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Chapter 1

INTRODUCTION TO RADIATIVE TRANSFER

[1]J. RONALD V. ZANEVELD, [2]MICHAEL J. TWARDOWSKI, [1]ANDREW BARNARD AND [3]MARLON LEWIS

[1]*WET Labs, Inc., 620 Applegate St., Philomath, Oregon, 97370 USA*
[2]*WET Labs, Inc., 165 Dean Knauss Dr., Narragansett, Rhode Island, 02882 USA*
[3]*Dalhousie University, Halifax, Nova Scotia B3H 4J1, CANADA*

1.0 Introduction

For the purpose of this paper, we will define Radiative Transfer as the study of the change in direction and intensity of radiation in the atmosphere and ocean, due to absorption, scattering, fluorescence, inelastic scattering, and air- sea interface effects. This is a very large topic and many books have been written on this subject. It is therefore impossible to cover the subject comprehensively in a short article such as this. For detailed presentations on various aspects of this subject the reader is referred to books such as Preisendorfer (1965, 1976), Jerlov (1974), Jerlov and Steemann-Nielsen (1974), Liou (1980), Shifrin (1988), Dera (1992), Mobley (1994), and Thomas et al.(1999) . A very readable introduction to the subject is the first half of Kirk (1994). Detailed definitions of optical and radiometric parameters and protocols for their measurement can be found in Mueller et al. (2003)

Energy is generated in the sun by thermonuclear reactions. Gamma rays resulting from these reactions diffuse toward the surface of the sun. On their way to the surface, the gamma rays are scattered, absorbed and reemitted by nuclei and electrons and are changed to x-rays, then to ultraviolet rays and finally emerge mainly as visible light from the surface. After approximately eight minutes of travel a small fraction of the solar photons reach the outer atmosphere of the earth. The amount of solar radiation reaching the top of the atmosphere depends on the motion of the earth around the sun. The planet orbits about the sun in a slightly elliptical orbit. It also rotates about its axis, which is tilted relative to the plane of the planet about the sun. These motions cause the irradiance incident on the top of the atmosphere to fluctuate on a daily and seasonal basis. Longer term effects such as precession, obliquity and eccentricity are also present; these give rise to the various well-known Milankovitch modes of variability in incident solar irradiance.

The atmosphere consists primarily of a mixture of gases, some of which are relatively permanent and some of which vary in concentration. Nitrogen and oxygen dominate the permanent gases, while water vapor and ozone dominate the variable constituents. These gases are distributed non-linearly as a function of height above the planet. Each of the gases has a particular absorption spectrum. Molecular scattering by the gases diffuses the collimated solar photons. In addition to the gases, aerosols (particles) are present. These particles have various origins, such as deserts, ice, fossil fuel burning, forest fires, etc. The aerosols also absorb and scatter light.

The light that arrives at the sea surface is already considerably modified from that which reached the top of the atmosphere. The intensity and spectrum are changed by the absorption and scattering of the gases and aerosols, and the directionality has been changed by the scattering. In addition to direct sun light there is now also a large amount of diffuse light, the intensity and spectrum of which depends on the solar zenith angle and the nature and distribution of aerosols and gases. The sea surface can be characterized by a directional slope spectrum of the waves. This spectrum determines the refraction and reflection patterns at the surface, and so the directional pattern of the light beneath the surface. In addition to the slope spectrum, the sea surface can generate white caps and bubbles, which also affect the redistribution of light.

The sea water itself has certain scattering and absorption characteristics. In addition, the water contains particles of various origins and dissolved materials, each with its own optical characteristics. A fraction of the light that reaches the sea surface and penetrates into the water is reflected back into the atmosphere. This upward light is again modified by the atmosphere and its constituents before some of it reaches spacecraft orbiting the earth. In shallow waters some light may reach the bottom and be reflected upwards.

The formidable task of radiative transfer is to quantify all the processes described above. The forward problem for remote sensing is to predict the spectral intensity of the light reaching the satellite sensor based on a quantitative description of all the absorption, scattering and reflection characteristics of the optical components in the atmosphere and ocean. The inverse problem is the determination of the concentration of the atmospheric and oceanic constituents (including the bottom) when the quantity and spectral characteristics of the light received at the space- or airborne sensor are known.

The above tasks require the exact definition of radiometric quantities, definitions of the inherent optical properties (IOP, the scattering and absorption of the various atmospheric and oceanic constituents) and apparent optical properties (AOP) that describe the attenuation characteristics of the light field. All of these parameters must then be combined with the incident radiance distribution in physical relations that allow mathematical calculations.

Here we have taken a narrative approach, introducing parameters as they are needed. The usual approach is to define all the radiometric quantities, and the inherent and apparent optical properties, and then proceed from there. We have taken the inverse approach for readability. A more complete discussion of the various properties is given at the end of the chapter.

2.0 The Equation of Radiative Transfer

In this section it is assumed that all parameters are measured at a single wavelength, λ, which is omitted for brevity. The intensity of a narrow beam of light emanating from a source can be described by its **radiance**: energy per unit area per unit solid angle (the cone into which the light radiates, or is measured, units of steradian, sr). **Units of radiance are therefore W/m^{2}sr, the symbol is L**. Imagine this narrow beam being attenuated by absorption and scattering in a medium. It makes sense for the loss of radiance ΔL due to attenuation over short distances Δr to be proportional to the distance traveled and radiance itself. We then get that: $\Delta L = -cL\Delta r$, where c is a proportionality constant, called the **attenuation coefficient**, **units of m^{-1}**. Taking the limit, we obtain $dL/dr = -cL$, so that $L(r) = L(0)\exp(-cr)$. This equation is used in attenuation meters which contain highly collimated sources to obtain the attenuation coefficient, c. This coefficient is both a function of wavelength and location; c should thus be properly given by $c(\lambda,\vec{\mathbf{x}})$, where $\vec{\mathbf{x}}$ is the position vector (x,y,z).

If we look at the attenuation of radiance in the case of sunlight in a scattering medium, we not only have attenuation of the radiance as it travels from one location $\vec{\mathbf{x}}_1$ to another, $\vec{\mathbf{x}}_2$, but light from other directions can be scattered into the direction specified by the two locations $\vec{\mathbf{x}}_1$ and $\vec{\mathbf{x}}_2$. We thus need to quantify this process. In general we can specify a direction by (θ,ϕ) , where θ is the solar zenith angle and ϕ the azimuthal angle in a given reference frame. If we look at the rate of change of L, along a path r in the direction (θ,ϕ), due to scattered light entering the beam, we would expect this increase to be proportional to the radiance coming from another direction (θ',ϕ') multiplied by a small solid angle $d\omega$, to get the energy per unit area perpendicular to the incoming beam. Thus, $dL/dr \;\alpha\; L d\omega$. We now define a proportionality function, the **volume scattering function, with units of $m^{-1}sr^{-1}$** to relate dL/dr to $Ld\omega$ **.** Clearly this function depends on location, the incoming direction, and the outgoing direction; it is designated by $\beta(\vec{\mathbf{x}},\theta,\phi,\theta',\phi')$:

$dL(\vec{\mathbf{x}},\theta,\phi)/dr = \beta(\vec{\mathbf{x}},\theta,\phi,\theta',\phi')\, L(\vec{\mathbf{x}},\theta',\phi')d\omega$.

This is the rate of increase of radiance in direction (θ,ϕ) at location $\vec{\mathbf{x}}$, with wavelength λ (not explicitly specified), due to scattered light from direction (θ',ϕ'), with the same location and wavelength. Clearly there can be light coming from all directions (θ',ϕ'), we thus need to integrate over all these directions (there are 4π steradians in a sphere) to get the total amount of radiance that is scattered from all directions into (θ,ϕ) :

$$dL(\vec{\mathbf{x}},\lambda,\theta,\phi)/dr = \int_0^{4\pi} \beta(\vec{\mathbf{x}},\lambda,\theta,\phi,\theta',\phi')\, L(\vec{\mathbf{x}},\lambda,\theta',\phi')d\omega'. \tag{1}$$

We thus have found the rate of increase in the radiance along a path due to scattering. Earlier, we had found the rate of decrease due to attenuation. We combine these two, losses due to attenuation and gains due to scattering to get:

$$dL(\vec{\mathbf{x}},\lambda,\theta,\phi)/dr = -c(\vec{\mathbf{x}},\lambda)\, L(\vec{\mathbf{x}},\lambda,\theta,\phi) + \int_0^{4\pi} \beta(\vec{\mathbf{x}},\lambda,\theta,\phi,\theta',\phi')\, L(\vec{\mathbf{x}},\lambda,\theta',\phi')d\omega'. \tag{2}$$

This is the equation of radiative transfer (ERT) without internal sources such as fluorescence or Raman scattering. Many books have been written regarding solutions to the ERT. The classical Legendre functon solution is by Chandrasekar (1960). Other solutions are given in Preisendorfer (1965). The most common approach in Oceanography is to assume that horizontal gradients in radiance and IOP are much smaller than vertical ones, so that horizontal structure is ignored. Taking $\cos\theta\, dr = dz$, leads to:

$$\cos\theta \, dL(z,\lambda,\theta,\phi)/dz = -c(z)\, L(z,\lambda,\theta,\phi) + \int_0^{4\pi} \beta(z,\lambda,\theta,\phi,\theta',\phi')\, L(z,\lambda,\theta',\phi')d\omega'. \quad (3)$$

This is the ERT for the so-called plane parallel assumption without internal sources and is widely applied. Discussions on various solutions to this equation can be found in Mobley (1994) and Thomas and Stamnes(1999). Mobley's numerical solution is commercially available (Hydrolight, Sequoia, Inc.) and is widely used. In the surface zone of the ocean, in the presence of waves, the plane parallel assumption is incorrect, however (e.g. Zaneveld et al., 2001), as it is in the atmosphere in the presence of clouds. In fact, waves form a formidable obstacle to the correct measurement of components for, and the verification of, solutions and inversions of the ERT. If we think of the **light field** at a given location as being made up of all radiances at that location, we can readily surmise that due to refraction at the surface the light field in the presence of waves is not the same as that for a flat surface. Horizontal variations in the light field occur on the scales of the smallest capillary waves which have wavelengths on the order of a cm to length scales of hundreds of meters for open ocean swell. We can horizontally average Eq.(2), leading to

$$\int_0^D \int_0^D dL(\vec{\mathbf{x}},\theta,\phi)dxdy\,/dr = -\int_0^D \int_0^D c(\vec{\mathbf{x}})\, L(\vec{\mathbf{x}},\theta,\phi)dxdy + \int_0^{4\pi} \int_0^D \int_0^D \beta(\vec{\mathbf{x}},\theta,\phi,\theta',\phi')\, L(\vec{\mathbf{x}},\theta',\phi')dxdyd\omega', \quad (4)$$

where D is an appropriate length scale. We can write an equivalent ERT (and use numerical solutions) for a horizontally averaged light field $\overline{L(\vec{\mathbf{x}},\lambda,\theta,\phi)}$ that is the equivalent of Eq. (3), only if $c(\vec{\mathbf{x}})$ and $\beta(\vec{\mathbf{x}},\theta,\phi,\theta',\phi')$ are constant horizontally over the length scale D:

$$\cos\theta \int_0^D \int_0^D dL(\vec{\mathbf{x}},\lambda,\theta,\phi)dxdy\,/dz = -\,c(z) \int_0^D \int_0^D L(\vec{\mathbf{x}},\lambda,\theta,\phi)dxdy + \int_0^{4\pi} \beta(z,\lambda,\theta,\phi,\theta',\phi') \int_0^D \int_0^D L(\vec{\mathbf{x}},\lambda,\theta',\phi')dxdy\, d\omega'. \quad (5)$$

Note that the attenuation and scattering parameters can now be functions of depth only. While it is likely that the attenuation and scattering properties are constant over a few cm, this is highly unlikely over hundreds of meters. For example, Farmer and Li (1995) found bubble clouds aligned with the wind caused by Langmuir circulation on the scale of tens of meters and Colbo and Li (1999) observed that Langmuir circulation affected horizontal and vertical distribution of other particles on these scales. Barth (1999), using a SeaSoar undulating towed device found sub-km scale variations in optical properties. It will be necessary to make a number of small scale transects of the inherent optical properties and radiance distribution in the ocean to properly assess the error due to the plane parallel assumption in different oceanic regimes.

The bottom can be included in equations such as the above. In that case one treats the bottom as a special case of scattering function $\beta(\vec{\mathbf{x}},\theta,\phi,\theta',\phi')$ in which all downward values are infinite. What one is left with is the **bi-directional reflectance distribution function, the BRDF** (Voss et al., 2000, Zhang et al., 2003)**.** The BRDF describes the transfer of radiance from a downwelling direction to an upwelling direction, where the amount of radiance transferred is a function of both the incoming and scattered directions. A common assumption for the directional scattering of the bottom is the Lambertian. In that case the scattered radiance is independent of direction.

In order to include fluorescence or Raman effects source terms must be added to the ERT. This poses no particular problem for numerical methods, such as Hydrolight. Thus fluorescence by phytoplankton pigments such as chlorophyll and phycoerythrin can be accommodated, in addition to Raman scattering which influences the radiance intensity (Marshall and Smith, 1990).

We have thus seen that plane parallel programs such as Hydrolight can be used in a horizontally average sense for the radiance, provided the IOP are constant horizontally. Refractive effects due to waves can then be included in an average sense. If one wants to investigate small scale effects of the full three dimensional ERT Eq.(2) one can use so-called Monte Carlo methods. Reviews are provided in Mobley (1994) and Thomas and Stamnes (1999). In these methods, in principle, one follows the paths of multiple photons from the time they are inserted into the ocean to the moment they are absorbed, or leave the ocean. One calculates the probability of occurence of a given scattering or attenuation event derived from the IOP. Using random numbers based on the probability distribution function of events, one obtains paths for the photons. After many calculations the photon density and direction provides the desired underwater or water leaving light field. This method has the advantage that any structure of IOP and interfaces can be used. In addition, internal sources such as fluorescence and Raman scattering can readily be handled. The method is computationally intensive, but with faster computers it is likely that this will become the dominant method in the future for obtaining solutions to the ERT.

The ERT provides a link between the radiance distribution and the IOP. In principle, if the radiance distribution at the boundary and the IOP are known, we can solve for the radiances in the interior. The radiance distribution is difficult to measure although it has been done (Smith et al, 1970, Voss, 1989, Voss et al., 2003). Similarly the volume scattering function is difficult to measure underwater in its entirety, although again, a few examples exist (Petzold, 1972, Lee and Lewis, 2003). Integration over the radiance distribution provides links between more readily measured parameters as we shall see below.

3. 0 Gershun's Equation

Integrated parameters are usually easier to measure as they contain fewer variables. If in Eq.(3) we assume the IOP to be homogeneous and we integrate over all directions, we obtain:

$$d[\cos\theta\, L(z,\theta,\phi)d\omega]/dz = -c(z) \int_0^{4\pi} L(z,\theta,\phi)d\omega +$$

$$\int_0^{4\pi} \int_0^{4\pi} \beta(z, \theta,\phi,\theta',\phi')\, L(z,\theta',\phi')\, d\omega d\omega'. \quad (6)$$

The above leads us to define the following radiometric quantities:

$$E(z) = [\int_0^{4\pi} \cos\theta\, L(z,\theta,\phi)d\omega], \text{ and}$$

$$E_0(z) = [\int_0^{4\pi} L(z,\theta,\phi)d\omega].$$

The first quantity, E(z), is called the **plane irradiance**, because it is a measure of the flux of energy through a plane perpendicular to the z –direction. **Its units are W/m^2**. Note that it consists of the difference between the downwelling planar irradiance, $E_d(z)$ and the upwelling plane irradiance, $E_u(z)$, which represent the weighted integration over the upper ($\theta< \pi/2$)and lower hemisphere ($\theta> \pi/2$) respectively. The second quantity, $E_0(z)$, called the **scalar irradiance**, measures the total energy flux through a point. **Its units are W/m^2**. Inserting these quantities into Eq. (6) leads to:

$$dE(z)/dz = -c(z)\, E_0(z) + b(z)\, E_0(z) \quad (7)$$

This is Gershun's equation. The attenuation coefficient c is the sum of the scattering and **absorption coefficients,(units m^{-1})** so that Eq.(7) can be rewritten as:

$$dE(z)/dz = -a(z)\, E_0(z) \quad (8)$$

This provides an interesting link between the radiometric quantities (based on radiance) and the IOP. Let us define a **diffuse attenuation coefficient**, **units of m^{-1}**, similar to the beam attenuation coefficient:

$$K(z) = -\, dE(z)/\, E(z)dz, \text{ so that}$$

$$E(z) = E(0)\exp[-\int_0^{z} K(z)dz] \quad (9a)$$

Similar attenuation coefficients can be defined for the radiances:

$$k(z,\theta,\phi)) = -\, d\, L(z,\theta,\phi)/\, L(z,\theta,\phi)\, dz, \text{ and}$$

$$L(z,\theta,\phi) = L(0,\theta,\phi)\exp[-\int_0^{z} k(z,\theta,\phi)dz] \quad (9b)$$

The similarity of the definition of K(z) and $k(z,\theta,\phi)$ to the inherent optical property, beam attenuation coefficient, leads us to call parameters such as K(z), **apparent optical properties (AOP,** see section 5.2**).** These AOP are differential properties of the light field that are used to indicate the rate of change of radiometric properties with depth. Substitution of Eq. (9a) into (8) leads to:

$$\frac{a(z)}{K(z)} = \frac{E(z)}{E_o(z)} = \bar{\mu}(z) \tag{10}$$

The quantity $\bar{\mu}(z)$ is called the **average cosine (dimensionless)** of the light field. This name follows from the definitions of E(z) and $E_0(z)$. The average cosine is a useful concept as it connects the IOP a(z) with the AOP K(z). In the case of optical remote sensing it is reasonable to assume that the light field above the sea surface is dominated by the sun. The light field in the ocean tends to be oriented nearly vertically since the maximum intensity is due to the refracted image of the sun, and refraction reduces the solar zenith angle in water compared to that in air.. The average cosine thus tends to vary in a narrow range of values from 0.7 to 0.9 An interesting aspect of the average cosine is that it converges to an **asymptotic value** $\bar{\mu}_\infty$ (Preisendorfer,1976, Højerslev and Zaneveld, 1977), that is a function of the IOP only. That is, as one goes down into the ocean the light field reaches a constant shape, and the absolute values of the radiances decrease exponentially with a coefficient of K_∞, the **asymptotic diffuse attenuation coefficient,** which is an IOP, since its value is independent of the radiance distribution at the surface**.** Zaneveld (1989) and Berwald et al. (1995) have provided functional dependencies of $\bar{\mu}_\infty$ on the IOP in terms of ω_o (**the single scattering albedo = b/c**, dimensionless).

4. 0 Inversions and Remote Sensing

In section 2 we have sketched the methods with which one can obtain the radiance distribution if the IOP are known. The IOP can be measured directly (see Twardowski et al., chapter 4, this volume). In the case of remote sensing, however, we have an airborne or spaceborne radiance detector, typically at several wavelengths. The inverse problem of radiative transfer is then to derive IOP from the remotely sensed radiance. This is a problem for which there is no exact solution, so various approximations must always be made. Here we will use several approaches to the inversion problem. First we will use an intuitive approach, to get an idea of the parameters involved. Secondly, we will use an inversion from the ERT to see how far we can take an analytical approach before we must resort to approximations. The value of inversions from remote sensing lies in the determination of IOP which can then be further translated, if desired, into other less exact parameters such as chlorophyll, particle concentration, etc.

First we will derive from first principles the remote sensing reflectance just beneath the sea surface. We wish to derive equations for the upwelling light just beneath the surface, if the downwelling plane irradiance and the IOP are known. Zaneveld and Pegau (1998) and Zaneveld et al. (1998) have derived steady state two flow equations that can be adjusted to the present problem. From first principles one can think of a downwelling stream of photons that is attenuated on the way down. At all depths some of the photons are scattered into an upward direction. These upwelling photons are further attenuated on the way up. Finally only a small fraction of the upwelling photons are moving in a direction that can be detected by the sensor. The contribution to the upwelling nadir radiance, L_u (z), (vertically upwelling, $\theta = \pi$, ϕ can be any value), at a given depth is the downwelling plane irradiance at that depth, E_d (z), multiplied by a weighted integral of the volume scattering function in the backward direction, $\beta_b(z)$. This scattered light is a small fraction F_b of the backscattered light given by the backscattering coefficient b_b, as the remote sensing detector typically has a small solid angle of detection. Thus $\beta_b(z) = F_b\, b_b$. The vertical attenuation coefficient for downwelling irradiance is given by $K_d(z)$ and the vertical attenuation of the upwelling nadir radiance is given by $k_u(z)$. Integrating over all depths gives the nadir radiance just below the sea surface:

$$L_u(0^-) = E_d(0^-) \int_0^\infty (\beta_b(z)\, e^{-\zeta(z)} dz, \tag{11}$$

$$\text{where } \zeta(z) = \int_0^\infty (K_d(z') + k_u(z'))\, dz'. \tag{12}$$

If the scattering and diffuse attenuation properties are assumed to be constant with depth,

$$L_u(0^-)/\, E_d(0^-) = R_{rs}(0^-) = \frac{\beta_b(z)}{K_d(z') + k_u(z')} \tag{13}$$

where $R_{rs}(0^-)$ is the remote sensing reflectance just below the surface, with units of sr^{-1}. We had already seen that $\beta_b(z) = F_b\, b_b$. In addition Eq. (10) showed that the diffuse attenuation coefficient was related to the absorption coefficient. We can thus reasonably expect the downwelling and upwelling diffuse attenuation coefficients to also be related to the absorption coefficient. Combining these arguments and applying them to Eq.(13) allows us to set:

$$L_u(0^-)/\, E_d(0^-) = R_{rs}(0^-) = F_R \frac{b_b(z)}{a(z)} \tag{14}$$

This relationship was first derived (using different approaches) by Gordon et al., 1975 and Morel and Prieur, 1977. We thus found that the remote sensing reflectance is proportional to the backscattering coefficient and inversely proportional to the absorption coefficient. The proportionality factor F_R (sr^{-1}) depends on how the backscattered light relates to the backscattering coefficient, and therefore to the details of the volume scattering function in the backward direction and the radiance distribution. The relationship between the diffuse attenuation coefficients and the absorption coefficient also depends on the details of the radiance distribution and the IOP. Much radiative transfer is thus contained in the factor F_R. This factor has been studied in detail (for example Gordon et al., 1975, Gordon et al., 1988; Morel and Gentili, 1996).

Based on extensive Monte Carlo calculations, Gordon et al. (1975) derived for a collimated beam of irradiance:

$$R = E_u(0^-)/E_d(0^-) = 0.33 \frac{b_b}{a + b_b}.$$

In order to relate the upwelling irradiance to the upwelling radiance, a commonly introduced factor is $Q = E_u(0^-) / L_u(0^-)$. Since typically the backscattering is much less than the absorption, this sets the parameter F_R in Eq. (14) equal to 0.33/Q. This parameterization is the starting point for many inversion algorithms, but it ignores the dependence of F_R on the shape of the volume scattering function and the radiance distribution. In order to obtain the dependence of F_R on these parameters, we must use a more analytical approach, which follows.

A more theoretical relationship (Zaneveld, 1982, 1995) can be derived from the ERT (in the form of Eq.(3)) for the nadir radiance, L_u, for which $\cos\theta = -1$, and for which we can define an attenuation coefficient k_u (as in Eq. (9b)):

$$k_u(z)\, L_u(z) = -c(z)\, L_u(z) + \int_0^{4\pi} \beta(z,\pi,0,\theta',\phi')\, L(z,\theta',\phi')d\omega'. \qquad (15)$$

We split the integration over $d\omega$ into integrations over θ and ϕ and split them into upwelling ($\theta > \pi/2$) and downwelling ($\theta < \pi/2$) components.

$$[k_u(z) + c(z)]\, L_u(z) = \int_0^{2\pi}\int_0^{\pi/2} \beta(z,\pi,0,\theta',\phi')L(\theta',\phi',z)\sin\theta' d\theta' d\phi' +$$

$$\int_0^{2\pi}\int_{\pi/2}^{\pi} \beta(z,\pi,0,\theta',\phi')L(\theta',\phi',z)\sin\theta' d\theta' d\phi' \qquad (16)$$

We now define shape factors f_b and f_L that relate the complex integral expressions above to simpler and more readily measured parameters.

$$f_b(z) = \left[\int_0^{2\pi}\int_0^{\pi/2} \beta(z,\pi,0,\theta',\phi')L(\theta',\phi',z)\sin\theta' d\theta' d\phi'\right] / \left[\frac{b_b(z)}{2\pi} E_{od}(z)\right] \qquad (17a)$$

$$f_L(z) = \left[\int_0^{2\pi}\int_{\pi/2}^{\pi} \beta(z,\pi,0,\theta',\phi')L(\theta',\phi',z)\sin\theta' d\theta' d\phi'\right] / \left[b_f(z)\, L_u(z)\right] \qquad (17b)$$

The parameters are chosen this way because in Eq. (17a), all the scattering is in the backward direction. If the scattering were constant in the backward direction, we could take it outside of the integral, which would then reduce to $E_{od}(z)$, the downwelling scalar irradiance, defined as the unweighted integral over all radiances in the upper hemisphere. In Eq.(17b) the radiances are all upward, and if they were constant, we could take it outside of the integral, which would then become $b_f(z)$, the forward part of the scattering coefficient. The parameter $f_b(z)$ thus to a large extent describes how uniform the backscattering function is, and $f_L(z)$ describes primarily how uniform the upwelling radiance distribution is. Values of unity indicate uniformity. Substitution of Eq. (17) into (16) then gives:

$$[k_u(z) + c(z)]\, L_u(z) = f_b(z)\, E_{od}(z)\frac{b_b(z)}{2\pi} + f_L(z)\, L_u(z) b_f(z)\text{ , or} \qquad (18)$$

$$L_u(z) / E_{od}(z) = \frac{f_b(z) \frac{b_b(z)}{2\pi}}{k_u(z) + c(z) - f_L(z) b_f(z)} \tag{19}$$

This is an exact expression as it is only a rewrite of the ERT. We note that it contains the scalar irradiance rather than the plane irradiance. This equation can be the platform from which to obtain approximations for the remote sensing reflectance based on observations. Zaneveld (1995) has described a series of approximations that lead from more exact to more approximate expressions. The more approximate expressions contain measurable parameters, however, and so are needed for experimental work.

Zaneveld (1995) and Weidemann et al. (1995) have calculated values for $f_b(0^-)$ and $f_L(0^-)$ based on observations for homogeneous media. From Eq. (17b) we see that f_L is an integration over the forward scattering function multiplied by the upward radiance distribution. The upward radiance distribution is approximately uniform so that it was found that generally f_L was within a few percent of 1. We can assume as well that the backscattering coefficient and (1- f_L b_f) are much less than the absorption coefficient, which would be the case in all but the most strongly scattering media. In addition, Zaneveld (1995) has argued, based on the observed nearly exact exponential shape of $L_u(z)$ near the surface, that $k_u(z)$ could be modeled as $a / \bar{\mu}_\infty$. Applying these approximations, Eq. (19) reduces to:

$$L_u(0^-) / E_{od}(0^-) \approx \frac{f_b}{2\pi\, (1 + 1/\bar{\mu}_\infty)} \frac{b_b}{a} .$$

Analogous to Eq. (10), we can define a **downwelling average cosine**, $\bar{\mu}_d\,(z) = E_d(z) / E_{od}(z)$. We can now obtain an expression analogous to Eq. (14):

$$L_u\,(0^-)/ E_d\,(0^-) = R_{rs}(0^-) \approx \frac{f_b}{2\pi\, \bar{\mu}_d\,(0^-)\,(1 + 1/\bar{\mu}_\infty)} \frac{b_b}{a} \tag{20}$$

Clearly, through multiple approximations, it is possible to get closer to measurable parameters when proceeding from the purely theoretical expression Eq.(19). We also need an expression for f_b in terms of measurable parameters. Eq. (17a) shows that f_b contains the backward part of the volume scattering coefficient and the forward part of the radiance distribution. We can assume that the downward radiance distribution near the surface forms a diffuse beam (due to forward scattering) along the refracted image of the sun. The refracted image of the sun and the refracted detection angle of the remote sensing detector, form an angle θ_m. We can then hypothesize that most of the backward scattering will be between the diffuse forward beam and the diffuse backward beam, with an average volume scattering function of $\beta(\pi-\theta_m)$. If there is a significant amount of light that is backscattered twice before reaching the remote sensor, this hypothesis would be incorrect. However, even in the case of multiple scattering, the most likely scattering event will be in the very near forward direction. Multiple scattered light that reaches the remote sensor will thus most likely have undergone multiple forward scattering events and a single backscattering event. Even considering multiple scattering, the dominant backscattering will still be due to $\beta(\pi-\theta_m)$ except when scattering strongly dominates diffuse attenuation. Considering this, we can then write Eq. (17a) as:

$$f_b(0^-) \approx 2\pi\, \beta(\pi-\theta_m) / b_b \tag{21}$$

Eq. (21) was tested extensively by Weidemann et al. (1995) and found to have typical errors of 5% and maximum errors of 12%. Note that this formulation makes no assumptions about the shape of the volume scattering function. Other formulations such as that by Morel and Gentili (1991, 1993) assume a constant shape for the particulate VSF. In any case the reader should be aware that the relative orientation of the sun and the sensor has a potentially large impact on the remote sensing reflectance. Substitution of Eq. (21) into Eq. (20) then leads to:

$$L_u(0^-)/E_d(0^-) = R_{rs}(0^-) \approx \frac{\beta(\pi-\theta_m)}{a\,\bar{\mu}_d(0^-)\,(1 + 1/\bar{\mu}_\infty)} \qquad (22)$$

Finally we note that the average cosine of the near surface light field can be given by the cosine of the refracted solar zenith angle, $\cos\theta_s$, and that Zaneveld (1989) showed that $a/\bar{\mu}_\infty \approx c\,(1- 0.52\omega_o - 0.44\omega_o^2)$. We now have a complete model of the dependence of the remotely sensed reflectance on the IOP. We want to be able to invert relationships such as Eq. (22) for the IOP. In these models we have not specifically indicated wavelength, but for the purposes of inverting remotely sensed signals one can assume that one has measured reflectances at N wavelengths. In principle one can then invert for N parameters. This is done by dividing the IOP into components (see section 6.1) and assigning spectral models for these component IOP, each with one or more parameters. One then minimizes the resultant N equations to obtain N or fewer parameters. Examples of such methods are Roesler and Perry (1995), Garver and Siegel (1997), Hoge and Lyon (1999), Lee et al. (2002), Roesler and Boss (2003) and can also can be found elsewhere in this volume.

5. Lidar

Light Detection And Ranging (Lidar) is a method whereby very short pulses of light from a laser are sent into the medium. By looking at the return as a function of time, one can determine profiles of optical properties. For oceanographic studies (Hoge, 1988), one typically uses a doubled YaG laser with a wavelength of 532nm.

5.1 ILLUMINATION AND DETECTION FOOTPRINTS

Here we will look at some of the oceanographic implications of illumination and detection footprints, or spot size. In a plane parallel situation there is a very large illumination spot size (ISS), and a limited detection spot size (DSS). This is the classical passive remote sensing arrangement. For the active arrangement we can assume a small ISS and a large DSS. By means of a reciprocity argument, we can immediately say that this will give the same result as the passive arrangement. This reciprocity can be verified by a thought experiment. In the passive case, if we envision the detector beam, we know that if a photon leaves the confines of the beam , another photon will, statistically, take its place on the other side of the beam. No net photons are lost from this system due to course changing caused by scattering. We can then see the system as closed.

In the case of the active beam let us imagine the illumination beam to be exactly the same size as the detector beam in the passive case. The active detection beam is made large. Let us have the same number of photons per unit area per unit time enter the water. The same number of photons will wind up heading back to the satellite sensor as in the passive case. A number of photons will have left the source beam, but this doesn't matter as the detector beam is large and will catch them. Thus the two systems are equivalent as we started with the same number of photons (per unit area per unit time) and we get the same number of photons returned. The depth integrated return from an active system with a 1 km ISS and a larger DSSshould then be the same as that of a passive sensor with a 1km DSS. This is a potentially very useful concept as we can now compare passive remote sensing signal strength such as that obtained with SeaWiFS or MODS with depth-integrated Lidar signal strength.

The above construct simplifies life as strange things happen in the active source beam. On the way down photons leave the beam, so that K_d is not uniform within the beam. In the middle, if the beam is wider than 1/beam attenuation coefficient, we have a nearly plane parallel situation, and K_d is the same as in the case of sunlight, K_{dsun} . Towards the edges light is leaking out of the beam, and $K_d > K_{dsun}$. Just outside of the beam, however, we have only upwelled light near the surface. As we go down, we have more and more downward photons outside of the beam due to scattering, thus in that area, the light can be increasing with depth, leading to negative K_d! The beams will be interesting to model. On the way up we will also have non-uniform K_u's at the same depth. Using reciprocity we can avoid all that and study the passive case instead. We can then also use known models for diffuse K_d and K_u for radiance.

5.1 BACKSCATTERING SIGNAL

In this case we are interested in the time dependent signal of the upwelling radiance, $L_u(0^-,t)$, just below the sea surface. In particular we wish to know how much energy is received in the time period (t, t+Δt). In that case only photons having traveled distances (z, z+Δz) = c_W(t, t+Δt), where c_W is the speed of light in water, will be counted. If t is the round trip time to depth z, we can then approximate the signal from the layer due to backscattering as follows:

$$S_b(z,\Delta z)= \int_t^{t+\Delta t} L_u(0^-,t)\, dt = E_d(0^-) \int_z^{z+\Delta z} \beta_b(z')\, e^{-\zeta(z')} dz' \quad ,$$

where $\Delta z = (\Delta t/2)\, c_W$. If we assume a homogeneous ocean (i.e. β_b and ζ are constant with depth), and a light illuminating at 532 nm, we get:

$$S_b(z,\Delta z)= \int_t^{t+\Delta t} L_u(0^-,t)\, dt = E_d(\lambda, 0^-)\, F_b \frac{b_b(\lambda)}{\zeta(\lambda)} \left[e^{-\zeta(532)z} - e^{-\zeta(532)(z+\Delta z)} \right] \quad (23)$$

5.2 STIMULATED FLUORESCENCE, EXCITATION WAVELENGTH 532 NM, EMISSION WAVELENGTH 685 NM.

In the case of chlorophyll fluorescence, we note that the fluoresced light is equal to a quantum efficiency of conversion, Q(532,685) multiplied by the light absorbed by pigments over the depth interval, less a small portion of the fluoresced light re-absorbed within the algal cells. Let us set the pigment absorption coefficient equal to $a_p(532)$. This fluoresced light must now be brought to the surface. Only half of the fluoresced light will go into upward directions and only a fraction of the fluoresced light is radiated into directions that can be detected by the satellite. We will call this fraction F_f (it should nearly be equal to $\Omega_s /2\pi$, where Ω_s is the detection solid angle of the satellite). The light that reaches the surface is attenuated by the diffuse attenuation coefficient for 685 nm light. The total signal at the surface due to fluoresced light in the interval (z, z+Δz) is then:

$$S_f(685, z,\Delta z) = E_d\,(532, 0^-)\, F_f\, Q(532,685)\frac{a_p(532)/2}{K_d\,(532) + K_u(685)}\left[e^{-\,(K_d(532) + K_u(685))z} - e^{-\,(K_d(532) + K_u(685))\,(z + \Delta z)}\right] \quad (24)$$

5.3 RAMAN SCATTERING, EX 532 NM, EM 651 NM

The emission is at around 651 nm (Marshall and Smith, 1990), so that we have to use the diffuse attenuation at 651nm for the upwelling light. We then get, similar to Eq.(5):

$$S_R(651, z,\Delta z) = E_d\,(532, 0^-)\, F_R\, Q(532,651)\frac{a_p(532)/2}{K_d\,(532) + K_u(651)}\left[e^{-\,(K_d(532) + K_u(685))z} - e^{-\,(K_d(532) + K_u(651))\,(z + \Delta z)}\right] \quad (25)$$

5.4 SIGNAL STRENGTH ESTIMATIONS

We can now make some first order calculations as to the signal strengths of the scattered, Raman, and fluoresced light intensities.

First let us assume that $F_b = F_f = F_R = 1/2\pi$. If the upwelled scattered light, Raman, and fluoresced light are assumed to be totally diffuse, this would be reasonable. Note that by using the ratio, the signal will be in units of W/m^2sr. The incoming light is an irradiance units of W/m^2. The ratio will thus have units of sr^{-1}, and can be thought of as photons in over photons out per unit solid angle.

We now need to model $\beta_b(z)$, $K_d(z)$ and $k_u(z)$ (and hence$\zeta(z)$) at the various wavelengths (for simplicity we ignore the (λ) notation in the next few paragraphs). It was already shown that $k_u(z)$ can be modeled as (Zaneveld,1995):

$$k_u(z) = a(z)/\ \bar{\mu}_\infty(z) \quad (26)$$

which was given as a function of $b(z)/c(z) = \omega_o(z)$ in the next section. This model is based on observations which show that in the backward direction the radiance has a constant attenuation coefficient with depth, even near the surface, whereas the forward radiance does not have a constant attenuation coefficient with depth.

Modeling $K_d(z)$ requires the inclusion of the depth dependence of the shape of the radiance distribution. This can be accomplished by using Gershun's equation (see section 3):

$$K(z) = a(z) / \bar{\mu}(z)\quad , \quad (27)$$

where a(z) is the absorption coefficient and $\bar{\mu}(z)$ is the average cosine of the light field. $K_d(z)$ differs from K(z) by only a few percent, so that we may set:

$$K_d(z) \approx a(z) / \bar{\mu}(z)\quad . \quad (28)$$

Berwald et al. (1995) have derived a parametric model for the dependence of $\bar{\mu}(z)$ on ω_o for a vertical sun in a black sky. This fits our problem exactly. Barnard et al. (1998) have given global spectral slope averages for the particulate absorption and scattering coefficients and the yellow matter absorption all referenced to 488nm. Using these averages, we relate the absorption and scattering coefficients at 532nm to those at 651 and 685nm. In a few cases (e.g. Twardowski et al., 2001), we have measured backscattering coefficients. In general, only the total scattering coefficient is measured. For modeling purposes, in the latter case, one can use Petzold's (1972) average of: backscattering = 0.015* total scattering coefficient. In all cases we must add the pure water contribution since backscattering sensors are calibrated to exclude the pure water contribution.

The quantum efficiency for fluorescence is 2- 5% (Sam Laney, pers.comm.) Marshall and Smith (1990) have determined that the Raman scattering coefficient is 2.6×10^{-4} m^{-1} at 488nm. Note that this is about one order of magnitude less than the scattering coefficient of pure water. Using a λ^{-4} dependence we calculate the Raman scattering coefficient at 532nm to be 8.2×10^{-5} m^{-1}.

We now have all the parameters in hand to model the range-gated backscattering, Raman and stimulated fluorescence signals.

6.0 Inherent, Radiometric, and Apparent Optical properties

6.1 INHERENT OPTICAL PROPERTIES

Two fundamental IOPs are a (m^{-1}) and b (m^{-1}), the rates of radiant intensity loss over a fixed pathlength due to the processes of absorption and scattering, respectively. The **beam attenuation coefficient**, c **(m^{-1})**, is defined by their sum:

$$c = a + b. \tag{29}$$

There are many ways to decompose total or integrated IOPs into constituent IOPs. Scattering, for example, can be partitioned with respect to its angular distribution, the **volume scattering function** ($\beta(\theta)$, VSF), **units of $m^{-1}sr^{-1}$**. The VSF is defined by

$$\beta(\theta) = \frac{dI(\theta)}{EdV}, \tag{30}$$

where dI (θ) is the **radiant intensity (w/sr)** emanating into a small solid angle when a small volume dV is illuminated by an irradiance E. The beam attenuation coefficient and the volume scattering function are the IOPs that appeared in the equation of radiative transfer, Eq (3), and so form the connection between the particulate and dissolved materials and the remotely sensed radiances.

If we integrate the light emitted over all directions, we obtain the **total scattering coefficient, b, units of m^{-1}.** The total scattering coefficient can be divided into forward, b_f, and backward, b_b, components:

$$b_f = 2\pi \int_0^{\pi/2} \beta(\theta) \sin\theta \, d\theta \quad \text{and} \quad b_b = 2\pi \int_{\pi/2}^{\pi} \beta(\theta) \sin\theta \, d\theta \tag{31}$$

$$b = b_b + b_f. \quad (32)$$

The theoretical aspects of light scattering are treated extensively in van de Hulst (1981). For the various semi-analytical and analytical remote sensing algorithms (see section 4), we now have defined the two key IOPs relevant to the remote sensing reflectance, a and b_b. These IOPs are then often separated into operationally defined components such as the dissolved and particulate fractions and water:

$$a_t = a_g + a_p + a_w, \text{ and} \quad (33)$$

$$b_{bt} = b_{bp} + b_{bw}, \quad (34)$$

which applies to Eq. (29) as:

$$c_x = a_x + b_x \qquad \text{(where subscript x = t, g, p, or w).} \quad (35)$$

The subscripts t, g, p, and w represent total, dissolved (historically called gelbstoff or gilvin), particulate, and water, respectively. Operationally, the dissolved fraction typically comprises all substances that pass through a 0.2 μm filter. Other commonly used parameters are a_{pg} and c_{pg}, defined as the quantities $(a_p + a_g)$ and $(c_p + c_g)$, respectively.

Eq. (34) assumes that scattering from dissolved molecules in seawater will be negligible compared to the other terms. Another common assumption with errors typically less than 1% (Twardowski and Donaghay 2001) is that $c_g \approx a_g$ because the total scattering from dissolved materials in natural waters, b_g, is sufficiently low relative to a_g. This may be disputed, however, in waters with a high content of fine clays, where colloidal material passing through a 0.2 μm filter may be detectable (Aas 2000).

For algorithms focusing on the absorption and backscattering by phytoplankton, an additional partitioning of the particulate component of Eqs. (33) and (34) is often made:

$$a_p = a_\phi + a_d, \text{ and} \quad (36)$$

$$b_{bp} = b_{b\phi} + b_{bd}, \quad (37)$$

where the ϕ and d subscripts represent the algal and non-algal components, respectively. The non-algal component is comprised of non-living particulate organic material, living particles such as bacteria, inorganic minerals, and bubbles. The relative contributions of these different particle groups to particulate backscattering is poorly known, but recent progress has been made (Stramski et al. submitted). All the IOPs in Eqs. (26)-(33) have wavelength dependencies, examples of which can be found throughout the books by Shifrin (1988), Kirk (1994) and Mobley (1994).

Fluorescence is also an IOP that can be detected with passive and active remote sensing techniques. Common fluorophores in the dissolved fraction include humic substances (humic and fulvic acids), proteins, and hydrocarbons. Fluorescent phytoplankton pigments include chlorophyll, phycoerythrin, and phycocyanin.

The IOP fractional components discussed in this section can be related to several biogeochemical parameters (Twarowski et al., **Chapter 4, Table 1,**). Algorithms exist to derive nearly all of these IOPs from passive remote sensing or active lidar platforms (for example, Garver and Siegel 1997; Hoge and Lyon 1999; Roesler and Boss 2003) and, as a consequence, remote sensing algorithms have been developed for many of these biogeochemical properties. Excellent IOP reviews, including components and some biogeochemical associations, are given in Jerlov (1976), Shifrin (1988), Dera (1992), Kirk (1994), and Mobley (1994).

6.2 RADIOMETRY AND APPARENT OPTICAL PROPERTIES

The fundamental radiometric property is the **radiance distribution** ($L(\theta,\phi,\lambda)$, units of **$W\ m^{-2}\ sr^{-1}$ or quanta $m^{-2}\ s^{-1}\ sr^{-1}$**), described as the radiant power in a specified zenith (θ) and azimuth (ϕ) direction per unit solid angle, per unit area normal to the incident beam at a given wavelength. The radiance distribution in the sea (and above it) can never be constant with depth, as it results from the modification of the incident radiance field by the sea-surface, the inherent optical properties of the ocean interior, and the reflectivity of the sea bottom.

All other radiometric quantities derive from this. In particular, the various irradiances are derived by weighted integration of the radiance field over defined solid angles. The **downwelling ($E_d(\lambda)$, units of $W\ m^{-2}$) and upwelling ($E_u(\lambda)$, units of $W\ m^{-2}$) irradiances** are given as the cosine-weighted integration of the radiance distribution over the upper (downwelling) and lower (upwelling) hemispheres, respectively (see section 3). These hemispheres are separated by a horizontal surface oriented normal to the local gravity vector. The **net downward irradiance ($E(\lambda)$, units of $W\ m^{-2}$**) represents the vertical component of the irradiance vector and is given by the difference between the upward and downward irradiances, or the cosine weighted integral over all solid angles. A further quantity of biogeochemical and physical interest is the **scalar irradiance ($E_o(\lambda)$, units of $W\ m^{-2}$**) which results from the unweighted integration of radiance over all hemispheres.

A particularly useful relationship results from the integration of the radiative transfer equation to yield the Gershun equation as derived in section 3. Another derived quantity of interest results from spectral integration over the wavebands active in photosynthesis, generally taken as the interval from 350 or 400 nm to 700 or 750 nm. All of the above irradiances can be spectrally integrated, to provide a measurement of the so-called "**Photosynthetically Available Radiation" (PAR, units of $W\ m^{-2}$ or quanta $s^{-1}\ m^{-2}$)**. This is typically given as the integrated scalar irradiance.

6.3 THE AIR-SEA INTERFACE

For remote sensing, measurements of radiance and irradiance taken in water must be related to remotely sensed above-water radiances. To do this requires consideration of two factors, first the propagation of measurements taken at depth to the surface, and second the propagation of radiance across the sea-air boundary. The second is more straightforward than the first:

$$L_w(0^+,\theta,\phi) = L_u(0^-,\theta',\phi')\,\frac{1-\rho(\theta,\theta')}{n^2} \qquad (38)$$

where the water-leaving radiance just above the water, $L_w(0^+,\theta,\phi)$, in a given direction (θ,ϕ) derives from an upwelling below water (0^-) radiance stream, $L_u(0^-,\theta',\phi')$, of direction (θ',ϕ'). The two streams are related through Snell's law, $\theta' = \sin^{-1}[\sin\theta / n]$. The index of refraction, n, is actually an inherent optical property (the real part of the complex refractive index) and is dependent on salinity and (weakly) on temperature and pressure (Austin and Halikas, 1976). The reflection of the air water interface is given by ρ; note that $\rho = 1$ for incident angles greater than the critical angle ~ 48° for n = 1.34. Note as well that this relationship presumes no transpectral scattering (e.g., water Raman effects).

Most of the historical work has assumed nadir viewing geometry ($\theta = \pi$). In this special case, $L_w(0^+,\pi,0,) \approx 0.55\, L_u(0^-,\pi,0)$. However, most remote sensing instruments view the ocean surface at angles removed from nadir. Furthermore, the Fresnel scattering of downward radiance from the ocean surface upward is a strong function of illumination and viewing geometry. These so-called bi-directional characteristics of the radiance field incident on the sensor on orbit are therefore considerably more complicated. A rather complete theoretical analysis of this can be found in Morel and Gentilli (1996) and Mueller (2003); full evaluation of the bi-directionality of the radiance field below and above the sea-surface will require the routine measurement of the full radiance field (e.g. Morel et al. 1995; Voss et al. 2003).

A more problematic situation occurs when radiance measurements taken at depth are required to be propagated to the sea-surface to estimate $L_u(0^-,\pi,0)$ (in practice, nadir viewing instruments are usually employed, but in principle, the full radiance distribution could be used as well). It is rarely possible or even feasible to measure L_u accurately near the sea-surface (i.e., 0^-), given the presence of surface waves of various scales, and typically, reliable measurements have only been made for depths z > ~1-2 meters in the open ocean. More recent instruments provide accurate statistics of upwelling radiance at depths ~ 10 cm, at least for moderate sea-states (see Twardowski et al., chapter 4).

Given an accurate measurement of radiance at a range of depths, the problem faced is the extrapolation to just below the sea surface. The usual approach is to assume homogeneity over the upper ocean in some sense, and compute $L_u(0^-,\pi,0)$ for $\theta=\pi$ in:

$$L(0^-,\theta,\phi) = L(z_0,\theta,\phi)\exp\left[\int_0^z k(z,\theta,\phi)dz\right] \tag{39}$$

where z_0 is a reference depth below the surface, and $k(z,\theta,\phi)$ is the diffuse spectral attenuation coefficient for radiance at depth z_0 (assumed constant over the interval 0^- to z_0)as defined in Eq. (9b). The diffuse attenuation coefficient is operationally derived from the derivative of the neperian log of the vertical radiance profile and is usually assumed to be constant over the interval 0^- to z_0; analogous terms can be computed for the various irradiances.

Note that the diffuse attenuation coefficient is derived from radiometric properties, that are never constant as a function of depth, and therefore is an AOP. The rapid modification with depth of the radiance distribution in the upper optical depth, even with constant IOPs, implies that the assumption of homogeneity is almost assuredly invalidated; careful measurements, well-resolved in the vertical and taken near the sea-surface minimize this error, at least in the absence of significant surface roughness. At longer wavelengths (>650 nm), the strong attenuation of water conspires with increased instrument shading, fluorescence and Raman scattering to render this extrapolation extremely tenuous.

It is also necessary to determine the downwelling irradiance just beneath the sea surface, in order to determine the reflectance for the ocean alone. This is even more difficult as the influence of waves is larger on the downwelling irradiance than the upwelling radiance (Zaneveld et al., 2001). Even for small waves horizontal gradients can be much larger than vertical ones. A thorough discussion of light fields beneath waves can be found in Walker (1994).

In addition to providing a means to extrapolate radiances (and irradiances) within the upper ocean, the diffuse attenuation coefficients themselves are of considerable interest. Often viewed as "quasi-inherent" optical properties (e.g., Morel 1988, Gordon et al. 1988; Morel and Maritorena 2001), the close correspondence between K and the absorption coefficient places variations in K central to a wide range of applications, including the computation of photosynthesis (regulates the penetration of irradiance available for photosynthesis, as well as light absorbed; e.g., Behrenfeld and Falkowski 1997), the computation of local heating rates due to absorption of solar radiation (e.g., Lewis et al. 1990), the photochemical degradation of organic matter (e.g., Johannessen et al., in press), lidar system performance (e.g., Allocca et al. 2002), and underwater visibility (e.g., Zaneveld and Pegau 2003).

Remote sensing applications often derive K as an output product from measurements of normalized water-leaving radiances through empirical and semi-empirical approaches. For example, *K* can, with some accuracy, be decomposed into component contributions as with the IOPs in Eqs.(32-34), and can be used as diagnostics for constituents in the ocean, in particular the derivation of chlorophyll concentrations in oceanic case I waters (see Morel 1988; Gordon et al. 1988; Morel and Maritorena 2001).

The above sections deal with variations in apparent optical properties in the ocean interior, and their propagation to and through the sea-surface for the estimation of the water-leaving radiances required for calibration and validation of sensors on orbit. As an alternative approach, measurements of upwelling radiance can be made above the sea-surface from ship, buoy or tower platforms. Such measurements are appealing in principle, as they provide a direct measurement of radiance leaving the ocean, and are free from errors in propagation in the upper layer. However, in addition to the desired water-leaving photons, such measurements suffer from the inclusion of photons reflected off the sea-surface. This Fresnel reflectance includes both radiance resulting from the direct reflection of the Sun, and from sky reflectance.

For all but the calmest of seas, the contribution from surface reflectance is complex, and can often overwhelm the water-leaving signal (see full discussion in Walker, 1994; and Mobley, 1994). For a flat sea-surface and uniform sky radiance distribution, it is straightforward to compute the Fresnel reflectance over a small subtended solid angle looking down at the sea-surface; it is the downwelling radiance at equivalent relative azimuth and at the complementary zenith angle, multiplied by the Fresnel reflectance, which varies with respect to zenith angle from ~0.02 for normal incidence to ~0.03 at 40^{o} and then increases strongly with increasing angle in a well-behaved, and well-known manner.

In practice, even the lightest of winds ruffle the sea-surface, and uniform sky conditions are rarely encountered, except in heavily overcast days which are not of relevance to remote sensing applications as the sea-surface cannot be viewed from above. The physics are known; the difficulty is in the measurement (or in reality, parameterization) of the convolution of the sea-surface slope spectra (relative to the field of view) with the full sky radiance distribution, and the appropriate time-integration of the resulting at-sensor radiance time-series. With careful attention to detail, and under conditions approaching ideal, correspondence between water-leaving radiances determined from in-water approaches and above water measurements can be as good as 5%; under most realistic conditions, deviations >20% are more common. Current measurement approaches to the estimation of water-leaving radiances from above water platforms and caveats are discussed in Mueller et al. (2003).

7.0 Conclusions

Radiative transfer as related to optical remote sensing is a complex field that requires physical understanding of the absorption and scattering processes in the atmosphere and oceans, as well as sea surface and bottom reflection characteristics. Given this complexity, as briefly described in this chapter, the success with which IOP and particulate and dissolved properties have been derived from remotely sensed radiance is impressive, but many improvements remain to be made. The remainder of this book describes many of these inversions. In order to advance the field it will be necessary to obtain more detailed descriptions of the IOP and AOP together with particulate and dissolved properties, particularly in coastal zones.

8 .0 Acknowledgments

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