

**MULTIANGULAR HYPERSPECTRAL POLARIMETRIC OBSERVATIONS IN
COASTAL WATERS**

by

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Abstract

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The objectives of this study are: (1) investigate the geometrical and spectral properties of the underwater polarized light field, in various atmospheric conditions and water compositions, but focusing mostly on coastal regions. This requires the development of suitable in situ instrumentation, to validate existing polarized radiative transfer models. (2) Correlate polarized radiance to the properties of the water components, i.e. inherent optical properties (IOPs), concentrations, size distributions, refractive indices, etc.

With this work we set out to fill the gaps in previous studies to obtain a comprehensive understanding of the changes that occur in the polarized light in coastal and open ocean

waters. Unlike previous multiband instruments, we will obtain underwater angularly resolved hyperspectral measurements of the degree of polarization (DOP) in coastal and open ocean environments illustrating how the variability of the DOP is connected to water constituents. In particular, we will propose a method to estimate particles' composition from the DOP of the water – leaving radiance. Measurements were also taken in different atmospheric conditions, to observe the effects of diffuse light, rather than direct sun light on the DOP. Finally, we assess the consistency between our measurements and theoretical polarized radiative transfer models regarding the influence of marine particles on the polarized signal with particular emphasis on the corresponding wavelength and angular dependence.

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1. Introduction

Polarization characteristics of underwater light contain useful additional information on IOPs (Inherent Optical Properties), concentrations and size distributions of water constituents when compared with standard reflectance data [Chami et al. 2001], [Chowdhary et al. 2006], [Sabbah et al. 2005], [Gilerson et al. 2006], [Ahmed et al. 2004]. In particular, information on the state of the water constituents can be obtained through analysis of the spectral and geometrical angular dependence of the polarized light components. In addition, this analysis can help assess visibility in underwater environments and provide interpretation of ocean LIDAR signals. These properties should also be taken into account in the studies of atmospheric aerosols above ocean which employ polarization properties of atmospheric particulates [Mishchenko et al. 2007]. Remote sensing of the aerosols or clouds extensively uses the polarized light field. The radiance observed from space should be corrected by the water – leaving radiance (whose contribution varies substantially with the amount of suspended particles).

Although many measurements of light scattering in the seawater have been made, the majority of them have not taken in consideration the changes that occur in the linear polarization of the light field. Despite the importance of polarization for marine applications, relatively few *in situ* observations of the oceanic polarization state of light have been carried out, owing to a lack of instrumentation and to the practical difficulties in achieving reliable measurements. Watermann [1954] was the first to measure the polarized light field under water, using visual or photographic instruments operated by divers, and showed its dependence on the solar zenith angle and the viewing azimuthal

angle. Watermann's pioneering work was followed by Ivanoff *et al.* [1961] and Timofeeva [1961], whose studies, combining both laboratory measurements and underwater measurements in natural environments, highlighted the effects of various particulate properties (both organic and inorganic) on the polarization of the underwater light field.

Here we should probably spend some more time reviewing the work of the first scientists involved in the polarization measurements in the sea. Ivanoff [1974] was the first one to notice that both the refractive index and the particle size distribution have a great effect on the polarized backscattered light. He designed the first photoelectric polarimeter [1957] (whose main limitation was the fact that it was only working in horizontal line of sight) and pointed out [Ivanoff et al. 1961] that the scattering coefficient measured at 90° scattering angle, β_{90° and the degree of polarization (DOP), also measured at the same angle, P_{90° , are correlated. Specifically, when the scattering coefficient increases, the DOP generally decreases. However, this wasn't always the situation (at some specific depths, Ivanoff noticed that β_{90° and P_{90° increased simultaneously, suggesting the presence of strongly polarizing particles). Moreover, large variations of P_{90° for the same value of β_{90° (Fig. 1.1) suggested that the simple relationship $\beta - \text{DOP}$ needed more study.

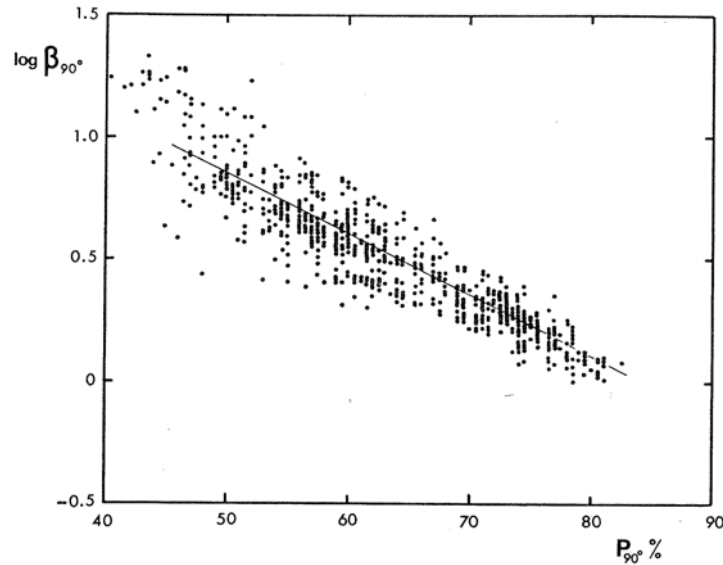


Fig. 1.1. Scattering coefficient at 90° vs. DOP at the same angle for waters near Bermuda [Ivanoff et al. 1961].

In the 1960's Timofeeva [Timofeeva 1962] performed several measurements both in the field and in the laboratory using milky solutions. Milk has a volume scattering function similar to that of seawater, but this doesn't necessary imply that the polarization properties are the same, in fact, as we mentioned before, the polarization properties of suspended particles cannot be explained only by means of their volume scattering functions. The real field situation is generally more complex than laboratory experiments. Timofeeva [Timofeeva 1961] analyzed the variations of the spatial distribution of the DOP with depth, considered equivalent to the optical thickness of a milky solution. Her studies revealed that the depth at which the DOP reaches its asymptotic value coincides with the depth at which asymptotic radiance distribution occurs. This means that (1) when the diffuseness of the submarine illumination increases, the DOP decreases and (2) the asymptotic condition is not only characterized by a stationary state of the radiance

distribution but also by a stationary angular distribution of the DOP. Not only has the value of the DOP changed with depth but also the angular position at which the maximum of the DOP occurs changes. The angle at which the DOP is maximal increases with depth and also reaches an asymptotic value (Fig. 1.2).

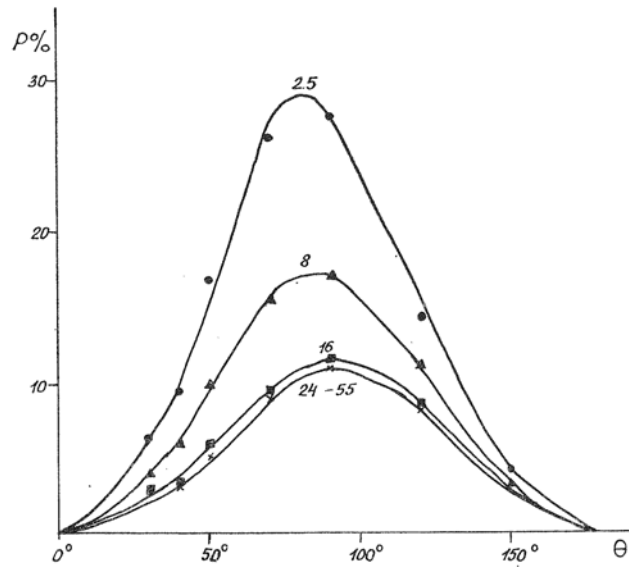
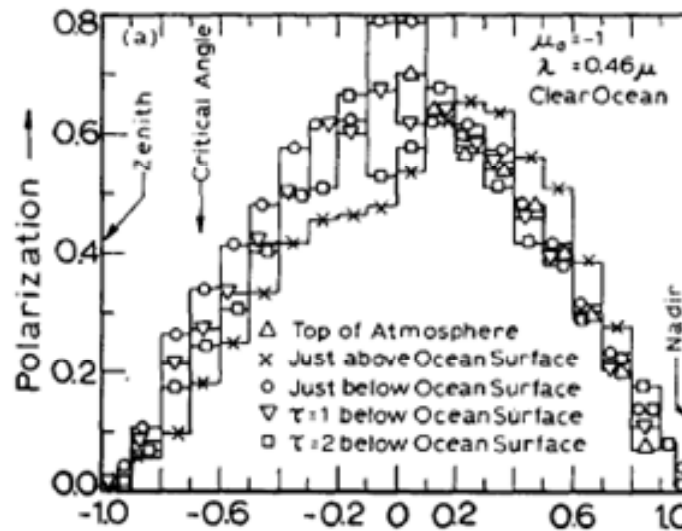


Fig. 1.2. Dependence of the DOP on the scattering angle and the optical depth of the milky medium [Timofeeva 1974].

In the following decades, a large amount of theoretical modeling of the underwater polarization was carried out. Much interest was focused on the calculation of Stokes vectors and Mueller matrices to determine the scattering properties of suspended particles and the effects of the air – water interface on the underwater polarized light [Kattawar et al. 1989], [Kattawar 1994]. Kattawar developed the first radiative transfer code for the earth's atmosphere – ocean system using the Monte Carlo method. In one of his first papers on the polarization of the ocean [Kattawar *et al.* 1973], he analyzed the angular

behavior of the DOP for a water body containing particles with refractive index $1.15 - 0.001j$, for a flat ocean surface. Fig 1.3 shows some of the results he obtained. Kattawar considered two situations: “Clear Ocean” and “Turbid Ocean”. Let σ_{ST} be the total scattering cross section and σ_{SR} the Rayleigh scattering cross section. If $\sigma_{SR} / \sigma_{ST}$ is the probability that a scattered photon undergoes Rayleigh (water) instead of Mie (hydrosol) scattering, $\sigma_{SR} / \sigma_{ST} = 0.218$ refers to the “Clear Ocean” situation and $\sigma_{SR} / \sigma_{ST} = 0.0454$ corresponds to the “Turbid Ocean” situation. The bottom panel of Fig. 1.3 shows the results for the “Turbid Ocean” situation. The most noticeable difference with the panel to the left is the peak at Brewster’s angle ($\mu = 0.599$). At this angle the reflected radiation is completely polarized and the contribution of the polarized reflected radiation increases with the turbidity. As we will find out, this is not exactly right.



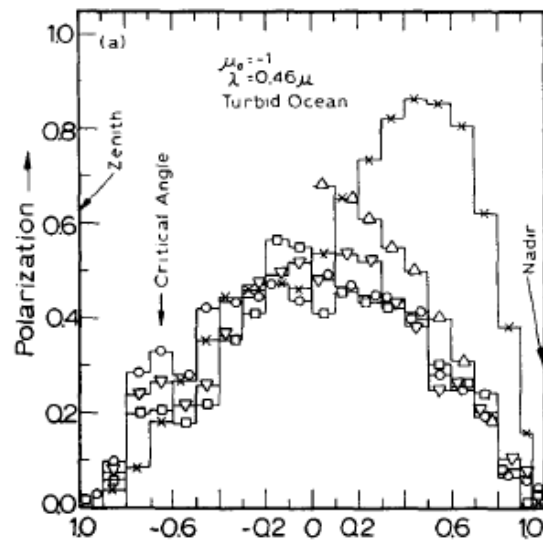


Fig. 1.3. Dependence of the DOP on $\mu = \cos\theta$ where θ is the viewing angle [Kattawar et al. 1973].

Noteworthy exceptions to the theoretical work are the experimental measurements obtained by Voss and Fry [1984], who developed their own electro – optic light scattering polarimeter to measure the Mueller matrices of water samples.

In the last 10 years, studies related to the underwater light field have been gradually emerging again. This definitely includes the work of Cronin and Shashar [Cronin et al. 2001], [Sabbah et al. 2005] who were the first to provide hyperspectral polarization measurements at various zenith and azimuthal viewing angles. Their investigations described the mechanisms of polarization vision in underwater environments focusing on visually mediated tasks performed by polarization sensitive animals. However, due to the fact that measurements were mainly taken in clear waters, the effect of suspended particles was not taken in consideration. Lerner *et al.* [2006], following the work of

Cronin and Shashar, reported polarization measurements not only in clear waters but also in presence of a phytoplankton bloom (moderately turbid waters). Reported data showed high instability of the percent polarization in the presence of Mie scatterers (as was the case of the phytoplankton bloom), see Fig. 1.4. The calculated values, while nicely reproducing the shape of the degree of linear polarization (DOP) in clear water conditions (Rayleigh scattering regime), didn't show the same instabilities. This is probably due to the fact that the instrument had to be manually adjusted to the desired azimuth and elevation positions therefore, it wasn't suitable for the rapidly changing underwater environment.

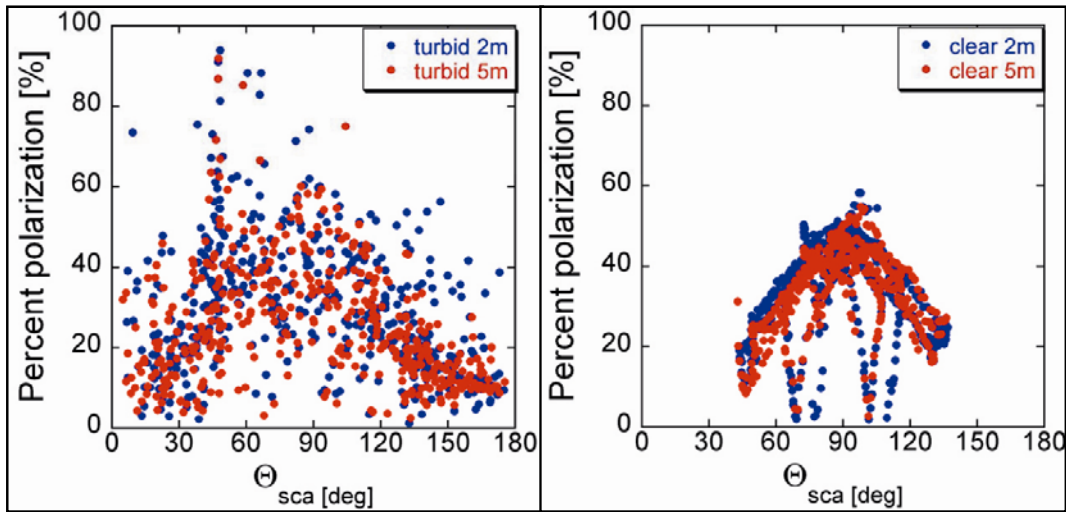


Fig. 1.4. Dependence of the DOP the scattering angle at two different depth, for two different loadings of suspended particles. Lerner mentioned a shift of the maximum DOP toward 100° , however this is not easily noticeable due to the high variability of the data points (left panel). The right panel shows, instead, the results corresponding to a clear water situation. As predicted by the Rayleigh theory of scattering, the maximum DOP is at 90° scattering angle [Lerner et al. 2006].

Remote sensing of ocean water provides information on suspended particles. Among the different types of suspended matter in ocean water, phytoplankton plays the primary role in global biological production in the ocean and, therefore, in the carbon cycle. Remote sensing measurements of ocean color are directly related to the water leaving spectral radiance which depends on the absorption and scattering properties of the suspended particles. The inverse problem of ocean color in coastal waters is ambiguous; in fact several combinations of IOP values can lead to a unique reflectance spectrum. The

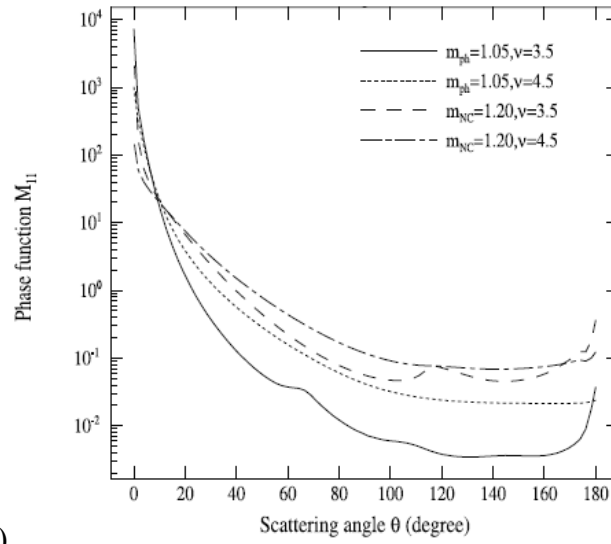
following equation is an example of relationship that is used as basis in many inversion algorithms [Defoin Platel et al. 2007]:

$$rrs = L_u / E_d = g \frac{b_{tot}}{a_{tot}}, \quad (1.1)$$

r_{rs} is the subsurface remote sensing reflectance, defined by the ratio of nadir – viewed upwelling radiance L_u (in $\text{W m}^{-2} \text{sr}^{-1}$) to downwelling irradiance E_d (in W m^{-2}) just beneath the sea surface, b_{tot} is the total backscattering coefficient (in m^{-1}), a_{tot} is the total absorption coefficient (in m^{-1}) and g is a proportionality factor (sr^{-1}). Because of their additive property, IOPs of each water component combined in different ways can lead to identical sum of IOPs and thus to similar reflectance values. Furthermore, a given value of the ratio bb_{tot} / a_{tot} can be obtained from different values of bb_{tot} and a_{tot} , thus also generating a similar reflectance value. For example the total absorption coefficient is given by the sum of the absorption coefficient of water, phytoplankton, NAP and CDOM. Different combinations of these coefficients can lead to the same a_{tot} .

Phytoplankton cells exhibit only weak polarization effects (because of the small index of refraction relative to water) while inorganic particles, which are strong backscatters, appreciably affect the polarization signal [Chami et al. 2001]. Having this in mind, Chami [2007a] investigated the influence of marine particles on the polarized radiation exiting the ocean. Using theoretical modeling (the OSOA radiative transfer code) he showed that an empirical – based inversion approach relying on the underwater polarized radiance could retrieve the concentration of inorganic particles regardless of the phytoplankton content in coastal waters. He also showed that the polarization of the underwater light field is very sensitive to the particle microphysics, in fact, as shown in Fig. 1.5, the

Mueller matrix element M_{11} (which represents the volume scattering function) and M_{12} (which, being normalized to M_{11} , represents the DOP) show pronounced differences for different refractive indices and size distributions of the particles. This suggests that polarization and multiangular views can help separate algal (phytoplankton) and nonalgal particles (NAP).



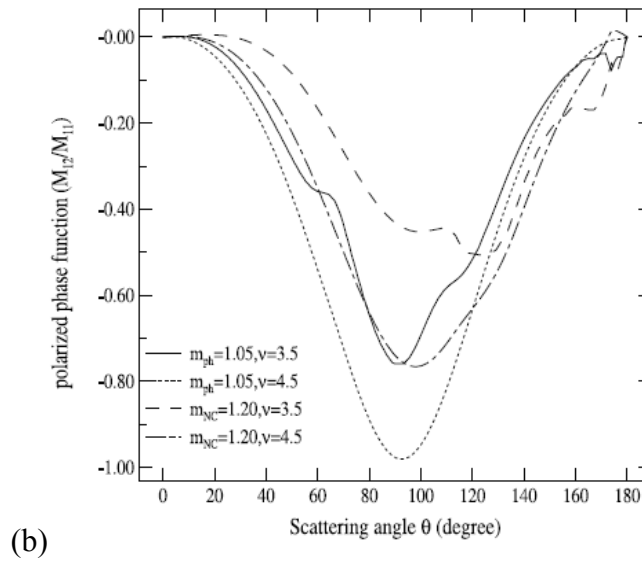


Fig. 1.5. (a) Particulate phase function (*i.e.* the element M_{11} of the Mueller matrix). (b) Polarized phase function normalized to the phase function (*i.e.* the ratio M_{12}/M_{11}). [Chami 2007a].

Chami *et al.* [2001] published these results in 2001 considering phytoplankton – like particles with refractive index equal to 1.05 and NAP with refractive index equal to $1.20 - 0.001j$. The ocean surface was modeled with a Cox – Munk wave slope distribution. Fig 1.6 shows one of the results obtained. He noticed that sediments can shift the position of the maximum of the DOP (from 90° to 100°) and change its maximal value. The DOP profile is, instead, insensitive to variations of the phytoplankton concentration (in water out of bloom conditions).

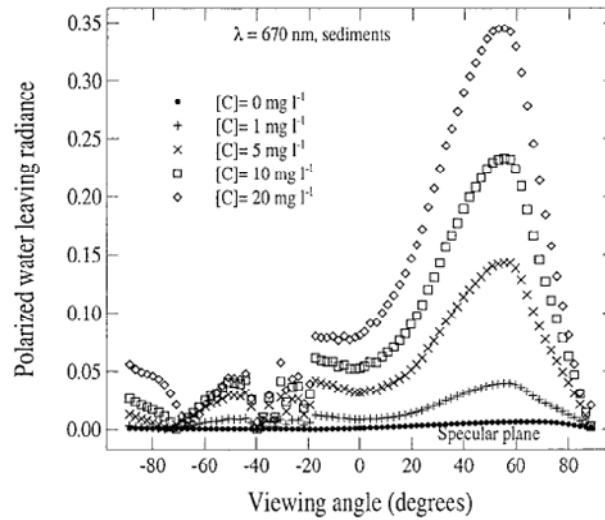
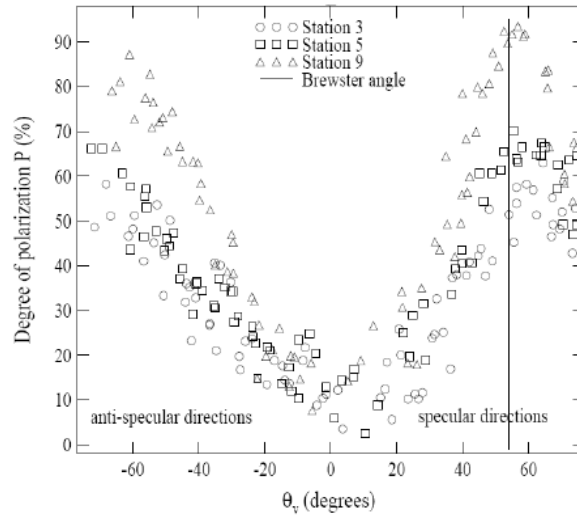


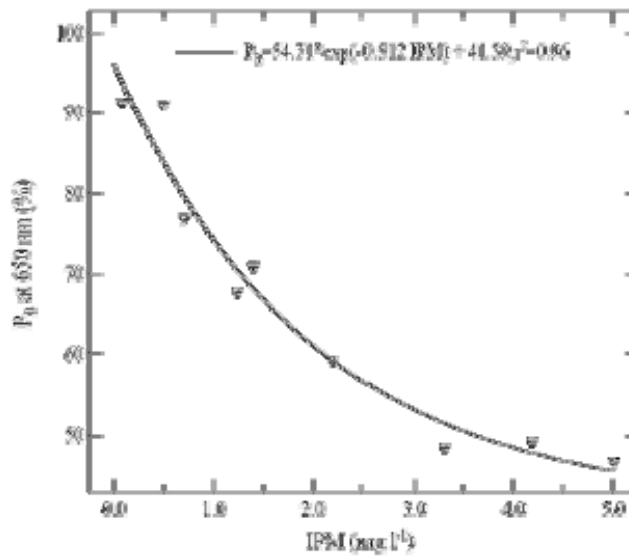
Fig. 1.6. Polarized water – leaving radiance corresponding to various concentrations of sediments. The observer position is just above the water. The solar angle is $\theta_0 = 30^\circ$ [Chami 2001].

On this basis, Chami et al. [2007b] also performed *in situ* measurements of the polarization state of underwater oceanic radiation with the purpose of having direct estimation of suspended inorganic matter concentration from remotely sensed data in coastal waters. REFPOL, the instrument used in these measurements, is a multispectral radiometer with only four channels centered at 450, 650, 850 and 1650nm together with polarizers which rotate in front of the detectors, allowing for successive (not simultaneous) measurements of radiance values. Fig. 1.7 shows the results of his efforts. The vertical line at $\approx 53^\circ$ indicates the position of the Brewster angle, at which the reflected light is completely polarized. Chami correlated the value of the measured DOP to the concentration of inorganic particulate matter (IPM), see Fig 1.7(b). Chami proposed that, when the turbidity increases, the vectorial sum of the light transmitted at

the water – air interface and the reflected skylight radiation reduces the polarization measured at the Brewster's angle. This approach, even if absolutely valid from a theoretical point of view, fails in real *in situ* situations, as we will illustrate in the following chapters.



(a)



(b)

Fig. 1.7. (a) Dependence of the DOP on the viewing angle. The instrumentation used, REFPOL, was positioned 10 m above the water surface. Results are

presented for a wavelength of 650 nm. (b) Dependence of the DOP measured at the Brewster angle on the concentration of inorganic particles [Chami 2007b].

2. Theoretical background

2.1 The Degree of Polarization

In this chapter we illustrate the needed definitions and geometrical angles and we discuss the principles of the relevant scattering processes. Fig. 2. 1 illustrates the sources of underwater polarization [Sabbah *et al.* 2005], *i.e.* (1) internal reflection, (2) single scattering, (3) multiple scattering, (4) refraction at the air – water interface, (5) reflection from the bottom, (6) forward scattering, and (7) attenuation of polarized light.

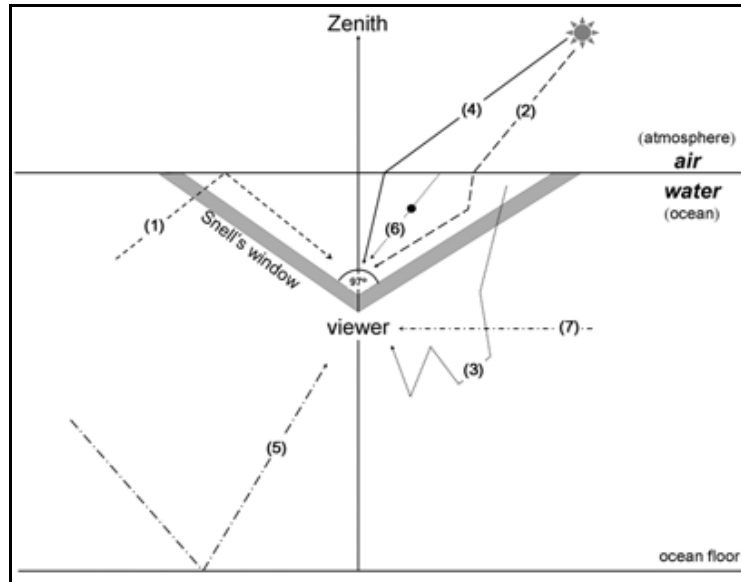


Fig. 2.1. Source of underwater polarization from [Sabbah *et al.* 2005]. The thick grey lines denote the borders of Snell's window (which is the circular region above an underwater observer with an aperture equal to twice the critical angle, $\theta_c \approx 48^\circ$).

The ocean contains many particles with dimensions on the order of the wavelength of sun light or smaller. These particles scatter the incoming sun light. Polarization by scattering is therefore the main source of underwater polarization. When a linearly polarized plane wave hits a particle, the electric field of the singly – scattered light and the oscillating dipole are coplanar (see Fig. 2.3). If the incident wave is unpolarized, it can be imagined as a superposition of two incoherent linearly polarized plane waves. In this situation (Fig. 2.3), the scattered light is also a superposition of the two phenomena shown in Fig. 2.2(a) and 2.2(b). The light scattered in the forward direction is completely unpolarized, while away from that axis is partially polarized and becomes more and more polarized as the angle increases. At 90° from the axis the scattered light is completely linearly polarized. As the concentration of the particles increases, some of the scattered light from one particle will probably hit a second particle and themselves be scattered again in all directions, and so forth. This situation is called multiple scattering.

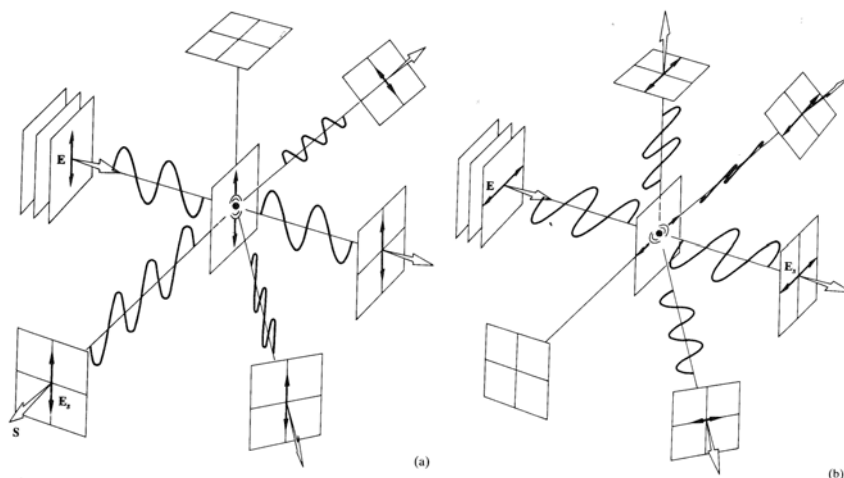


Fig. 2.2. Scattering of polarized light [Hect 1998]. This is valid for scattering by very small particles compared to the wavelength of light. Scattering by larger

particles with large refractive index (*e.g.* sediments in coastal waters) shows different DOP.

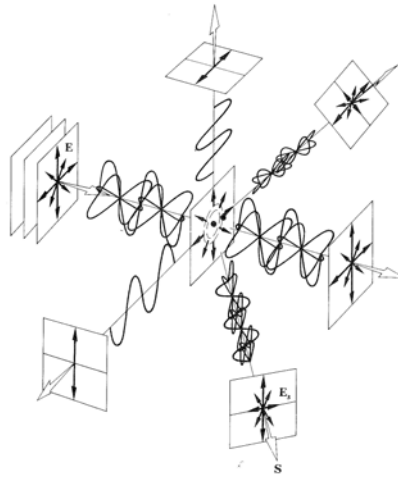


Fig. 2.3. Scattering of unpolarized light [Hecht 1998].

Every time light is scattered the percent of linear polarization at a certain angle (measured relatively to the incident unscattered light) might increase or decrease, depending on the specific angle. The light observed at this specific angle is a superposition of the light that has been singly scattered, doubly scattered, triply scattered, etc. The percent of linear polarization of this light, since it contains both high and low values of percent of linear polarization, is generally reduced from the singly scattered value in the same direction [Sabbah et al. 2005].

The light field just above the water surface is also polarized. Part of the observed polarization is due to the scattering in the underwater region and part is due to the phenomenon of polarization by surface reflection/transmission. Let's consider an incident

wave that is linearly polarized in the direction perpendicular to the plane of incidence (Fig. 2.4(a)). The dipoles of the dielectric surface (which would be the ocean surface in our situation) vibrate in a direction parallel to the electric field of the incoming wave and they reradiate. Note that dipoles do not radiate in the direction of their axes. In this situation both the refracted and reflected light are polarized in the direction perpendicular to the plane of incidence. If the incident wave is, instead, polarized in a direction parallel to the plane of incidence, something different happens. In this case, the intensity of the reflected light, $[E_r]_{||}$, is very low, since it makes a small angle, θ , with the dipole axis. In the case in which $\theta = 0$ (or, equivalently, the transmitted wave is perpendicular to the reflected wave, *i.e.* $\theta_t + \theta_r = 90^\circ$), the intensity of the reflected light is zero. This means that, for an incoming unpolarized wave, only the component whose polarization is perpendicular to the plane of incidence (and, therefore, parallel to the surface) will be reflected. This situation happens only for a particular angle of incidence (and, since $\theta_i = \theta_r$, only for one particular viewing angle), called the Brewster's angle, θ_B , which for the air – water interface is $\approx 53^\circ$.

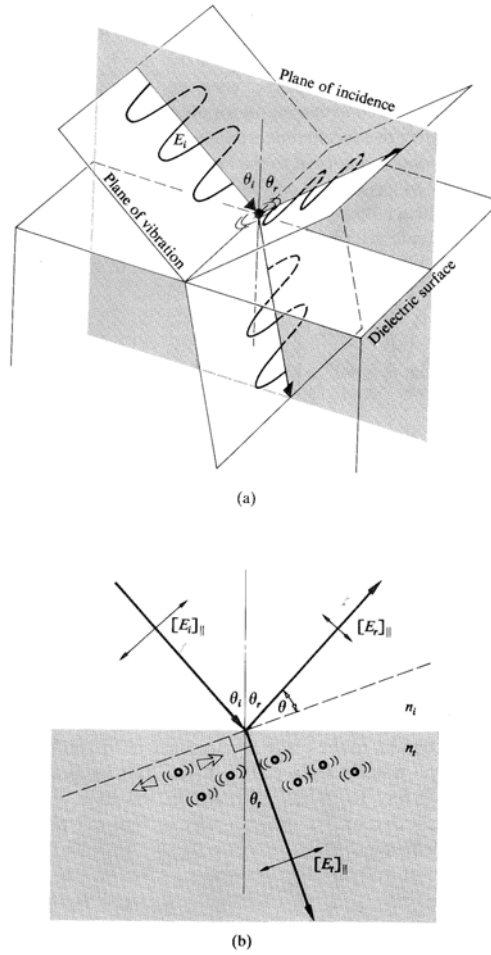


Fig 2.4. Reflection and refraction of light at the interface [Hecht 1998].

The polarization state of a light wave is quantified using the Stokes vector $S = [I \ Q \ U \ V]$, where I represents the energy flux ($\text{W m}^{-2} \text{ nm}^{-1}$), Q and U describe the linearly polarized component of this flux, while its circular polarized component is described through the Stokes parameter V . Natural sunlight is an example of completely unpolarized light and could be represented by $S = [I \ 0 \ 0 \ 0]$. The Stokes representation is advantageous as all elements possess the same units and are real and measurable quantities. Except for circular/elliptical polarization just outside the borders of Snell's window, for large zenith

angle of the sun and away from the main scattering plane [Ivanoff et al. 1958], the underwater light field is essentially linearly polarized [Ivanoff et al. 1961]. The DOP is a measure of the percentage of linear polarization and can be expressed as:

$$DOP = \frac{\sqrt{Q^2 + U^2}}{I}. \quad (2.1)$$

If we use three sensors, each of which has a polarizer in front of it, and the transmission axes of the polarizers are oriented at 90° (horizontal), 0° (vertical) and 45°. If I_{90} , I_0 , and I_{45} are the intensity values recorded by the sensors, then the Stokes components can be calculated as:

$$\begin{aligned} I &= I_{90} + I_0 \\ Q &= I_{90} - I_0 \\ U &= 2I_{45} - I \end{aligned} \quad (2.2)$$

If U is much smaller than Q , an alternative definition (which preserves the sign of Q) of the DOP is employed,

$$DOP = -\frac{Q}{I}, \quad (2.3)$$

When light interacts with matter, its polarization state changes. The Stokes vector of the incident light is transformed into the Stokes vector of the transmitted, scattered, reflected, or refracted light by multiplying it by a 4x4 matrix called the Mueller matrix:

$$\begin{bmatrix} I' \\ Q' \\ U' \\ V' \end{bmatrix} = \begin{bmatrix} M_{11} & M_{12} & M_{13} & M_{14} \\ M_{21} & M_{22} & M_{23} & M_{24} \\ M_{31} & M_{32} & M_{33} & M_{34} \\ M_{41} & M_{42} & M_{43} & M_{44} \end{bmatrix} \begin{bmatrix} I \\ Q \\ U \\ V \end{bmatrix} \text{ or } \mathbf{S}' = \mathbf{M} \cdot \mathbf{S}, \quad (2.4)$$

Where $\mathbf{S}$ is the Stokes vector of the incident light $\mathbf{S}'$ is the Stokes vector of the transmitted, scattered, reflected or refracted light. Oceanic water, especially in the coastal areas, contains many suspended particles (inorganic and organic particles, which include living and nonliving varieties and even air bubbles). Scattering processes clearly modify the polarization state and the angular features of the polarized light (*i.e.* their dependence on the scattering angle, the angle between the incoming light and the direction of observation), are strongly related to the size, shape and composition (*i.e.* refractive index) of their scattering sources.

When the scatterers are isotropic, homogeneous spheres, the scattering matrix reduces to four independent elements:

$$\begin{bmatrix} M_{11} & M_{12} & 0 & 0 \\ M_{12} & M_{22} & 0 & 0 \\ 0 & 0 & M_{33} & M_{34} \\ 0 & 0 & -M_{34} & M_{33} \end{bmatrix}, \quad (2.5)$$

Two important types of scattering for which this simplification is applicable are Rayleigh and Mie. Although non – spherical particles are frequently encountered in the ocean (*e.g.* phytoplankton), the scattering theory governing them is beyond the scope of this study. In order to conserve energy upon a scattering event, one component of the scattering matrix must be normalized. This constraint is expressed as:

$$\int_0^{4\pi} \frac{M_{11}}{4\pi} d\Omega = 1, \quad (2.6)$$

Where the component M_{11} , also known as the scattering phase function, relates scattered radiance to incident radiance. In the real atmosphere, Rayleigh scattering by molecules is

the dominant process affecting the observed polarization and the contribution of the aerosols can be considered as a first-order correction to the pure Rayleigh atmosphere [Coulson 1988]. The angular shape of the DOP for Rayleigh scattering is well known:

$$DOP = DOP_{MAX} \frac{\sin^2(\theta_{sca})}{1 + \cos^2(\theta_{sca})}, \quad (2.7)$$

where θ_{sca} is the scattering angle and DOP_{MAX} the maximum value of the DOP. In the case of single Rayleigh scattering by spherical particles, the DOP of scattered light is minimal (actually zero if the incoming beam is 100% unpolarized) at scattering angles of 0° and 180° . In contrast, the scattering in the orthogonal plane (90° and 270°) eliminates the in – plane polarization component so that the DOP is maximized with a value of 1 ($DOP_{MAX} = 1$). Multiple scattering, however, depolarizes the signal, causes a decrease of the DOP and the appearance of neutral points in the DOP distribution. However, in the ocean environment, scattering by water and suspended particles is the dominant process. Assuming the shape of each scattering sources in the ocean to be spherical, and the composition of each scattering source to be represented by a single value of the refractive index, the scattering properties of these particles can be computed relatively easily using Mie theory, assuming that the size distributions and the refractive indexes are known [Mie 1908]. In the case of Mie scattering, the DOP of scattered light is also minimal in the forward and backward directions; however, increasing concentrations of Chlorophyll affect the spectral behavior of the DOP (for specific spectral intervals, hence the necessity of a hyperspectral analysis), and multiple scattering effects, due to increasing concentrations of minerals, can shift the maximum of the DOP toward 100° scattering

angles [Chami et al. 2001]. Fig. 2.5 shows the relevant scattering and geometric angles we use in describing our measurements.

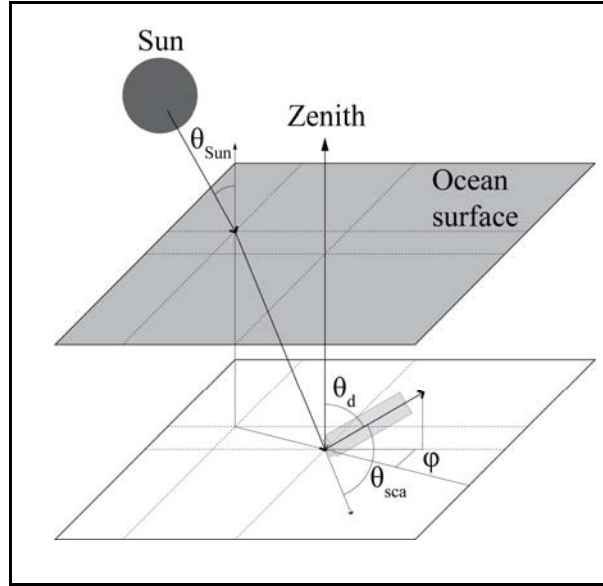


Fig. 2.5. Geometry of observation. θ_{Sun} is the solar zenith angle; θ_d is the detector, or viewing zenith angle; θ_{sca} is the scattering angle; φ is the detector azimuthal angle.

2.2 The e – vector orientation

According to the classic theory of scattering of light, the e – vector, *i.e.* the electric field of an electromagnetic wave, is normal to the scattering plane. Therefore, as long as a scattering plane can be identified (which, in ocean optics means when the sun is shining), in case of Rayleigh scattering, the e – vector orientation depends only on geometrical factors, and the optical properties of water are not considered. The e – vector orientation, χ , [Waterman 1954] relative to the horizontal direction (0 and 180° mean horizontal, $\pm 90^\circ$ mean vertical direction) is given by:

$$\tan(\chi) = \frac{\tan(\tilde{\theta}_{Sun}) \sin(\varphi)}{\sin(\theta_d) - \tan(\tilde{\theta}_{Sun}) \cos(\theta_d) \cos(\varphi)}, \quad (2.8)$$

Where $\tilde{\theta}_{Sun}$ is the solar zenith angle when viewed from underwater, θ_d is the detector, or viewing zenith angle, φ is the detector azimuthal angle (see Fig. 2.5). At noon, according to this model and as observed by Waterman [1954], in clear waters the e – vector orientation is horizontal at all zenith and azimuth viewing angles, (see Fig. 2.6). At sunrise, instead, the situation changes and the e – vector orientation depends both on the zenith and azimuth angles as can be seen in Fig. 2.6.

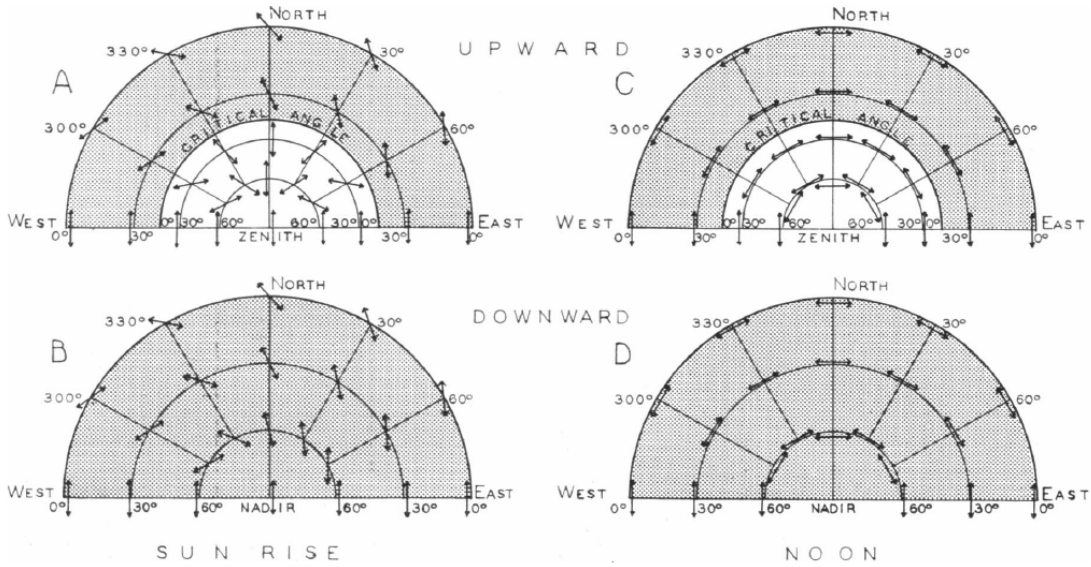


Fig. 2.6. The underwater e – vector orientation as it would be seen by an observer diving 2–3 m underwater. The diagrams are plotted on polar coordinates with azimuth angle represented by different radii, and zenith angles as concentric circles. The outermost circle (0°) in each case is the underwater horizon. A and C illustrate the polarization pattern when the observer is looking up, toward the zenith. When the critical angle is passed, the sky's polarization determines what

the underwater observer sees (unshaded areas in the A and C). The arrows indicate the e – vector tilt respect to the horizontal direction, the angle between an arrow and the radius that it intersects is the angle between the e – vector and the vertical direction [Waterman 1954].

Multiple scattering events alter the simple Rayleigh pattern predicted by Eq. (2.7) and makes possible any orientation of the e – vector. Kattawar *et al.* [Kattawar 1973], using a Monte Carlo radiative transfer code, showed that the orientation of the e – vector can be obtained from:

$$\tan(2\chi) = U/Q, \quad (2.9)$$

From the equation above, it can be seen that the e – vector in the main scattering plane ($\varphi = 0^\circ$) is identically equal to zero, since $U = 0$ in this plane [van de Hulst 1957].

3. Instruments and methods

3.1 The polarimeter

Here we describe the operating principles of the hyperspectral polarimeter. Polarization measurements were taken using a hyperspectral and multiangular polarimeter developed by the Optical Remote Sensing Laboratory at the City College of New York, NY. The instrument consists of three Satlantic Hyperspectral radiance sensors (recording intensity at the wavelengths 350 – 800 nm, 8.5° in-water field of view) mounted on a scanning system controlled by an underwater electric stepper motor as shown in Fig. 3.1. The heads of the sensors are positioned on the pivot point of the motor, in this way measurements are always taken at the same depth. The motor rotates the sensors in a vertical plane in a specific angular range which was adjusted according to the solar zenith angle in order to cover the full 0–180° range of scattering angles. Measurements were taken every 5°. Linear polarizers are attached in front of the sensors with orientations at 0° (vertical), 90° (horizontal) and 45°. From these measurements, the analysis of total radiance and DOP is obtained [Shashar et al. 2004]. Briefly, if I_0 , I_{90} , I_{45} are the intensity values recorded by the three Satlantic sensors, then the total intensity (which corresponds to the Stokes parameter I) is given by:

$$I = I_0 + I_{90} . \quad (3.1)$$

while the DOP is given by:

$$DOP = \frac{\sqrt{(I_0 - I_{90})^2 + (2I_{45} - I_{90} - I_0)^2}}{I} . \quad (3.2)$$

By rotating the sensors relative to the nadir direction, the instrument scanned the angular features of the underwater DOP in a vertical plane defined by its azimuth angle relative to the sun. The azimuth angular orientation of the instrument can be easily controlled by hand with an accuracy of 5° . The initial azimuth angular position usually corresponds to the principal plane, which is defined by the sun and the zenith. If measurements are taken in the principal plane (in which the Stokes parameter $U = 2I_{45} - I_{90} - I_0 = 0$), only the outputs of the sensors with polarizers oriented at 0° and 90° are needed for the calculation of the DOP and a simplified version of Eq. (3.2) can be used:

$$DOP = \frac{I_0 - I_{90}}{I}. \quad (3.3)$$

Data are acquired through a customized Labview program which automatically controls the rotation of the electric stepper motor synchronized with the data acquisition of the hyperspectral sensors. The method of data collection allows us to measure simultaneously and in real-time the hyperspectral radiance values recorded by all the sensors as the stepper motor rotates. The total angular sampling time was approximately 10–15 minutes depending on the integration time required to obtain noise-free measurements. Even though the duration of the total acquisition time should be decreased to reduce the strong variability in the optical properties of particles that can be found in highly dynamic coastal ecosystems, we didn't notice breaks in the recorded data, therefore we assumed no major changes in the water conditions during the sampling time. Typical values of the integration time ranged between 0.2 and 2.0 s per recording, depending on scattering angles and atmospheric conditions. Several recordings were taken for each angle to obtain the averaged value of the intensity.

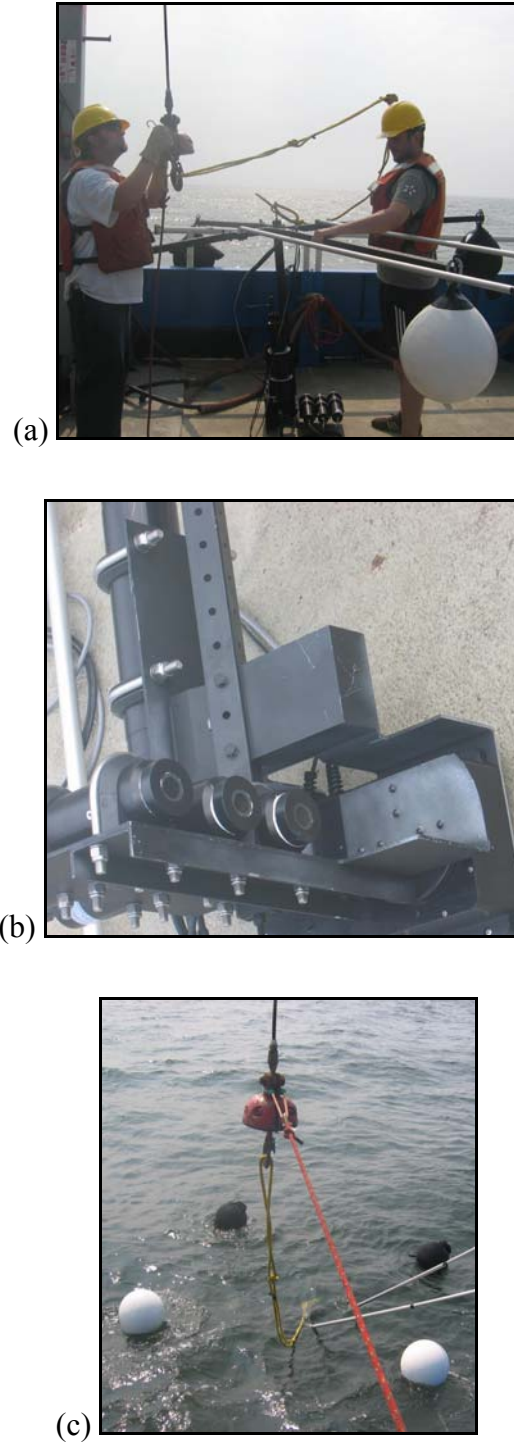


Fig. 3.1. The underwater instrument developed by the Optical Remote Sensing group at City College of New York. (a) The instrument on the deck of R/V “Connecticut” during a field trip in the coastal areas of New York Harbor –

Sandy Hook, NJ region (USA), on July 21–23 2008, (b) a detail of the Atlantic Hyperspectral sensors, (c) the instrument under water.

3.2 WET Labs AC – s

Absorption and attenuation of all in – water constituents except water were measured with an ac – s (WET Labs) The ac – s is composed of two pressure housings separated by a frame. The lower housing holds the source optics which include two incandescent lamps that couple into a single scanning filter spectrometer. Light output from the spectrometer is collimated and then propagates into the volume separating the two housings. Within this volume two continuous flow cells surround the optical paths. The absorption path is surrounded by a reflective quartz tube which constrains scattered light within the diameter of the tube. The attenuation path tube has blackened walls which absorbs scattered light from the attenuation beam path. Light reaching the end of the absorption tube is collected by a large area detector. Light reaching the end of the attenuation path propagates into the second receiver housing and is refocused upon a small aperture and detector. The receiver housing also holds the acquisition and control electronics for the sensor.

3.3 Sequoia Scientific LISST 100X

The LISST 100X is a multi – parameter system for *in situ* observations of particle size distribution. Scattering at 32 angles is the primary information that is recorded. This primary measurement is mathematically inverted to get the size distribution, and also scaled to obtain the volume scattering function (VSF). The size distribution is presented

as concentration ($\mu\text{l/l}$) in each of 32 log – spaced size bins. Optical transmission, water depth and temperature are recorded as supporting measurements. Fig. 3.2 shows the deployment of the LISST 100X attached to the cage of the WET Labs ac – s.

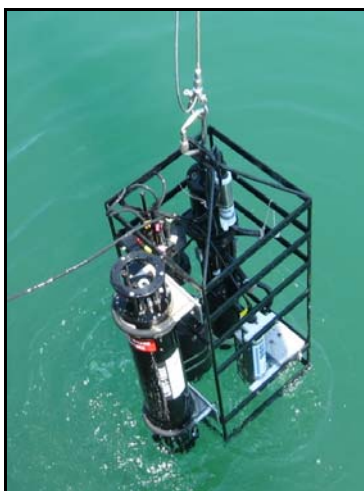


Fig. 3.2. The LISST 100X particle size analyzer attached to the cage of the WET Labs package. This picture was taken during a cruise in the Atlantic Ocean, approximately 20 – 25 miles away from the shore of Norfolk, VA (USA).

3.4 Bio – optical model

In ocean environments, particulate attenuation spectra $c_p(\lambda)$ can be very well approximated by a power law for a suspension of particles despite the fact that absorption spectra normally present peaks. This is because the peaks in an absorption spectrum are compensated by the inverse of such features in the scattering spectrum. Resulting c_p spectra ($a_p + b_p$) in oceanic waters therefore decrease monotonically with increasing wavelength following a hyperbolic model, $c_p(\lambda) = A_c \lambda^{-\gamma}$ where γ is the hyperbolic slope

(or shape) of the attenuation spectrum and A_c is the corresponding amplitude. Volz [1954] pointed out that the PSD and c_p are linked according to the simple relationship:

$$\gamma \approx \xi - 3. \quad (3.4)$$

Morel [1973] noted this relationship as a result of computations and Diehl and Haardt [1980] verified it using Mie theory and its approximations. Eq. (3.4) is exact for non-absorbing spherical particles, with constant index of refraction, assuming a Junge-type PSD and with diameters spanning from $r_{\min} = 0$ to $r_{\max} = \infty$. Boss *et al.* [2001] quantified the limits of applicability of Eq. (3.4) and proposed a correction to the above mentioned model, the correction is given by $\gamma = \xi - 3 + 0.5 \exp(-6 \gamma)$. In most oceanic situations $\xi > 3$ and the deviation between these two models is negligible. Boss *et al.* also showed that perturbations of the assumptions of classical theory lead to a deviation of O(2%). Maximum deviation was observed when including absorbing particles (many types of oceanic particles absorb light). Twardowski *et al.* [2001] pointed out that the only region where the $\gamma \approx \xi - 3$ relationship is maintained, regardless of changes in the complex refractive index, is for ξ values between 3.5 and 4.0.

In earlier studies [Bukata *et al.* 1988], a four component model is used to generate IOPs (Inherent Optical Properties) of the bulk water. Specifically, absorption, scattering and attenuation are described as:

$$\begin{aligned} a(\lambda) &= a_w(\lambda) + a_{phyto}(\lambda) + a_y(\lambda) + a_{NAP}(\lambda) \\ b(\lambda) &= b_w(\lambda) + b_{phyto}(\lambda) + b_{NAP}(\lambda) \\ c(\lambda) &= c_w(\lambda) + c_{phyto}(\lambda) + c_{NAP}(\lambda) + c_y(\lambda) \end{aligned} \quad (3.5)$$

CDOM (colored dissolved organic matter), or yellow substance, may have a local origin from the degradation of phytoplankton cells and other organic particles or may be

advected from a distant source. Fitting of the CDOM absorption spectrum, based on an exponentially decreasing function [Bricaud et al. 1995], is used to obtain the value of CDOM absorption at $\lambda_0 = 412$ nm, *i. e.* a_{y0} :

$$a_y = a_{y0} e^{-S_y(\lambda - \lambda_0)} . \quad (3.6)$$

Normalized absorption spectra of phytoplankton can be reconstructed with a linear combination of two spectra representing complementary contributions of the smallest (picoplankton) and largest (microplankton) cell size typically founded in phytoplankton communities [Ciotti et al. 2002], so that:

$$a_{phyto}^*(\lambda) = S_f a_{pico}^*(\lambda) + (1 - S_f) a_{micro}^*(\lambda), \quad (3.7)$$

where $a_{phyto}^*(\lambda) = a_{phyto}(\lambda) / [Chl]$ ($[Chl]$ is the concentration of Chlorophyll *a*, in $\mu\text{g/L}$), assuming that phytoplankton absorption is mainly due to Chlorophyll *a* absorption. Chlorophyll *a* is a photosynthetic pigment found in microscopic plants and phytoplankton that absorbs strongly in the blue part of the spectrum. Hence $[Chl]$ affects the color of the ocean and can be used as a proxy for the amount of phytoplankton biomass.

NAP represents all non – algal colored particulate matter, *i.e.* mainly mineral (terrigenous) particles, such as finely ground quartz sand, clay minerals and others, and detritus (biogenous) particles. Bacteria and viruses are excluded from the present investigation. Fitting of nonalgal particles is based on another exponentially decreasing function [Roesler *et al.* 1995], *i. e.*:

$$a_{NAP} = a_{NAP0} e^{-S_{nap}(\lambda - \lambda_0)}, \quad (3.8)$$

where $a_{NAP0} = 0.05$ [NAP].

The absorption spectrum of all particulate components can be obtained from the total absorption spectrum, recorded with the ac – s, or the ac – 9, a previous multispectral version of the ac – s (WET Labs), once the absorption due to dissolved matter (*i. e.* CDOM) is subtracted. Concentrations of Chlorophyll and nonalgal particles can then be obtained by fitting the total absorption spectrum with (3.7) + (3.8).

The attenuation spectrum, as stated above, can be fitted with the following equation: $c_p(\lambda) = A_c \lambda^{-\gamma}$ where γ is the hyperbolic slope (or shape) of the attenuation spectrum and A_c is the corresponding amplitude. The only piece of information of the attenuation spectrum that will be used is its slope, *i.e.* γ .

In situ measurements have shown b_p to be systematically correlated with [Chl], the concentration of Chlorophyll *a*, and [NAP], the concentration of non – algal particles. Following Bukata *et al.* [1988], fitting of the scattering spectrum, b_{Phyto} of phytoplankton is based on the following equation:

$$\begin{aligned} b_{\text{phyto}} &= c_{\text{phyto}} - a_{\text{phyto}} \\ c_{\text{phyto}} &= 0.3[\text{Chl}]^{\gamma} \left(\frac{550\text{nm}}{\lambda} \right)^{\gamma_1}, \end{aligned} \quad (3.8a)$$

Following Mobley *et al.* [1993], scattering of NAP can be modeled as:

$$b_{\text{NAP}} = b_{\text{NAP}}^* [\text{NAP}] \left(\frac{550\text{nm}}{\lambda} \right)^{\gamma_2}, \quad (3.8b)$$

The scattering spectrum of all particulate components can be obtained from the total attenuation spectrum, recorded with the ac – 9 (WET Labs), once the total absorption spectrum is subtracted.

3.5 The Adding – Doubling radiative transfer code

Experimental results were also compared with the results of a vector radiative transfer model [Chowdhary et al. 2006]. It's a coupled atmosphere – ocean polarized radiative transfer model which uses the adding – doubling method to solve the radiative transfer equation. The simulations are based on a four – component water model: water itself, CDOM, phytoplankton-like particles and non-algal-like particles.

The atmosphere – ocean system used in these computations is specified as follows. The atmosphere is divided into two purely molecular layers; the molecular layer at the bottom has an optical thickness a thousand times smaller the top layer. The optical thickness of the upper layer is 0.13 at 510 nm. The observer is positioned above the bottom layer, to simulate a point of view just above the water surface. For the ocean surface we adopt the wind speed and wind direction dependent slope distribution measured by Cox et al. [1954]. In these calculations we didn't include shadowing effects, wind directionality and scattering by oceanic foam, however, they can be easily added. The ocean body is assumed to be a homogeneous mixture of pure seawater and particulate components. The effect of CDOM is also taken into consideration.

The scattering properties of a mixture of phytoplankton and mineral particles are calculated according to Mie theory. The theory assumes the shape of each particle to be spherical, and the composition of each particle to be represented by a single value of refractive index. The results are then used as input for the radiative transfer part of the model, which provides the Stokes parameters for the upwelling radiance of the water body specified in the input file.

Scattering by clean oceanic water is modeled as Rayleigh scattering with a depolarization factor of 0.09 [Morel 1974] and with the scattering and absorption coefficients given by Hale and Querry [1972]. A Junge size distribution was assumed with the slope calculated using equation (3.4) for both types of particles. Values for $r_{\min}$ and $r_{\max}$, the minimum and maximum particle sizes, were 0.1 and 50 μm , respectively. CDOM absorption at $\lambda_0 = 412$ nm, and concentrations of Chlorophyll and nonalgal particles were obtained by fitting *in situ* measurements (WET Labs), as described in the bio – optical model. The number fractions of suspended particles, proportional to the number of particles per unit volume, were obtained using the particle compositions of Stramski *et al.* [2001]. The scattering properties of the suspended particles were also obtained from *in situ* measurements (WET Labs), as described in the bio – optical model.

The real part of the refractive index, relative to water, of phytoplankton was kept constant at 1.04, while the real part of the refractive index of NAP was varied until the modeled DOP matched the measured one. The refractive index of NAP is, indeed the only unknown and depends very much on the specific location. Refractive indices of minerals commonly found in the ocean range from 1.14 to 1.26. Amorphous silica, used for diatom exostructures, is the exception with a refractive index of 1.08. Away from river inputs, most nonalgal particles are derived from atmospheric deposition. The bulk refractive index of minerals in aerosols might therefore be close to what would be expected for minerals found in the ocean. The bulk refractive index of NAP is the average of the individual indices of refraction of all the particles that are not algal particles weighted by their scattering cross section.

4. *In situ* under water measurements in Chlorophyll – rich waters

In this chapter we describe the first test of the underwater polarimeter. We exploited all the features of our instrumentation (*i.e.* the possibility of measuring at different azimuth angles and at different depths) during its initial testing in Chlorophyll rich waters. The site under investigation is the marina next to Kingsborough College, NY (USA) as it can be seen in Fig. 4.1. Fig. 4.2 shows the absorption spectra of all water components. By fitting the absorption spectra (for both particulate and dissolved matter) with the equations of the bio – optical model previously described, we obtained that $[Chl] = 56.2 \mu\text{g/L}$, $[NAP] = 8.9 \text{ mg/L}$ and $a_{CDOM}(400 \text{ nm}) = 0.652 \text{ m}^{-1}$.

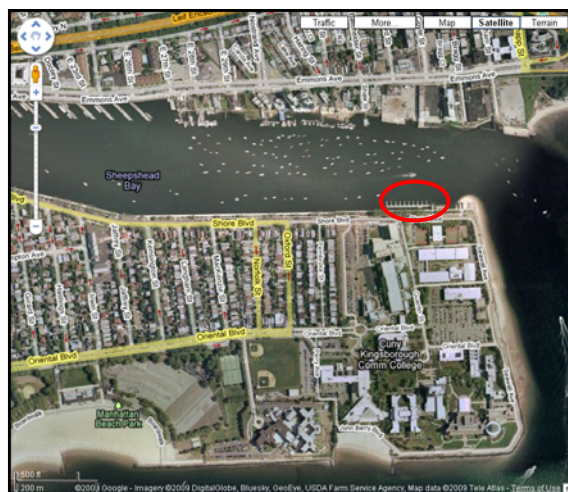


Fig. 4.1. The location used as test for the underwater polarimeter. It corresponds to the marina of Kingsborough College, NY (USA).

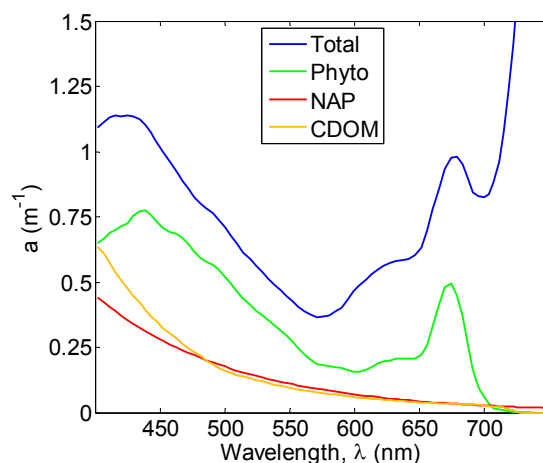


Fig. 4.2. Absorption spectra. The concentration of phytoplankton was very high in this site, as it can be seen from the high absorption peak in the red region of the spectrum.

DOP measurements were taken at two different depths and at various azimuth angles. Fig. 4.3 shows the DOP plots at 1 m depth. What can be seen from the plots is that the DOP seems to be insensitive to small variation of the azimuthal angle when moving away from the main scattering plane. Features clearly noticeable in the plot are the peaks close to 0° scattering angle. We initially thought they were artifacts; however, since they appear always in the same angular position and they disappear when the polarimeter is lowered at two meters depth, it is more likely that they are due to surface effects. Further investigation is still required to clarify this aspect.

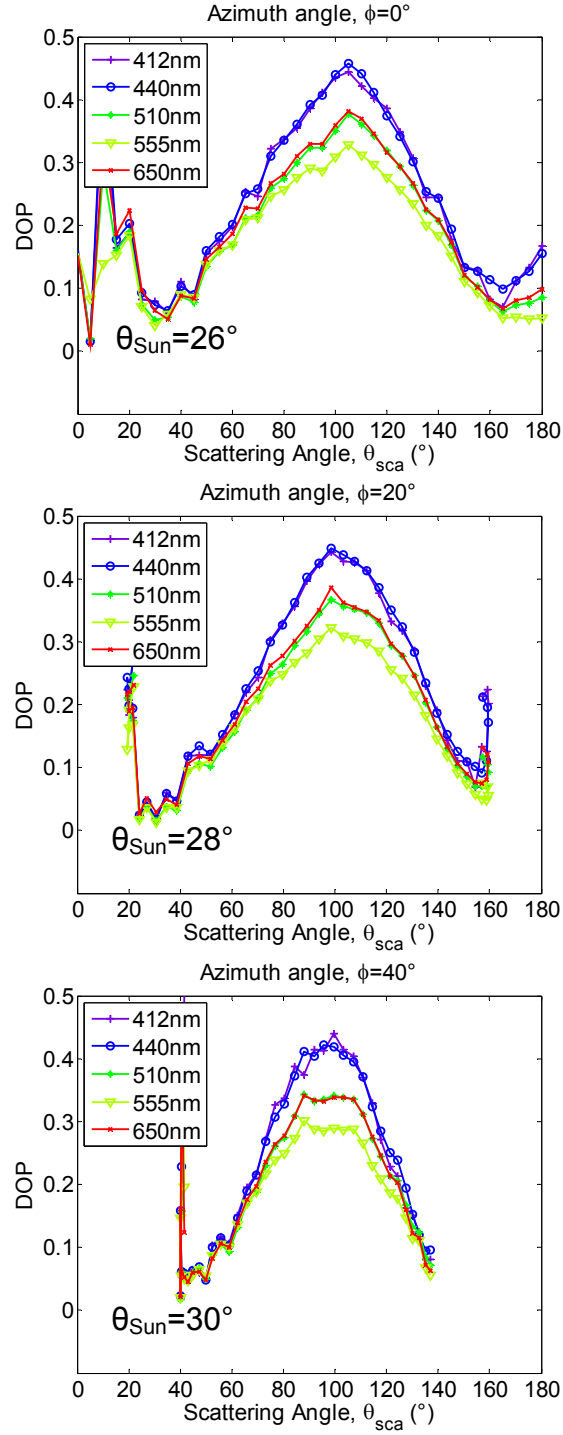


Fig. 4.3. Plots of the DOP vs. scattering angle for three different azimuthal angles. The polarimeter is positioned 1 m below the surface. The scattering angular range for the plots corresponding to measurements away

from the main scattering plane is smaller when compared with the ones recorded in the main plane.

Two meters below the surface the situation changes, as it can be seen in Fig. 4.4. In the main scattering plane, the position of the maximum of the DOP shifts toward 120° and there is an overall decrease of the DOP of about 30% at all wavelengths. At 40° away from the main scattering plane, the position of the maximum is around 90° , the curves are noisier and the wavelength dependence is reduced.

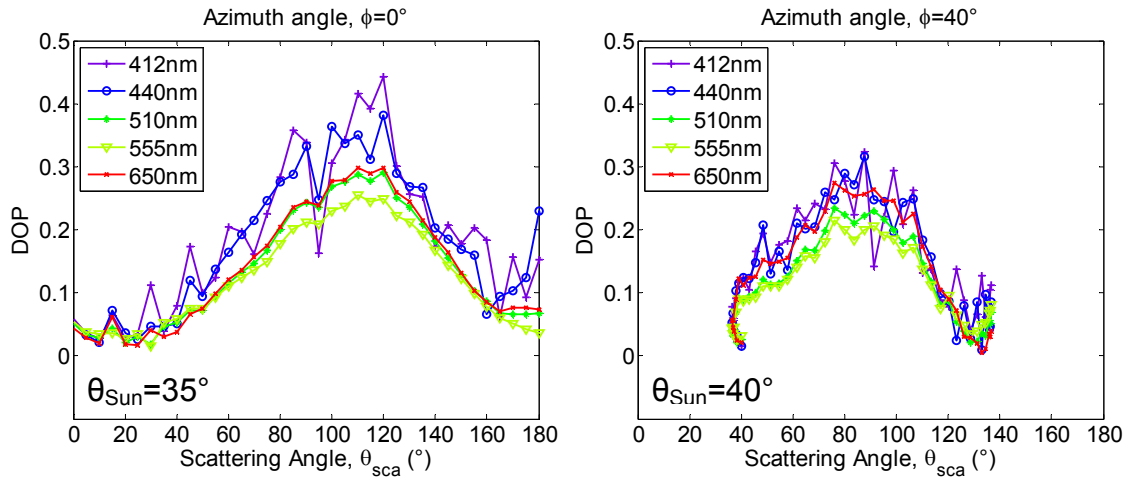


Fig. 4.4. Plots of the DOP vs. scattering angle for two different azimuthal angles. The polarimeter is positioned 2 m below the surface.

Fig. 4.5 shows the spectral dependence of the DOP. The pronounced dip in the red region of the spectrum, due to the fluorescence emission, highlights the high concentration of phytoplankton in this water. We will get back to the effects of fluorescence on the DOP later in the text.

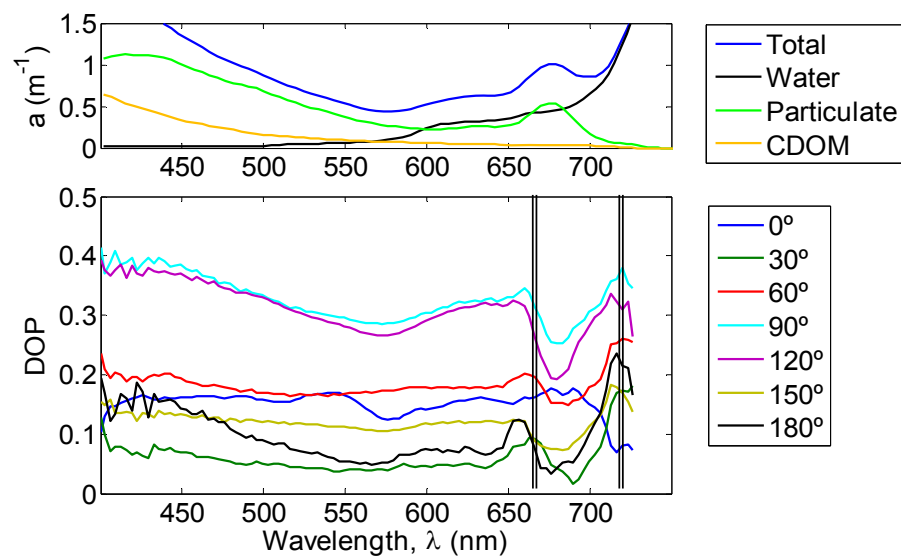


Fig. 4.5. Spectral dependence of the DOP for few relevant scattering angles. The absorption spectra of the various water constituents are also shown. The polarimeter is positioned 1 m below the surface, in the main scattering plane.

5. *In situ* above water measurements and comparison with adding – doubling Radiative transfer model

In this section we describe the first set of above – water measurements of the DOP. The purpose of this effort was to recreate the field experiment described in Chami *et al.* [2007b] in order to have a better understanding of the physics involved. Shipborne above – water angularly resolved measurements of the DOP were taken in the Hudson River on September, 26 2007 (between 11AM and 2PM). A linear polarizer was positioned in front of a hyperspectral radiometer (HyperOCR, Satlantic), and alternatively switched from the vertical to the horizontal position. The hyperspectral radiometer scanned continuously in the range 350 – 800 nm and it was mounted on a scanning apparatus which was rotated up to 70° (5° steps), either side of nadir viewing, in the main scattering plane (*i.e.* $\phi = 0^\circ$), see Fig. 5.1. The scanning system was mounted at the extremity of the back deck of the boat (approximately 1 m above water, yielding a footprint of approximately 0.2 m). Simultaneous measurements of downwelling irradiance (E_d) were also taken for normalization purposes.



Fig 5.1. The Satlantic HyperOCR sensor on the scanning apparatus used for above water measurements.

The physical quantity experimentally measured is the remote sensing reflectance (R_{rs}) which is defined as:

$$R_{rs}(\theta, \phi, \lambda) = \frac{L_w(\theta, \phi, \lambda)}{E_d(\lambda)}, \quad (5.1)$$

where θ and ϕ are the viewing zenith and azimuth angles, respectively, λ is the wavelength, L_w is the water-leaving spectral radiance and E_d is the downwelling spectral irradiance incident onto the sea surface. However the detector measures L_t , total radiance, which is given by the sum of L_w and L_r , the reflected sky radiance, *i.e.* $L_t(\theta, \phi, \lambda) = L_r(\theta, \phi, \lambda) + L_w(\theta, \phi, \lambda)$. Using the physical quantities experimentally retrieved, we can define the DOP as follows:

$$DOP(\theta, \phi, \lambda) = \frac{R_{rs}^H(\theta, \phi, \lambda) - R_{rs}^V(\theta, \phi, \lambda)}{R_{rs}^H(\theta, \phi, \lambda) + R_{rs}^V(\theta, \phi, \lambda)}, \quad (5.2)$$

Where R_{rs}^H is the above water remote sensing reflectance measured when the polarizer, positioned in front of the sensor, was in the horizontal position and R_{rs}^V is the above water remote sensing reflectance measured when the polarizer, positioned in front of the sensor, was in the vertical position. Here it should be noticed that having the possibility to measure only two directions of polarization, our viewing geometry is restricted to the principal plane. In this plane it is legit to calculate the DOP with the above equation.

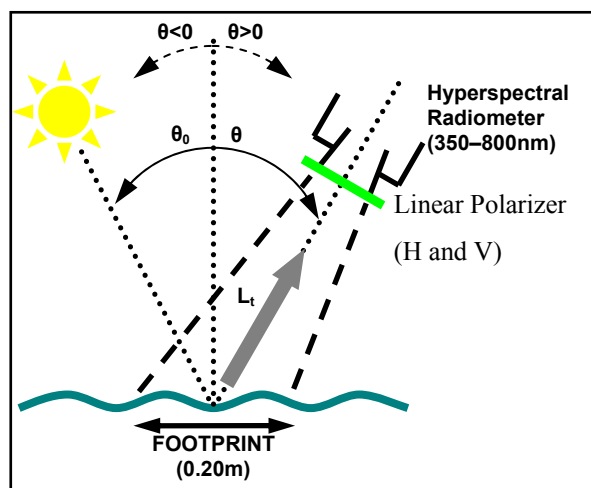


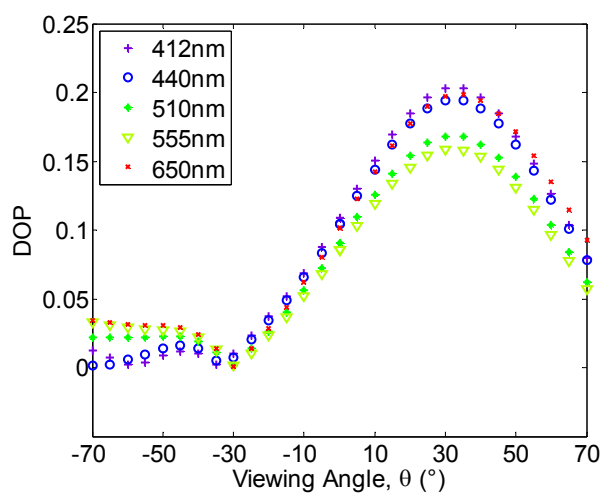
Fig 5.2. Geometry of observation of the hyperspectral radiometer in the principal plane.

Total reflectance spectra (GER spectroradiometer), as well as water absorption and attenuation spectra (WET Labs AC-S instrument) were also measured at the same stations together with concentrations of Chlorophyll [Chl] and organic and inorganic components of total suspended sediments (TSS) from the water samples. For analysis of phytoplankton pigments and particulate concentrations, suspended particles were filtered from water samples onto Whatman glass-fiber filters (GF/F). The filters were then stored in a cooler containing dry ice before analysis in the laboratory. [Chl] was calculated from absorption spectra. The concentration of total suspended sediments (TSS) was also calculated. Results are presented for Station 2 and 3, which showed the lowest and highest concentration of mineral particles, respectively (see Table 5.1). These values were used as input for the radiative transfer computations.

Table 5.1. Results of the biogeochemical characterizations.

	[Chl]	[NAP]	a_{CDOM} (400nm)
Station 2 12:50PM, $\theta_0=42^\circ$	3.67 μg /L	5.20mg /L	0.8155 m^{-1}
Station 3 1:40PM, $\theta_0=44^\circ$	3.48 μg /L	18.2mg /L	0.8302 m^{-1}

Fig. 5.3 shows the results of the calculations of the DOP of the water-leaving radiance for Station 2 (top panel) and the corresponding in-water WET Labs AC-S measurements (bottom panel).



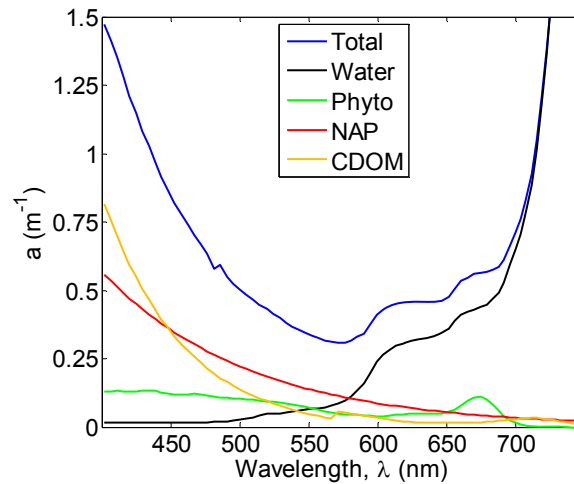


Fig 5.3. Station 2. Dependence of the simulated DOP (water-leaving radiance) on the viewing angle (top). In-water absorption spectra measured with the WET Labs package (bottom).

The decreases in the magnitude of the DOP at 510 and 555 nm (if compared with the one at 440 and 650 nm) are due to the increased level of scattering from beneath the surface). This is due to larger subsurface illumination levels thanks to less absorption at those wavelengths.

Fig. 5.4 shows the results of the calculations of the DOP of the water-leaving radiance for Station 3 (top panel) and the corresponding in-water WET Labs AC-S measurements (bottom panel).

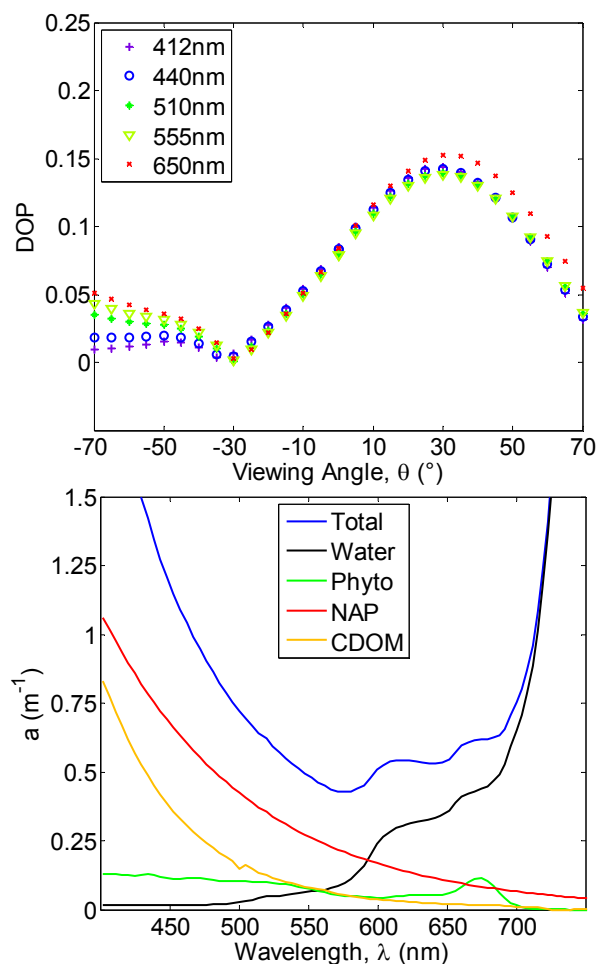
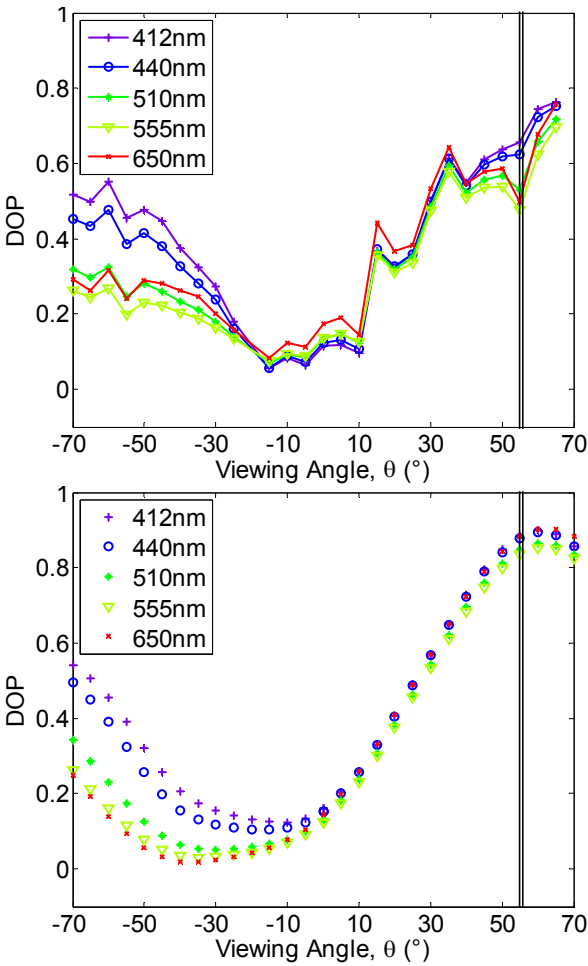


Fig 5.4. Station 3. Dependence of the simulated DOP (water-leaving radiance) on the viewing angle (top). In-water absorption spectra measured with the WET Labs package (bottom).

Station 3 contained a higher loading of NAP particles. High NAP concentrations cause multiple scattering and result in depolarizing the water-leaving radiance, which translates into lower values of the DOP. The effect of the subsurface particles is enhanced at 650 nm since water molecules absorb in that region and their contribution to the DOP is

reduced. At 412 and 440 nm, the suspended particles absorb and their contribution to multiple scattering is weak.

Following Chami *et al.* [2007b], we thought that, in case of high loadings of NAP, the polarized water-leaving radiance would have competed vectorially with the above-water reflected skylight (which is completely polarized at $\theta_B \approx 53^\circ$ for the air – water interface), thus inducing a significant depolarization of the upwelling radiation. The DOP measured at the Brewster's angle could, therefore, be related to the NAP concentration. Unfortunately, this didn't happen, as it can be seen in Fig. 5.5.



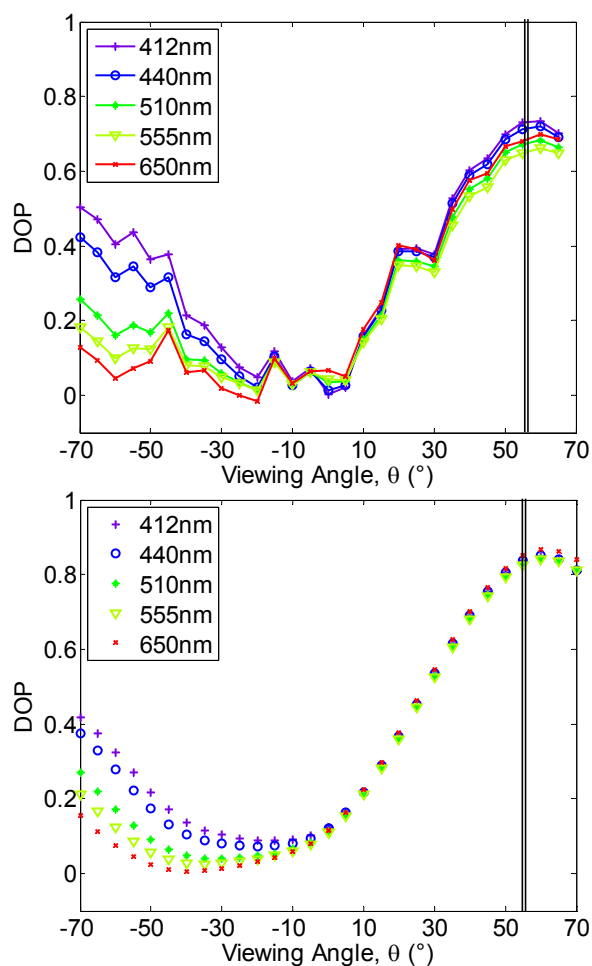


Fig 5.5. From top to bottom: measured DOP for Station 2, measured DOP for Station 3, simulated DOP for Station 2, simulated DOP for Station 3. The vertical lines indicate the position of the Brewster's angle. The two stations can be compared because the sun's position was approximately the same, therefore the position of the maximum of the DOP doesn't change.

The reflected skylight (also containing the sunglint) becomes dominant over the water-leaving radiance and the DOP doesn't contain any information on suspended particles.

The tiny difference in the simulated DOP plots near the Brewster's angle is completely lost in the experimental measurements due to the noise introduced by the surface waves.

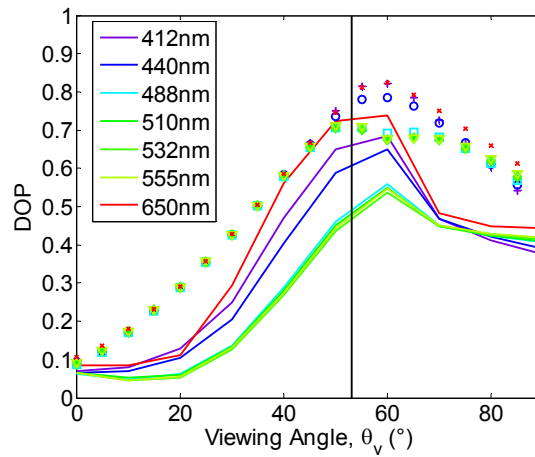
This result provided an incentive for considering other viewing geometries, *i.e.* away from the principal plane. In fact, as Chowdhary *et al.* [2006] showed, for sensor orientations slightly outside the principal plane, the DOP can approach maximal values observed in the principal plane (or even higher, depending on the concentration of suspended particles), thereby minimizing undesirable sun glint effects. Measurements away from the principal plane require collecting simultaneously three directions of polarization (horizontal, vertical and 45°) but our equipment at that time didn't allow for such measurements.

On August 17 – 20 2009, measurements were taken in the Chesapeake Bay and in the Atlantic Ocean, approximately 30 miles away from the shore of Norfolk, VA. This time our polarimeter was ready. Fig. 5.5 shows a picture of the polarimeter above water, before being immersed in the ocean.



Fig 5.6. The polarimeter above water. At this station the research vessel was approximately 30 miles away from shore.

For Station 1 ($73^\circ 30.9554$ W, $36^\circ 53.7833$ N, $\theta_{Sun} = 38^\circ$, 11:00), the concentration of phytoplankton is $1.2 \mu\text{g/L}$ while the concentration of NAP in this case is negligible, $\xi = 3.9$. Fig. 5.7 shows a comparison between the DOP measured above water (in the principal plane) and results of radiative transfer calculations. Station 3 was located in the Chesapeake Bay ($75^\circ 52.4573$ W, $36^\circ 53.7833$ N, $\theta_{Sun} = 27^\circ$, 14:00), the concentration of phytoplankton was $11.3 \mu\text{g/L}$, while the concentration of NAP was 3.0 mg/L , $\xi = 4.2$.



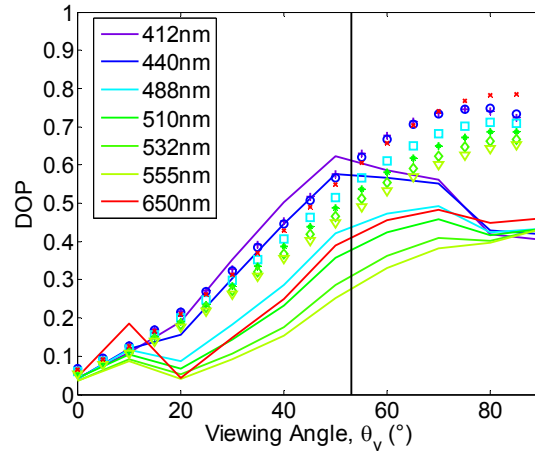
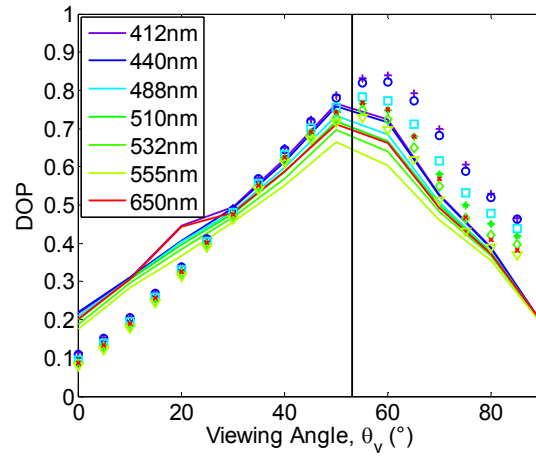
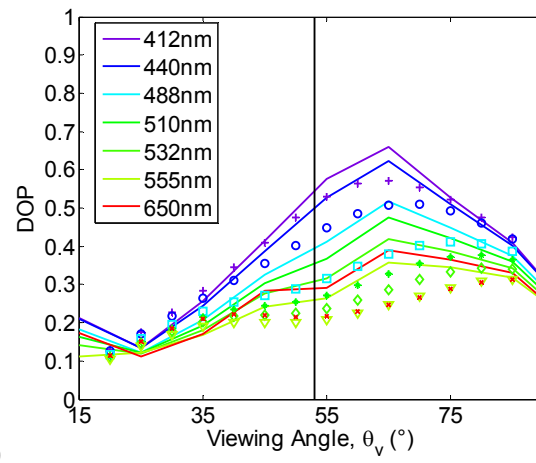


Fig 5.7. Measured (solid lines) and simulated (dotted lines) DOP for Station 1 (left) and Station 3 (right). The vertical line indicates the position of the Brewster's angle.

In these two cases the comparison with the simulation is not as good as before, but still satisfactory. However, even if differences between the plots can be easily noticed, the mismatch experimental / simulated results is too big to allow any possible discussion on the data. The instrumentation was sometimes oscillating back and forth, thus introducing uncertainties in the angular position. During this trip we also took above water measurements of the DOP for azimuth angle away from the principal plane. At Station 4 ($76^{\circ} 20.8552$ W, $36^{\circ} 58.5331$ N, $\theta_{Sun} = 51^{\circ}$, 3:30PM) the concentration of phytoplankton was $32.2 \mu\text{g/L}$ while the concentration of NAP was 11.3 mg/L , a very turbid water (Fig. 5.8).



(a)



(b)

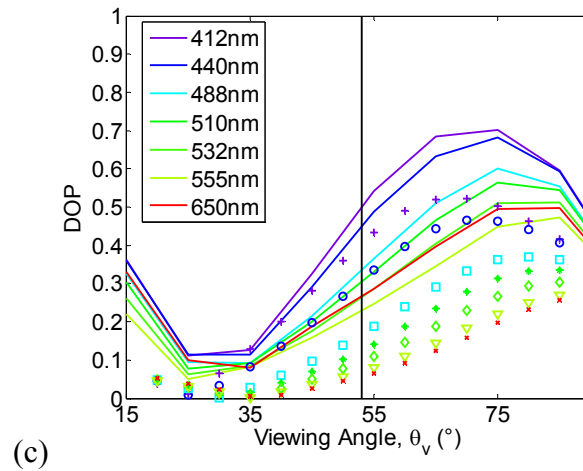


Fig 5.8. Measured (solid lines) and simulated (dotted lines) DOP for Station 4, in the main scattering plane (a), 45° (b), and 90° (c) away from the main scattering plane. The vertical line indicates the position of the Brewster's angle.

Simulations satisfactorily resemble the experimental measurements in this case also. For a quantitative analysis of the above – water measurements, though, we still need to improve substantially our methodology of recording data.

6. In situ under water measurements and comparison with Monte Carlo radiative transfer model

In this chapter field measurements obtained in the coastal areas of New York Harbor – Sandy Hook, NJ region (USA) will be described. Measurements of the DOP will be presented and discussed together with corresponding *in situ* standard optical measurements, including total reflectance spectra (GER Spectroradiometer) and water absorption and attenuation spectra (WET Labs AC-9 instrument). This will illustrate how the variability of the DOP is connected to water constituents. Measurements were also taken in different atmospheric conditions, to observe the effects of diffuse light, rather than direct sun light on the DOP and a comparison between above and underwater measurements is explored. Field data are also compared with simulations obtained using a Monte Carlo radiative transfer code of the atmosphere–ocean system. Both radiance and DOP in the principal plane obtained with the Monte Carlo simulations are presented. Because the solar angle, surface wind speed and detector depths were recorded along with the radiance and polarization value, to perform these simulations, only a few assumptions need to be made about the scattering characteristics of the water.

6.1 Field measurements methodology

Data were collected during a recent cruise on the R/V “Connecticut” in the coastal areas of New York Harbor – Sandy Hook, NJ region (USA), on July 21–23 2008. The data reported in this section were collected at eight stations. The coordinates of each station, the corresponding solar zenith angles (θ_{Sun}) and the time of the day are given in Table 6.1.

Table 6.1. Coordinates and solar zenith angles of the sampling stations.

Station	Longitude	Latitude	θ_{Sun}	Wind speed	Starting time
1	73°53'658"W	40°26'978"N	21±1°	2.5m/s	12 ³⁰
2	73°53'066"W	40°27'975"N	57±2°	3.8m/s	16 ¹⁵
3	73°52'202"W	40°23'509"N	36±2°	5.5m/s	9 ⁴⁰
4	73°54'002"W	40°26'905"N	22±2°	3.2m/s	12 ⁴⁵
5	73°47'387"W	40°29'646"N	56±2°	4.8m/s	16 ¹⁰
6	74°08'727"W	40°29'758"N	28±2°	6.4m/s	10 ³⁰
7	74°08'727"W	40°38'023"N	21±1°	8.1m/s	12 ³⁰
8	75°03'445"W	41°30'843"N	46±2°	8.0m/s	15 ¹⁵

The instrument was lowered from a winch extending from the side of the ship (approximately 3–4m from the side of the ship) so that shadow effects were minimized. The principal plane was approximately normal to the ship axes and the sun was on the side of the instrument (Fig. 6.1(a)). The entire assembly was kept 1m below the water surface using four arms with buoys attached (Fig. 6.1(c)). This configuration also allowed the scanning system to always rotate in a plane perpendicular to the water surface. For normalization purposes, downwelling irradiance was also recorded with a Satlantic Hyperspectral irradiance sensor positioned on the front deck of the ship. During this cruise all measurements were recorded only in the principal plane (*i.e.* $\varphi = 0^\circ$). The wind speed according to the ship instrument measurements was in the range 2.5 to 8.1m/s and

the ocean surface wave amplitude did not exceed 1.2 m. The sky was clear blue with no clouds during the data acquisition time for the first two days (Stations 1–5) and overcast during the last day (Stations 6–8). Water optical properties were measured at the same stations by an AC–9 instrument (WET Labs, Inc.). Attenuation and absorption data are available for Stations 1, 4, 5 and 7. Concentrations of mineral particles and Chlorophyll were not measured directly in the field, but they can be estimated by analyzing absorption and attenuation spectra measured with our AC–9 instrument. Specifically, the total scattering coefficient at 555nm, which is defined as the difference between total attenuation and absorption coefficients, is a good approximation for TSS (Total Suspended Sediments) based on previous studies [Gould et al. 2002]. Minerals concentration then equals TSS concentration multiplied by 0.76. Chlorophyll concentration, on the other hand, can be estimated as the elevation from the baseline at 675 nm on the absorption spectrum divided by an *in vivo* specific absorption for Chlorophyll of $0.0146\text{m}^2\text{mg}^{-1}$ [Sullivan et al. 2005]. Understanding that these assumptions can strongly differ for different regions and waters, we used them only as a rough approximation for Chlorophyll and minerals concentrations.

6.2 Experimental results

Characterization of water compositions

Attenuation (c) and absorption (a) curves with and without water are shown in Fig. 6.1, minerals and Chlorophyll concentrations estimated using the above approximations are summarized in Table 6.2. In Fig. 6.2 total absorption spectra (a_{tot}) are shown which include water absorption together with the spectra from Fig. 6.1(a). As will be shown

below, *a priori* knowledge of the total absorption and attenuation spectra are important in the spectral analysis of the DOP.

Table 6.2. Minerals and Chlorophyll concentrations estimations.

	St1	St4	St5	St7
Minerals (mg/l)	2.5	3.2	2.0	3.9
Chlorophyll (μg/l)	4.8	3.9	1.3	2.1

We also recorded total reflectance spectra just below the water surface, using a fiber optic probe attached to a GER spectroradiometer (Fig. 6.3). As mentioned earlier, the sky was clear with no clouds at Stations 1–5 while at Stations 6–8, the sky was overcast resulting in mostly diffuse incident light. Station 7 was located at the entrance of the Hudson River and Station 8 was located further up the river which explains the strong increase of reflectance (especially for Station 8) due to an increase of mineral scattering.

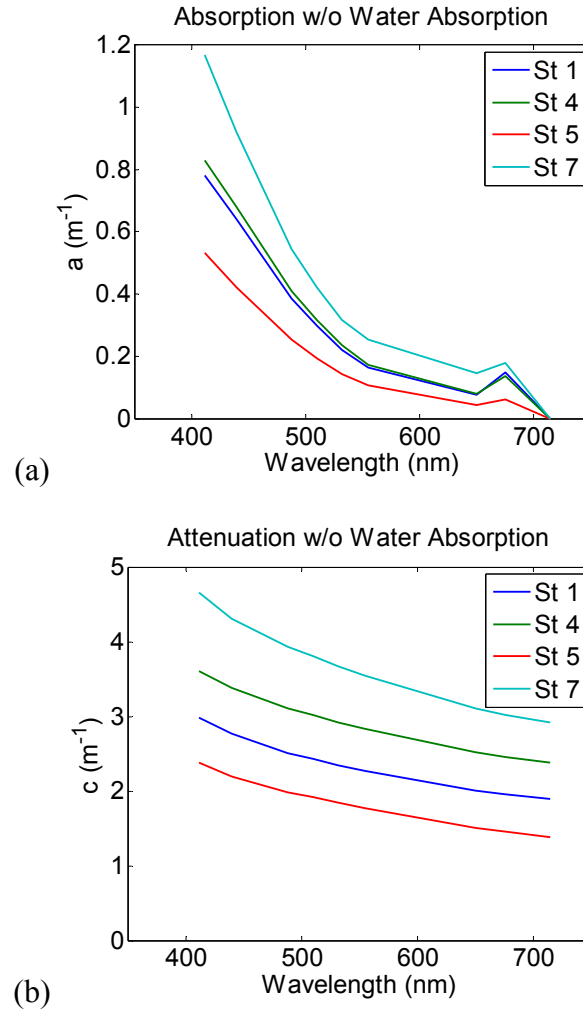


Fig. 6.1. Absorption and attenuation spectra recorded with the WetLabs package.

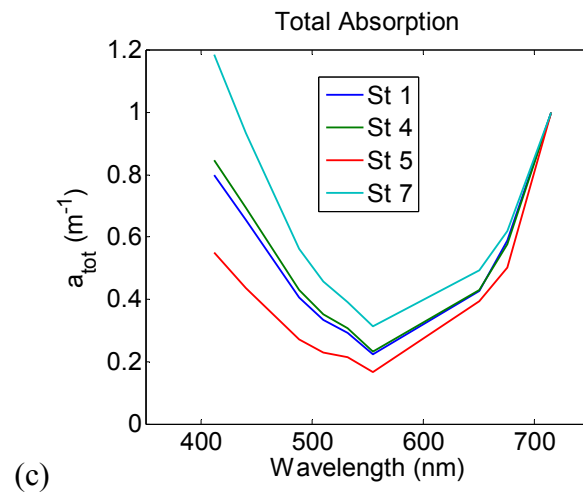


Fig. 6.2. Total absorption spectra recorded with the WetLabs package.

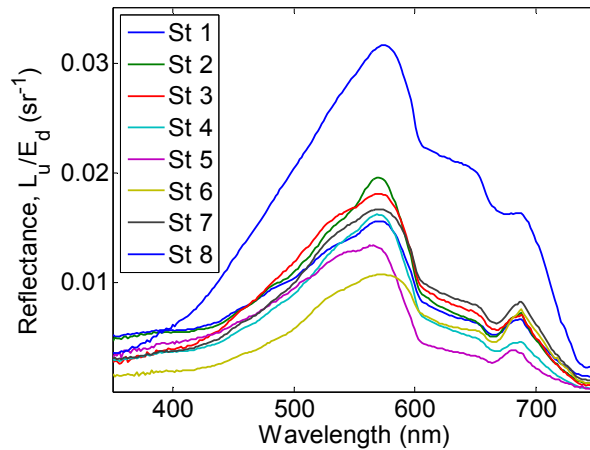


Fig. 6.3. GER total reflectance spectra.

Measurements of particle phase functions

Along with the absorption and attenuation coefficients, the volume scattering function (VSF) is one of the fundamental IOPs which govern the propagation of polarized light in aquatic environments. VSF measurements were made with a custom device called the MASCOT [Sullivan et al. 2009]. It uses a 658 nm laser diode source and 17 independent detectors to measure volume scattering from 10 to 170° in 10° increments. Sampling rate for all detectors is 20Hz. A wedge depolarizer installed in front of the source provides unpolarized incident irradiance. Independent silicon diode detectors allow resolution of the VSF without any moving parts, with each detector gain optimized for its specific angular measurement. Based on curvature in the VSF at specific angles, relative nominal scattering intensities, and the need to avoid stray light contamination, detector field-of-views (FOVs) are set at 0.8°, 2°, 3° and 4° for the 10°, 20°, 30° and 40° measurements, respectively. Detectors at other angles have a FOV of 5°. Calibration and validation of MASCOT VSF measurements were carried out with microspherical beads traceable to

NIST standards. Angular weighting functions were computed numerically for each angle measurement. Beam attenuation was used to correct for light losses along the optical path of the MASCOT scattering measurements (all pathlengths nominally $\sim 20\text{cm}$). Normalized particle phase functions (PPFs) are calculated from the VSFs values divided by the scattering coefficient at 650nm (obtained from the difference between attenuation and absorption at 650nm measured with our AC-9 instrument). The PPF for Station 1 is compared with the standard Petzold's phase function [Petzold 1972]. Specifically, the three phase functions corresponding to each water type are averaged to produce a representative particulate phase function, as described by Mobley [1994]. Fig. 6.4 shows the comparison between MASCOT and Petzold's phase functions. Similar results are obtained for Stations 4, 5 and 7, due to similar water compositions. The agreement between MASCOT measurements and Petzold's functions, justifies the choice of the latter to represent the phase functions of particulate matter in the model.

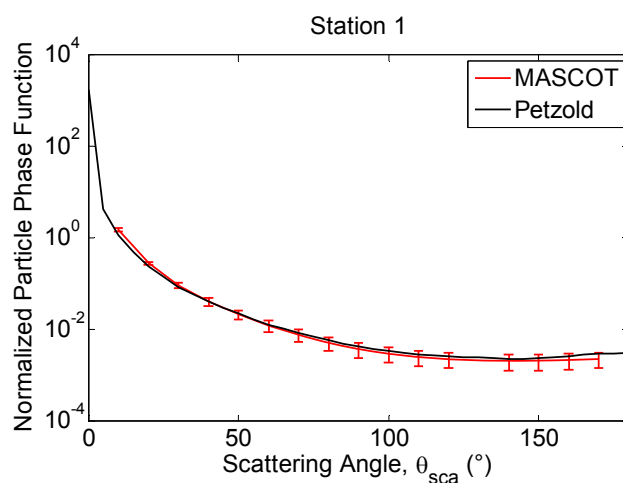
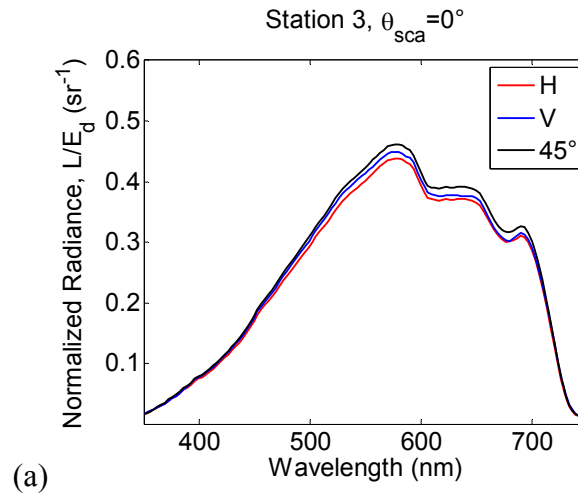


Fig. 6.4. Comparison of MASCOT measurements and standard Petzold functions.

Results of polarization measurements

To begin, we wish to verify the polarization state of the water under specialized illumination conditions. Fig. 6.5(a) is an example (obtained when scattering angle is 0° , which corresponds to the sensors looking directly at the underwater sun light) where the DOP is effectively 0 since there's no difference in readings between the three polarization directions. In this case, the impact of atmospheric particles and air–water interface on underwater polarization was minimal, while for scattering angles away from 0 and 180° (90° , for example) the situation changes dramatically (Fig. 6.5(b)).



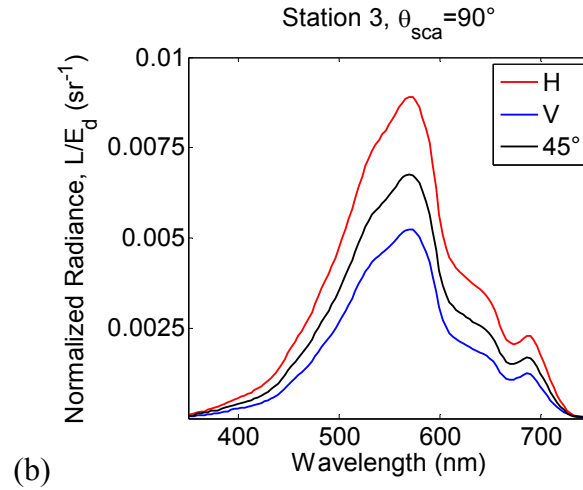


Fig. 6.5. Spectral dependence of the signal recorded by the Satlantic Hyperspectral sensors when the scattering angle is 0° (a) and when it's 90° (b). The instrument is positioned in the main scattering plane, 1m below water.

All recorded DOP values are based on measurements of radiances with different polarizers in front of the sensors. The dependence of these radiances (*i.e.* vertical, horizontal and 45°) as well as of the total radiance normalized to the downwelling irradiance as a function of the scattering angle is shown in Fig 6.6. The maximum differences are around 100° which corresponds to the maximum of the DOP. Relationships of these dependencies with the volume scattering functions should be further investigated.

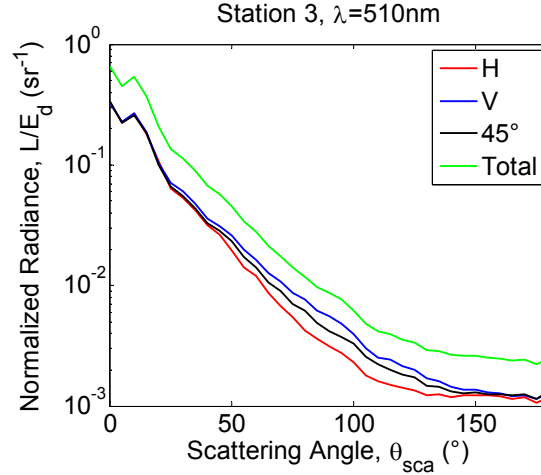


Fig. 6.6. Dependence of normalized radiance components on the scattering angle.

In assessing our polarization measurements for stability, we took multiple measurements when possible. Clearly measurement errors can occur due to the changing ocean environment and can be amplified when DOP values are small. Fig. 6.7 shows the DOP for all data points acquired for a set of representative measurements, corresponding to Station 1. Note the low variability of the experimental data points for most of angles underlining the high accuracy of the detection method. We attribute this to the positioning of the entire assembly on four buoys, which kept the sensors approximately at the same depth even in conditions of high waves. For scattering angles inside Snell's window (*i.e.* between 0 and 65° scattering angle, for Station 1), the largest fluctuations induced by surface waves were observed.

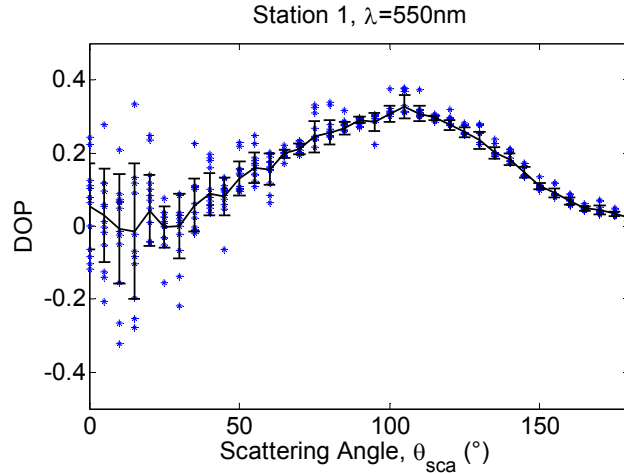
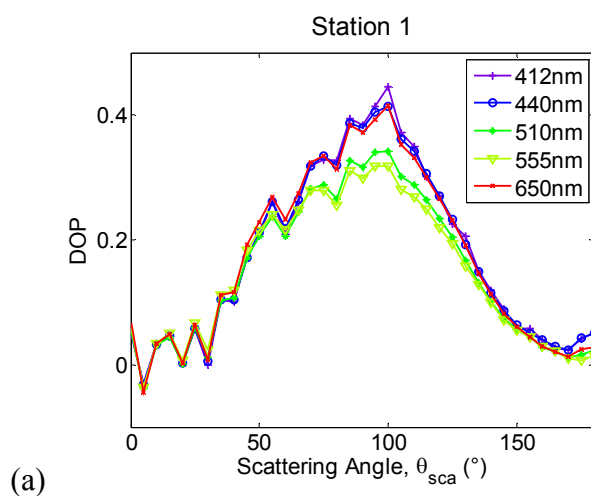
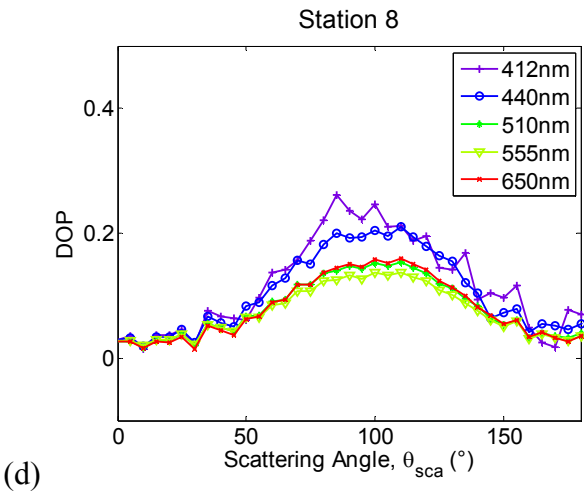
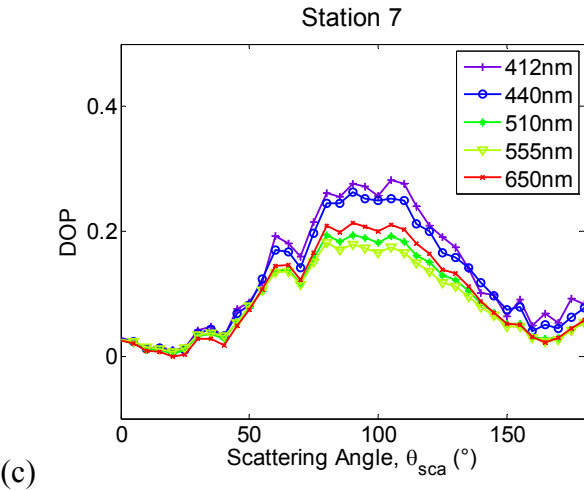
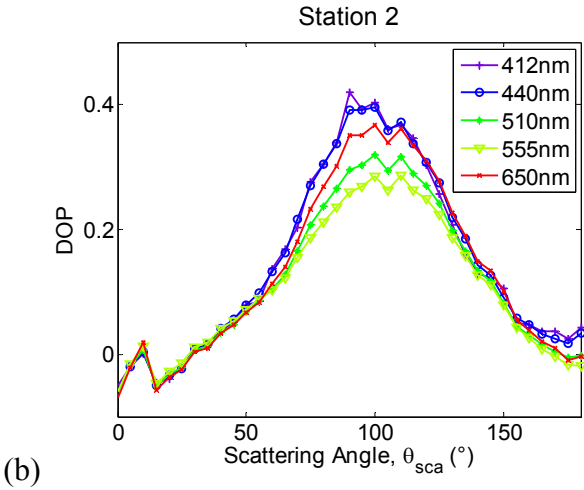


Fig. 6.7. Spread of the data points acquired during a set of measurements. Data are shown for Station 1, $\lambda=550\text{nm}$.

To assess our measurements, we first note that the DOP for ocean waters typically makes a bell-shaped distribution as function of the scattering angle with the maximum around 90° and going to zero in proximity of 0 and 180° . Typical plots of the DOP vs. scattering angle, recorded in the main scattering plane at 1m depth are presented in Fig. 6.8. The maximum of the DOP is lower than half the value predicted by Rayleigh theory (*i.e.* it reaches a maximum value of approximately 0.4 at 410nm). Results for Stations 1–5 are very similar due to similar water compositions and are represented by Station 1 and 2 in Fig. 6.8. However, we note a significant reduction of the DOP at Station 7 and 8 (Fig. 6.8(c) and (d)). These can be traced to the diffuse illumination from clouds as well as an increase in mineral concentrations (Station 8, located in the Hudson River). When the sky is cloudy the DOP decreases, according to the general rule that the underwater light polarization decreases as the diffuseness of the illumination increases. At the same time the angular distribution of the DOP tends to become symmetrical around the vertical axis.

This has been observed by several investigators [Ivanoff 1957]. For reference, downwelling irradiance spectra for the same stations are shown in Fig. 6.8(e). In particular, an increased amount of suspended particles results in increased number of multiscattering events, further depolarization and in turn lower DOP. We also note the shift of the maximum of the DOP towards 100° scattering angle. Chami *et al.* [2001] predicted this effect and suggested its use to allow discrimination between biological and non-biological constituents which should be further verified. In fact, as the real part of the refractive index increases (mineral particles) the absolute maximum of the DOP decreases in value and shifts towards higher scattering angles.





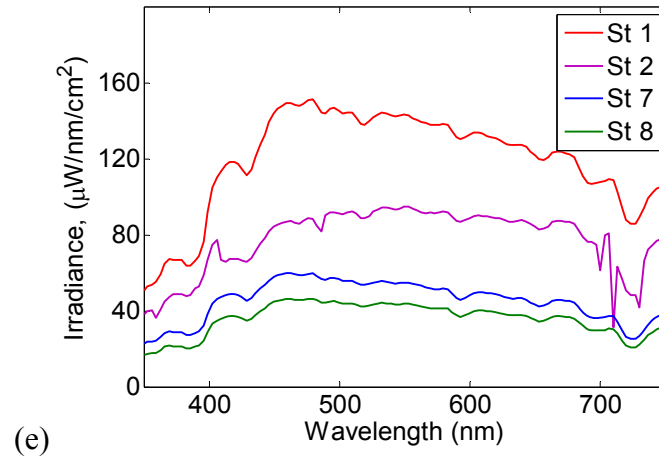
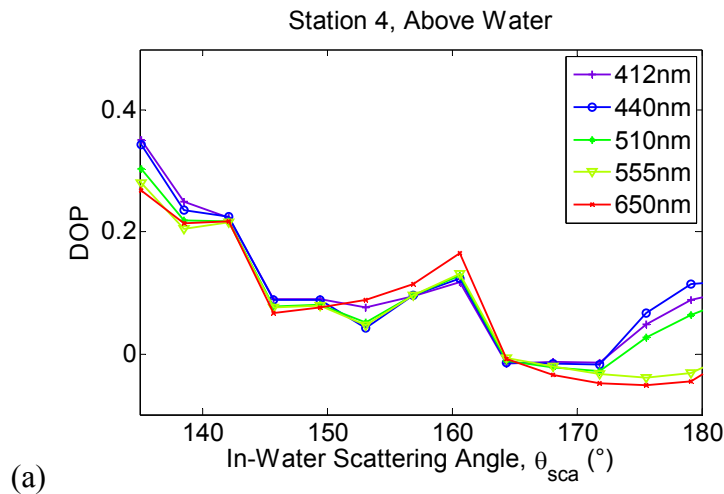


Fig. 6.8. (a)–(d) Plots of the DOP vs. scattering angle. The instrument is located in the principal plane 1m below water. (e) Downwelling spectral irradiance recorded at the same stations.

It's worth noting that for remote sensing purposes only scattering angles in the range 140–160° should be realistically considered. We recorded above water measurements for Station 4 and the data are presented in Fig. 6.9 in comparison with underwater measurements. For angles corresponding to in–water scattering angle less than 145° (since specular reflection of sun light occurs at ~147° in–water scattering angle, for Station 4) abnormally high values of DOP were observed, due to the contribution of the polarized reflected sky light. These values are not related to the characteristics of the water components; hence they are not shown in Fig. 6.9(a). On the other hand, for in–water scattering angle more than 145°, the reflection coefficients for the in–plane and out–of–plane polarizations are similar [Hecht 1988]; therefore the observed DOP is not due to reflection effects but to the polarization properties of the water–leaving radiance. From this it is clear that the DOP corresponding to the range 140–160° will not exceed

0.2. Taking into account the fact that Station 4 presented a relatively low concentration of minerals and the DOP decreases for higher concentration of minerals, we can assume this value of the DOP to be close to the upper limit which can be expected for remote sensing measurements in coastal waters. Values of DOP measured above and below water surface (Fig. 6.9(a) and (b)) for this range of angle well correspond to each other. The effect of absorption is not noticed here since the optical thickness of the water medium wasn't high enough to decrease the magnitude of the DOP measured above water [Ramella-Roman 2005a], when compared with the underwater DOP. Swell and wind ruffling of the water surface affect much more the above water signal than the underwater measurements, as shown by the comparison of Fig. 6.9(a) and (b) curves. The former contain random sun glint reflections not present underwater. Above water recordings were performed while the instrument was held by the winch cable approximately 2–3m outboard from the ship, and 1m above water.



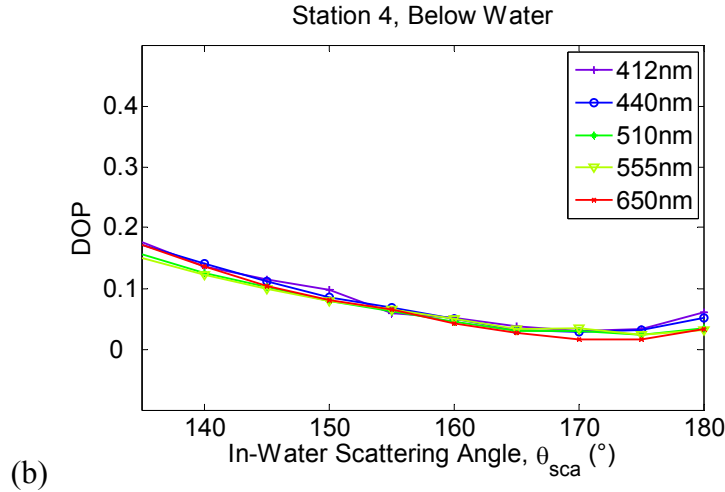


Fig. 6.9. Plots of the DOP vs. scattering angle for Station 4: (a) 1m above water, (b) 1m below water. The instrument is located in the principal plane.

The DOP was also found to vary with wavelength. Comparisons between spectral dependences for measurements of the DOP taken at different stations are presented in Fig. 6.10. In Fig. 6.10(a) (which corresponds to Station 1), we observe a maximum in the DOP at lower wavelengths. This region is dominated by Chlorophyll and CDOM absorptions, as can be seen in Fig. 6.2 and in the normalized absorption spectrum (normalized to the maximum absorption) reported in the top part of Fig. 6.10(a). In the top part of Fig. 6.10(a) the total absorption spectrum (a_{tot}) divided by the total attenuation spectrum (c_{tot}) is also shown. Note that $a_{tot}/c_{tot} = 1 - \omega$, where ω , the single scattering albedo, is a good measure of the amount of multiple scattering. If ω decreases, multiple scattering events are reduced and the DOP increases. Fig. 6.10(a) also shows that the DOP reaches maximum values in the range 0.4–0.5 at 410 and 440nm. On the other side of the spectrum (*i.e.* 700–750nm), another maximum appears. This behavior is consistent with the absorption spectrum. After 700nm, water absorption starts increasing,

minimizing again elastic scattering. The relative maximum between 600–650nm is also consistent with the absorption curve. On both sides of this relative maximum, two minima occur and the DOP reaches minimum values around 0.3. The first minimum is consistent with the minimum in the absorption; absorption decreases and multiple scattering events increase, depolarizing the underwater light field. The second minimum, however, cannot be directly related to the absorption curve. This dip in the DOP is due to the Chlorophyll fluorescence in this spectral interval, which occurs in addition to elastic scattering and it's unpolarized. This hypothesis is confirmed by the data recorded at Station 7 (Fig. 6.10(b)). In particular, within the Hudson River, the concentration of Chlorophyll is very low, and the dip in the DOP is barely noticeable. In addition, the effect of Chlorophyll fluorescence on the DOP is further confirmed by the comparison of experimental data with Monte Carlo simulations, which will be shown in the next section. It's worth noticing that the spectral variations of the DOP and the effects of Chlorophyll fluorescence are more pronounced at scattering angles where the DOP is close to maximum and weaker at other angles. In addition, below 400nm and above 700nm, because of the increasing water absorption, the absolute values of radiance become very small, leading to big uncertainties in the calculation of the DOP.

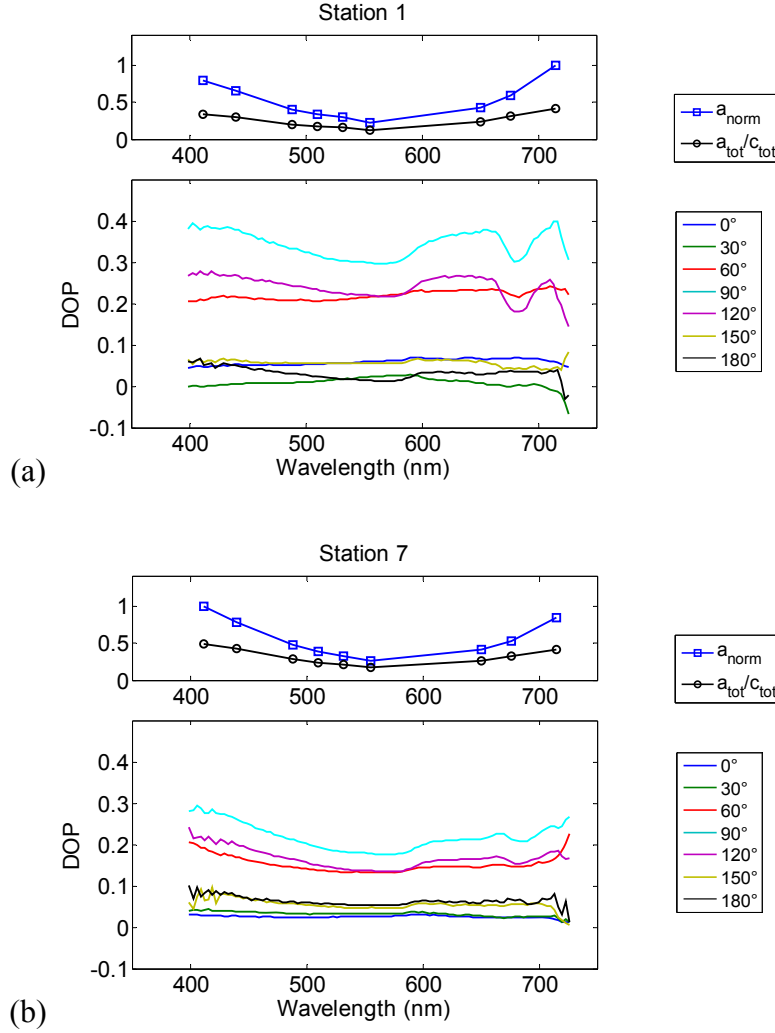


Fig. 6.10. Spectral dependence of the DOP: (a) Station 1 and (b) Station 7. The normalized total absorption spectrum (a_{norm}), normalized to the maximum absorption, and the total absorption spectrum divided by the total attenuation spectrum (a_{tot}/c_{tot}) are also shown.

6.3 Comparison of modeled and measured data

To assess the results beyond simple qualitative experiments, the polarimeter measurements were compared to the results of a Monte Carlo vector radiative transfer model. This is an updated version of the code used in Adams and Kattawar [1997]. It is a

plane – parallel model of the coupled atmosphere – ocean system and allows computation for a number of differing layers in both the atmosphere and ocean. The specific details can be found in the above reference, and we will only give a brief summary here. The atmosphere is modeled as a two–layer system. The top layer is assumed to be Rayleigh scattering only, while the bottom layer is a marine haze model comprised of Rayleigh plus aerosol scattering. The aerosol layer above the ocean surface is modeled using the MAR–I model published by the International Association for Meteorology and Atmospheric Physics [World Climate Research Program 1984]. The sea surface is modeled using the sea–slope statistics of Cox et al. [1954]. In the water column, we use the dissolved and particulate absorption coefficients, the particulate scattering coefficient, and the absorption and scattering coefficients of pure ocean water. Light scattering in both the atmosphere and ocean is described by the elastic scattering Mueller matrix for each constituent. Atmospheric Rayleigh scattering is described using the Rayleigh matrix including depolarization factors. The aerosol matrix is found using the aerosol constituent parameters in a Mie code to produce the Mueller matrix. In the water, molecular scattering is again described using the Rayleigh matrix and the depolarization ratio for water. For particulate scattering, we use the averaged Petzold’s particulate volume scattering phase function reported by Mobley [1954]. To find a Mueller matrix for particulate scattering, we first define a reduced Muller matrix by dividing each element by the (1,1) element, the phase function. The particulate Mueller matrix is then obtained by assuming a hyperbolic particle size distribution with a slope of -4 and a constant relative index of refraction of $n=1.09+0.002i$, and calculating the reduced matrix using a Mie code. The result is that the reduced matrix is not much different than a reduced

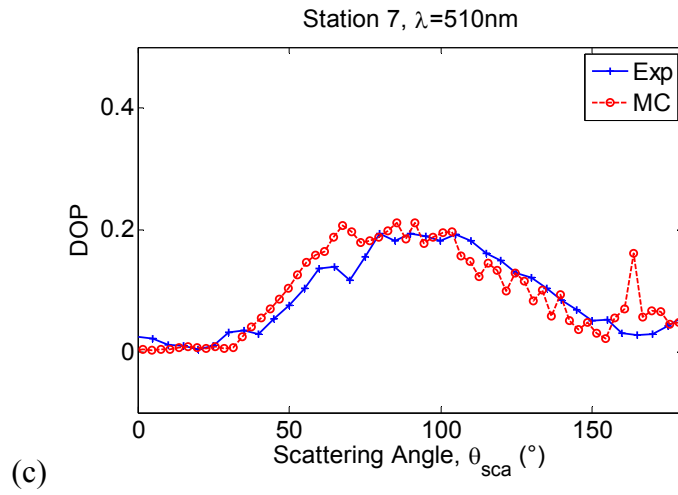
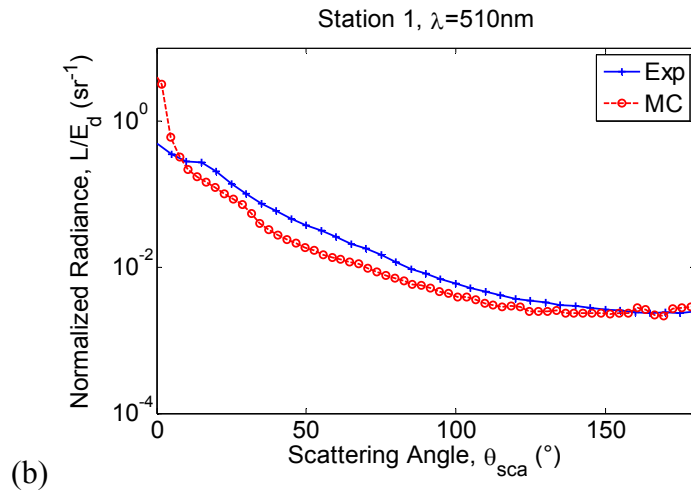
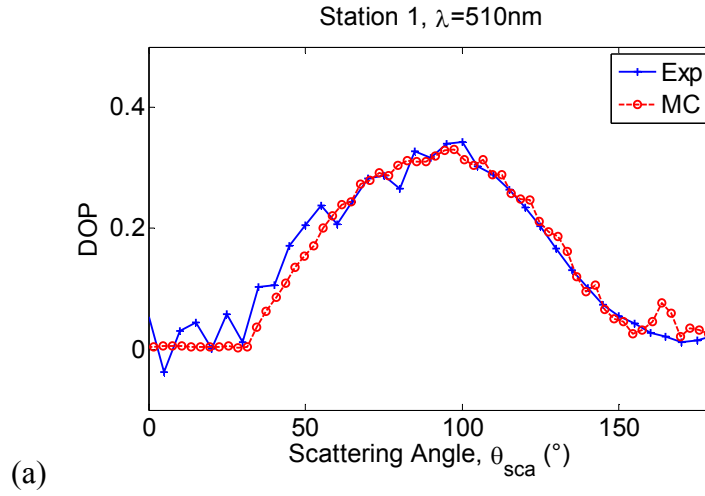
Rayleigh matrix, similar to that reported in Voss and Fry [1984]. In all cases, atmosphere and ocean, the phase function and Mueller matrices are assumed to be independent of wavelength.

The model was run for each wavelength of the AC-9 in-water measurements. The measured values of the dissolved and particulate values of the absorption and attenuation coefficients were used for the water column. The measured wind speed was used for the waveslope statistics (Station 1: 2.5 m/s; Station 7: 8.1 m/s). The largest unknown is the atmosphere. To determine the level of aerosols, we turned to the surface irradiance measurements shown in Fig. 6.8(e). The ratio of the measured surface irradiance to the extraterrestrial irradiance was found for each wavelength. In the model, the optical depth of the aerosol haze layer was varied until the modeled surface to top of the atmosphere ratio matched the measured one. Once this aerosol optical depth was found, the code was run for each wavelength and the Stokes vector parameters were found for a depth of 1m below the surface. These values were then divided by the calculated surface irradiance values to produce the normalized radiance reflectance.

Fig. 6.11 shows a comparison between the measured and modeled values at Station 1 and 7, at a wavelength of 510nm for the DOP and the normalized radiance. Figs 6.11(a) and (b) compare the DOP and radiance at Station 1. The atmospheric optical depth at this wavelength was calculated to be 1.34, and values for the other wavelengths ranged from 1.0 to 2.0. The radiance values agree well, but there are differences, especially in the direction of the sun. The model tends to over predict the radiance in this direction, and then under predict it at larger angles. The most probable cause of this discrepancy is the generally unknown state of the atmosphere. By looking at the shape of the modeled

radiance curve we guess that the incident skylight was more diffuse than the model predicts. The agreement between the magnitudes of the measured and modeled DOP is very good. At this shallow depth, one of the largest sources of uncertainty is the effect of the wind-blown surface. Though generally small in other azimuthal planes, in the principal plane, on the side containing the direct beam of the sun, increased wind speed will cause the direct solar beam to spread into more angles and in general reduce the maximum DOP. Modeled results for the same optical properties show that a smooth sea surface results in a maximum DOP approximately 5–10% higher than the wind-blown surface at Station 1.

Figs 6.11(c) and (d) show the comparisons at Station 7, also at a wavelength of 510nm. This day was overcast, and the atmospheric optical depth was calculated to be 8.98, with similar values for other wavelengths. Now it is unlikely that the atmospheric haze model correctly predicts the true nature of the atmosphere under these conditions, but we hypothesize that the light field incident on the ocean will be sufficiently diffuse that it doesn't make a great difference. The radiance values again show good agreement, though this time the model tends to under predict the radiance at angles toward the sun. The DOP again agrees very well, though there is quite a bit of statistical noise in the Monte Carlo simulations due to the large atmospheric depth and very windy conditions.



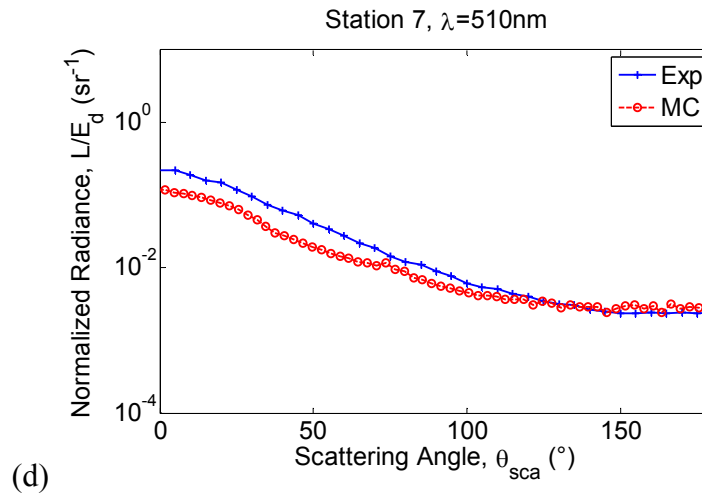


Fig. 6.11. Comparison of modeled and measured data for 510 nm, Station 1, (a) DOP, (b) normalized radiance. Station 7, (c) DOP, (d) normalized radiance.

The peak in the DOP curves near 180° is an artifact of the statistical nature of the Monte Carlo method. This is a well known issue when the solid angles become very small near the zenith direction, and there are insufficient upwelling photons traveling into these solid angles to resolve the DOP sufficiently. The peak observed in the Station 7 data is particularly noticeable.

Similar comparisons for a wavelength of 676nm are also presented in Fig. 6.12. The radiance curves match well, similar to the 510nm wavelength with similar discrepancies. But now the DOP match is poor. The Monte Carlo radiative transfer code didn't include Chlorophyll fluorescence which, as hypothesized above, would decrease the value of the DOP. This probably explains the significant difference in the measured and simulated values. Fig. 6.13 summarizes the spectral comparisons of modeled and measured DOP

for four relevant scattering angles for Station 1. Again, the match is quite good except in the region of Chlorophyll fluorescence as discussed before.

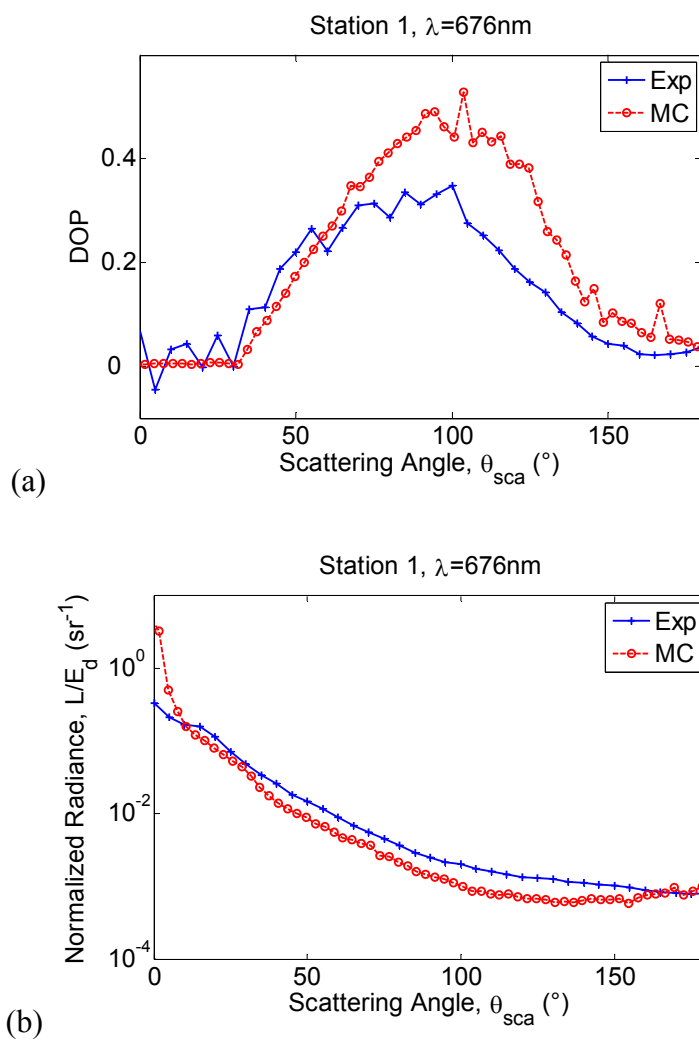


Fig. 6.12. Comparison of modeled and measured data for 676nm, Station 1, (a) DOP, (b) normalized radiance.

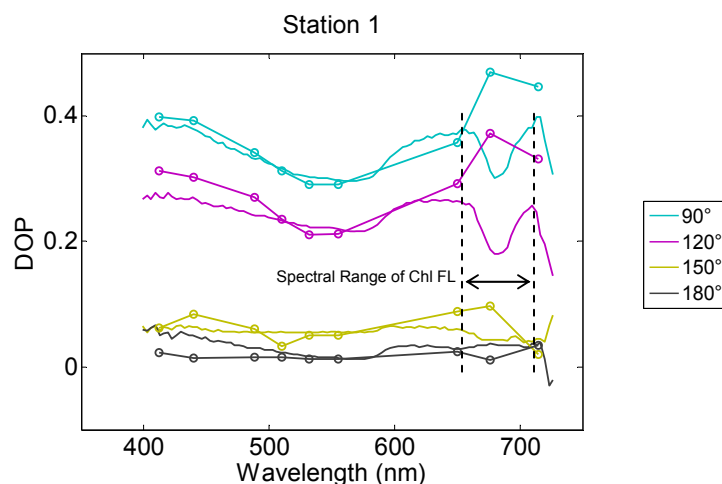


Fig. 6.13. Spectral comparison of modeled (circles) and measured (solid lines) DOP for four relevant scattering angles, Station 1.

6.4 Conclusions

Measurements of the polarization characteristics of coastal waters were acquired during a field experiment in the coastal areas of New York Harbor – Sandy Hook, NJ region (USA), using a newly developed hyperspectral and multiangular polarimeter. An analysis of the angular (0–180°) and spectral (400–750nm) variations of the degree of polarization was performed. It was observed that maximum values of the DOP, which occurred approximately at scattering angle of 100°, did not exceed 0.4 for all stations. In overcast conditions however, the light was still partially polarized but with the maximum DOP reduced to approximately 0.2.

Spectral dependence of DOP very well correlates with the measured water absorption and single scattering albedo (ω) spectra: increase of absorption (decrease of ω) corresponds to the decrease of the number of the scattering events which means less depolarization.

These effects are more pronounced at scattering angles close to 90° where the DOP has its maximal values and almost unnoticeable at angles closer to 0 and 180° . In the spectral area of Chlorophyll fluorescence we observed significant decrease of DOP which is explained by the depolarizing effect of fluorescence. Despite windy conditions at some stations where waves were up to 1.2m high, the angular profiles of the DOP and its maximal values remained similar to the values at the stations with the same water composition but lower wind speeds.

For remote sensing applications, when scattering angles are in the range of 140° – 160° , the DOP does not exceed 0.2. In addition, we note that the measured values above water correctly correspond to the results of underwater measurements. This result suggests that, despite the effect of the water surface on the polarization of light, the contribution of the underwater polarized light field is sufficiently significant to affect the above water signal. This makes the study of polarization in the ocean promising for future improvement of retrieval algorithms in complex waters, such as those founded in the coastal areas. In addition, by performing these measurements underwater, further noise sources due to the wind-roughened state can be eliminated.

The agreement between the Monte Carlo results and the experimental data are also shown and clearly demonstrates the success of radiative transfer simulations applied to the transmission and scattering of light in an atmosphere–ocean system. The shape of both the radiance distribution and the DOP has been correctly reproduced for different atmospheric and water conditions.

It is expected that the ability of our polarimeter to provide information about the characteristics of the underwater polarized light field has great potential for application in

radiative transfer problems in the earth–ocean system, especially if used in combination with other polarization–sensitive instruments recently developed; hyperspectral and multiangular data can be collected very accurately and in a relatively short time, thus changes both in the water and in the atmosphere can be avoided.

7. Polarization characteristics of coastal waters and their impact on in – water visibility

This section focuses on the use of polarization to improve underwater visibility. In order to numerically reproduce the polarized images for underwater horizontal imaging system the measured typical underwater polarized radiance was used to estimate the polarized components of the background veiling light and the blurring effects were modeled by point spread functions obtained from the measured volume scattering functions (recorded during the cruise on the R/V “Connecticut” in the coastal areas of New York Harbor – Sandy Hook, NJ region (USA), on July 21–23 2008) and other typical oceanic environments. It is shown that the visibility can be improved for unpolarized target by placing a polarizer oriented orthogonally to the partially polarized direction of the veiling light in front of the camera. The blurring effects strongly depend on the small angle scattering in the forward directions. For polarized targets, the Monte Carlo simulation of slab geometry for polarized pencil light shows that the scattering medium with high asymmetry factor, g , value has a very strong ability to retain the polarization status of the incident light, which can be utilized to improve the image contrasts for targets with very different polarized reflection properties.

Study of the underwater polarized light field can lead to significant improvements of underwater visibility. Polarization-sensitive vision is well documented and serves in navigation and enhancement of target detection for many marine species [Sabbah et al. 2005]. Knowledge of the underwater polarization pattern can help us better understand the mechanisms involved in the sensitivity of animals to the polarization of light and the

manners in which these animals utilize underwater polarization. Inspired by biological visual systems, many studies have applied various polarization techniques in underwater imaging systems to improve the visibility of selected target features in a scattering medium [Tyo et al. 1996], [Kattawar et al. 1999], [Schechner et al. 2004], [Chang et al. 2003]. Since the underwater scattered light, or *veiling light*, is the dominant cause of image degradation, assessing the effectiveness of these polarization techniques under natural illumination also relies on the analysis of the spectral and geometrical angular dependence of the underwater polarized light field. When compared with standard radiance intensity data, the polarized upwelling radiance is more sensitive to the intrinsic nature of in-water particulates such as particle size, shape and refractive index [Chami et al. 2001], [Chami et al. 2007a]; therefore it carries more information on the background veiling light.

In order to determine the effects of a scattering medium on the quality of underwater imaging, the point-spread function, obtained from the measured VSF, and the measured polarized pattern are combined to numerically reproduce the image of a standard Air Force Target under natural illumination conditions. Results obtained by using different VSFs measured in various oceanic environments were then compared, focusing on the impact on underwater visibility.

7.1 Image modeling for underwater visibility assessment for unpolarized targets

Under natural illumination an imaging system measures light from the target and additional light (veiling light) from the intervening medium. In addition, light from the

target experiences scattering when it passes through the scattering medium between the target and the camera, hence the image gets blurred. This blurring effect can be fully described by the PSF of the scattering medium [Hou et al. 2008]. In this section an image model will be first introduced to combine together blurring and veiling light in order to simulate the imaging quality in the water scattering medium for an unpolarized target. Then the underwater visibility will be assessed and discussed by applying various particulate phase functions as well as various polarization techniques.

Let's consider an underwater object illuminated by sun light and a camera taking pictures looking horizontally (as shown in Fig. 7.1). On the image plane, the irradiance signal from the target S is given by the convolution of its reflection map $r(x,y)$ and the PSF of the water medium $p(x,y)$, *i.e.*:

$$S = r(x,y) \otimes p(x,y) \quad (7.1)$$

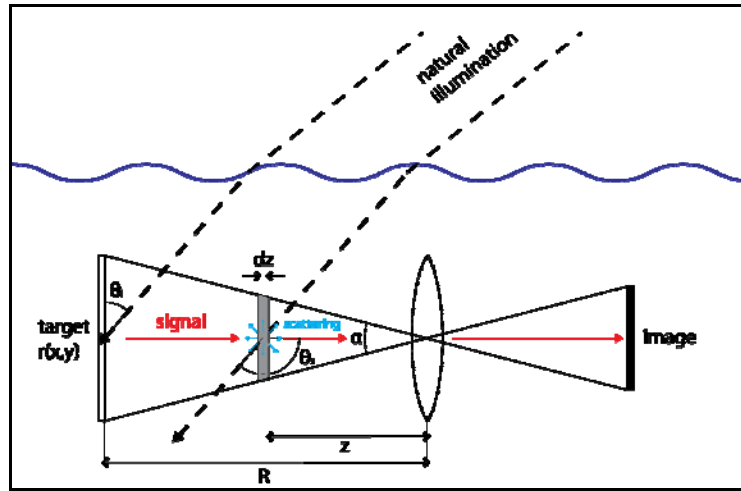


Fig. 7.1 Diagram of imaging geometry for image modeling.

Note here that the target is assumed to completely depolarize the incident light and diffusely reflect the sun light acting like a lambertian surface. The scattered light from each layer dz between the target and the camera reaches the camera experiencing an attenuation equal to $\exp(-cz)$. The total veiling light is then obtained integrating over all the layers between 0 and Z (the distance between the target and the camera). Therefore approximated veiling light irradiance should have the following form, assuming that single scattering is dominant:

$$V(Z) = V(\theta, \infty)(1 - \exp(-cZ)) \quad (7.2)$$

The term $V(\theta, \infty)$ is the background light without the target or with the target extended to infinity, which can be estimated from our angular underwater radiance measurement, $L(\theta)$:

$$V(\theta, \infty) = 2\pi \int_{\theta_{\min}}^{\theta_{\max}} L(\theta) \sin \theta d\theta \quad (7.3)$$

$\theta_{\min}$ and $\theta_{\max}$ are, respectively, the minimum and maximum angle of the particulate scattering extended to the camera aperture. Assuming that the scattered light keeps its polarization when passing through the medium, the linear polarized component of $V_{0/90}(\theta, \infty)$ has a form similar to (7.4), if $L(\theta)$ is replaced by its corresponding measured linear polarized components for 0 and 90°, which correspond to horizontal and vertical polarization:

$$V_{0/90}(\theta, \infty) = 2\pi \int_{\theta_{\min}}^{\theta_{\max}} L_{0/90}(\theta) \sin \theta d\theta \quad (7.4)$$

Then the total signal collected by the camera is the sum of the blurred image and the veiling light, $E_T = S + V$.

A Monte Carlo simulation is performed by using the measured phase function $\beta(\theta)$ to obtain the point spread function at optical depth $\tau = 5$ and single scattering albedo $\omega = 0.9$. In addition to the measured $\beta(\theta)$ from our cruise, which resembles the standard Petzold function (shown in Fig. 7.2), $\beta(\theta)$ values obtained from other typical ocean environments were also used for comparison purposes. Specifically, $\beta(\theta)$ values were collected in Monterey Bay, in waters with mostly biological particulates, in Puget Sound with mostly sediment particles and in the Hudson Plume which belongs to oceanic environments [Gray et al. 2008]. The phase functions and corresponding PSFs were plotted in Fig. 7.2 together with the Petzold functions. The asymmetry factors g , defined as the cosine value weighted by phase function, are also calculated:

$$g = \int_{d\Omega} \cos(\theta) \beta(\theta) d\Omega \quad (7.5)$$

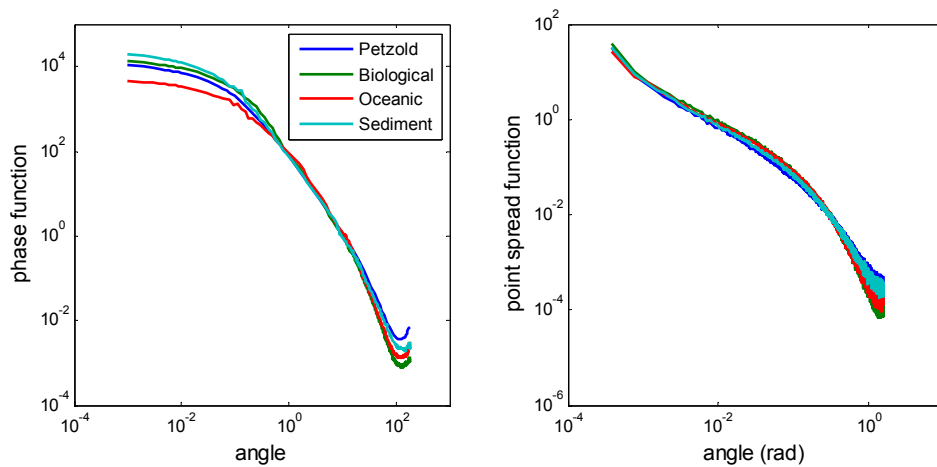


Fig. 7.2. (a) Phase functions and (b) PSF ($\tau = 5$ and $\omega = 0.9$) for four types of typical oceanic waters.

The g values for Petzold, biological, oceanic and sediment cases are 0.91, 0.97, 0.96, and 0.94, respectively. Although the four types of phase functions have very close g values (above 0.9), the oceanic case is the least peaked in the near forward direction. They are angularly flat in the backward direction but with different backscattering levels because of the different refractive indices of the dominant particles. For example, the low refractive index of the biological particles leads to the lowest level of backscattering. Typical polarized angular radiances (corresponding to the angles at which the DOP reaches its maximal value) obtained from field measurements at 510nm were used in Eq. (7.2) to estimate the veiling light background.

When acquiring images of an unpolarized target looking in the horizontal direction (under sun illumination), a polarizer can be placed in front of the camera to remove all or part of the veiling light depending on the DOP of the veiling light itself. This is due to the fact that, in this geometry, the scattered light at an angle corresponding to the horizontal direction (around 100° scattering angle) is partially linearly polarized, as it was shown in previous section. Fig. 7.3 displays the results of the modeled image for the four types of phase functions shown in Fig. 7.3, the target has a maximum reflection equal to 5%.

The blurred images under oceanic environment shows the worst contrast among the four types of water conditions. This is related to the fact that its phase function is the least forward peaked among these four environments. The vertically polarized images always show a better contrast than the intensity and the horizontally polarized ones. The images of the DOP are negatives of the intensity images because the more unpolarized light from

object represents lower total DOP. Of course the effectiveness of image improvement increases with higher DOP of the veiling light.

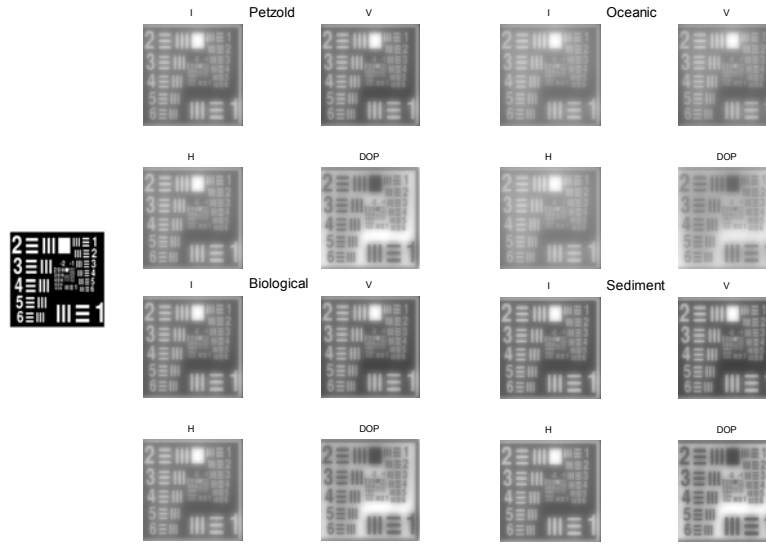


Fig. 7.3. Image modeling results of a standard Air Force Target using four types of phase functions with $\tau = 5$ and $\omega = 0.9$).

7.2 Polarized light traveling through a scattering medium

The utilization of polarized techniques to enhance underwater detection requires that the polarized light produced by illumination sources or interactions such as scattering, surface reflection and transmission keep its polarization when it travels through a scattering medium. The visibility improvement based on polarization is determined by how far the polarization information can be carried through the scattering medium. In order to examine how linear polarized light spreads and depolarizes when travelling through the scattering medium, a Monte Carlo simulation based on slab geometry (Fig. 7.4) is performed. The polarized light is incident along the z axis and the detector array is

placed in the xy plane. The Stokes vector referring to the xz plane was calculated for each array element.

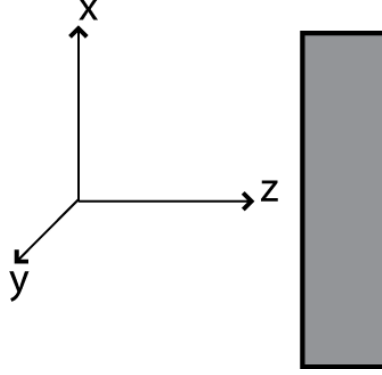


Fig. 7.4. Slab geometry.

The phase function is obtained based on Mie theory with Junge size distribution $f(r) = r^{-\gamma}$ and refractive index 1.15 (relative to water). The following three cases were considered:

$$\begin{aligned}
 (1) & r_{\min} = 0.1 \mu m, r_{\max} = 10 \mu m, \gamma = 3, g = 0.92 \\
 (2) & r_{\min} = 0.1 \mu m, r_{\max} = 10 \mu m, \gamma = 5, g = 0.82 \\
 (3) & r_{\min} = 0.01 \mu m, r_{\max} = 10 \mu m, \gamma = 5, g = 0.71
 \end{aligned} \tag{7.6}$$

The DOP of the transmitted light is plotted in Fig. 7.5 with increasing optical depth for horizontally polarized incident light for the above three cases. The depolarization rate depends heavily on the g factor of the phase function [Ramella – Roman et al. 2005]. The smaller the g factor is, the faster the DOP decreases with the optical depth. It is shown that the forward scattering has a very strong ability to retain its original polarization status even for optical depth higher than 10 when multiple scattering is dominant. Taking a typical g value of 0.91 for example, DOP is over 90% after a polarized light travelling

through 10 optical depths. This can be used to distinguish objects with very different polarization reflectance properties.

The polarized images recorded keeping the same polarization as the incident light spread to larger angles with increasing optical depths as shown in Fig. 7.6. Its circularly symmetrical pattern resembles that of unpolarized incident light, implying limited resolution improvement by using polarized light illumination.

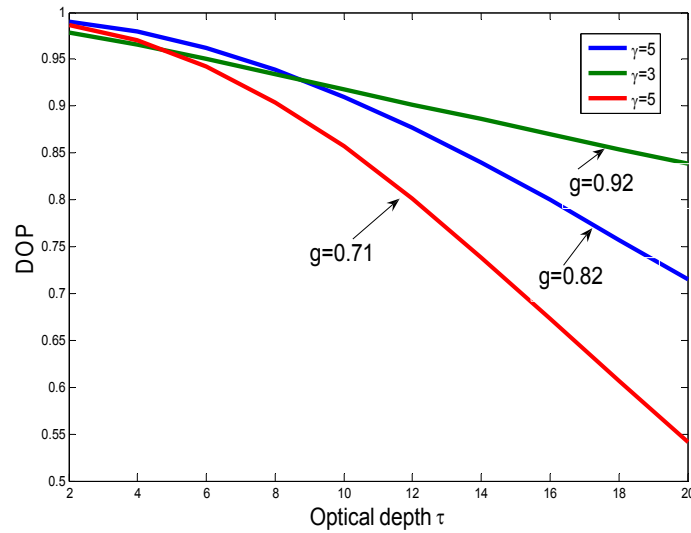


Fig.7.5. DOP as a function of optical depth for three g values.

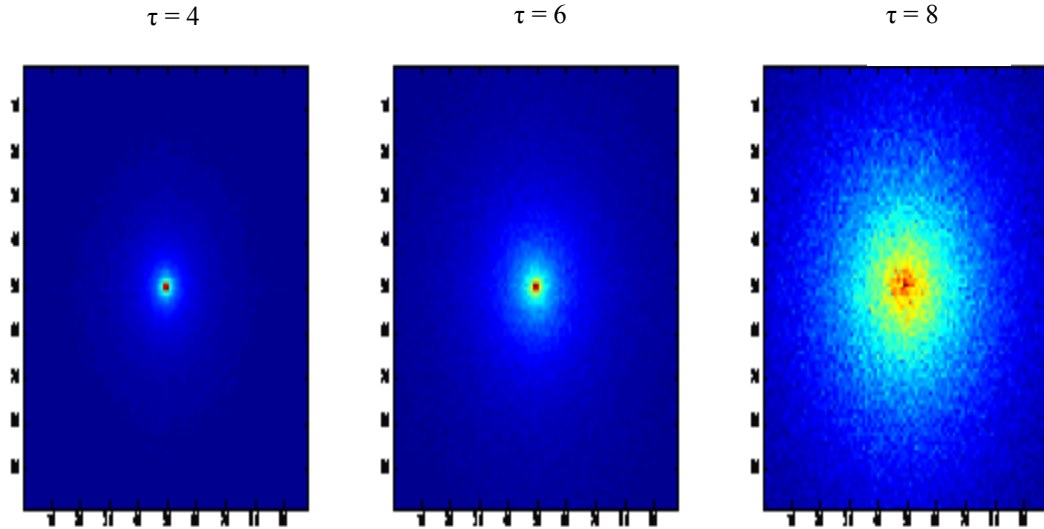


Fig.7.6. The spreading images of the horizontal polarized incident light along z axis with the polarizer oriented horizontally before the detector array.

7.3 Conclusions

The measured polarized radiance and volume scattering data were utilized to model the underwater images of unpolarized targets. The veiling light level is estimated from the measured polarized radiance and the PSF is obtained through a Monte Carlo method using the measured VSF. Taking advantage of the targets' lack of polarization and partially horizontal polarization of the background radiance, the veiling light can be reduced by placing a vertically oriented polarizer in front of the imaging system and thus improving the visibility. Four types of VSFs under typical ocean environments were applied to model the blurred images. At the same optical depth the quality of the underwater images strongly depends on the small angle scattering. The ocean condition with the least peaked forward scattering, the oceanic case, has the worst image quality.

On the other hand, Monte Carlo simulations of the slab geometry for polarized pencil light shows that the scattering medium with high g value has a very strong ability to retain the polarization status of the incident light, which can be utilized to improve the image contrasts for targets with very different polarized reflection properties. Evaluations of imaging the polarized targets by using polarization techniques should be the next steps of research.

8. Estimating particle composition from the polarized water leaving radiance

The focus of this chapter is on the dependence of *in situ* hyperspectral and multiangular polarized data on the size distribution and refractive index of the suspended particles. The degree of polarization (DOP) of the underwater light field in coastal environments was measured and the water-leaving polarized radiance was derived. In-water optical properties were also measured with an AC-9 / AC-s (WET Labs). Absorption and attenuation spectra are then used to derive information on the dissolved and suspended components in the water medium which are used in a vector radiative transfer code which provides the upwelling polarized radiance. The model was run for various values of the refractive index of NAP until the modeled DOP matched the measured one. The relationship between the intensity of the maximum of the DOP and both the refractive index of the NAP and the shapes of their size distributions is analyzed in detail.

Remote sensing measurements of ocean color are directly related to the water leaving spectral radiance which depends on the absorption and scattering properties of the suspended particles. Phytoplankton cells exhibit only weak polarization effects while NAP appreciably affect the polarization signal because of the large index of refraction relative to water¹. It is the real part of the refractive index that primarily determines the scattering properties of the particles. Refractive indices of NAP commonly found in the ocean range from 1.12 to 1.26 [Lide 1994].

The shape of the particle size distribution (PSD) is also very important in determining the scattering properties of a polydispersion. PSDs are usually described using a Junge-

type model, where the distribution is fit with an hyperbolic curve with the slope parameter ξ describing the shape. For marine particles, ξ typically varies between 2.5 and 5, with most values falling within 3.5 – 4.0 [Bader 1970].

In this work, the spectral and geometrical angular dependences of the DOP is related to the slope parameter, ξ , and to the NAP bulk refractive index, n_{NAP} . The results of vector radiative transfer computations are used to estimate n_{NAP} .

8.1 Optical instruments and methods

Underwater polarization measurements were previously described in detail, focusing on the characteristics of the hyperspectral and multiangular polarimeter developed by the Optical Remote Sensing Laboratory. Absorption and attenuation of all in–water constituents except water were measured with an AC–9 (WET Labs). The AC–9 measures spectral attenuation and absorption at 9 wavelengths in the visible and NIR. Data analyzed in this section were collected during a cruise on the R/V “Connecticut” in the coastal areas of New York Harbor – Sandy Hook, NJ region (USA), on July 21–23 2008. The data were collected at eight stations. We selected Station 4 (73°54’002”W, 40°26’905”N, $\theta_{\text{Sun}} = 22.23^\circ$, 12:45 PM) as a case study, since absorption and attenuation of all in–water constituents were available for this station, and the sky was clear blue with no clouds during the polarized data acquisition time.

8.2 Data processing and analysis of underwater measurements

Absorption and attenuation spectra obtained with the WET Labs package are here used to derive information on the dissolved and suspended components in the water medium. Specifically, the quantities derived are: slope of the Junge size distribution, considered as

a model for marine particle size distributions (PSDs), CDOM absorption at $\lambda_0 = 412$ nm, and concentrations of Chlorophyll and nonalgal particles.

Once these values are derived, they are used as input for a vector radiative transfer code, which provides the Stokes parameters for the upwelling radiance of the water body specified in the input file. The model was run while varying the real part of the refractive index of NAP between 1.12 and 1.26 until the modeled DOP matched the measured one.

Relationship between spectral attenuation and size distribution

As we stated before, the PSD and c_p are linked according to the simple relationship:

$$\gamma \approx \xi - 3. \quad (8.1)$$

The derived value of ξ for Station 4 is 3.9, since this value falls in the range of applicability of the above equation, the $\xi - 3$ relationship was adopted in our model (Fig.8.1).

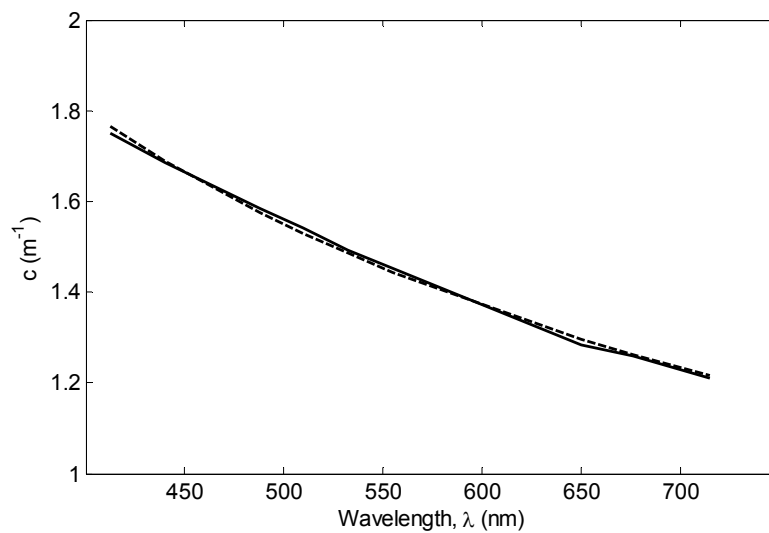


Fig. 8.1. Particulate attenuation spectrum. The solid line represents the *in situ* c_p measurements, the dashed line is the calculated spectrum with $\xi = 3.9 \pm 0.2$.

Fitting of the CDOM absorption spectrum, based on the exponentially decreasing function described in the bio – optical model is used to obtain the value of CDOM absorption at $\lambda_0 = 412$ nm, *i. e.* a_{y0} . Fig. 8.2 shows the fitting result for Station 4.

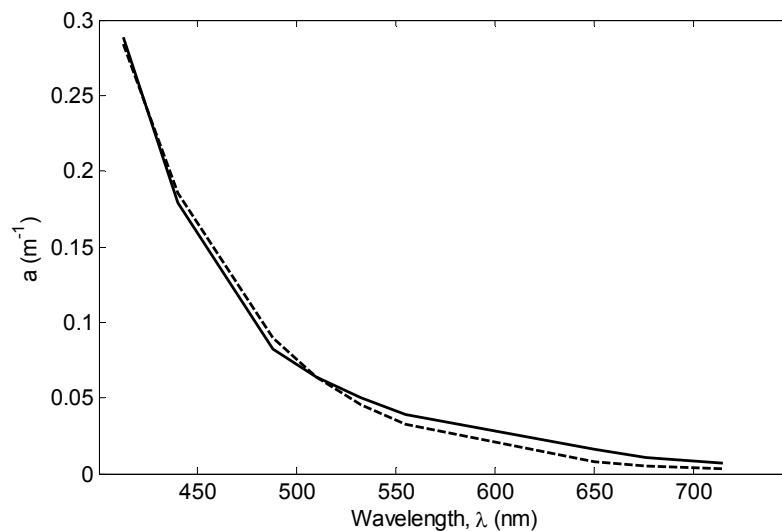


Fig. 8.2. CDOM absorption spectrum. The solid line represents the *in situ* CDOM measurements, the dashed line is the calculated spectrum with $a_{y0} = 0.280 \pm 0.004$ and $S_y = 0.015 \pm 0.002$.

The absorption spectrum of all particulate components can be obtained from the total absorption spectrum, recorded with the AC-9 (WET Labs), once the absorption due to dissolved matter (*i. e.* CDOM) is subtracted. Concentrations of Chlorophyll and nonalgal particles can then be obtained by fitting the total absorption spectrum with the equations referring to particulate absorption described in the bio – optical model, Fig. 8.3.

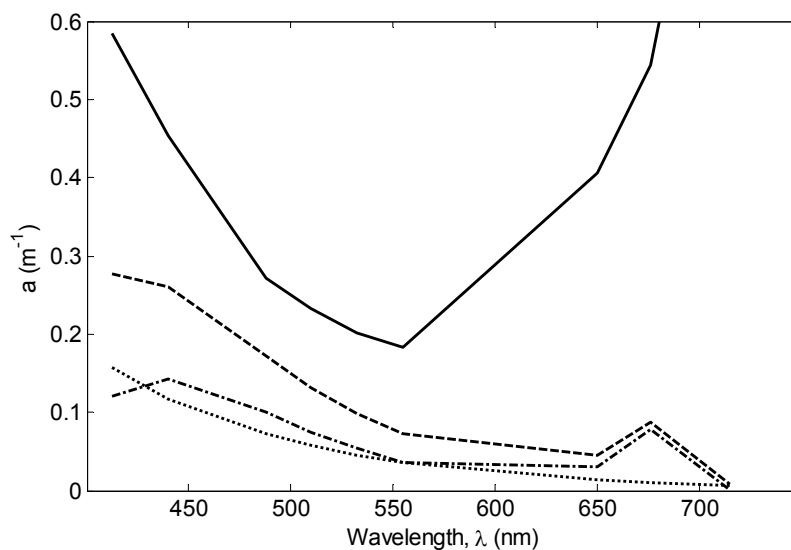


Fig. 8.3. Absorption spectra. The solid line represents the *in situ* total absorption measurements (including water absorption and CDOM absorption), the dashed line is the calculated total particulate absorption spectrum, which is the sum of phytoplankton absorption (dash-dotted line), with $[\text{Chl}] = 8.1 \pm 0.3 \mu\text{g/L}$ and nonalgal particles absorption (dotted line), with $[\text{NAP}] = 3.1 \pm 0.3 \text{ mg/L}$ and $S_{\text{NAP}} = 0.010 \pm 0.003$.

Underwater polarization measurements

Typical plots of the DOP vs. scattering angle, recorded in the main scattering plane at 1m depth are presented in Fig. 8.4. Results are shown for Station 4. The DOP exhibits a bellshaped curve as a function of the scattering angle. The maximum value of this curve and its position are 0.4 and 100° , respectively.

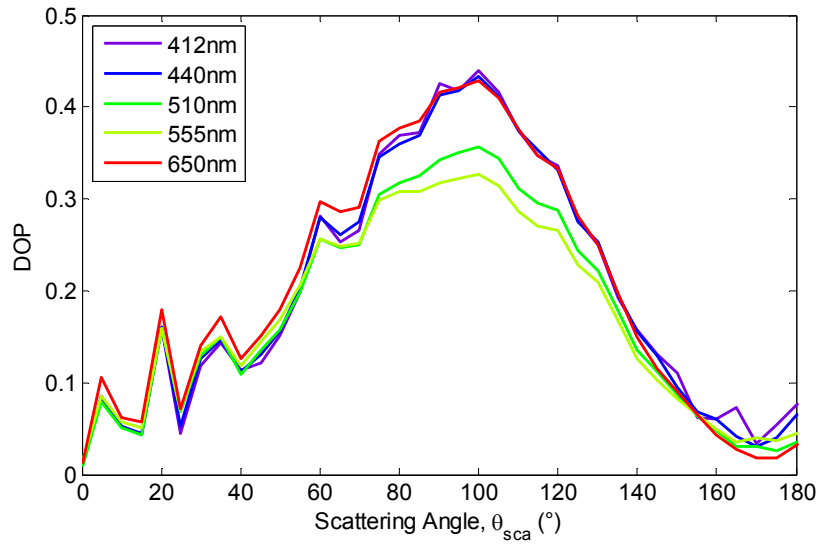


Fig. 8.4. Plots of the DOP vs. scattering angle. The instrument is located in the principal plane 1m below water.

Relationship between DOP and n_{NAP}

Snell's window is the circular region on the water surface above an underwater observer with an aperture angle of approximately 97.5° , within which the entire celestial hemisphere above the water is compressed due to refraction. This area is often seen as a bright circle at the water surface by submerged observers.

The change in the state of polarization of the underwater light field refracted at the air–water interface and transmitted through Snell's window (which is the circular region above an underwater observer with an aperture equal to twice the critical angle, $\theta_c = 48.6^\circ$) of a flat water surface can be described by the following Mueller matrix [Chowdhary 1999]:

$$\mathbf{M} = \frac{1}{4m\mu} \begin{bmatrix} (t_p^*)^2 + (t_r^*)^2 & (t_p^*)^2 - (t_r^*)^2 & 0 & 0 \\ (t_p^*)^2 - (t_r^*)^2 & (t_p^*)^2 + (t_r^*)^2 & 0 & 0 \\ 0 & 0 & 2t_p^* t_r^* & 0 \\ 0 & 0 & 0 & 2t_p^* t_r^* \end{bmatrix}, \quad (8.2)$$

where:

$$\begin{aligned} t_p^*(\mu) &= \frac{2m\mu}{\mu + m\mu_t} \\ t_r^*(\mu) &= \frac{2m\mu}{m\mu + \mu_t} \\ \mu_t^*(\mu) &= \sqrt{1 + (\mu^2 - 1)m^2} \\ \mu &= |\cos(\theta_i)| \end{aligned}, \quad (8.3)$$

where $m = 1.33$ is the index of refraction of water (supposed to be constant with wavelength) and $\theta_t = \arcsin(m \sin(\theta_i))$, with θ_i the angle of incidence measured from the vertical. The elements of this Mueller matrix originate in the Fresnel theory of polarization of light transmitted through the air–water interface. $t_p^*(\mu)$ and $t_r^*(\mu)$ are the Fresnel transmission coefficients. The asterisk denotes illumination from below (see Chowdhary [1999] for details). The assumption of a flat water surface is justified by the fact that the effect of waves on the polarized water leaving radiance is weak [Kawata et al. 1994]. If the Stokes vectors of the incident and transmitted (*i. e.* refracted) light are $\mathbf{S}_i$ and $\mathbf{S}_t$, then:

$$\mathbf{S}_t = \mathbf{M} \cdot \mathbf{S}_i. \quad (8.4)$$

The DOP of the transmitted light, DOP_t , is then given by:

$$DOP_t = \frac{\sqrt{Q_t^2 + U_t^2}}{I_t}. \quad (8.5)$$

Here, the measured value of the DOP recorded at 1 m depth are considered good approximations of the DOP at 0^- , *i. e.* just below water. This assumption is true if the DOP doesn't change significantly when light travels through 1 m of water. This is indeed the case, as reported by Zhou *et al.* [2009]. Plots of the calculated DOP_t vs. viewing angles for Station 4 are shown in Fig. 8.5 (solid lines).

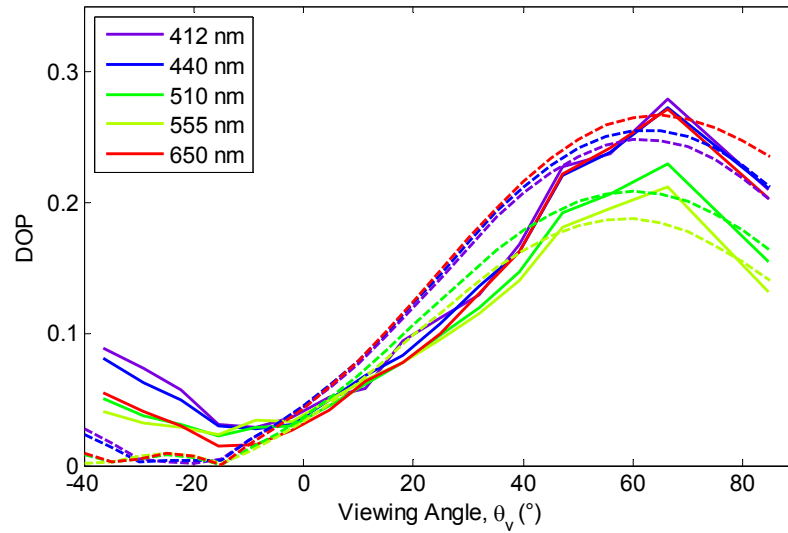


Fig. 8.5. Plots of DOP_t vs. viewing angle, where 0° corresponds to the observer looking down. The solid lines represent the values of DOP_t calculated from *in situ* underwater measurements. The dashed lines are the results of the vector radiative transfer program with $n_{NAP} = 1.16$.

Fig. 8.5 also shows the calculated spectra [Chowdhary et al. 2006] obtained through a coupled atmosphere – ocean polarized radiative transfer model which uses the adding – doubling method to solve the radiative transfer equation, as described in Section 3.5. CDOM absorption at $\lambda_0 = 412$ nm, and concentrations of Chlorophyll and nonalgal

particles were obtained by fitting *in situ* measurements as described in Section 3.5. The scattering properties of suspended particles were also obtained from *in situ* measurements (WET Labs). For Station 4, the number fraction for phytoplankton is 0.067 ± 0.001 , the number fraction for NAP is 0.933 ± 0.002 . The real part of the refractive index, relative to water, of phytoplankton was kept constant at 1.04, while the real part of the refractive index of NAP was varied between 1.12 and 1.26 until the modeled DOP matched the measured one. The plots in Fig 8.5 correspond to $n_{\text{NAP}} = 1.16$.

Figs. 8.6 show the individual results of the simulations, for $\lambda = 555$ nm (Fig. 8.6(a)) and 650 nm (Fig. 8.6(b)). Each curve represents a different NAP refractive index ranging between 1.12 and 1.26 at increments of 0.01. At 650 nm the values of the DOP are higher and more sensitive to particle compositions, when compared with the values at 555 nm²⁰. This can be traced to the fact that, at 650 nm, the contribution of water molecules to the DOP is reduced (*i. e.* the scattering of water is very weak) and the effect of suspended particles is enhanced.

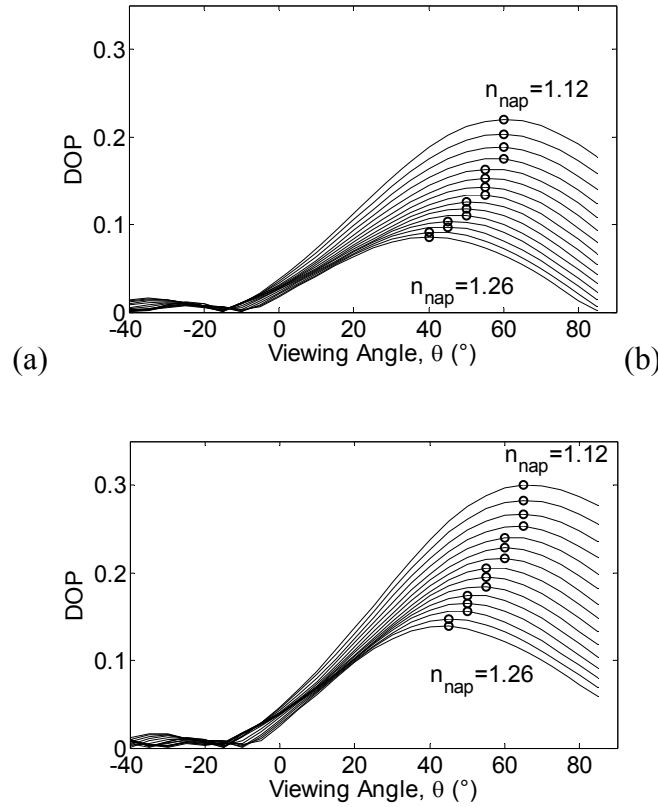


Fig. 8.6. Plots of simulated DOP_t vs. viewing angle, where 0° corresponds to the an observer looking straight down. (a) $\lambda = 555$ nm, (b) $\lambda = 650$ nm.

At both wavelengths, the maximum of the DOP, DOP_{MAX} , (marked with a circle in Fig. 8.6) decreases and shifts toward smaller viewing angles with increasing values of the real part of the refractive index of NAP. The depolarization effect of Mie scatterers with a large value of refractive index is a well known effect reported in several studies regarding optical imaging through biological tissues [for example Jaques and Roman 2000, 2002]. This means that polarized reflectance at 650 nm might allow the determination of the composition of suspended particles. Note, however, that an appropriate viewing geometry

should be adopted (*i. e.* at the angle where the polarization is maximal). This means that multiangular measurements of the polarized reflectance are mandatory.

Fig. 8.7 shows the plots of the intensity of the maximum of DOP as a function of the hyperbolic slope of the differential particle size distribution. In Fig. 8.7, ξ varies between 2.5 and 5, which is typical for marine particles populations. A smaller value of ξ corresponds to a bigger number of large particles. The majority of ξ values falls within 3.5 – 4.0, in this range the maximum of the DOP shows high sensitivity to both ξ and the NAP refractive index, n_{NAP} . For values of $\xi \geq 4$, DOP_{MAX} is almost exclusively function of refractive index. This implies that DOP_{MAX} may be used for obtaining n_{NAP} with reasonably good accuracy under these conditions. For n_{NAP} between 1.14 and 1.22 the all the curves intersect at $\xi \approx 3$. For this value of the Junge slope DOP_{MAX} is independent of n_{NAP} . Plots of DOP_{MAX} vs. ξ for different concentrations of NAP and phytoplankton particles still intersect at $\xi \approx 3$, but the corresponding value of DOP_{MAX} varies.

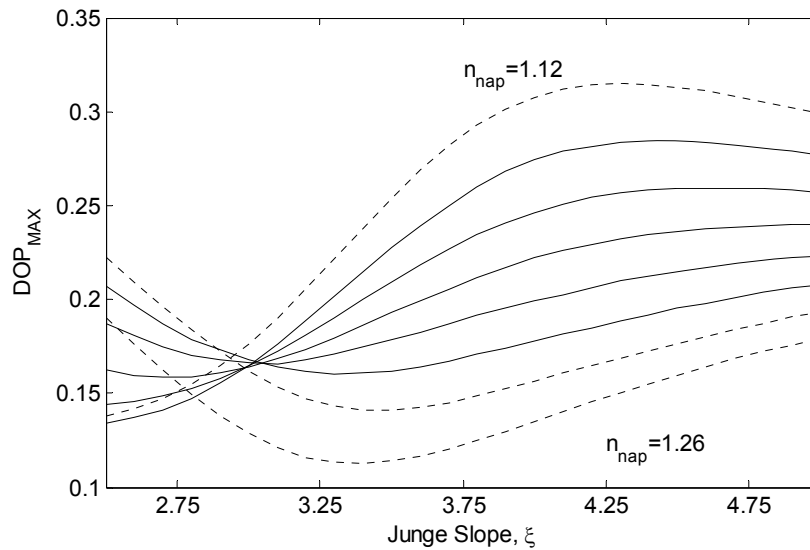


Fig. 8.7. The relationship between the intensity of the maximum of the DOP, DOP_{MAX} , and the hyperbolic slope, ξ , for refractive indices of NAP

ranging between 1.12 and 1.26, at increments of 0.02. The solid lines correspond to $n_{\text{NAP}} = 1.14$ to 1.22.

8.3 Conclusions

In this section the sensitivity of the above water polarized reflectance in the visible spectrum to particle compositions (*i. e.* refractive index) and size distributions was examined for case 2 waters. The first part of the study deals with *in situ* optical measurements, specifically absorption, attenuation and polarized underwater radiance measurements. The information gained is used as input in the adding–doubling computations which provides the water–leaving polarized radiance. The simulations revealed that the polarized reflectance is sensitive to both the refractive index of the NAP, n_{NAP} , and the shapes of their size distributions, ξ . Specifically, the angular position and the intensity of DOP_{MAX} , the maximum of the DOP, are functions of n_{NAP} and ξ .

The simulations showed that multiangular polarized water leaving radiance collected in the visible is a powerful tool that could be systematically used to gain auxiliary and complementary information on suspended particles. The main advantages of the method proposed here are as follows. (1) Only optical measurements are required, and (2) the DOP approach is weakly dependent on calibration, since the DOP itself is a ratio.

9. Polarimetric observations in turbid waters (New York Harbor, NY 2009)

9.1 *Estimating particle composition from the polarized water leaving radiance*

Data analyzed in this section were collected during a recent cruise in the coastal areas of New York Harbor, NY (USA), on July 10 2009. The data reported in this section were collected at three stations. We selected Station 1 (74°02'596''W, 40°37'020''N, $\theta_{Sun} = 63^\circ$, wind speed = 3 m/s, 10:06 AM) as a case study, since absorption and attenuation of all in-water constituents (measured with an AC-s, WET Labs), and PSD (measured with a LISST 100X particle size analyzer, SEQUOIA Scientific) were available for this station, and the sky was clear blue with no clouds during the polarized data acquisition time.

As mentioned before, PSDs are usually described using a Junge-type model, where the distribution is fit with a hyperbolic curve with the slope parameter ξ describing the shape. If $N(x)$ is the number of particles per unit volume with diameter (therefore assuming spherical particles) bigger than x , then the Junge-type *cumulative* size distribution is then [Bader 1970]:

$$N(x) = k \left(\frac{x}{x_0} \right)^{-m}, \quad (9.1)$$

where k gives the scale, x_0 is a reference diameter, and $-m$ represents the slope of the particle size distribution when plotted in a log – log graph.

In ocean optics the quantity most relevant is not the *cumulative* size distribution, but the *number* size distribution $n(x)$. the number size distribution is such that $n(x)dx$ represent the number of particles per unit volume in the diameter interval $(x, x + dx)$. The relationship between the cumulative and the number Junge size distributions is:

$$n(x) = |dN(x)/dx| = kmx_0^m x^{-m-1} \equiv Kx^{-\xi}, \quad (9.2)$$

Where $K = kmx_0^{-m}$ and $\xi = m + 1$. Usually, the quantity of interest is the slope of the plot of $\log n(x)$ vs. $\log(x)$. As mentioned before, marine particles usually have ξ values between 2.5 and 5, with $\xi = 3.5 - 4.0$ being typical. Fig. 9.1 shows the number distribution obtained from LISST and the fitted curve, considering the slope obtained from the attenuation spectrum (Fig. 9.2). The only unknown is, therefore, K . The measured PSD shows that the number of small particles is actually slightly smaller than what obtained from the analysis of the attenuation spectrum. Nonetheless, the two curves are parallel in most for most of the particles sizes, meaning that the slope of the PSD obtained from the attenuation spectrum is a good proxy for the slope of the actual PSD. For simplicity, we will, therefore, keep using a Junge-type size distribution with the slope calculated from the attenuation spectrum.

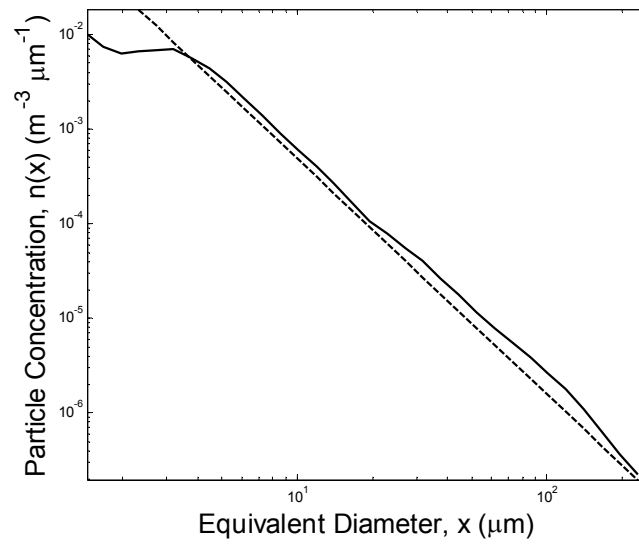


Fig. 9.1. PSDs for Station 1. The solid curve is obtained from LISST measurements. The dashed line is the result of the fitting using Eq. (9.2).

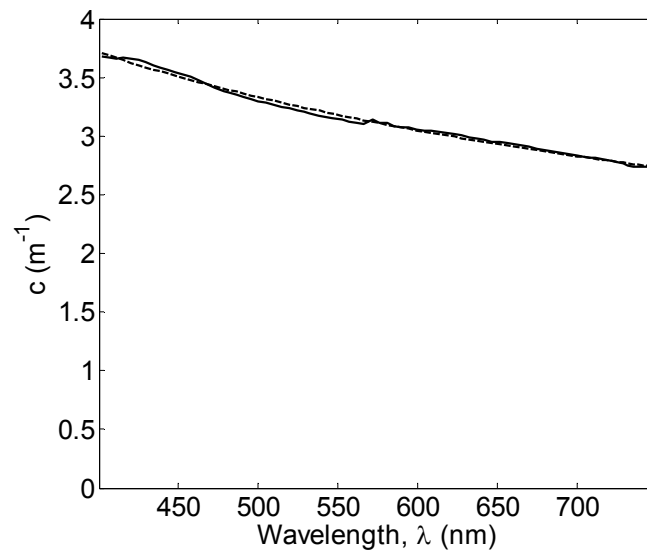


Fig. 9.2. Particulate attenuation spectrum. The solid line represents the *in situ* c_p measurements, the dashed line is the calculated spectrum with $\xi = 3.5$.

Fig. 9.3 and 9.4 show the fitting result for Station 1. Fig. 9.5 shows all the contributions to the absorption spectrum. Fig. 9.6 shows the result of the fitting of the scattering spectrum. It can be seen that the contribution of phytoplankton to the total scattering is very small, while nonalgal particles dominate the spectrum.

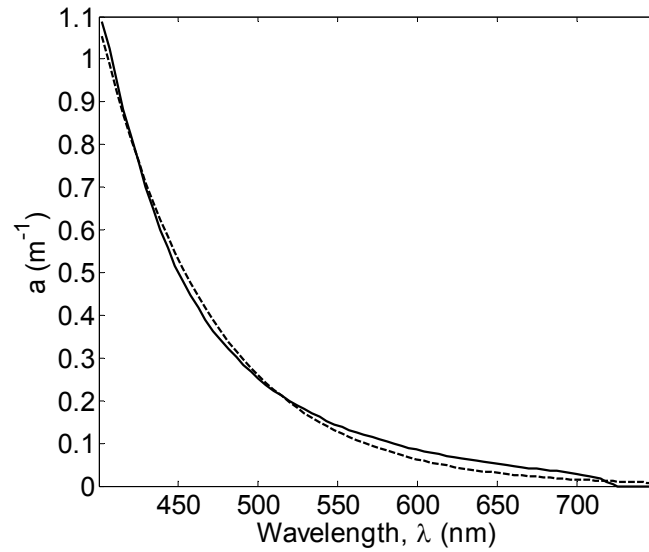


Fig. 9.3. CDOM absorption spectrum. The solid line represents the *in situ* CDOM measurements, the dashed line is the calculated spectrum with $a_{y0} = 1.08$ and $S_y = 0.0143$.

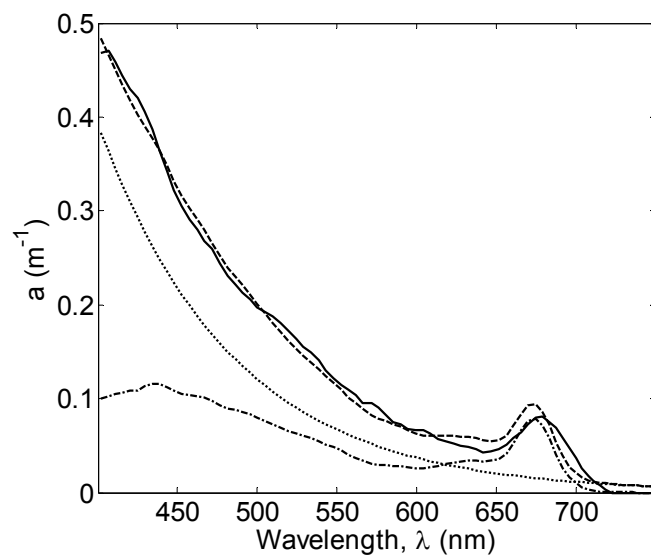


Fig. 9.4. Absorption spectra. The solid line represents the *in situ* total particulate absorption measurements, the dashed line is the calculated total particulate absorption spectrum, which is the sum of phytoplankton absorption (dash-dotted line), with $[\text{Chl}] = 8.1 \mu\text{g/L}$ and nonalgal particles absorption (dotted line), with $[\text{NAP}] = 8.4 \text{ mg/L}$.

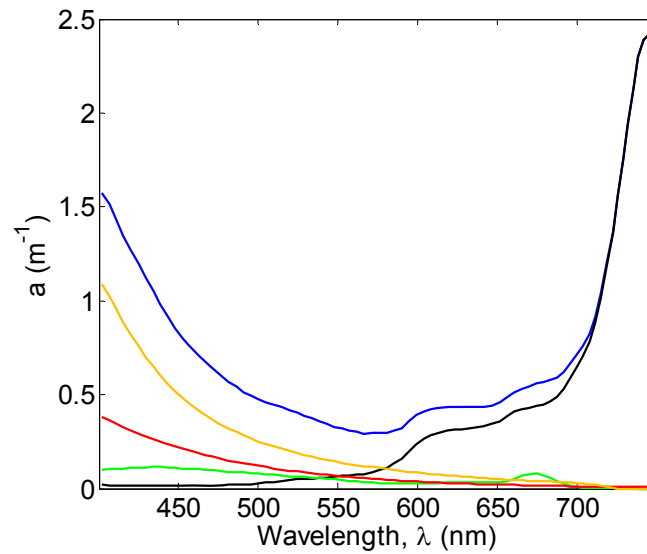


Fig. 9.5. Absorption spectra. The green line represents the phytoplankton absorption spectrum, the red line is the NAP's contribution to the absorption and the yellow – orange line is the CDOM's. The black line is the water absorption and the blue line is the total absorption.

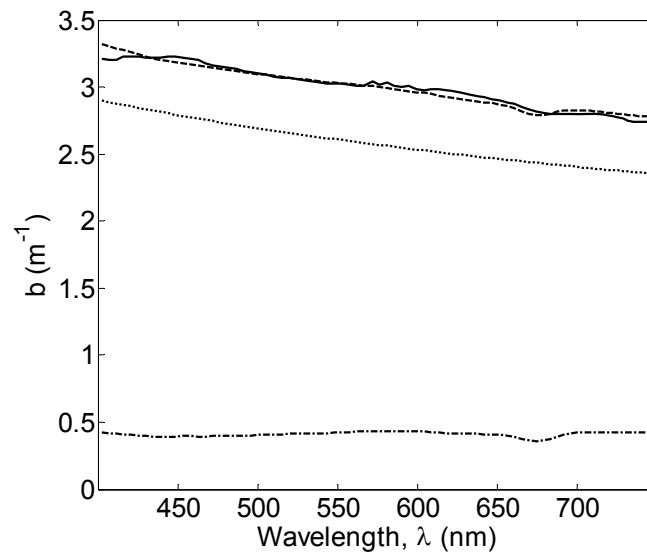


Fig. 9.6. Scattering spectra. The solid line represents the total *in situ* scattering measurements given by $c_p - a_p$, the dashed line is the calculated

total scattering spectrum, which is the sum of phytoplankton scattering (dash-dotted line) and nonalgal particles scattering (dotted line).

Plots of the calculated DOP_t vs. viewing angles for Station 1 are shown in Fig. 9.7 (solid lines) together with the results of the coupled atmosphere – ocean polarized radiative transfer model. The improvements respect to the previous set of measurements consists in the fact that, now, the above – water viewing angles are equally spaced (5°), allowing for a better comparison with the radiative transfer calculations. This emphasizes the accuracy of the *in situ* measurements, in fact the sensors were rotated only of fractions of degrees near the edge of Snell's window but the changes in the DOP_t are still noticeable. As before, the real part of the refractive index of NAP was varied until the modeled DOP matched the measured one. The plots in Fig. 9.7 correspond to $n_{NAP} = 1.13$.

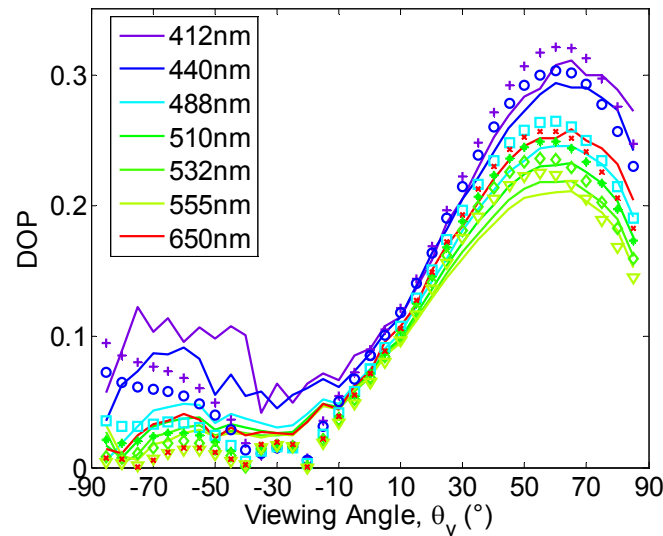


Fig. 9.7. Plots of DOP_t vs. viewing angle, where 0° corresponds to an observer looking down. The solid lines represent the values of DOP_t

calculated from *in situ* underwater measurements. The dotted lines are the results of the vector radiative transfer program with $n_{\text{NAP}} = 1.13$.

9.2 Spectral behavior of the DOP in the Chlorophyll fluorescence region

The next step consists in analyzing the spectral behavior of the DOP_t . Plot of the maximum DOP_t (corresponding to viewing angle equal to 60°) transmitted through the air – water vs. wavelength is shown in Fig. 9.8. The calculated values nicely reproduce the spectral features of the DOP_t . We extended the analysis even further, looking at the region of Chlorophyll fluorescence more in detail. Between 650 and 725 nm radiative transfer calculations were carried out at one nanometer resolution. Our goal was to analyze the effect of Chlorophyll fluorescence in the DOP_t . In fact, since the radiative transfer program didn't include fluorescence, the spectral differences between the measured and the calculated spectra will give us directly the contribution of fluorescence to the DOP_t . Fig. 9.9 shows this spectral comparison only in the range of interest, *i.e.* between 650 and 725 nm.

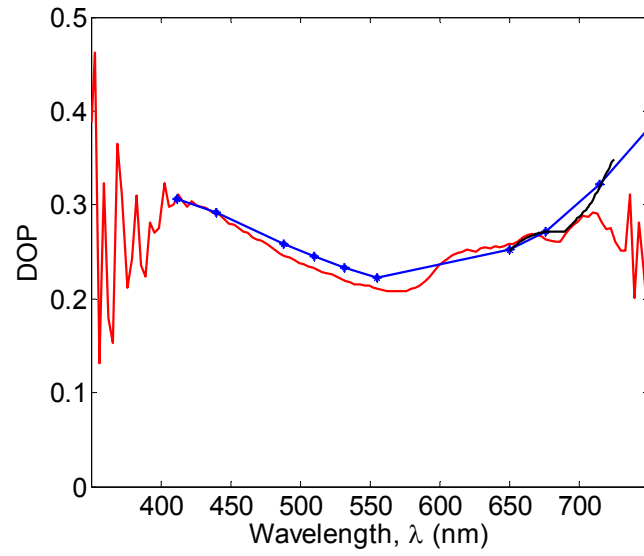


Fig. 9.8. Spectral behavior of the measured (red) and simulated (blue) DOP_t . The blue line represents simulations performed at wavelengths correspondent to the AC-9 wavelengths. The black line represents the result of simulations carried out at one nanometer resolution, in the spectral range 650 – 725 nm.

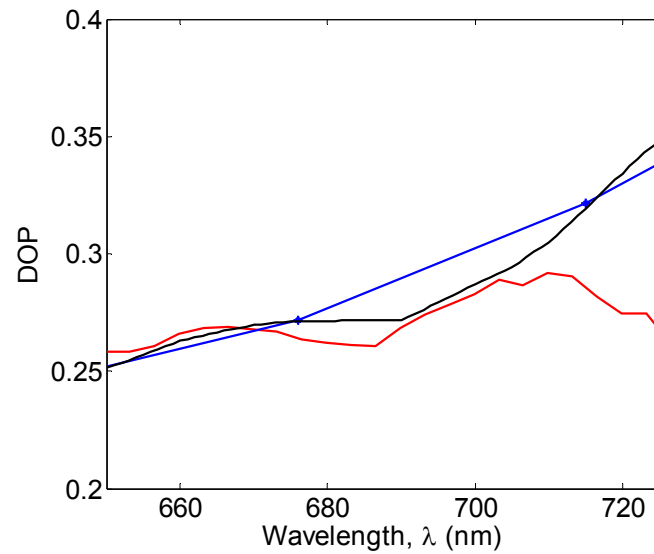


Fig. 9.9. Same as Fig. 9.8 but only the wavelength range 650 – 725 nm is considered.

The model predicts that if fluorescence wasn't present, the DOP_t would be essentially flat in the spectral region 670 – 690 nm. After 690 nm, because of the increasing water absorption, the absolute values of radiance become very small, leading to big uncertainties in the calculation of the DOP_t . Even if the model predicts an increase of the DOP_t after 690 nm, this feature can hardly be measured.

9.3 Out – of – plane polarimetric observations

As reviewed by Mobley [Mobley, 1999], a detector with a finite FOV pointing toward a wavy sea surface is detecting not only the water – leaving radiance but also reflected light from a very large part of the sky (a significant number of rays can be reflected into directions that are several degrees away from the specular direction, especially when the wind speed increases). If this part of the sky contains the sun, then the measurements will

be inevitably affected by Sun glitter. Having this in mind, we explored the variations of the DOP with azimuth angle, ϕ . Fig. 9.10 shows the behavior of the DOP_t for azimuth viewing angles equal to 30° and 60° , respectively.

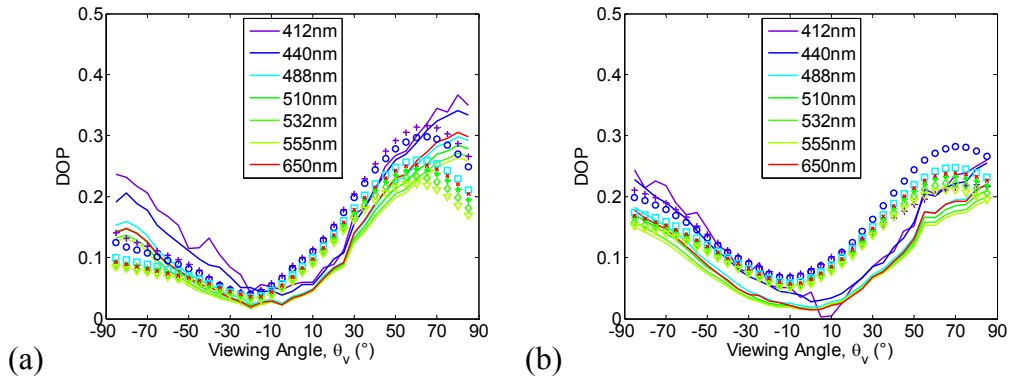


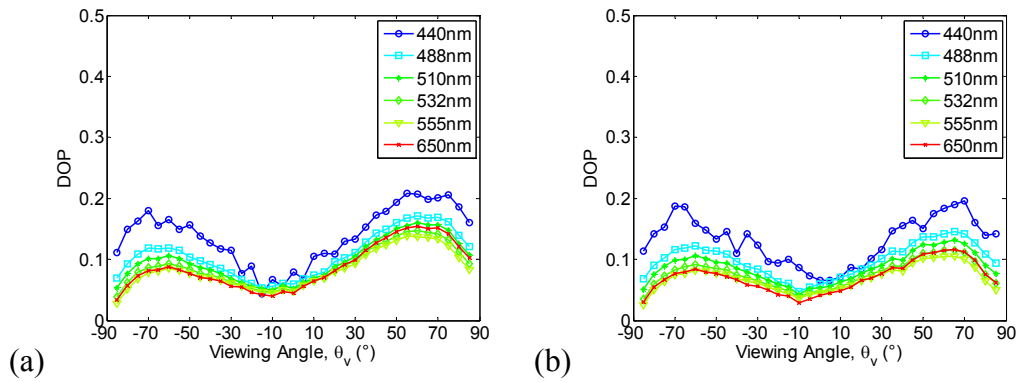
Fig. 9.10. Plots of DOP_t vs. viewing angle, where 0° corresponds to an observer looking down. The left panel is for $\phi = 30^\circ$ and the right panel is for $\phi = 60^\circ$. The solid lines represent the values of DOP_t calculated from *in situ* underwater measurements. The dotted lines are the results of the vector radiative transfer program with $n_{NAP} = 1.13$.

The results imply that polarimetric analysis might be extended to off – principal plane viewing angles without losing the wavelength dependence found in principal plane observations. Moreover, absolute values of the DOP_t are comparable to those found in the principal plane, thus corroborating the results of Chowdhary *et al.* [Chowdhary *et al.* 2006].

Fig. 9.10 also shows the comparisons between radiative transfer simulations and experimental measurements for the last two viewing geometries. The match in these cases is satisfactory but not perfect. Measurements taken at viewing angles off – principal

plane require a longer acquisition time when compared with in – plane measurements. Errors in the orientation of the instrumentation inevitably affect the result much more in this case. The plan for the future is to implement an automatic heading system for the instrumentation.

Fig. 9.11 shows plots of the DOP_t vs. viewing angle for two other stations during the same cruise. Both stations present high loads of particle concentrations; for Station 2 (Fig. 9.11(a)), $[Chl] = 16.5 \mu\text{g/L}$, $[NAP] = 24.8 \text{ mg/L}$; for Station 3 (Fig. 9.11(b)), $[Chl] = 23.3 \mu\text{g/L}$, $[NAP] = 17.0 \text{ mg/L}$. Fig. 9.11 also shows plots of the DOP_t vs. viewing angle for azimuth angle $\phi = 30^\circ$ and 60° for Station 3. The comparison between Fig. 9.11(b) and (c), (d) reveals that there's very little change in the magnitude of the DOP_t when moving away from the main scattering plane.



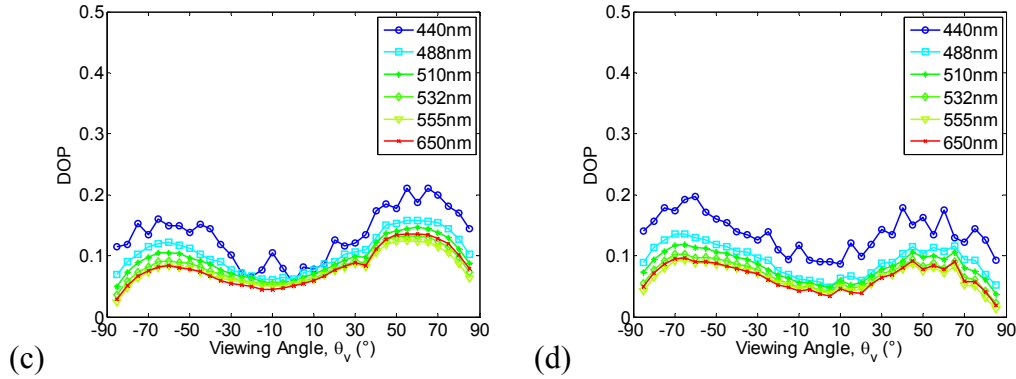


Fig. 9.11. Plots of DOP_t vs. viewing angle, where 0° corresponds to an observer looking down. The top left panel is for Station 2 and the top right panel is Station 3. The bottom panel refers to measurements taken 30° and 60° outside the main scattering plane, Station 3. The curve corresponding to 412 nm is not shown for these stations; it is very noisy due to the high absorption in the blue part of the spectrum.

9.4 The e – vector orientation

During a second trip in the same area of the New York harbor, approximately at the same stations, on July 09 2009 another set of measurements was taken. This time the main goal was to scan the whole range of scattering angles (from 0 to 180°). In fact, we will focus our attention on the determination of the e – vector orientation. For Station 2 ($74^\circ 02' 600''W$, $40^\circ 40' 980''N$, $\theta_{Sun} = 20^\circ$, wind speed = 2.5 m/s, 12:30 PM), $[Chl] = 12.3 \mu g/L$, $[NAP] = 9.5 mg/L$. In Fig. 9.14 results are presented for three different azimuth angles, 0° , 30° and 60° , respectively.

What the theory predicts is that χ is identically equal to zero in the main scattering plane, Fig. 9.14(a). Our measurements show that the e – vector orientation is approximately zero but not everywhere. For most of the scattering angles the angle between the e – vector and the horizontal direction is very small. The fact that it is not exactly zero may be attributed to the small uncertainties in the azimuthal orientation of the polarimeter (an error of $5 - 10^\circ$ might be not uncommon). Since the sensors were not exactly oriented with the main scattering, the Stokes vector component U is not exactly zero (which is how it should be in the main scattering plane), and χ is, consequently, not exactly zero. As we move away from the main scattering plane, U increases and the e – vector orientation deviates from the horizontal direction. At $\varphi = 60^\circ$ (Fig 9.12(c)) the average e – vector orientation is $\approx 25^\circ$. There's also something interesting happening at high scattering angles, at least for $\varphi = 0^\circ$, and for $\varphi = 30^\circ$: the angle that the e – vector makes with the horizontal direction increases up to $\approx 30^\circ$. This wasn't expected and needs further investigation, we will analyze the behavior of the e – vector in other water conditions, to see if multiple scattering is responsible for these deviations.

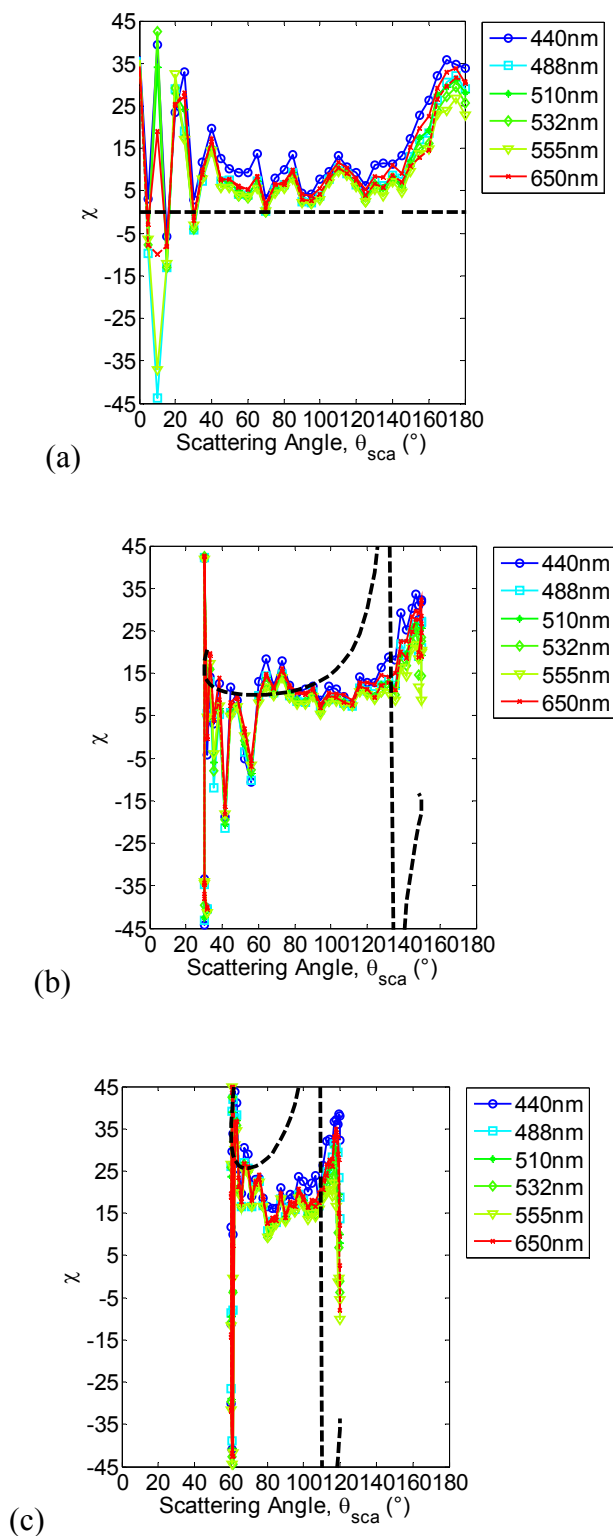
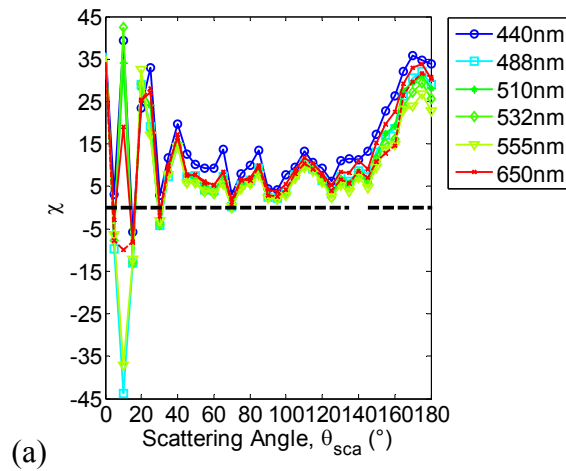


Fig. 9.12. Plots of the e – vector orientation vs. scattering angle, for Station 2. The dashed thick black line is the e – vector orientation predicted by

Rayleigh theory. Fig. (a) corresponds to $\varphi = 0^\circ$, Fig. (b) to $\varphi = 30^\circ$, Fig. (c) to $\varphi = 60^\circ$.

For another station during the same field trip, *i.e.* Station 3 ($74^\circ 00.578$ W, $40^\circ 45.329$ N, $\theta_{Sun} = 27^\circ$, wind speed = 2.5 m/s, 2:30 PM), [Chl] = 9.3 $\mu\text{g/L}$, [NAP] = 11.0 mg/L we analyzed the behavior of the e – vector. Station 3 was geographically located further up the Hudson River, just north of the George Washington Bridge. The concentration of suspended particles is very similar to the one for Station 2; this allows us to compare the two stations. Fig. 9.13 shows the results of the calculation of the e – vector orientation for Station 3. The results for the two different stations are consistent.



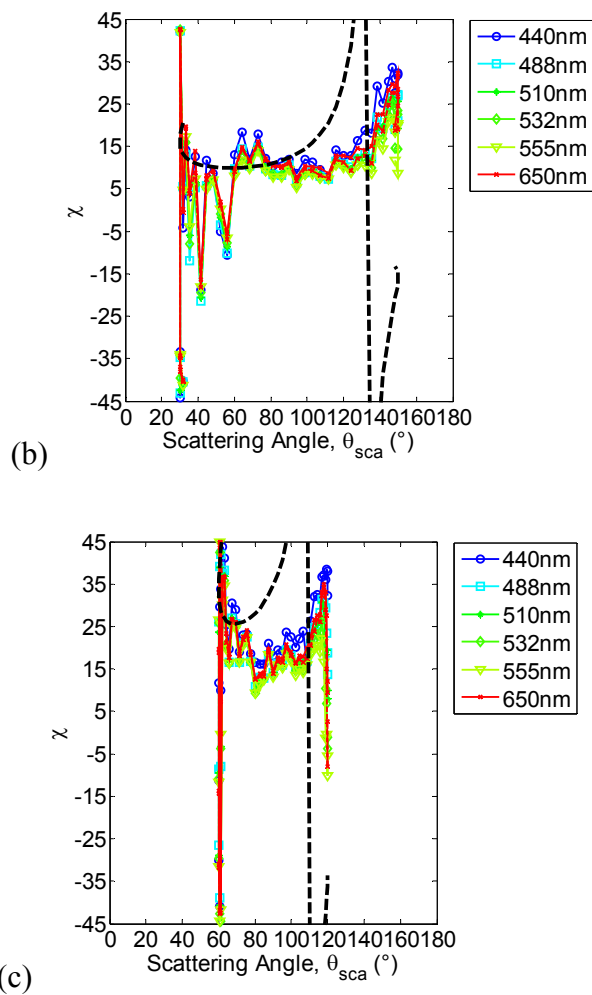


Fig. 9.13. Plots of the e – vector orientation vs. scattering angle, for Station 3. The dashed thick black line is the e – vector orientation predicted by Rayleigh theory. Fig. (a) corresponds to $\phi = 0^\circ$, Fig. (b) to $\phi = 30^\circ$, Fig. (c) to $\phi = 60^\circ$.

10. Polarimetric observations in clear waters (Norfolk, VA 2009)

10.1 Estimating particle composition from the polarized water leaving radiance

On August 17 – 20 2009, measurements were taken in the Chesapeake Bay and in the Atlantic Ocean, approximately 20 – 25 miles away from the shore of Norfolk, VA (USA). For Station 1 ($73^{\circ} 30.9554$ W, $36^{\circ} 53.7833$ N, $\theta_{Sun} = 38^{\circ}$, 11:00), the concentration of phytoplankton is $1.2 \mu\text{g/L}$ while the concentration of NAP in this case is negligible, $\xi = 3.9$. Fig. 10.1 (left) shows how clear this water was. Fig. 10.1 (right) shows a picture of the water at Station 2, where the concentration of Chlorophyll was higher.

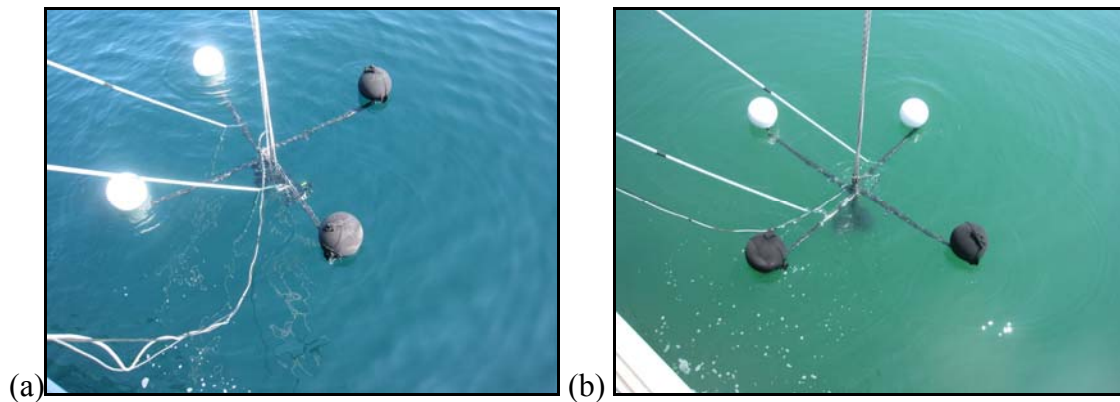


Fig. 10.1. The polarimeter at Station 1 (left) and Station 2 (right).

Here we report the results of the absorption measurements recorded with the WET Labs package for Station 1, to have an idea of the water conditions. Fig 10.2 shows the

absorption spectra of CDOM and the suspended particulate matter (which, in this situation, is all phytoplankton). The figures also show the results of the fitting with the equations of the bio – optical model described before. Fig. 10.3 shows all the contributions to the absorption spectrum

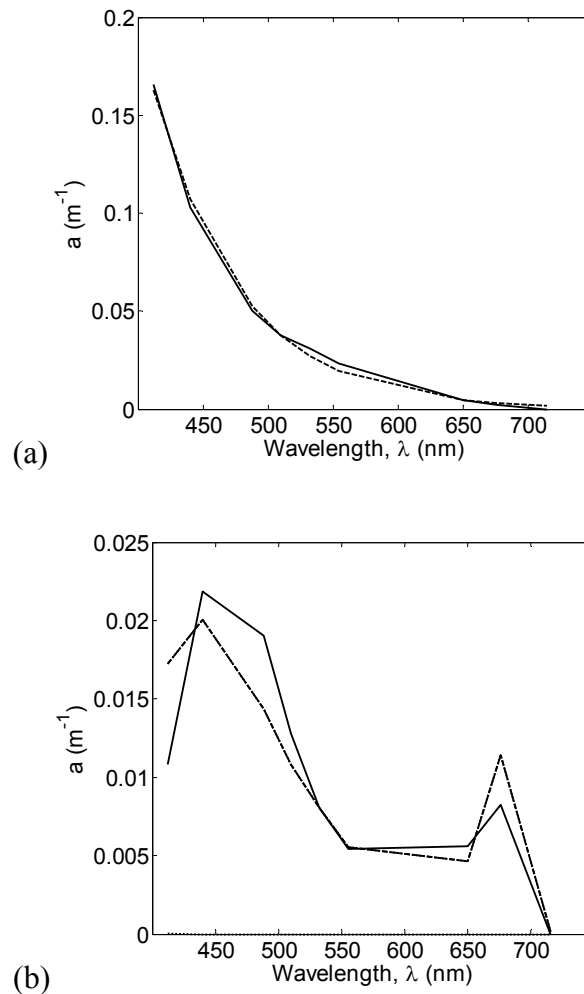


Fig. 10.2. (top) CDOM absorption spectrum. The solid line represents the *in situ* CDOM measurements, the dashed line is the calculated spectrum. (bottom) The solid line represents the *in situ* total particulate absorption measurements, the dash–dotted line is the calculated total particulate

absorption spectrum, which is given only by the phytoplankton absorption with $[\text{Chl}] = 1.2 \mu\text{g/L}$. NAP absorption is negligible at this station.

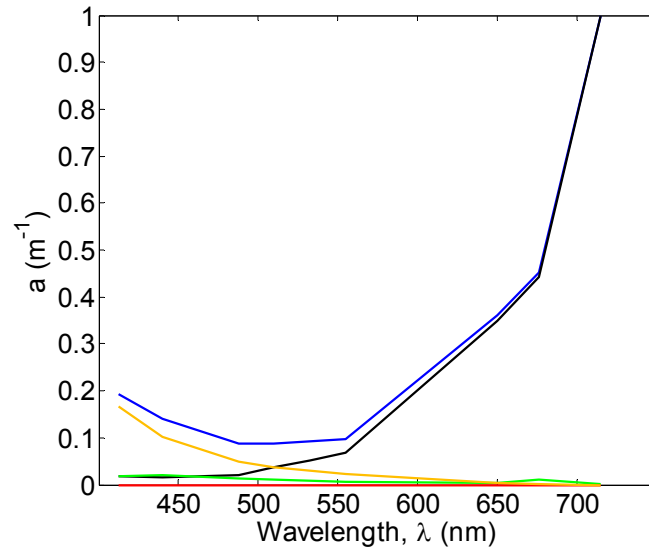


Fig. 10.3. Absorption spectra. The green line represents the phytoplankton absorption spectrum, the red line is the NAP's contribution to the absorption and the yellow – orange line is the CDOM's. The black line is the water absorption and the blue line is the total absorption.

Fig. 10.4 shows the number distribution obtained from LISST and the fitted curve, considering the slope obtained from the attenuation spectrum, Fig. 10.5, (recorded with the WET Labs package, as well). The measured PSD shows that the number of large particles is actually bigger than what obtained from the analysis of the attenuation spectrum. Nonetheless, the two curves are parallel for most of the particles sizes, meaning that the slope of the PSD obtained from the attenuation spectrum is a good

proxy for the slope of the actual PSD. For simplicity, we will, therefore, keep using a Junge-type size distribution with the slope calculated from the attenuation spectrum.

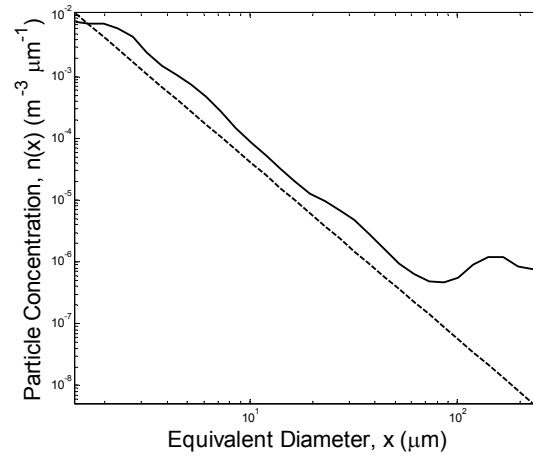


Fig. 10.4. PSD for Station 1. The solid curve is obtained from LISST measurements. The dashed line is the result of the fitting.

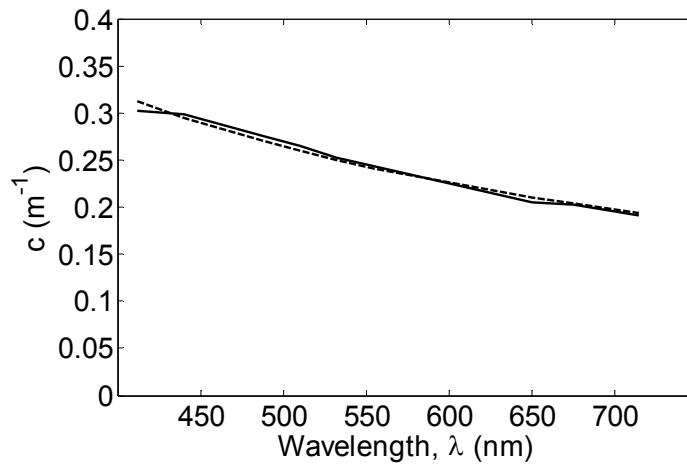
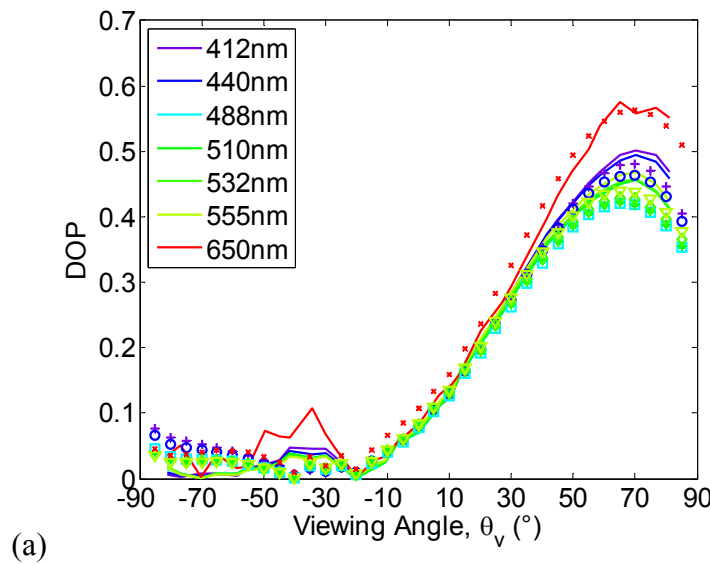


Fig. 10.5. Scattering spectra. The solid line represents the total *in situ* scattering measurements given by $c_p - a_p$, the dashed line is the calculated total scattering spectrum, which is given only by the phytoplankton scattering.

Fig. 10.6 shows a comparison between DOP_t and results of radiative transfer calculations. Also in this case the match is very good. Our impression was that Rayleigh scattering was the dominant process here, being the concentration of suspended particles very small. We therefore, plotted the following equation:

$$DOP = DOP_{MAX} \frac{\sin^2(\theta_{sca})}{1 + \cos^2(\theta_{sca})}, \quad (10.1)$$

considering $DOP_{MAX} = \max [DOP (650 \text{ nm})]$, vs. the scattering angle, θ_{sca} , Fig. 10.7 shows that our guess was correct. Notice that when Rayleigh scattering is the dominant process, the peak of the bell shaped curve is exactly at $\theta_{sca} = 90^\circ$. The only relevant differences between the well known Rayleigh scattering curve and the measured DOP appear at small scattering angles. In this region, when the sensors are looking directly at the incoming sun light, the surface waves play the major role.



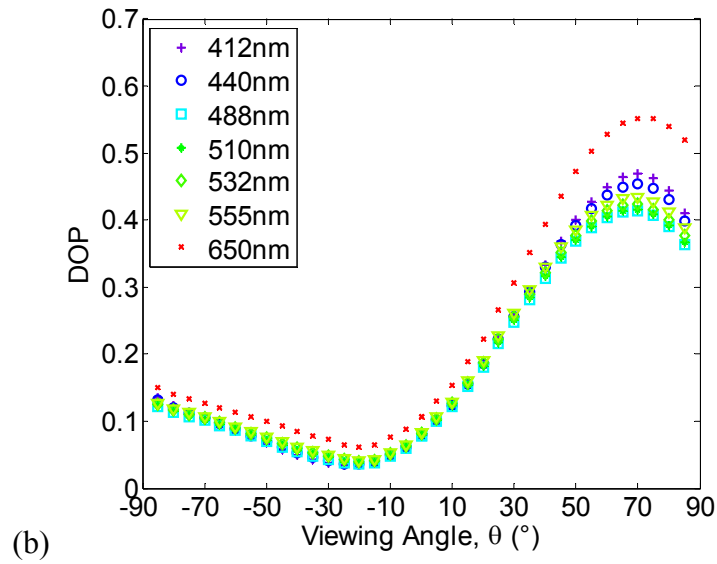


Fig. 10.6. (a) Plots of DOP_t vs. viewing angle, where 0° corresponds to the an observer looking down. The solid lines represent the values of DOP_t calculated from *in situ* underwater measurements. The dotted lines are the results of the vector radiative transfer program with $n_{\text{phyto}} = 1.04$. (b) Same as (a) but the azimuthal angle (ϕ) is now 30° .

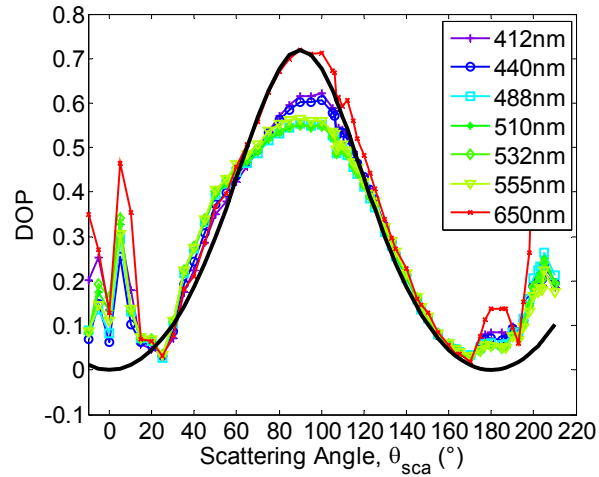


Fig. 10.7. Plots of the DOP vs. scattering angle. The instrument is located in the principal plane 1m below water. The solid black line represents Rayleigh scattering at 650 nm.

Fig. 10.8 shows the spectral behavior of the underwater DOP for some relevant scattering angles. As expected, the DOP reaches its maximal values over the entire spectrum at 90° scattering angle.

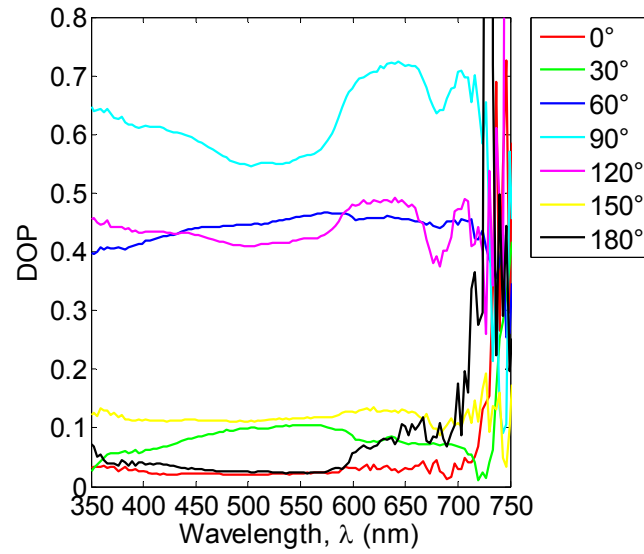


Fig. 10.8. Spectral behavior of the underwater DOP for few relevant scattering angles.

Station 3 was located in the Chesapeake Bay ($75^{\circ} 52.4573$ W, $36^{\circ} 53.7833$ N, $\theta_{Sun} = 27^{\circ}$, 14:00), the concentration of phytoplankton was $11.3 \mu\text{g/L}$, while the concentration of NAP was 3.0 mg/L , $\xi = 4.2$. Here we report the results of the absorption measurements recorded with the WET Labs package for Station 1, to have an idea of the water conditions. Fig. 10.9 shows the absorption spectra of CDOM and the suspended particulate matter. The figures also show the results of the fitting with the equations of the bio – optical model described before. Fig. 10.10 shows all the contributions to the absorption spectrum.

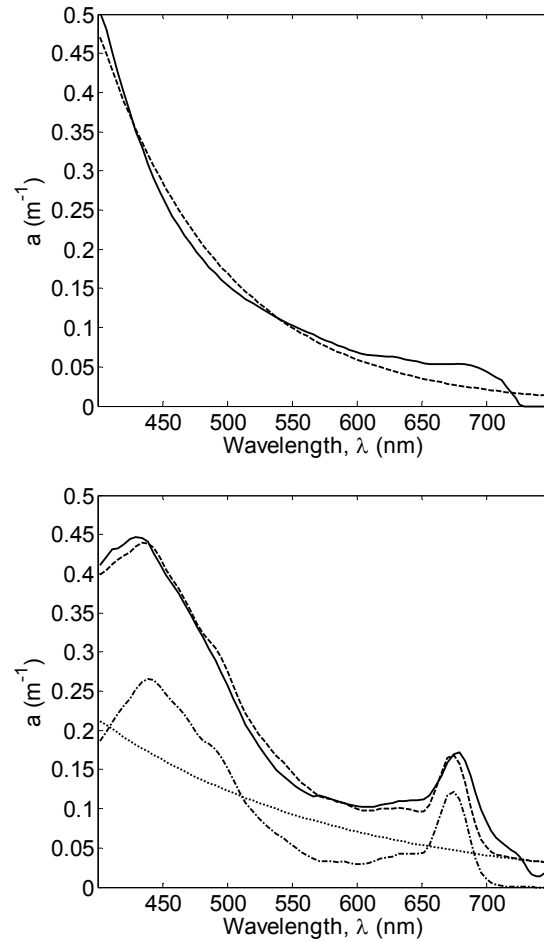


Fig. 10.9. (top) CDOM absorption spectrum. The solid line represents the *in situ* CDOM measurements, the dashed line is the calculated spectrum with $a_{y0} = 0.4807$ and $S_y = 0.0104$. (bottom) The solid line represents the *in situ* total particulate absorption measurements, the dash-dotted line is the calculated total particulate absorption spectrum, which is given only by the phytoplankton absorption with $[\text{Chl}] = 11.3 \mu\text{g/L}$ and nonalgal particles absorption (dotted line), with $[\text{NAP}] = 3.0 \text{ mg/L}$.

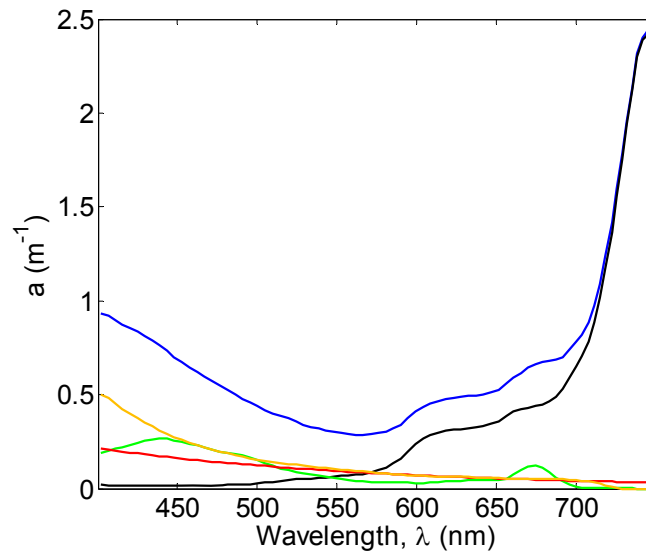


Fig. 10.10. Absorption spectra. The green line represents the phytoplankton absorption spectrum, the red line is the NAP's contribution to the absorption and the yellow – orange line is the CDOM's. The black line is the water absorption and the blue line is the total absorption.

Fig. 10.11 shows the number distribution obtained from LISST and the fitted curve, considering the slope obtained from the attenuation spectrum (recorded with the WET Labs package, as well). The measured PSD shows that the number of large particles is actually higher than what obtained from the analysis of the attenuation spectrum. Nonetheless, the two curves are parallel for most of the particles sizes, meaning that the slope of the PSD obtained from the attenuation spectrum is a good proxy for the slope of the actual PSD. For simplicity, we will, therefore, keep using a Junge-type size distribution with the slope calculated from the attenuation spectrum.

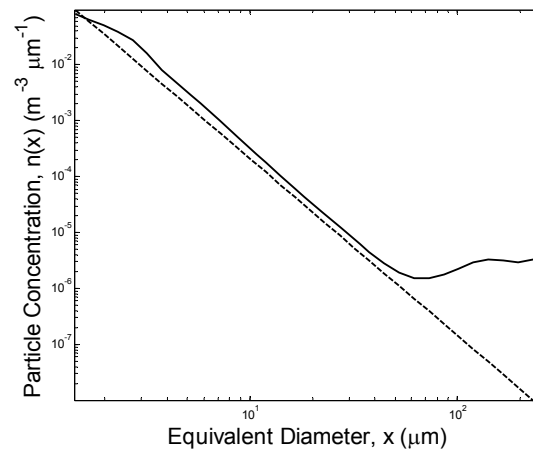


Fig. 10.11. PSD for Station 3. The solid curve is obtained from LISST measurements. The dashed line is the result of the fitting of the attenuation spectrum.

Fig. 10.12 shows a comparison between DOP_t and results of radiative transfer calculations. Also in this case the match is very good, assuming a refractive index for NAP equal to 1.17. Fig. 10.13 shows the behavior of the DOP vs. scattering angle. If we plot on the same graph the equation describing Rayleigh scattering (in the same way we did for Station 1), we notice that this time there's a big discrepancy. In fact, molecular scattering is not the dominant process here.

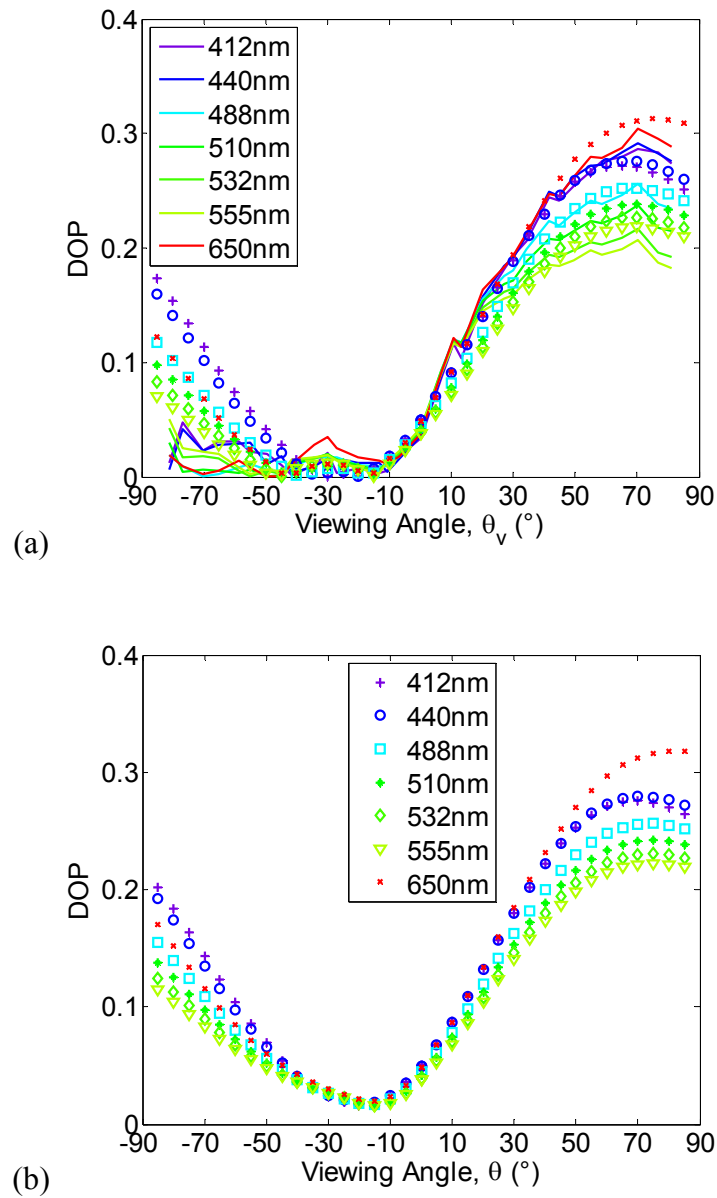


Fig. 10.12. (a) Plots of DOP_t vs. viewing angle, where 0° corresponds to an observer looking down. The solid lines represent the values of DOP_t calculated from *in situ* underwater measurements. The dotted lines are the results of the vector radiative transfer program with $n_{\text{phyto}} = 1.17$. (b) Same as (a) but the azimuthal angle (ϕ) is now 30° .

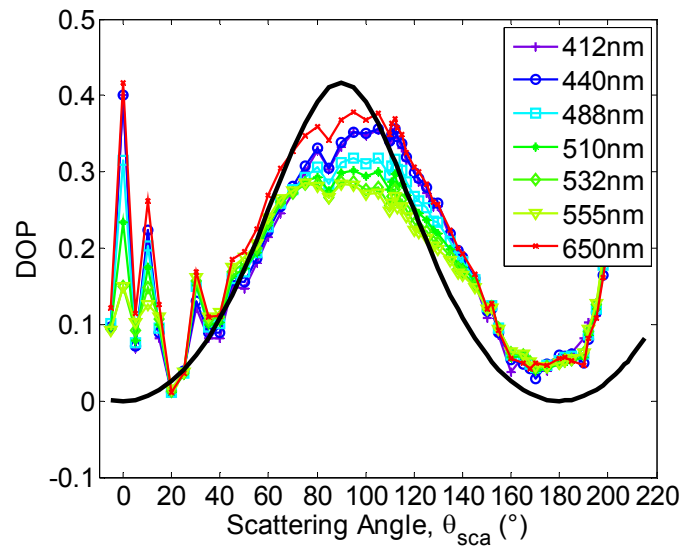


Fig. 10.13. Plots of the DOP vs. scattering angle. The instrument is located in the principal plane 1m below water. The solid black line represents Rayleigh scattering at 650 nm.

Fig. 10.14 shows the spectral behavior of the underwater DOP for some relevant scattering angles. Also in this case, the DOP reaches its maximal values over the entire spectrum at 90° scattering angle.

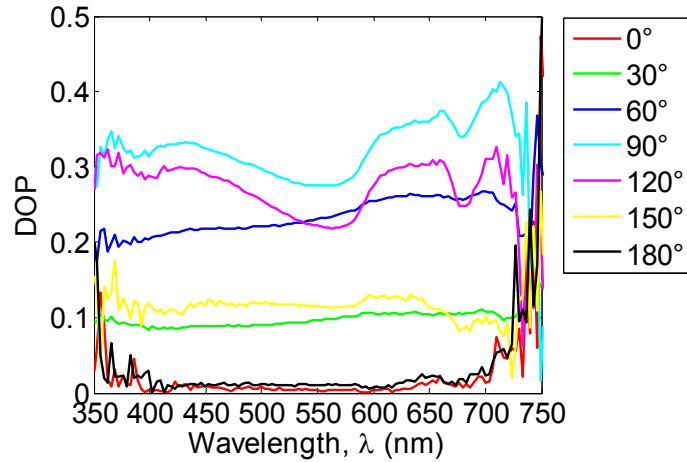


Fig. 10.14. Spectral behavior of the underwater DOP for few relevant scattering angles.

10.2 Spectral behavior of the DOP in the Chlorophyll fluorescence region

Plot of the maximum DOP_t transmitted through the air – water vs. wavelength is shown in Fig. 10.15. The calculated values nicely reproduce the spectral features of the DOP_t outside the region of Chlorophyll fluorescence. Between 650 and 725 nm radiative transfer calculations were carried out at one nanometer resolution. Fig. 10.16 shows the spectral comparison only in the range of interest, *i.e.* between 650 and 725 nm.

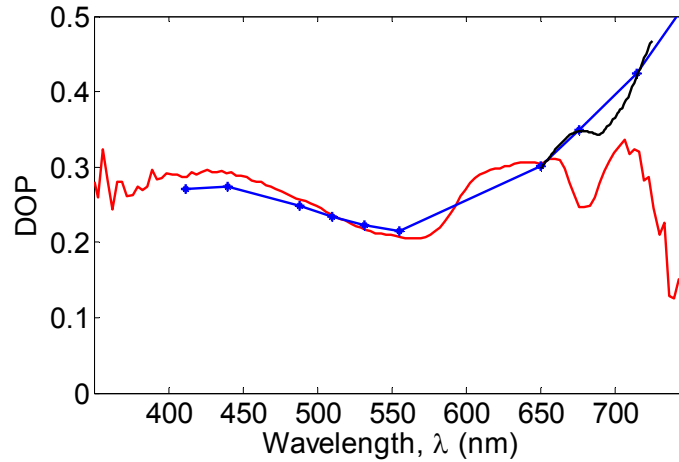


Fig. 10.15. Spectral behavior of the measured (red) and simulated (blue) DOP_t for Station 3. The blue line represents simulations performed at wavelengths correspondent to the AC-9 wavelengths. The black line represents the result of simulations carried out at one nanometer resolution, in the spectral range 650 – 725 nm.

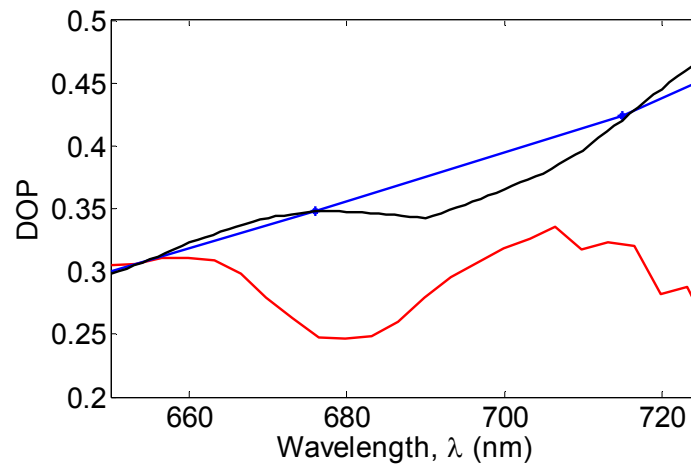
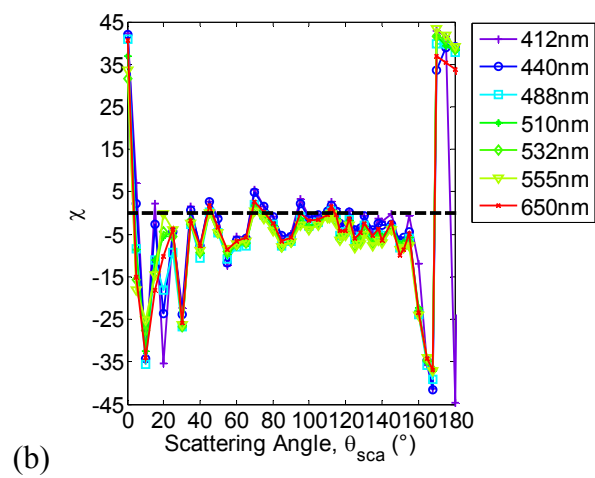
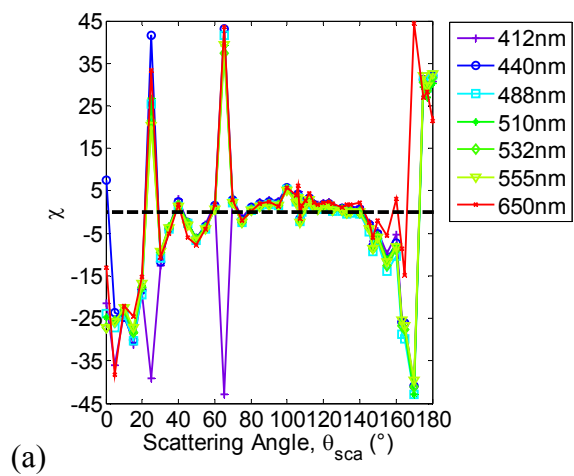


Fig. 10.16. Same as Fig. 10.16 but only the wavelength range 650 – 725 nm is considered.

The model predicts that if fluorescence wasn't present, the DOP_t would not be flat in the spectral region 670 – 690 nm, as was the case for Station 1 in the New York Harbor. Extracting Chlorophyll fluorescence directly from the analysis of the spectral variations of the DOP_t is not as simple as we expected.

10.3 The e – vector orientation

Let's now analyze the behavior of the e – vector for this station and for other stations which were part of the same field trip. Fig. 10.17 shows the plot of the e – vector orientation *vs.* scattering angle for Station 1, Day1 ($\theta_{Sun} = 38^\circ$, 11:00AM, [Chl] = 1.2 $\mu\text{g/L}$, [NAP] = 0); Station 3, Day 1 ($\theta_{Sun} = 27^\circ$, 14:00, [Chl] = 11.3 $\mu\text{g/L}$, [NAP] = 3.0 mg/L); Station 4, Day 3 ($\theta_{Sun} = 34^\circ$, 3:00PM, [Chl] = 32.2 $\mu\text{g/L}$, [NAP] = 11.3 mg/L), respectively. All plots refer to measurements taken in the principal scattering plane. One aspect of the plots that is easily noticeable is the spike around 180° scattering angle. We observed this feature at other stations too (see the previous section). The stations under consideration have particle loading very different one from the other, nonetheless, the plots of the e – vector orientation are very similar: the orientation is horizontal for most of the scattering angles apart from the ones around the backscattering direction.



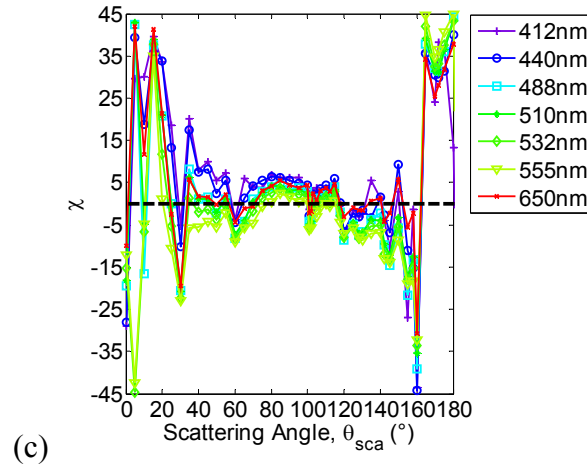


Fig. 10.17. Plots of the e – vector orientation vs. scattering angle, for (a) Station 1, Day 1; (b) Station 3, Day 1; (c) Station 4, Day 3. The dashed thick black line is the e – vector orientation predicted by Rayleigh theory.

10.4 Conclusions

In this study we investigated the geometrical and spectral properties of the underwater polarized light field, in various atmospheric conditions and water compositions, with particular attention to the coastal regions. We developed appropriate *in situ* instrumentation to compare our results with computational results. We also correlated the polarized water leaving radiance to the properties of the water components.

- (1) The first test of our instrumentation was in Chlorophyll rich waters. During this initial testing we confirmed the theoretical prediction that for sensor orientations slightly outside the principal scattering plane, the DOP can approach the maximum values observed in the principal scattering plane.
- (2) We also performed above – water measurements of the DOP. When measurements are taken in the main scattering plane, the reflected skylight (also

containing the sunglint) becomes dominant over the water-leaving radiance and the DOP doesn't contain any information on suspended particles. The tiny difference in the simulated DOP plots near the Brewster's angle is completely lost in the experimental measurements due to the noise introduced by the surface waves. This result provided an incentive for considering other viewing geometries, *i.e.* away from the principal plane. Simulations carried out using the adding – doubling radiative transfer code matched our experimental measurements for both in – plane and out – of – plane above water measurements. However, for a quantitative analysis of the above – water measurements, we still need to improve substantially our methodology of recording data.

- (3) Subsequent measurements of the underwater DOP revealed that, in coastal areas, maximum values of the DOP, which occurred approximately at scattering angle of 100° , did not exceed 0.4. In overcast conditions however, maximal values of the DOP reduced to approximately 0.2. Spectral dependence of DOP very well correlates with the measured water total absorption: increase of absorption corresponds to the decrease of the number of the scattering events which means less depolarization. These effects are more pronounced at right scattering angles, where the DOP reaches its maximal values and almost unnoticeable at angles closer to 0 and 180° . In the spectral area of Chlorophyll fluorescence we observed significant decrease of DOP which is explained by the depolarizing effect of fluorescence. Experimental measurements were then compared with a Monte Carlo radiative transfer model, showing excellent agreement. The sensitivity of the above water polarized reflectance in the visible spectrum to particle

- compositions (*i. e.* refractive index) and size distributions was examined for various types of waters.
- (4) We showed that the polarized reflectance at 650 nm might allow the determination of the composition of suspended particles. Note, however, that an appropriate viewing geometry should be adopted (*i. e.* at the angle where the polarization is maximal). This means that multiangular measurements of the polarized reflectance are mandatory.
- (5) It is expected that the ability of our polarimeter to provide information about the characteristics of the underwater polarized light field has great potential for application in radiative transfer problems in the earth–ocean system, especially if used in combination with other polarization–sensitive instruments recently developed; the simulations (using both the Monte Carlo and the Adding – Doubling method) showed that multiangular polarized water leaving radiance collected in the visible is a powerful tool that could be systematically used to gain auxiliary and complementary information on suspended particles. The main advantages of the method proposed here are as follows. (i) Only optical measurements are required, and (ii) the DOP approach is weakly dependent on calibration, since the DOP itself is a ratio.

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