

## Optical Properties of Water (MS 69)

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### Synopsis

Light is critical in structuring aquatic ecosystems. Seasonal variations in the heat energy provided by sunlight are responsible for thermal stratification and mixing regimes of aquatic systems. Sunlight used in photosynthesis (both terrestrial and aquatic) forms the foundation of all but the simplest microbial communities of aquatic ecosystems.

This article addresses the fate of light in aquatic ecosystems. It starts with a discussion of the nature of light and the spectral output of the sun. Reflectance, scattering, and absorption in the atmosphere are addressed, along with the role of latitude, in regulating the incident flux of light at the surface of aquatic systems.

Once light penetrates the air-water interface it can either be scattered or absorbed. The inherent optical properties of absorptance, scatterance, and volume scattering function are defined as they pertain to aquatic systems. Wavelength specific differences in scattering and absorption are discussed in the context of optically significant components of the water column such as dissolved organic carbon and phytoplankton.

The diffuse attenuation coefficient is introduced as a parameter in defining the transparency of aquatic systems. Wavelength specific differences in this parameter are discussed and placed in the context of a system's inherent optical properties. Finally, the environmental variability of transparency is illustrated using data from 5 western hemisphere lakes.

## The Nature of Light

Light is critical in structuring all aquatic ecosystems. Seasonal variations in the heat energy provided by sunlight are responsible for thermal stratification and mixing regimes of aquatic systems. Sunlight used in photosynthesis (both terrestrial and aquatic) forms the foundation of all but the simplest microbial communities of aquatic ecosystems.

The term “light” is commonly referred to as that portion of the electromagnetic spectrum which is visible to the human eye. In reality, visible light accounts for a very narrow portion of an electromagnetic spectrum that ranges from cosmic rays to radio waves. Although some electromagnetic radiation originates from outside our solar system (indeed some as remnants of the “Big Bang”), an overwhelming majority of the electromagnetic radiation reaching Earth originates from the Sun.

All electromagnetic radiation should be viewed as having 2 fundamental properties which include both a physical entity (photons/quanta) and wave properties. Thus, electromagnetic radiation is a transverse wave of energy that behaves as a particle with a defined mass. The photon carries energy in the form of a wave; each discrete photon has a distinct wavelength and frequency associated with it. The wavelength ( $\lambda$ , m) and frequency ( $\nu$ , cycles  $\text{sec}^{-1}$ ) of electromagnetic radiation is inversely related by

$$\lambda = c/\nu \quad (1)$$

where  $c$  is the speed of light ( $3.0 \times 10^{10}$  cm  $\text{sec}^{-1}$  in a vacuum). When dealing with light, it is more typical (and convenient) to express wavelength in units of nanometers ( $10^{-9}$  meters).

The energy content of discrete photons (in Joules) of electromagnetic radiation is inversely related to wavelength via the Planck equation

$$E = h\nu = hc/\lambda \quad (2)$$

Where  $h$  is Planck’s constant ( $6.63 \times 10^{-34}$  J s).

The bulk of solar radiation spans wavelengths from approximately 100-3,000 nm (Figure 1). A relatively small proportion of this spectrum corresponds to “visible” light (400-700 nm). Wavelength bands within the visible spectrum are perceived as colors by the human eye. The perceived primary colors of blue (450-500 nm), green (500-550 nm), yellow (550-600 nm), and red (650-700 nm) are all associated with distinct wavelength bands within the visible spectrum. Ultraviolet radiation has wavelengths shorter than the visible spectrum (200-400 nm) while infrared radiation is longer (700-3000 nm).

Planck’s equation (Eq. 2) can be used to calculate the energy content of photons associated with any of these wavelengths. Relative to photons with a wavelength of 665 nm (a major absorption peak for chlorophyll a), infrared photons at 1000 nm contain only 66% as much energy. In contrast, photons associated with UV-b radiation at 300 nm contain 221% the energy at 665 nm. This explains why UV-b radiation can cause erythema (sunburn) while infrared radiation produces only a gentle warming sensation.

Even more ecologically significant than the phenomenon of vision, light provides the energy necessary for photosynthesis. The energy utilized in photosynthesis is restricted to a relatively narrow spectrum referred to as “photosynthetically active radiation” (PAR). Interestingly, and probably not coincidentally from an evolutionary perspective, PAR and visible light are essentially identical (400-700 nm). The upper limit of PAR is defined because photons at a wavelength greater than 700 nm have too little energy to overcome the biochemical threshold for production of  $e^-$  by PS-I and PS-II. In contrast, photons with wavelengths less than 400 nm are so energetic that they may potentially damage photosystems either directly, or via the production of free radicals. While vision and photosynthesis utilize only a narrow band of the solar spectrum, the region from 400-700 nm contains almost 45% of the energy output of the sun (at Earth’s surface). Evolution of vision and photosynthesis to use this narrow band resulted from a balance between available solar radiation and the optimum energy content per photon for biological systems.

As a result of the nature of electromagnetic radiation, the radiant flux ( $\Phi$ ) of light may be quantified in one of 2 ways: By counting quanta, or measuring energy. Each method has utility with respect to the study of aquatic systems.

Full summer sunlight at the surface of Earth delivers a flux density ( $\Phi$ ) of approximately  $1.2 \times 10^{21}$  quanta  $m^{-2} s^{-1}$  of visible light. In the past,  $\Phi$  was expressed in terms of moles of photons (1 mole =  $6.02 \times 10^{23}$  photons). According to recent international standards, 1 mole of photons is given the unit of einsteins (E). Thus, a “full sun” value of  $\Phi$  for a typical summer day is approximately 2,000  $\mu E$  of visible quanta  $m^{-2} s^{-1}$ . Similarly, the flux density may also be measured in units of energy. A “full sun” value for the flux density reported in energy units is approximately  $430 J m^{-2} s^{-1}$  or  $430 W m^{-2}$  ( $1 J m^{-2} s^{-1} = 1 W m^{-2}$ ).

If the distribution of quanta or energy is known across the solar spectrum it is possible to convert between the 2 sets of units using Planck’s equation (Eq. 2). Often instrumentation to precisely measure the spectral distribution of quanta or energy in sunlight is not available. In these cases a conversion factor may be applied based on a typical spectral distribution of sunlight at Earth’s surface. Morel and Smith (1974) found that for solar radiation in the visible (and PAR) spectra, a conversion factor of  $2.77 \times 10^{18}$  quanta  $s^{-1} W^{-1}$  was accurate to within a few percent over a variety of sky conditions for measurements made in air.

## **Role of the Atmosphere**

Solar flux just outside Earth’s atmosphere is referred to as the “solar constant” and has a value of approximately  $1373 W m^{-2}$ . Reflectance, scattering, and absorption of light in the atmosphere can reduce this amount by 15-80% before reaching Earth’s surface. Scattering and absorption do not influence all wavelengths equally, thus significant changes in the spectral distribution of sunlight will occur as light penetrates the atmosphere.

Scattering of light by gas molecules in the atmosphere is proportional to  $1/\lambda^4$  (as described by Rayleigh’s Law). Some of this light is back scattered into space while some is scattered forward and reaches Earth’s surface as skylight. Since scattering is inversely related to  $\lambda^4$ , the shortest wavelengths are preferentially scattered in the atmosphere,

causing a cloudless sky to appear blue. Although not visible, large fluxes of short wavelength UV radiation can reach the surface of Earth as scattered skylight (in contrast to direct sunlight). In addition to gas molecules, particulate material and water vapor (as clouds) in the atmosphere can also contribute to light scattering.

The absorption of light by the atmosphere is largely a result of naturally occurring oxygen, ozone, carbon dioxide, and water vapor. Anthropogenic air pollution may also have significant effects in some regions. Ozone strongly absorbs high energy UV-B radiation and essentially prevents light of wavelengths shorter than 300 nm from reaching the surface of Earth. Water vapor and carbon dioxide strongly absorb some specific bands of infrared radiation and are significant as “greenhouse” gases. Figure 1 compares the spectral distribution of light above the atmosphere and at sea level.

Cloud cover can be effective in both reflecting and absorbing light in the atmosphere. The transmissivity of thick cumulus clouds may be as little as 10% while that of high stratus clouds may be as high as 70%. The transmissivity of total solar radiation can be as high as 86% under dry, cloudless, and “clean” atmospheric conditions. According to Gates (1962), the average annual transmissivity of the atmosphere over a region in the northern hemisphere was 47%. Reflectance and back scattering of light into space accounted for a loss of 34% while 19% was lost via atmospheric absorption.

### **The Role of Latitude**

Latitude plays a critical role in determining solar flux at Earth’s surface ( $E_0$ ) by its influence on sun angle and day length (Figure 2). Insolation generally declines with increasing latitude through its influence on sun angle. However during summer, this phenomenon is countered by increasing day length. The net result is that during mid-summer, high latitude regions may actually receive *greater* daily insolation than equatorial regions. Thus, the *maximum* daily insolation received at the surface of Earth does not vary significantly with latitude (disregarding any variations in atmospheric conditions). The most profound influence of latitude on insolation is in determining the *minimum* daily insolation. During winter, increasing sun angle is accompanied by decreasing day length at high latitudes. In fact, daily insolation falls to zero in polar regions during the winter.

Reflectance, scattering, and absorption of light by the atmosphere also increase with sun angle. Loss of light by scattering and absorption are both a function of the amount of atmosphere traversed. The atmospheric pathlength is proportional to the cosine of the solar angle. Thus, the atmospheric pathlength is roughly doubled when the sun is 30° above the horizon compared to being directly overhead. For this reason, the transmissivity of the atmosphere will vary greatly according to both latitude and time of day.

### **Reflectance at the Air-Water Interface**

Light at the surface of a water body may be reflected. Unless re-reflected or scattered back to the aquatic system from the atmosphere or the surrounding landscape, this light is lost from the system. The amount of light reflected from the water’s surface

is related to the incident angle of the sun (angle from the vertical, often referred to as the zenith angle) as described in the Fresnel Equation

$$r = \frac{1 \sin^2(\Theta_a - \Theta_w)}{2 \sin^2(\Theta_a + \Theta_w)} + \frac{1 \tan^2(\Theta_a - \Theta_w)}{2(\tan^2(\Theta_a + \Theta_w))} \quad (3)$$

where  $r$  is the reflection of un-polarized light as a fraction of the incident light,  $\Theta_a$  is the zenith angle of the sun, and  $\Theta_w$  is the angle of refraction (i.e. the change in the angle of a light beam as it penetrates the surface-water interface). Reflectance for a flat body of water with the sun directly overhead (i.e.  $\Theta_a = 0^\circ$ ) is approximately 2% and remains relatively low for zenith angles up to about  $60^\circ$  ( $r < 6\%$ ). As the zenith angle approaches  $90^\circ$ , reflectance increases rapidly (e.g.  $r \sim 60\%$  at  $80^\circ$ ). At small zenith angles, surface disturbances (i.e. small waves) have very little influence on reflectance but at more extreme zenith angles ( $>60^\circ$ ) reflectance can be increased by as much as 20%. Large waves in combination with high zenith angles may reduce reflectance by decreasing the effective zenith angle between the water's surface and the sun. Reflectance of light from frozen water bodies has been estimated to be as much as 75-95% however this value depends on the quality of the ice or snow.

This discussion of reflectance deals solely with direct solar radiation. Reflection of diffuse solar radiation striking the surface of the water as a result of scattered skylight is difficult to determine precisely because the angular distribution of light is unknown.

### **Inherent Optical Properties**

Once photons penetrate the air-water interface of aquatic systems they may only be scattered or absorbed. The extent to which either of these phenomena may occur is governed by 3 parameters: absorption coefficient, scattering coefficient, and the volume scattering function. These 3 parameters are often referred to as “inherent optical properties” because they only depend on constituents of the aquatic medium and not on the geometry of the light field.

In order to define these parameters it is necessary to consider an infinitesimally thin layer of medium being illuminated at right angles by a parallel beam of light (Figure 3). The fraction of flux that is absorbed across the medium relative to the incident flux is defined as the absorptance (A) and the fraction of flux which diverges from the parallel beam of incident flux is referred to as the scatterance (B)

$$A = \Phi_a / \Phi_0 \quad (4)$$

$$B = \Phi_b / \Phi_0 \quad (5)$$

where  $\Phi_0$  is the incident flux of the parallel beam of light,  $\Phi_a$  is the flux absorbed and  $\Phi_b$  is the flux scattered by the thin medium. If we represent the change in absorptance ( $\Delta A$ ) and scatterance ( $\Delta B$ ) with infinitesimal changes in the thickness of the medium ( $\Delta z$ ) we can estimate the absorption coefficient (a) and the scattering coefficient (b)

$$a = \Delta A / \Delta z \quad (6)$$

$$b = \Delta B / \Delta z \quad (7)$$

The beam attenuation coefficient ( $c$ ), an additional inherent optical property, is simply an expression of the amount of flux lost from the beam of light as a result of both absorption and scattering, per unit thickness of the medium

$$c = a + b \quad (8)$$

The absorption coefficient, scattering coefficient, and beam attenuation coefficient each have units of 1/length and are typically expressed as  $m^{-1}$ .

Another inherent optical property of the aquatic media is the volume scattering function ( $\beta$ ). As light is scattered within a media, the photons may be dispersed in variety of different directions. The volume scattering function defines the angular distribution of the resulting scattered flux. This function differs according to the medium but generally describes a radially symmetrical cone of distribution around the direction of the incident beam.

## Scattering

Scattering of light in aquatic ecosystems can be thought of as the deflection (reflection) of photons at a large number of angles from water and its dissolved and particulate constituents. While scattered photons are largely distributed in a radially symmetrical cone around the direction of the incident beam, there is a finite statistical probability that photons may be deflected in *any* direction. Scattered photons that are deflected back may leave the aquatic ecosystem along with their energy content. Scattering can vary with wavelength according to differential absorption and scattering characteristics aqueous media. In very clear aquatic ecosystems the greatest amount of scattering occurs in blue wavelengths (these systems appear blue because backscattered light is detected by the eye). Systems with large concentrations of dissolved organic carbon appear yellow/brown because blue photons are strongly absorbed and thus not available to be backscattered to the eye of the observer.

Scattering may be greatly influenced by the nature of particulate material in an aquatic ecosystem. For instance fine glacial “flour” may be highly reflective and can enhance scattering at the mouth of rivers draining glaciers. White calcium carbonate (marl) at the bottom of lakes may also enhance scattering (reflectance) in shallow systems or in the water column of deeper systems if sediments are re-suspended by wind driven mixing. In contrast, organic particulate material (colloids of humic substances, detritus, etc.) may strongly absorb rather than scatter photons.

The net result of scattering is that it can greatly increase the pathlength ( $\Delta z$ ) taken by a photon as it passes through an aqueous media. For instance, it is not uncommon for the “average” photon to travel a path of 1.2 meters (as a result of multiple scattering events) before reaching a depth of one meter in an aquatic ecosystem. Scattering also has important implications for aquatic photosynthetic organisms because scattered photons

are available from all directions (including the bottom) and not just that oriented toward the downward welling light.

## Absorption

Photons of light may interact with the electron shell of molecules and be absorbed. During this process, the photon is destroyed but its energy is transferred to the molecule that absorbed it. This energy is utilized to transition electrons from the ground state to higher energy levels. Molecules can have very distinct absorption spectra because the energy content of some photons (Eq. 2) match exactly the energy required for the electron transition. The energy of absorbed photons may be utilized to break chemical bonds or generate electrons (in the case of photosynthesis). However, due to requirements for conservation of energy, the energy of all absorbed photons is eventually released in aquatic ecosystems as heat.

Light absorption can be measured in a spectrophotometer which is an instrument specifically designed to measure the absorbance of light by a medium of known thickness. Values obtained by spectrophotometers are known equivalently as absorbance (A) or optical density (OD). Using a spectrophotometer, absorbance (A) can be calculated as

$$A = \log_{10} \frac{I_0}{I_z} \quad (9)$$

where  $I_0$  is the intensity of incident light and  $I$  is the intensity of light transmitted through the medium. If we arrange the system so that any light scattered in the medium is also included in our measurement of  $I$ , we can use Eq. 9 to determine the absorption coefficient ( $a$ ) of the medium

$$a = \frac{2.303A}{Z} \quad (10)$$

where 2.303 is the constant to convert base 10 to natural logarithms and  $Z$  is the pathlength of the medium. The absorption coefficient is given in units of 1/pathlength (typically  $\text{m}^{-1}$ ).

In aquatic ecosystems photons may be absorbed by the water itself, dissolved substances, and particulate material. Pure water is not a colorless liquid. As is obvious by observing very deep, clear systems, water is a blue liquid. This color implies that there is significant spectral variation in absorption over the visible region. As seen in Figure 4, the absorption of light by pure water is lowest in the blue range of the visible spectrum and increases substantially from 430 nm to 760 nm. Following a small decline centered around 810 nm, absorption increases markedly through the infrared region, peaking at about 975 nm. A gradual increase in absorbance is also evident through the UV-a and UV-b regions, although this may actually be an artifact of small amounts of dissolved organic carbon contamination present in even the most carefully purified water. The strong absorbance peak in the IR region attests the importance of infrared radiation in influencing the heat budgets of aquatic ecosystems.

Dissolved organic matter (DOM) plays the most critical role in influencing the absorption spectrum of natural waters. The colored fraction (typically yellow/brown and referred to as gelvin or gelbstoff) of DOC is referred to as chromophoric dissolved organic matter (CDOM). This material consists mainly of humic and fulvic compounds originating from the decomposition of terrestrial plant material under anaerobic conditions. The global mean concentration for DOM in freshwater systems is approximately  $5 \text{ mgC l}^{-1}$  but can vary widely ( $\ll 1 \text{ mgC l}^{-1}$  in very clear systems to  $>100 \text{ mgC l}^{-1}$  in bogs). As seen in Figure 5, the presence of CDOM causes an exponential increase in absorption from the short visible wavelengths through the UV-b region of the spectrum. The slope of this exponential curve ( $S, \ln(a) \text{ vs. } \lambda_{280-400\text{nm}}$ ) has been measured for a variety of freshwater systems and averages  $-0.0181 \text{ m}^{-1} \text{ nm}^{-1}$ .  $S$  typically ranges between  $-0.020$  and  $-0.010$  and seems to be related to qualitative differences in CDOM (likely due to CDOM source or photobleaching). CDOM has the greatest influence on short wavelength UV radiation but also strongly affects the absorption of visible light (especially blue). The optical property often referred to a “color” is typically measured as the absorption coefficient at 440 nm and is a reflection of CDOM concentration for a particular aquatic ecosystem. In systems with high CDOM concentrations, absorption of PAR can significantly reduce light available for photosynthesis. Due to the shape of the absorption spectrum, much smaller concentrations of CDOM can effectively absorb damaging solar UV-b radiation, producing UV refugia in aquatic systems. Above approximately 600 nm, CDOM has little influence on the absorption spectrum. Other dissolved constituents of aquatic systems (e.g. nitrate) may also absorb light but their specific absorption coefficients are very low compared to that of CDOM. Thus, these compounds will only be optically significant in systems with very low concentrations of CDOM.

Suspended particulate matter may also absorb light in aquatic systems (Figure 6). Phytoplankton evolution has produced photosynthetic pigments that are highly efficient at absorbing light in the PAR and UV spectrum. These cells often contain a variety of accessory pigments (e.g. carotinoids, chlorophyll *b*, and chlorophyll *c*, etc.) as well as chlorophyll *a*. As seen in Figure 6, the absorbance peaks of chlorophyll *a* (425 and 665 nm) are often apparent in the spectral analysis of seston over and above that of accessory pigments. Systems with high concentrations of CDOM may also have significant amounts of suspended organic colloids composed of humic matter. The chemical composition and absorption spectrum of this particulate matter is often similar to that of CDOM (compare Figures 5& 6). Figure 6 also shows an absorption scan for mineral clay particles. In contrast to either phytoplankton or colloidal CDOM, kaolin exhibits very little absorption above 400 nm.

## Optical Properties of Aquatic Ecosystems

The transparency of aquatic systems is represented by a value referred to as the extinction coefficient or diffuse attenuation coefficient ( $K_d$  or  $\eta$  in older literature). This value may be determined empirically for an aquatic system using the following:

$$E_{dz} = E_{d0} e^{-K_d z} \quad (11)$$

where  $E_{dz}$  is the downward welling irradiance at depth  $z$ ,  $E_{d0}$  is the downward welling irradiance just below the surface, and  $K_d$  is the extinction coefficient.  $K_d$  has units of reciprocal depth (typically  $m^{-1}$ ).

As demonstrated in the previous section, the extinction coefficient in aquatic systems will be influenced by absorption and scattering by pure water ( $K_{water}$ ), by materials dissolved in the water ( $K_{dissolved}$ ), and by particulate material ( $K_{particulate}$ ). Thus, the total measured diffuse attenuation coefficient for a given system will be the sum of each of the partial diffuse attenuation coefficients.

As illustrated by data presented in Figures 4-6, each of these partial diffuse attenuation coefficients have a strong spectral component. Thus, the *relative* contribution of each of these partial diffuse attenuation coefficients to the total diffuse attenuation coefficient will vary substantially across the solar spectrum and between systems. For instance, due to the spectral distribution of CDOM absorption,  $K_{dissolved}$  in lakes comprised on average 68% of  $K_{total}$  at 305 nm compared to only 33% for PAR. In contrast,  $K_{water}$  contributes a much larger fraction (>95%) of  $K_{total}$  in the infrared portion of the spectrum compared to the other partial diffuse attenuation coefficients. For lakes in general, the concentration of CDOM is the single most important determinant of  $K_d$  in the UV region of the spectrum while a combination of both CDOM concentration and phytoplankton biomass are significant determinants of  $K_d$  for PAR.

Plots showing the extinction of light in an aquatic ecosystem are shown for Lake Giles, PA (Figure 7). The top panel demonstrates that light diminishes exponentially with depth as one would expect from the form of Eq.11. The incident irradiance (i.e. at a depth of 0 m) follows from the data presented in Figure 1, where the highest value of radiant flux ( $\Phi$ ) is for PAR and a progressive decrease in flux is observed through each of the wavelengths in the UV spectrum. Light of each of the wavelengths diminishes through the water column with a characteristic slope. In fact, the value of  $K_d$  for each of these wavelengths can be derived as the negative slope of the plot of  $\ln(E_d)$  vs. depth (Figure 7, bottom panel).

Since  $K_d$  is an exponential rate constant, it is often difficult for the general public to embrace it as an estimate of water column transparency. A more generally understandable parameter is the 1% attenuation depth for light. This parameter is the depth in the water column at which 99% of the surface irradiance has been attenuated. This parameter can easily be calculated as

$$z_{1\%} = \frac{4.61}{K_d} \quad (12)$$

where  $z_{1\%}$  is the 1% attenuation depth (m) and  $K_d$  is the diffuse attenuation coefficient ( $m^{-1}$ ). The value 4.61 is simply  $\ln(100\%)$ . Lower values of  $K_d$  correspond to higher water column transparency. In Lake Giles (as well as many other aquatic systems) the progressive decrease in transparency with decreasing  $\lambda$  is due to the presence of CDOM ( $\sim 1.5 \text{ mg Cl}^{-1}$  in Lake Giles). The relationship between  $K_d$  and  $\lambda$  for Lake Giles will generally follow the relationship between the absorption coefficient ( $a$ ) and  $\lambda$  observed in for different concentrations of CDOM (Figure 5).

The transparency of PAR in aquatic ecosystems can vary tremendously. Figure 8 shows the vertical penetration of PAR in 5 lakes that span a range of transparency

typically found in aquatic systems. In extremely clear Lake Correntoso,  $z_{1\%}$  is in excess of 45 m while 99% of PAR is attenuated in the first meter of Little Echo Lake which has moderate concentrations of chlorophyll  $a$  ( $\sim 6 \mu\text{g l}^{-1}$ ) and a high concentration of CDOM ( $\sim 23 \text{ mgC l}^{-1}$ ).

One method of estimating the transparency of visible light (PAR) in aquatic systems is the Secchi disk. First used in the 1860s, the black and white Secchi disk (20 cm diameter) is lowered through the water column on a calibrated line. The depth at which the disk disappears is termed the Secchi depth ( $z_{\text{sd}}$ ). Although the Secchi depth can vary between users, light conditions, and surface conditions, it is generally thought that  $z_{1\%}$  for PAR corresponds to approximately 2-3 times  $z_{\text{sd}}$ . The value of Secchi measurements lies not in its accuracy, but in its ease of use and the extremely long records of transparency based on this device.

## **Conclusion**

In conclusion, light is critical in structuring aquatic ecosystems. Seasonal variations in the heat energy provided by sunlight are responsible for thermal stratification and mixing regimes so fundamental to aquatic systems. The energy of sunlight used in photosynthesis forms the foundation of all but the simplest microbial communities on Earth. Thus, an understanding of the nature of light, and how it interacts with the atmosphere and the water column, is fundamental to our understanding of how aquatic ecosystems function.

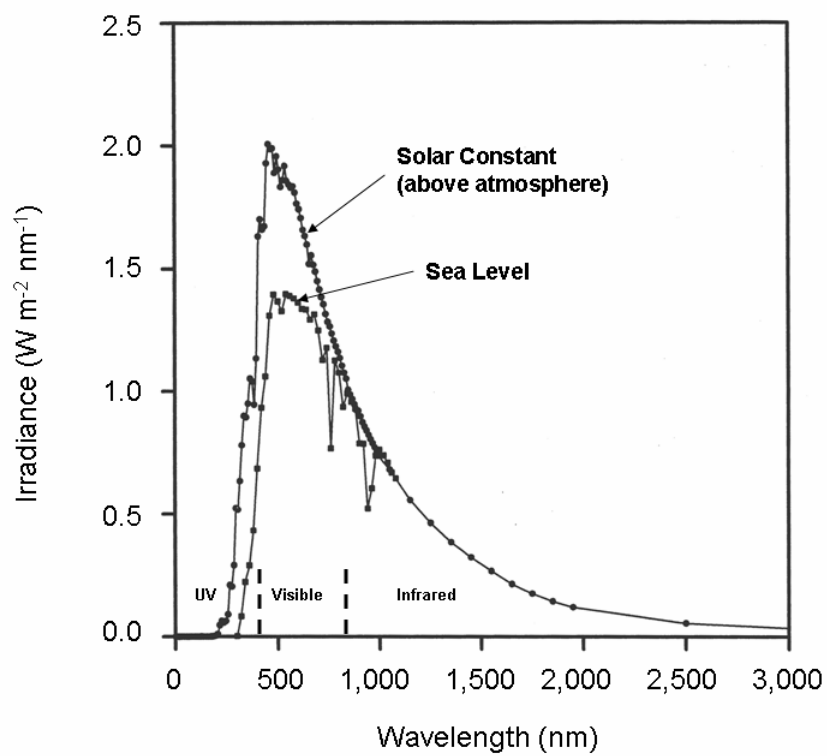


Figure 1. Spectral distribution of sunlight above the atmosphere and at sea level.

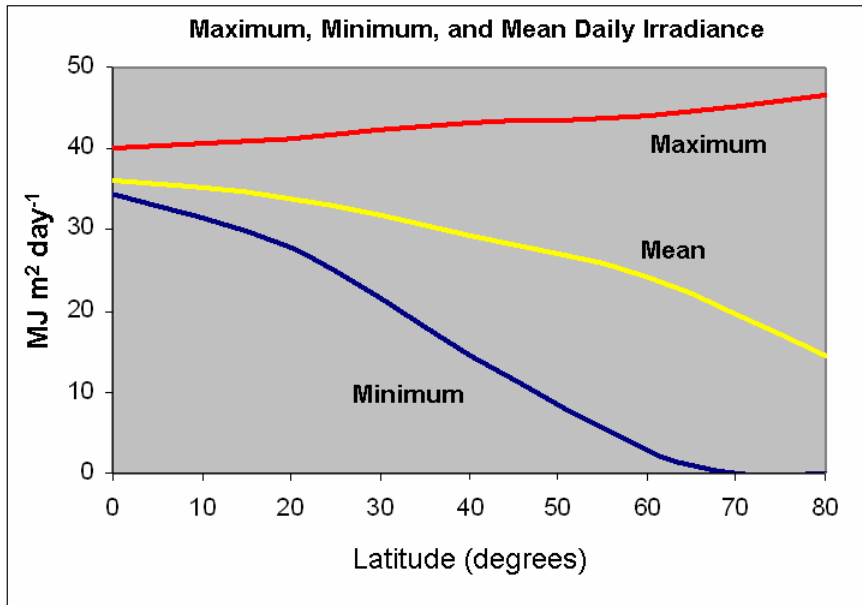


Figure 2. Maximum, minimum, and mean daily irradiance at different latitudes at the surface of Earth. Data do not account for regional differences in cloud cover.

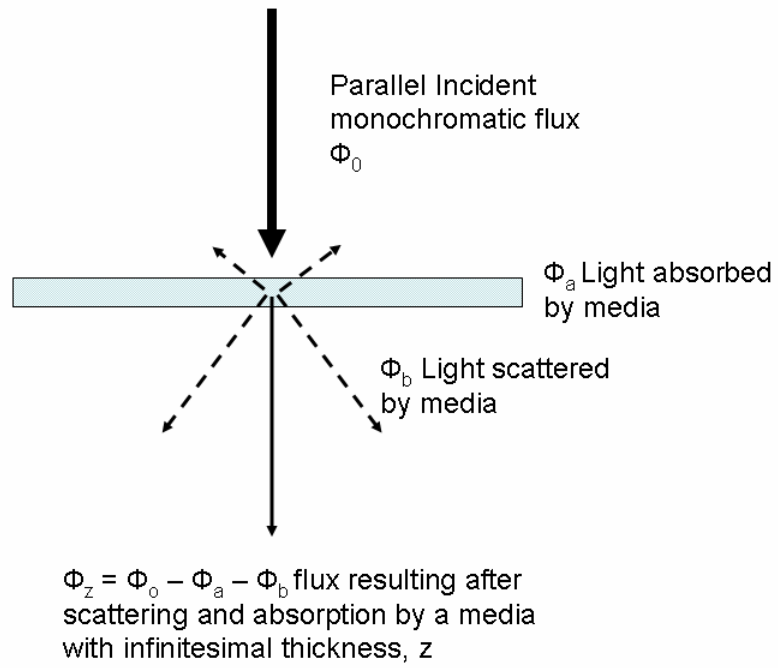


Figure 3. Schematic diagram illustrating absorption and scattering of light by an infinitesimally thin layer of water.

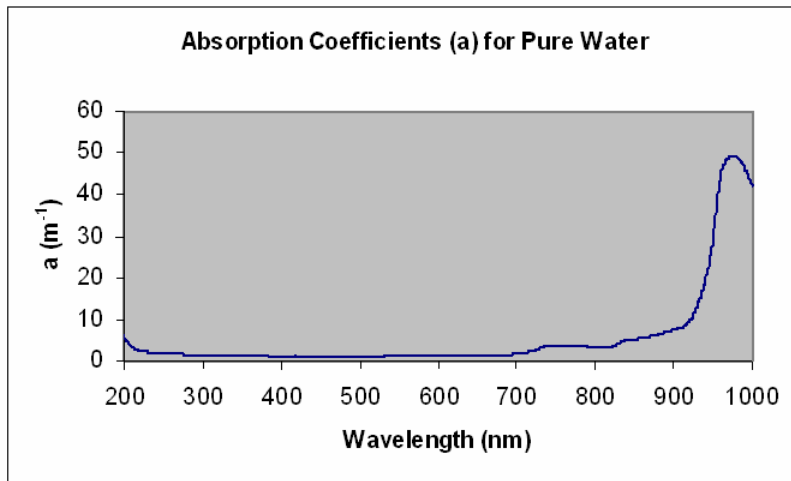


Figure 4. Spectral distribution of the absorption coefficient for pure water. Measurements were made with a spectrophotometer referenced to air.

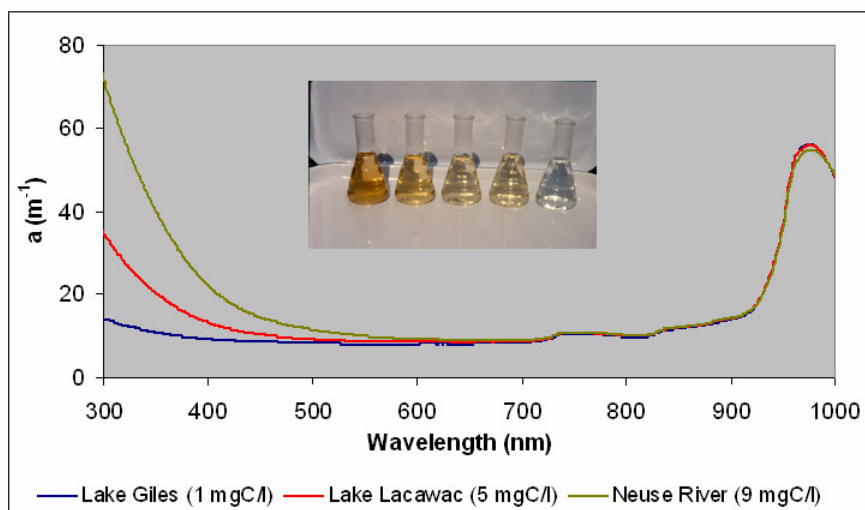


Figure 5. Spectral distribution of absorption for filtered water from lakes with different concentrations of CDOM. The inset shows the apparent color of 4 lake water samples ranging from approximately 1-15  $\text{mgC l}^{-1}$  of dissolved organic carbon. Measurements were made with a spectrophotometer referenced to air.

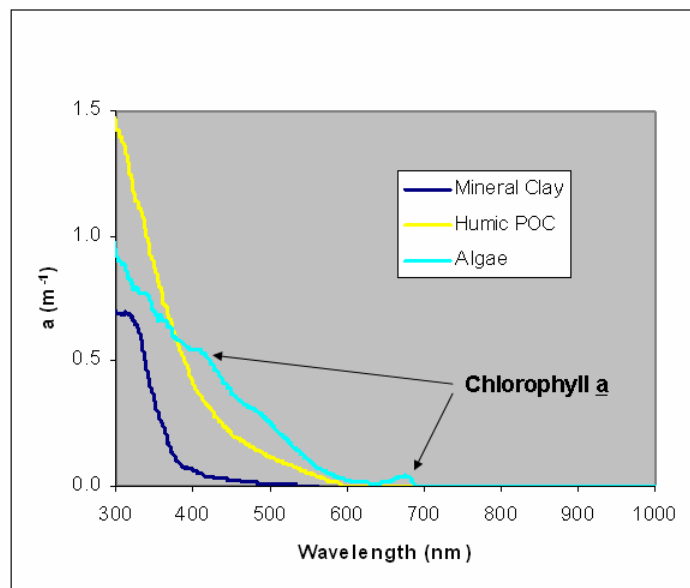


Figure 6. Spectral distribution of absorption for 3 categories of particulate material commonly found in aquatic ecosystems. Absolute size of the peaks will depend upon the concentration of these materials in the water column. Remnants of the chlorophyll a absorption peaks (665 and 425 nm) can be seen in the sample of algae. Data were obtained using spectrophotometry and the quantitative filter pad technique.

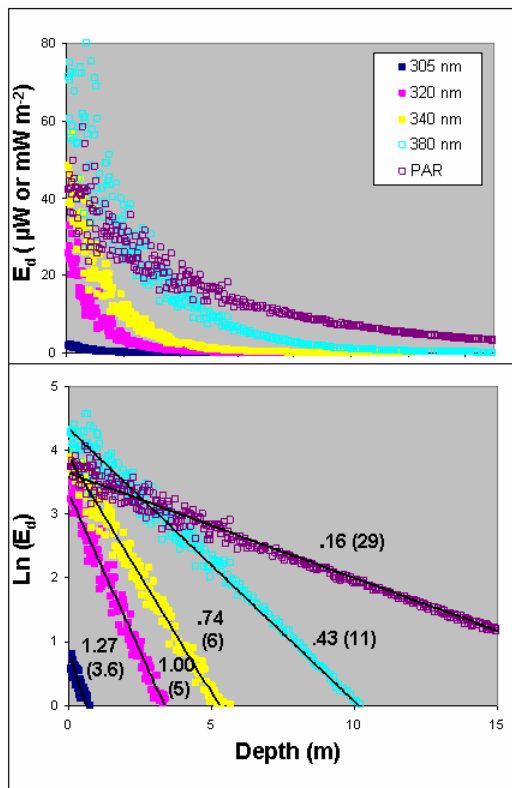


Figure 7. Attenuation of downward welling irradiance ( $E_d$ ) in Lake Giles, PA. In the top panel the units for PAR is  $\text{mW m}^{-2}$ . All other wavelengths are expressed as  $\mu\text{Wm}^{-2}$ . In the bottom panel the line of best fit is plotted through the data points of  $\ln(E_d)$  vs. depth. The negative slope of this line is the value of  $K_d$  for each wavelength. The values of  $K_d$  ( $Z_{1\%}$ ) are provided. Data were obtained using a Biospherical PUV-500 submersible radiometer.

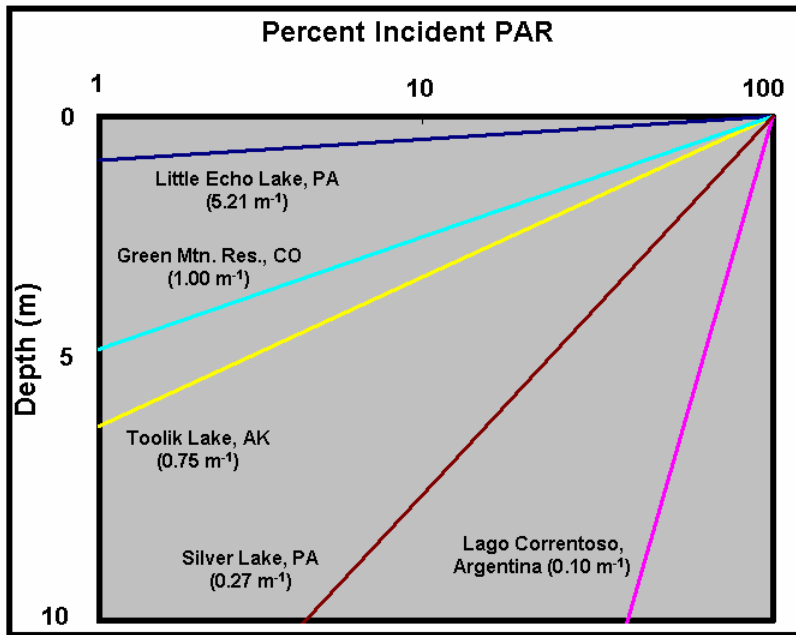


Figure 8. Penetration of PAR for lakes with widely differing transparency. Diffuse attenuation coefficients are provided for each lake ( $K_d$ ).

## Further Reading

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## List of special symbols

$\mu$  p. 2, Figure 7 caption.

$\Phi$  p. 2, 4, 8, Figure 3

$\Theta$  p. 4

$\Delta$  p. 4, 5

$\beta$  p. 5

$\lambda$  p. 1, 2, 7, 8

$\epsilon$  p. 1

$\eta$  p. 7