albedo as defined above.

ANALYSIS OF THE CALCULATIONS

In considering the radiance as measured at satellite altitude at two wavelengths $\lambda = 0.46$ and $0.54\,\mu$m and forming ratios of the resulting geometric albedos a set of three parameters are fundamental in describing the atmospheric effects. The variation of the atmospheric effect may be described by:

θ_o - the solar zenith angle
τ_A (0.46) - the aerosol optical depth measured at 0.46μm
τ_A (0.54) - the aerosol optical depth measured at 0.54μm

The first factor may be determined in a straightforward manner from the geometry of the experiment. The second and third parameters are related to one another by the total columnar number of aerosols per unit normal area of surface and the size distribution of the aerosol particles in the column as noted in equation 3.

From the calculations it is of interest to determine the accuracy to which the color ratio can be determined as a result of natural mean level of aerosol optical depth and the variation about this mean. Further, with the assumption that by some process the aerosol optical depth can be determined to a given degree of accuracy, we are interested in how well the color ratio can be measured. Analysis of the theoretical calculations were performed with both of the above considerations in mind.

The minimum values of radiance at the two wavelengths of concern are the radiance values at zero aerosol optical depth. This is the case of air light contribution from a pure molecular atmosphere. Figure 7 presents the calculated radiance as viewed from a nadir looking satellite for various values of the solar zenith angle. The three curves correspond to the same surface level albedo and differing aerosol optical depths. The middle curve is for an average value of the aerosol optical depth, $\tau_A = 0.2$, and the uppermost curve is for an optically thick aerosol atmosphere. Because the contribution of radiation from the surface level is small, the radiance at the top of the atmosphere depends quite strongly on the total number of aerosol particles in an atmospheric column. This property of the scattered radiance or conversely the geometric albedo is more evident if we plot the geometric albedo versus aerosol optical depth. As was noted in equation 3 the total

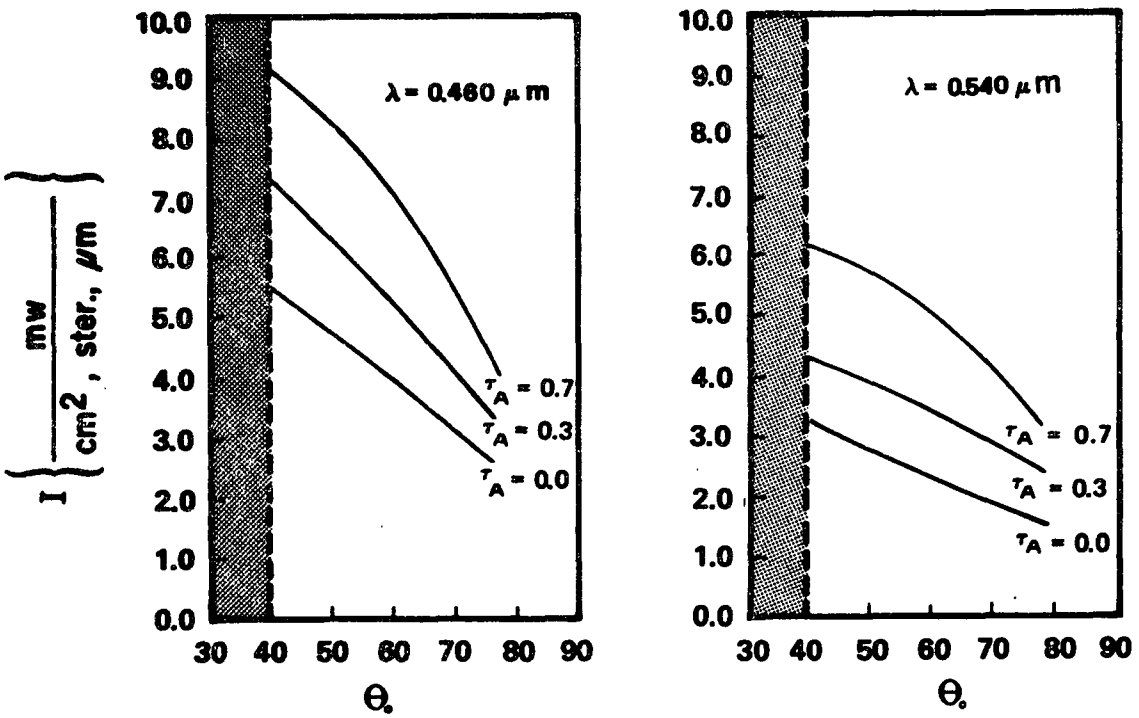

Figure 7. Representative calculated values for the nadir radiance at the top of the atmosphere. Calculations were made for solar zenith angles greater than 40° in order to avoid sun glint problems.

optical depth is a measure of the integrated number of particles in an atmospheric column. Figure 8 shows the relationship between the geometric albedo and the optical depth for various solar zenith angles for the two wavelengths of interest. For the calculations presented the surface level albedo was assumed to be zero in order to emphasize the effect of the atmosphere alone.

For the expected variation in the clear air optical depth as mentioned earlier in connection with figure 6, that is aerosol optical depths between 0.1 and 0.4, the geometric albedo is seen to vary between 0.05 and 0.21 for angles assumed to be typical for measurement. This variation in albedo is much larger than that expected for the ocean surface alone with no atmosphere present. The effect of the particulates in the atmosphere, therefore, causes a very strong obscuration of the ocean color features and this obscuration depends quite strongly upon the optical depth of the aerosol component of the atmosphere.

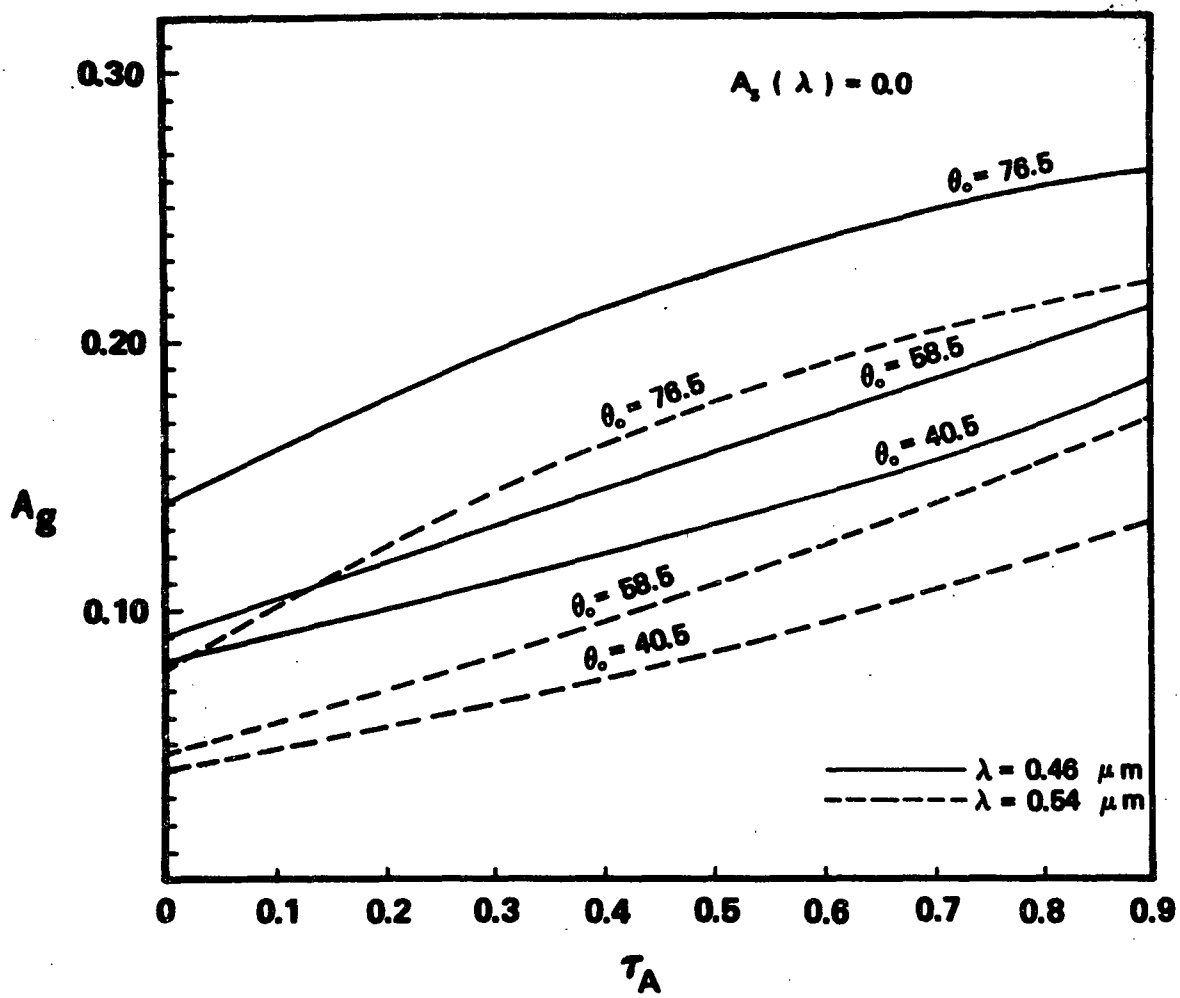

Figure 8. Geometric albedos as calculated at the top of atmosphere for nadir viewing angles. Three solar zenith angles are shown for the two wavelengths of interest. The surface level albedo equals zero, in this case, to emphasize the effects of the atmosphere.

POSSIBLE ACCURACY OF OCEAN MEASUREMENTS

If we can assume that the geometry of the ocean color measurement is known to a precision much greater than that of the other variables of this problem, and that the noise equivalent radiance of the satellite borne detector is such that the accuracy of the radiance measurement is much greater than the variations in radiance caused by changes in the optical properties of the atmosphere, then the accuracy to which the ocean color measurement can be made is determined solely by the optical depth of the atmosphere. The quantitative value to be used for ocean color is the color ratio which is defined as the ratio of albedo at two different wavelengths.

The ocean surface color ratio R_s was defined as:

$$R_s = \frac{A_s\ (0.54\ \mu m)}{A_s\ (0.46\ \mu m)} \tag{6}$$

where the subscript refers to surface quantities. The color ratio as measured at the top of the atmosphere R_t will be defined in terms of the geometric albedo $A_g\ (\lambda)$ at the top of the atmosphere as:

$$R_t = \frac{A_g\ (0.54\ \mu m)}{A_g\ (0.46\ \mu m)} \tag{7}$$

The geometric albedos are defined similar to (1) with the exception that the irradiance quantity appearing in the denominator of that expression is merely the incident solar irradiance at the top of the atmosphere. The wavelength dependent geometric albedo for the earth-atmosphere system is then defined as:

$$A_g\ (\lambda,\ \tau_T,\ A_s) = \frac{\pi\ I\ (\lambda,\ \tau_T,\ A_s)}{\mu_o\ F_o\ (\lambda)} \tag{8}$$

where $F_o\ (\lambda)$ is the wavelength dependent solar irradiance incident at the top of the atmosphere. For the calculations the total optical depth is measured relative to the total optical depth at $\lambda = 0.50\,\mu m$ which is midway between the two wavelengths of interest.

A transfer function may be considered which relates the surface level color ratio to the color ratio as measured at the top of the atmosphere. For the range of surface level albedos encountered and for ocean observations at the wavelengths of interest, this transfer function is quite linear. Assuming a linear variation in the surface level color ratio between the low and high values of chlorophyll concentration, the color ratio transfer function may be calculated from the atmospheric scattering models. Figures 9—11 show the transfer function for various values of total aerosol optical depth and solar zenith angle and constant size parameter. The obscuring effect of the large particle scatterers is evidenced by the decreasing slope of the transfer function with increasing aerosol optical depth.

The diminishing slope of the transfer function with increasing aerosol optical depth means that large variations in the surface level color ratio are

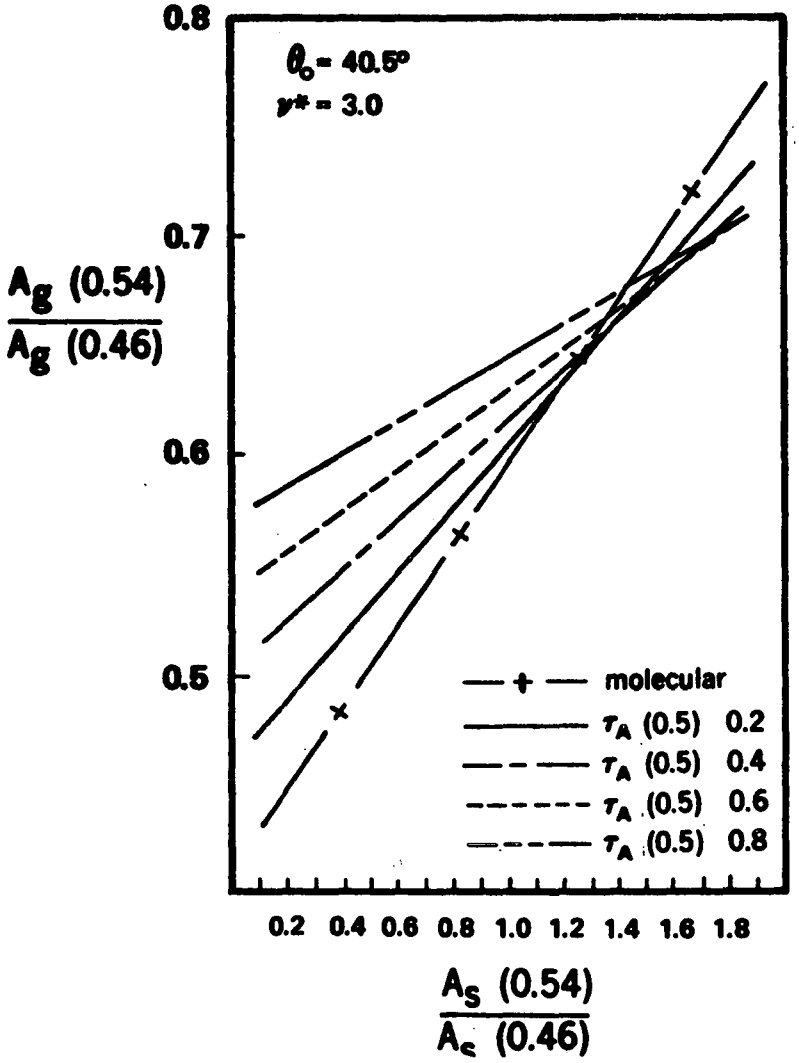

Figure 9. Relationship between surface level color ratio and color ratio at the top of the atmosphere for a solar zenith angle 40.5°.

measured at the top of the atmosphere as relatively small variations in color ratio. As the aerosol optical depth approaches very large values the slope of the transfer function approaches zero and thus in this limit the color ratio as measured at the top of the atmosphere contains no information as to the color ratio at surface level. This corresponds physically to the case of optically thick clouds whose opacity completely obscures vision of the surface level.

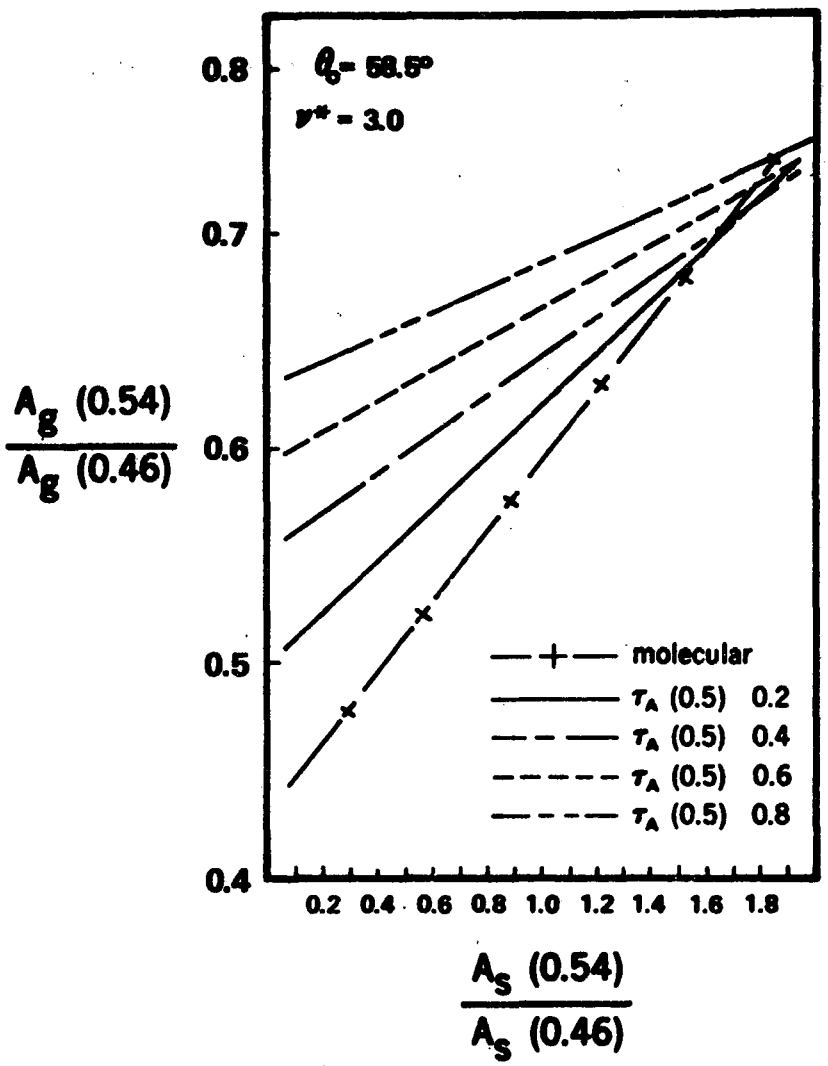

Figure 10. Relationship between surface level color ratio and color ratio at the top of the atmosphere for a solar zenith angle 58.5°

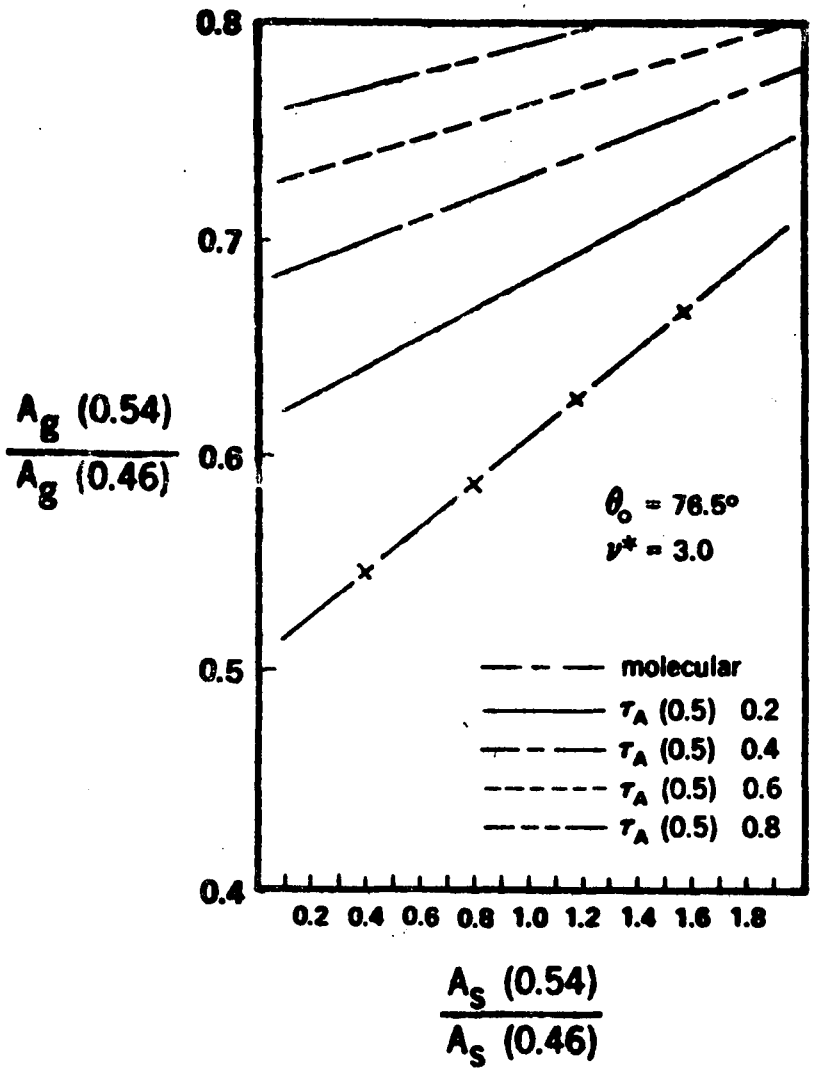

Figure 11. Relationship between surface level color ratio and color ratio at the top of the atmosphere for a solar zenith angle 76.5°

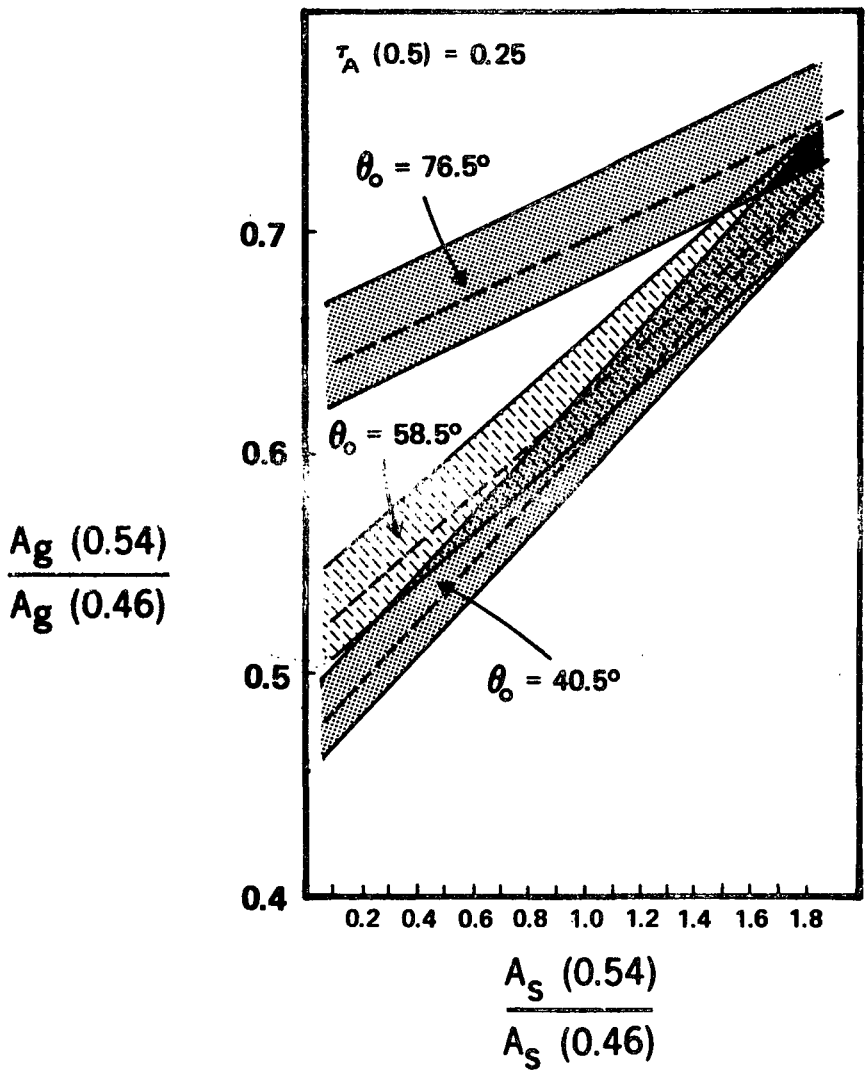

Figure 12. Variation in the transfer function relating the color ratio at the top of the atmosphere to that at surface level for varying values of the aerosol particle size distribution parameter. The shaded areas indicate the color ratio transfer function for size parameter values intermediate to the limiting values of $\nu^* = 2.0$ and $\nu^* = 4.0$.

The transfer function relating the color ratio at the top of the atmosphere to the color ratio at sea level is also dependent upon the size parameter in a manner as shown in figure 12. It may be seen in this figure that the slope of the transfer function does not vary with the size parameter. This is reasonable since the effect of the size parameter, at a given surface color ratio, is to alter the geometric albedos in such a way as to keep the aerosol optical depth at 0.50 μm a constant. Because these alterations to the albedos are small the effect upon the color ratio at the top of the atmosphere is quite linear.

If we assume that the solar zenith angle is known to any required accuracy then the unknown variable in the determination of ocean color is the aerosol optical depth and its wavelength dependence. A statistical average may be derived for the aerosol optical depth as a function of position on the earth's surface from available visibility data. Further, the possibility exists that measurements of geometric albedo at longer wavelengths, where the ocean is essentially black, may enable one to determine the mean aerosol optical depth at the position of a satellite color measurement. By obtaining an average value for this optical depth and knowing the solar zenith angle, a transfer function relating the color ratio at the top of the atmosphere to that at the surface level may be formed similar to those described in figures 9—11. Thus the measured color ratio at the top of the atmosphere may be related to the color ratio at sea level giving a mean value for the surface level color ratio.

The expected deviation of the aerosol optical depth from the mean value may be related to the expected deviation of the ocean color ratio from its mean value. These standard deviations may be related in the following manner:

$$\sigma_{R_s} = \left(\frac{\partial R_s}{\partial \tau_A}\right) \sigma_{\tau_A} \tag{9}$$

where σ_{R_s} is the uncertainty expressed as one standard deviation in the color ratio at the sea surface level and σ_{τ_A} is one standard deviation in the aerosol optical depth (the aerosol optical depth τ_A is here measured at 0.50 μm). Relating the partial derivative to quantities which are easily derived from calculations we may expand the partial derivative in the following manner:

$$\frac{\partial R_s}{\partial \tau_A\,(0.50)} = \left(\frac{\partial R_s}{\partial R_t}\right) \cdot \left(\frac{\partial R_t}{\partial \tau_A\,(0.50)}\right) \tag{10a}$$

$$= \left(\frac{\partial R_s}{\partial R_t}\right) \cdot \left[\frac{\partial R_t}{\partial \tau_A\,(0.46)} \quad \frac{\partial \tau_A\,(0.46)}{\partial \tau_A\,(0.50)}\right. \tag{10b}$$

$$\left. + \frac{\partial R_t}{\partial \tau_A\,(0.54)} \cdot \frac{\partial \tau_A\,(0.54)}{\partial \tau_A\,(0.50)}\right]$$

From the calculations made for a mixed aerosol-molecular atmosphere the partial derivatives of equation (10b) were evaluated as a function of several solar zenith angle, size parameter, (ν^*), and aerosol optical depth. The resultant uncertainties in the surface color ratio for differring accuracies in the determination of the aerosol optical depth are displayed in figures 13–15.

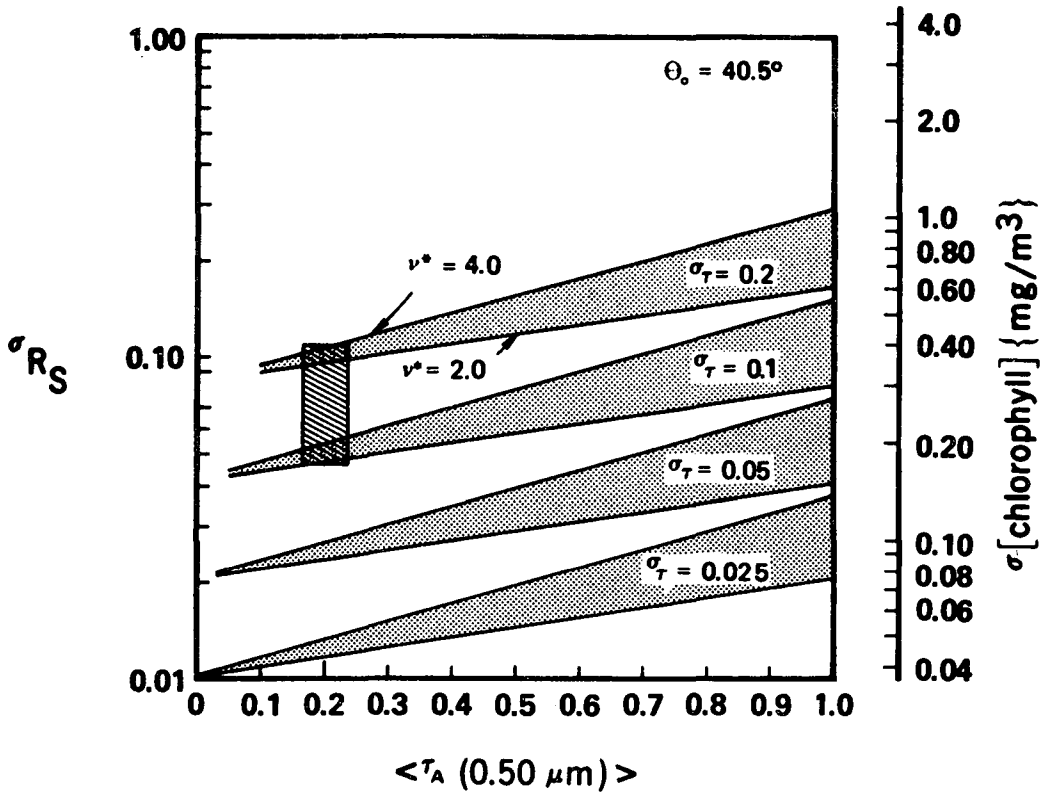

Figure 13. Relationship between one standard deviation error in surface color ratio and one standard deviation error in aerosol optical depth for differing mean values in aerosol optical depth. The shaded areas contain the error relationship between the size parameter $\nu^* = 2.0$ and $\nu^* = 4.0$. The region of the diagram indicated by the diagonal lines for the expected mean and standard deviations of the natural aerosol optical depth. This region indicates the expected uncertainty in the surface color ratio as measured from the top of the atmosphere for natural variation in particulates. The scale to the right uses the values of Figure 3 to relate the uncertainty of the surface color ratio to the uncertainty in chlorophyll concentration.

In general it may be noted that the determination of surface color ratio becomes less certain with both increasing solar zenith angle and increasing mean aerosol optical depth. For very large solar zenith angles the effects of the aerosol scatterers begin to dominate and for mean aerosol optical depths greater than 0.4 the rate of change of the color ratio at the surface to that at the top is almost constant. This tendency of the transfer function

for a solar zenith angle of 76.5° may be noted in figure 11. The rate of change of the color ratio at the top of the atmosphere to the rate of change of mean aerosol optical depth, in general, decreases to small values as the aerosol optical depth increases. When combined (first and second factor of the right hand side of equation 10a) these effects tend to decrease slightly with increasing mean aerosol optical depth as indicated in figure 11.

For the expected values of aerosol optical depth and derived standard deviations of the natural aerosol background level, a minimum level of uncertainty may be established for the determination of the surface level color ratio. The diagonally shaded areas in figure 13 through 15 indicate one standard deviation in the aerosol optical depth about a mean background level. If the relationship assumed between the surface level color ratio and

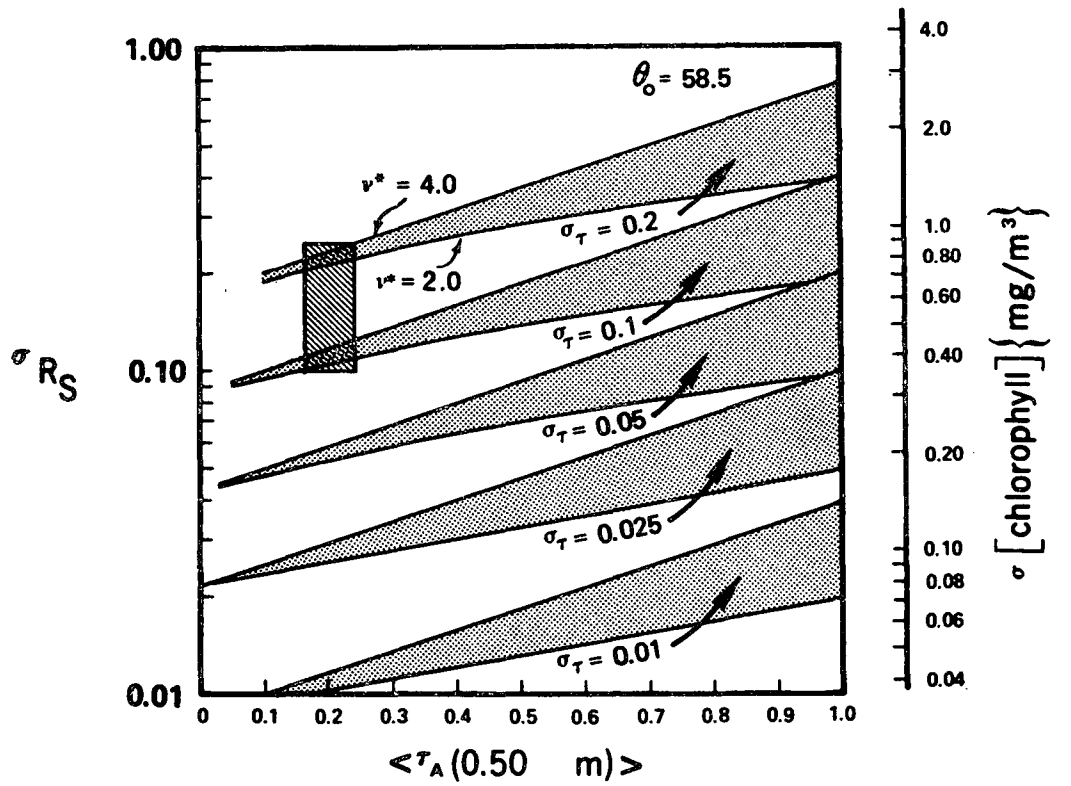

Figure 14. Relationship between one standard deviation error in surface color ratio and one standard deviation error in aerosol optical depth for differing mean values in aerosol optical depth. The shaded areas contain the error relationship between the size parameter $\nu^* = 2.0$ and $\nu^* = 4.0$. The region of the diagram indicated by the diagonal lines for the expected mean and standard deviations of the natural aerosol optical depth. This region indicates the expected uncertainty in the surface color ratio as measured from the top of the atmosphere for natural variation in particulates. The scale to the right uses the values of Figure 3 to relate the uncertainty of the surface color ratio to the uncertainty in cholorophyll concentration

chlorophyll concentration may be trusted then the chlorophyll concentration may be determined to within one standard deviation of from 0.5 to 2.5 milligrams per cubic meter, depending upon the solar zenith angle at the time of measurement and no prior knowledge of the optical depth of the atmosphere.

COMPARISON BETWEEN THEORETICAL AND SOME OCEAN COLOR MEASUREMENTS

To obtain representative values for spectral radiance at satellite altitudes, Hovis and Fraser (1971) have made ocean color measurements from high altitude aircraft above the Pacific Ocean in the vicinity of Los Angeles. These measurements of upwelling radiation were made at both 3000 and 49,500 feet

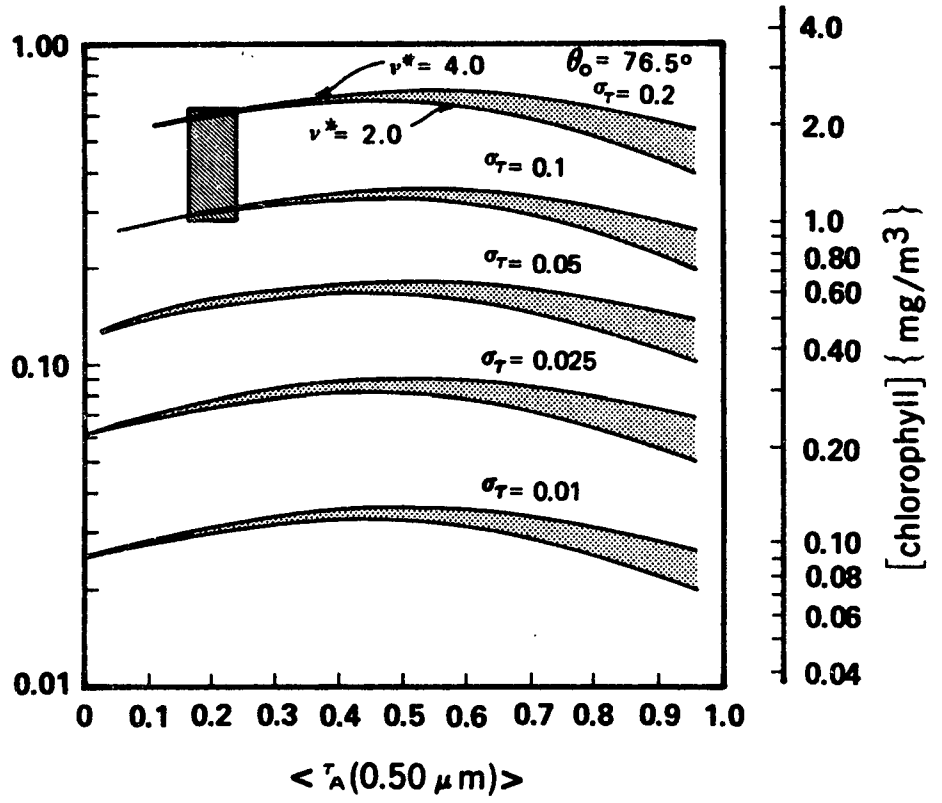

Figure 15. Relationship between one standard deviation error in surface color ratio and one standard deviation error in aerosol optical depth for differing mean values in aerosol optical depth. The shaded areas contain the error relationship between the size parameter $\nu^* = 2.0$ and $\nu^* = 4.0$. The region of the diagram indicated by the diagonal lines for the expected mean and standard deviations of the natural aerosol optical depth. This region indicates the expected uncertainty in the surface color ratio as measured from the top of the atmosphere for natural variation in particulates. The scale to the right uses the values of Figure 3 to relate the uncertainty of the surface color to the uncertainty in cholorophyll concentration.

altitude. At the same time as the ocean color measurements, ground truth data were collected consisting of the chlorophyll concentration and aerosol optical depth. In general, two values of chlorophyll concentration were encountered, and the aerosol optical depth measured at Catalina Island was estimated at a value of 0.08. Using the spectral radiance measured at 3000 feet a spectral surface albedo was calculated. The surface albedo at the wavelengths of interest for these calculations, together with the measured optical depth, was used to predict the nadir radiance at 49,500 feet.

The radiance prediction also included the small effect of ozone absorption in 0.54 μm channel and the correction to the solar irradiance due to the earth's orbital position. The values for the solar irradiance adopted for these calculations were those due to Thekaekara and Drummond (1971). Comparison between the measured and predicted nadir radiances for the two measured chlorophyll concentrations is shown in the top part of Table 1. The uncertainty introduced by the absolute calibration of the spectroradiometer is about 5 percent. The solar irradiance for the wavelengths given here differ by 5 to 10 percent depending upon which data is used (Thekaekara and Drummond (1971), Arveson, et al. (1969)). Although the theoretically predicted radiances at the top of the atmosphere are slightly larger than the measured values, the percent difference between the two is well within the error limits of the experiment.

Table 1

Comparison of theoretical and experimental values of Ocean Color

	(Chlorophyll) <0.01 mg/m^3		(Chlorophyll) $= 0.23$ mg/m^3	
λ	0.46 μm	0.54 μm	0.46 μm	0.54 μm
$A_s(\lambda)$	0.0218	0.00914	0.0180	0.0101
I (theor.)*	3.90	1.68	3.83	1.69
I (meas.)*	3.88	1.56	3.77	1.52
% difference	0.513	7.50	1.58	11.22
R_s	0.419		0.562	
R_t (meas.)	0.439		0.440	
R_t (theor.)	0.470		0.484	

*Units of radiance are given in mw/cm^2, ster., μ and are evaluated for 49,500 feet altitude.

In Table 1 are also given the color ratio at the ocean surface level and the theoretical and measured color ratios at the top of atmosphere. The correspondence between the measured and theoretical color ratios at the top of atmosphere is less than would be desired. The theoretical color ratios calculated for the top of the atmosphere are made at two purely monochromatic wavelengths and do not depend upon the solar spectral distribution. However, if the spectral resolution of the spectroradiometer measuring the ocean color is different than that used to measure the solar spectrum then the nadir radiance and solar irradiance are averaged over different wavelength intervals and thus may combine to give an erroneous value for the color ratio. The ocean color data used here were measured with a spectral resolution of 0.0015 μm and the solar irradiance used are given at a resolution of 0.01 μm. This disparity between the measured and calculated color ratios points out the fact that it is highly advantageous in the experimental arrangement to have the same instrument measure both the ocean color and the solar spectrum.

CONCLUSION

Clarke et al. (1970) have expressed the hope of measuring the chlorophyll concentration to an accuracy of 0.1 mg/m^3. If the assumptions which have been made in this work are valid then the aerosol optical depth must be measured to an accuracy of $\pm$ 0.01. It seems feasible to measure the optical depth to an accuracy of several hundredths using measurements made at several wavelengths longer than 0.50 μm, in conjunction with the albedo measurements made at the wavelengths near 0.50 μm. With a regression approach to this data it is hoped that the accuracy of determining the aerosol optical depth may be pushed to $\pm$ 0.01 at 0.50 μm.

The basic assumption between the color ratio and chlorophyll concentration to the ocean surface level is most dubious for small chlorophyll concentration. It may be imagined that the chlorophyll amount per individual organism varies both with the species of organism and the geographic location of the organism. Therefore, it is possible that the relation expressed in figure 2 may fluctuate with the variables affecting the biological organisms. These fluctuations will set a lower limit to the degree of accuracy to which the chlorophyll measurements can be made.

If the basic parameter is merely the color ratio of the ocean surface then models used appear to be on firm ground. Two criticisms of the theoretical models may be made concerning their relationship to the physical reality. The first is the possibility that the phase functions theoretically developed may not accurately reflect the actual phase functions often found

in nature. Experimental evidence of this fact has been presented by Barteneva (1960). The phase function directly affects the albedo measured at the top of the atmosphere and thus a comparison should be made between theoretical results and measured albedos.

The assumption of the Lambert reflecting surface may also be questioned. It is felt that this assumption may yield results close to reality for nadir looking radiometers and small solar zenith angles (avoiding specular reflection). A careful analysis of the radiation transfer between an ocean atmosphere system is necessary in order to test this assumption.

These calculations indicate that although the atmosphere does obscure the quantative measurement of ocean color on the average, useful measurements may be made at satellite altitude. Ocean color measurements may be used to indicate strong gradients in chlorophyll concentration. The location of chlorophyll rich and chlorophyll poor areas may be recorded on a synoptic scale to delineate times and positions of upwelling, nutrient rich water.

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