

A spectral light absorption meter for measurements in the sea

Abstract—The light absorption coefficient of a medium such as seawater is an inherent optical property that can be determined in absolute values by means of the uncalibrated in situ light absorption meter described here. The method is based on the assumptions that the optical conditions remain steady within the measuring period, that the horizontal divergence of the vector equals zero, and that the natural fluorescent light energy within the visible part of the spectrum can be ignored. The light absorption meter measures simultaneously the sum and the difference of the scalar and vector irradiance.

Direct spectral in situ measurements of absorption coefficients in the sea are few compared with measurements of light scattering and transmission. Accordingly, values of light absorption coefficients are often obtained indirectly from measurements of scattering and transmission. Other approaches to obtain absorption coefficients are measurements of the vertical attenuation coefficient of the downwelling irradiance in the upper few meters at high solar elevations (Jerlov 1961, 1968; Jerlov and Nygård 1969; Højerslev 1973, 1974*a,b*) (for lower solar elevations corrections are required) or approximate optical models only valid in the near-asymptotic region (Tyler et al. 1972).

There are two principal types of in situ light absorption meters. The most common one is equipped with a stabilized light source producing a collimated beam (Bauer et al. 1971; Gilbert et al. 1969). Such a meter takes advantage of the pronounced forward light scattering in seawater which, together with the direct transmitted light, impinges on a light receiving unit. It is independent of the daylight conditions, to a certain extent of the state of the sea, and is in principle not limited to any depth. However, in clear ocean waters it cannot measure the absorption coefficients at the most transparent wavelengths reliably because of its relatively short measuring path-length.

The other type of absorption meter measures the scalar and vector irradiance as a

function of depth and wavelength (Højerslev 1973; Tyler 1961). Insertion of these two parameters into Gershun's (1936) equation allows straightforward calculation of the absorption coefficient. This meter is commonly restricted to measurements in the upper 100 m of the sea, and stable surface light conditions are necessary. The method assumes that the divergence of the horizontal vector irradiance equals zero (Gershun 1936). Practically speaking, this is always the case when the meter is away from fixed boundaries (ship, bottom, etc.). The underwater fluorescent light energy formed by the shortwave part of the daylight is always small compared with the daylight energy at the same wavelength and depth. Accordingly, determinations of absorption coefficients in the wavelength range 372–670 nm (the measuring range of the light absorption meter) are practically unaffected by the fluorescent properties of natural seawater. The parameters measured by the spectral light absorption meter are important in the theory of radiative transfer (Højerslev 1972; Preisendorfer, 1959; Zaneveld and Pak 1972), including effects of varying solar elevation on the vertical transmission of daylight (Højerslev 1974*a*; Jerlov and Nygård 1969), optical water mass classification (Jerlov 1968), and photosynthetic studies (Højerslev 1973, 1974*a*; Smith and Wilson 1972). Finally, rather accurate absorption values can be obtained in clear ocean water because this instrument is not confined to small measuring distances.

For both types of absorption meters calibration errors are inevitable and calibration often cumbersome. As a consequence a new spectral absorption meter, using daylight but not requiring calibration, has been designed and built.

To avoid confusion the light collectors described below are named so that they correspond to the quantity collected. The light collectors are all assumed to be ideal, which in practice is never the case. However, techniques concerning optimization of

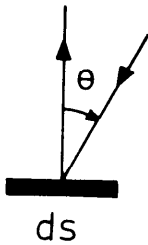


Fig. 1. Light collector appropriate for measurements of the downwelling irradiance. Notation: E_d -collector.

real light collectors will be excluded here since they can be found elsewhere (e.g. see Smith 1969). It suffices to say that when dealing with measurements of the downwelling light field at high to moderately high solar elevations most light collectors serve their purpose within an acceptable degree of accuracy, but when measuring the upwelling daylight optimal collectors are desirable.

Figure 1 symbolizes the classical E_d -collector. (Notation is given in Table 1.) Reversed it becomes an E_u -collector. Consequently, an E -collector consists of two flat collectors with opposite orientation.

Figure 2 depicts similarly the classical E_o -collector having the receiving flat collector ds placed horizontally. Ideally, the area of the spherical collector and the ratio of the area of ds to that of the spherical collector are both infinitesimal. Ignoring reflection, the light flux from the direction (θ, ϕ) falling on the sphere is given by

$$dF = \pi r^2 \times L \, d\omega, \quad (1)$$

where r = radius of the sphere. The total light flux from all direction is easily obtained by integration:

$$F = \pi r^2 \int_{4\pi} L \, d\omega = \pi r^2 \times E_o. \quad (2)$$

From this it can be proved—as done in many optical textbooks—that the receiving collector ds obtains an irradiance proportional to E_o .

Figure 3 shows an E_{od} -collector. The hatched area refers to an infinite absolute black horizontal plane at the same level as ds . The light flux from the direction $(\theta,$

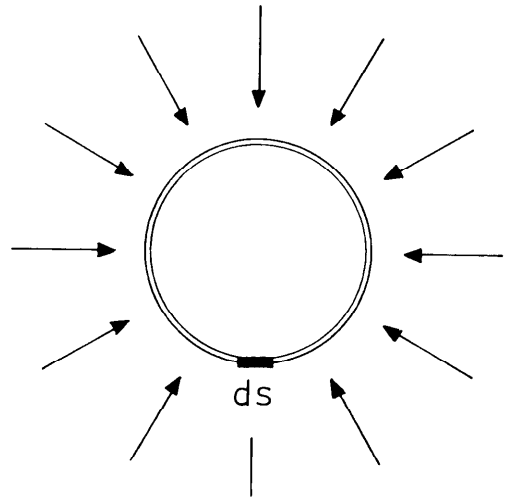


Fig. 2. Light collector appropriate for measurements of the scalar irradiance. Notation: E_o -collector.

$\phi)$ falling on the sphere can be given the same expression as in Eq. 1, but integration with respect to the polar angle θ can be performed only for the upper hemisphere.

Accordingly,

$$F = \pi r^2 \int_{2\pi} L \, d\omega = \pi r^2 E_{od}. \quad (3)$$

Inverted it becomes an E_{ou} -collector.

Figure 4 shows a $\frac{1}{2}(E_{od} + E_d)$ -collector. The horizontal black plane extends from the equator of the sphere to infinity, and the lower hemisphere is absolute black except at the receiving flat collector ds . From arguments similar to those above

$$dF = \frac{1}{2}\pi r^2 L \, d\omega + \frac{1}{2}\pi r^2 L \cos \theta \, d\omega, \quad (4)$$

and

$$F = \pi r^2 \times \frac{1}{2}(E_{od} + E_d). \quad (5)$$

Inverted it becomes a $\frac{1}{2}(E_{ou} + E_u)$ -collector.

Figure 5 shows a $\frac{1}{2}(E_o + E)$ -collector. As in the case of the $\frac{1}{2}(E_{od} + E_d)$ -collector, only the upper hemisphere is exposed to the submarine light field but the infinite horizontal black plane is removed. Again the light fluxes collected are obtained from integration of Eq. 4 but now with the limits of integration being 0° and 180° . Accord-

Table 1. Definitions of terms used. The notation follows the recommendations of IAPSO (1964). [The equation for E_u should read $E_u = \int_{-2\pi} L |\cos \theta| d\omega$.]

$L(z, \theta, \varphi, \lambda)$	= radiance	as a function of depth, polar- and azimuthal angle and wavelength
$E_d = \int_{2\pi} L \cos \theta d\omega$		downwelling irradiance, where $\int_{2\pi} d\omega$ means integration over all solid angles in the upper hemisphere
$E_u = \int_{-2\pi} L \cos \theta d\omega$		upwelling irradiance, where $\int_{-2\pi} d\omega$ means integration over all solid angles in the lower hemisphere
$E = \int_{4\pi} L \cos \theta d\omega = E_d - E_u$		vector irradiance
$E_{od} = \int_{2\pi} L d\omega$		downwelling scalar irradiance
$E_{ou} = \int_{-2\pi} L d\omega$		upwelling scalar irradiance
$E_o = \int_{4\pi} L d\omega$		scalar irradiance
$K_E = -\frac{1}{E} \frac{dE}{dz}$		vertical attenuation coefficient for the vector irradiance (m^{-1})
$K_o = -\frac{1}{E_o} \frac{dE_o}{dz}$		vertical attenuation coefficient for the scalar irradiance (m^{-1})
Other vertical attenuation coefficients are defined in a similar way below:		
$K_d = -\frac{1}{E_d} \frac{dE_d}{dz}$		(m^{-1})
$K_u = -\frac{1}{E_u} \frac{dE_u}{dz}$		(m^{-1})
$K_1 = -\frac{1}{E_o + E} \frac{d(E_o + E)}{dz}$		(m^{-1})
$K_2 = -\frac{1}{E_o - E} \frac{d(E_o - E)}{dz}$		(m^{-1})
a	= absorption coefficient	(m^{-1})
b	= scattering coefficient	(m^{-1})
c	= $a + b$	attenuation coefficient (m^{-1})
$R = \frac{E_u}{E_d}$		irradiance reflectance

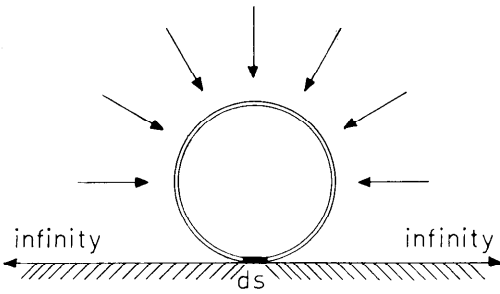


Fig. 3. Light collector appropriate for measurements of the downwelling scalar irradiance. Notation: E_{oa} -collector.

ingly, $F = \pi r^2 \frac{1}{2}(E_o + E)$. When the $\frac{1}{2}(E_o + E)$ -collector is turned downward it becomes a $\frac{1}{2}(E_o - E)$ -collector.

Figure 6 illustrates two collectors often termed as 2π -collectors. The receiving collector ds lies symmetrically in the equator plane of the opalized sphere and the hatched areas correspond to black surfaces. Such collectors are incapable of measuring any optical parameter in a well defined way because the quantities they measure are dependent on the radiance distribution. This can be seen from the following arguments: The radiance distribution $L(z, \theta, \phi)$ in water causes a certain radiance distribution $L_{col}(z, \theta, \phi)$ on the innerside of the hemispherical collector dependent on its diffusing and reflecting properties. Unless $L(z, \theta, \phi)$ is a constant in the upper hemisphere and zero in the lower, $L(z, \theta, \phi) \neq \text{constant} \times L_{col}(z, \theta, \phi)$ for all θ, ϕ . The irradiance on the receiving area ds equals

$$\int_0^{2\pi} \int_0^{\pi/2} L_{col}(z, \theta, \phi) \cos \theta \, d\omega$$

$$\neq \frac{1}{\text{constant}} \int_0^{2\pi} \int_0^{\pi/2} L(z, \theta, \phi) \cos \theta \, d\omega$$

$$= E_a / \text{constant}.$$

Similarly, it can be shown that the above mentioned collector neither collects a quantity proportional to E_{oa} nor any other defined parameter.

It is assumed that 1) the ocean is horizontally stratified, 2) steady daylight conditions prevail at the sea surface, and 3) the underwater light is unpolarized. From

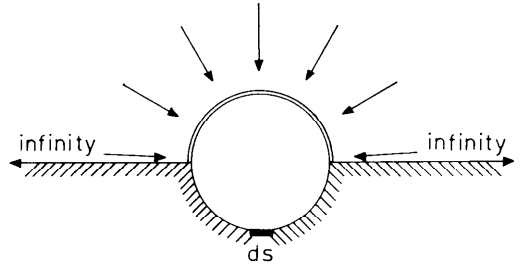


Fig. 4. Light collector appropriate for measuring a quantity proportional to the sum of the downwelling scalar irradiance and the downwelling irradiance. Notation: $\frac{1}{2}(E_{oa} + E_a)$ -collector.

these assumptions the classical equation of radiative transfer at a fixed wavelength λ_0 is derived (Table 2). Integration of Eq. 6 over all solid angles gives:

$$dE(z, \lambda_0) / dz = -a(z, \lambda_0) E_o(z, \lambda_0)$$

$$+ 4\pi \int_{300}^{\lambda_0} E_o(z, \lambda)$$

$$\cdot f(z, \pi/2, \lambda, \lambda_0) \, d\lambda. \quad (17)$$

It should be stressed that light absorption is defined as transformation of light energy forms into other energy like sensible heat, fluorescence, and chemical energy.

According to the previous definition for the vertical attenuation coefficient for the vector irradiance Eq. 17 can be written:

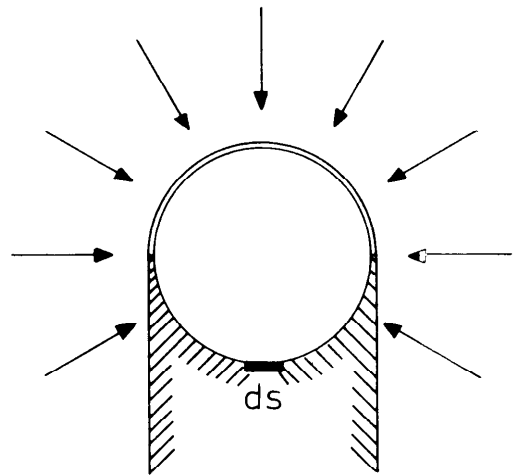


Fig. 5. Light collector appropriate for measuring a quantity proportional to the sum of the scalar irradiance and the vector irradiance. Notation: $\frac{1}{2}(E_o + E)$ -collector.

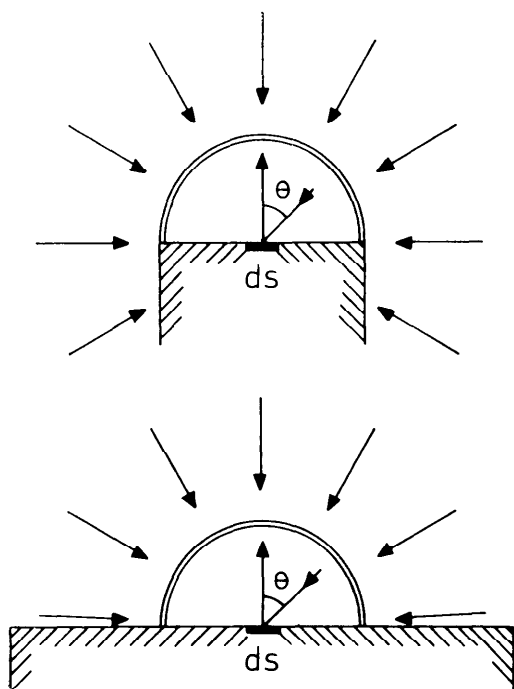


Fig. 6. Two examples of commercial light collectors that collect light in an ill defined way.

$$a(z, \lambda_0) = K_E(z, \lambda_0) \frac{E(z, \lambda_0)}{E_o(z, \lambda_0)} + \frac{4\pi}{E_o(z, \lambda_0)} \int_{300}^{\lambda_0} E_o(z, \lambda) \cdot f(z, \pi/2, \lambda, \lambda_0) d\lambda. \quad (18)$$

Since $a(z, \lambda_0)$ represents the true absorption coefficient and $K_E(z, \lambda_0) \times E(z, \lambda_0)/E_o(z, \lambda_0)$, the measured absorption coefficient, Eq. 18 can be written

$$a_{\text{true}} - a_{\text{meas}} = 4\pi/E_o(z, \lambda_0) \int_{300}^{\lambda_0} E_o(z, \lambda) \cdot f(z, \pi/2, \lambda, \lambda_0) d\lambda \geq 0. \quad (18a)$$

Accordingly, for all depths and wavelengths the true $a(z, \lambda_0)$ is always greater than the measured $a(z, \lambda_0)$ when applying the method described here.

As far as I know it has never been possible to demonstrate a significant difference between a_{true} and a_{meas} experimentally (e.g. Høgerslev 1973; 1974b) and in all optical models only a_{meas} is treated (e.g. Tyler et al. 1972). A significant difference between

a_{true} and a_{meas} is difficult to observe because: 1) Only daylight at wavelengths less than say 450 nm can produce fluorescent light of importance (Traganza 1969). This means that the upper limit in Eq. 18 can be replaced by 450 nm. 2) In clear ocean water where $E_o(z, \lambda)$ is not necessarily small for wavelengths less than 450 nm, $f(z, \pi/2, \lambda, \lambda_0)$ is extremely small for all (λ, λ_0) , minimizing the integral in Eq. 18. 3) In turbid coastal waters $E_o(z, \lambda)$ decreases rapidly versus depth for $\lambda < 450$ nm which makes $E_o(z, \lambda)$ extremely small from a few meters depth and below. 4) For the wavelength λ_0 in the UV region, integration is only performed over a small wavelength interval minimizing the integral in Eq. 18. 5) For the wavelength λ_0 situated in the red part of the spectrum, the integral attains its maximum. However, here the absorption coefficient is also great, irrespective of the concentration of suspended particles and dissolved organic substances in natural seawater, which will cause a relative error in the determination of the absorption coefficient to become small. 6) The light absorption meter measures only at wavelengths < 670 nm, i.e. the fluorescent light emission from chlorophyll when excited by ultraviolet daylight can never be observed.

Ignoring the last term in Eq. 17, Ger-shun's (1936) equation is obtained:

$$dE/dz = -aE_o, \quad (19)$$

which leads directly to

$$a = K_E E/E_o. \quad (20)$$

Equation 20 shows that a is completely determined when the E —and E_o —fields are known. Practically, the absorption coefficient is calculated from the expression

$$a(z_n) = \frac{1}{z_{n+1} - z_{n-1}} \frac{E(z_n)}{E_o(z_n)} \ln \left[\frac{E(z_{n-1})}{E(z_{n+1})} \right], \quad (21)$$

where the z_n s increase downward and are equidistantly spaced. Snell's law can be applied for the direct sun rays:

$$\sin(90 - h_s)/\sin \theta_s = n/n_o, \quad (22)$$

Table 2. Derivation of the classical equation of radiative transfer.

$$\cos \theta \frac{\partial L(z, \theta, \varphi, \lambda_0)}{\partial z} = -c(z, \lambda_0) L(z, \theta, \varphi, \lambda_0) + L_*(z, \theta, \varphi, \lambda_0) + F_*(z, \lambda, \lambda_0) \quad (6)$$

where L_* is the path function

$$L_* = \int_{4\pi} \beta(z, \theta, \varphi, \theta', \varphi', \lambda_0) L(z, \theta', \varphi', \lambda_0) \sin \theta' d\theta' d\varphi' \quad (7)$$

and F_* is the source function of fluorescent light at λ_0

$$F_* = \int_0^\infty E_0(z, \lambda) f(z, \pi/2, \lambda, \lambda_0) d\lambda. \quad (8)$$

Equation 6 can be proven from the following considerations:

The volume scattering function $\beta(z, \theta, \lambda_0)$ is defined by the relation

$$\beta(z, \theta, \lambda_0) = \frac{dI_{\text{scat}}(z, \theta, \lambda_0) d\lambda_0}{E_v(z, \lambda_0) d\lambda_0 dV} \quad (9)$$

where $dI_{\text{scat}}(z, \theta, \lambda_0) d\lambda_0$ is the intensity of the light scattered in the direction θ from the incident beam, and $E_v(z, \lambda_0) d\lambda_0$ is the irradiance at wavelength λ_0 on the volume element dV at depth z . The total scattering coefficient b is found by integration of $\beta(z, \theta, \lambda_0)$ over all solid angles:

$$b(z, \lambda_0) = \int_{4\pi} \beta(z, \theta, \lambda_0) d\omega. \quad (10)$$

In case of isotropic scattering, equations 10 and 7 give:

$$\beta(z, \theta, \lambda_0) = \frac{1}{4\pi} b(z, \lambda_0) = \beta(z, \pi/2, \lambda_0) \quad (11)$$

and

$$L_*(z, \theta, \varphi, \lambda_0) = E_0(z, \lambda_0) \beta(z, \pi/2, \lambda_0). \quad (12)$$

Since daylight acts like a broad spectral light source, emission of fluorescent light energy takes place in the sea when excited by the shortwave part of the daylight.

Table 2. Continued.

A volume scattering function of fluorescent light

$f(z, \theta, \lambda, \lambda_0)$ is introduced by the relation:

$$f(z, \theta, \lambda, \lambda_0) d\lambda_0 = \frac{d^2 I_{\text{fluor}}(z, \theta, \lambda, \lambda_0) d\lambda_0}{E_V(z, \lambda) d\lambda \cdot dV} \quad (13)$$

where $d^2 I_{\text{fluor}}(z, \theta, \lambda, \lambda_0) d\lambda_0$ is the intensity of fluorescent light at wavelength λ_0 excited by daylight at wavelength λ and emitted in the direction θ from the incident beam. $E_V(z, \lambda) d\lambda$ is the daylight energy at wavelength λ impinging upon the volume element dV at the depth z . The total scattered intensity of fluorescent light at wavelength λ_0 in the direction θ from the incident beam is obtained from integration of equation 13 over all wavelengths

$$dI_{\text{fluor}}(z, \theta, \lambda_0) = \int_0^\infty E_V(z, \lambda) f(z, \theta, \lambda, \lambda_0) d\lambda dV. \quad (14)$$

Obviously, fluorescent light energy is scattered isotropically which means that

$$f(z, \theta, \lambda, \lambda_0) = f(z, \pi/2, \lambda, \lambda_0). \quad (15)$$

The source function of fluorescent light F_* can now be derived analogous to the derivation of the pathfunction I_* in the case of isotropic scattering - see equation 9 and 12:

$$F_* = \int_0^\infty E_0(z, \lambda) f(z, \pi/2, \lambda, \lambda_0) d\lambda. \quad (16)$$

The scalar irradiance $E_0(z, \lambda)$ can be measured by means of the light absorption meter and the volume scattering function of fluorescent light $f(z, \pi/2, \lambda, \lambda_0)$ by aid of a conventional fluorimeter (Parker 1968) measuring through 90° the excitation spectrum at the emission wavelength λ_0 of a water sample taken at depth z . In practice it suffices to integrate the expression in equation 16 between 300 nm (lower limit of incoming daylight energy) and λ_0 .

where h_s is the solar elevation in air, θ_s the solar angle in water measured from zenith, and n and n_o are the refraction indices of seawater and air.

Inserting the definitions for E and E_o into 20 yields

$$a(z) = K_E(z) \frac{\int_{4\pi} L \cos \theta \, d\omega}{\int_{4\pi} L \, d\omega} \equiv K_E(z) \langle \cos \theta \rangle \quad (23)$$

valid at all depths. The average function $\langle \cos \theta \rangle$ is approximately equal to $\cos \theta_s$ because of the strong forward scattering and provided the depth is small and the solar elevation is moderate to high.

Differentiating Eq. 19 once and rearranging gives

$$d \ln a/dz = (d \ln K_E/dz) + K_o - K_E. \quad (24)$$

Equation 24 gives the relative variation of a versus depth. The parameters K_E and K_o are determined from relative daylight measurements. Since 23 approximately gives the surface value of a , the vertical distribution of a can be obtained. Unfortunately, this method is to some extent impractical.

From the definitions and Eq. 19,

$$K_1 = K_E \frac{K_o + a}{K_E + a} \quad (25)$$

and

$$K_2 = K_E \frac{K_o - a}{K_E - a}. \quad (26)$$

Since $K_E > a$ when $b > 0$, Eq. 26 states that $K_2 = 0$ implies $K_o = a$. From 25 and 26 it can be seen that $K_1 \begin{matrix} > \\ < \end{matrix} K_2$ implies $K_E \begin{matrix} > \\ < \end{matrix} K_o$, respectively. (27)

In the upper few meters of the ocean the common situation is that the water is optically homogeneous due to the strong mixing, the ratio $E : E_o$ decreases with depth (or K_E increases), and $K_u > K_d$ in the case of visible daylight (Jerlov 1968; Høgerslev 1973, 1974b). This leads to the inequalities

$$K_u > K_d > K_E > K_1 > K_o > K_2, \quad (28)$$

when applying the definitions, Eq. 24, 25, and 26. At great depths all the K s become

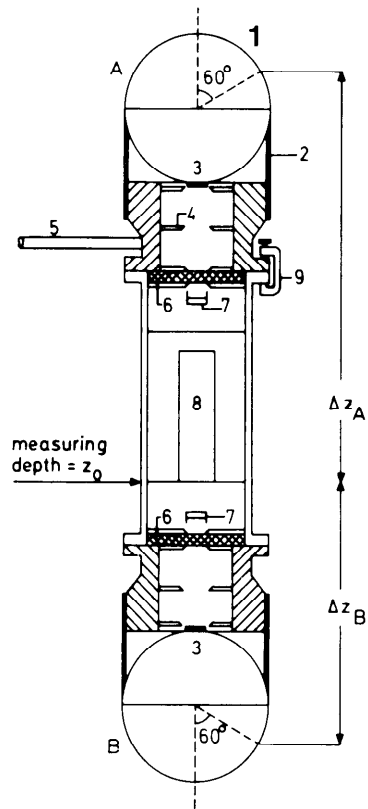


Fig. 7. The spectral light absorption meter consisting of one $\frac{1}{2}(E_o + E)$ -collector (A) and one $\frac{1}{2}(E_o - E)$ -collector (B), for an upward orientation of the a -meter as depicted in the figure. The color filters applied are double interference filters (Schott und Genossen) having half-value widths between 7.5 and 17 nm. The light sensitive detectors are PIN diodes (UDT 500/UV; MWO 2948 United Detector Tech. Inc.). Determination of depths is obtained by a pressure gauge (Bell and Howell Ltd. Type 4-393-L102-01MO) having a measuring range from the surface down to about 100-m depth. The numbers refer to: 1—opal-flashed globe; 2—black shield; 3—receiving flat opal-flashed collector; 4—diaphragms; 5—pin for suspension of the meter; 6—color filter; 7—PIN diode; 8—pressure gauge; 9—clamp.

the same, assuming optically homogeneous water (Preisendorfer 1959; Zaneveld and Pak 1972; Høgerslev 1972).

The light absorption meter is shown in Fig. 7. It measures the two quantities $\frac{1}{2}(E_o + E)_A$ and $\frac{1}{2}(E_o - E)_B$ simultaneously. The index letters A and B refer to the two measuring units, consisting of collector, dia-

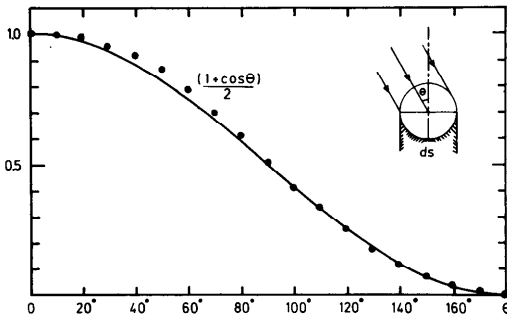


Fig. 8. The collecting properties of the light collectors mounted on the spectral light absorption meter when in water and illuminated by a parallel beam of light. Both collectors had the same collecting properties; furthermore no change in collecting properties with wavelength in the range 370–670 nm could be observed.

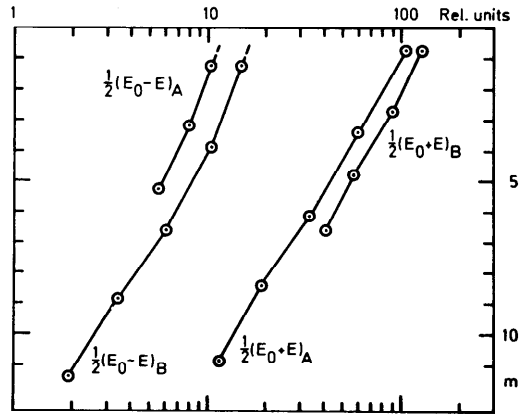


Fig. 9. Daylight measurements at 532 nm in the Danish Sound, north of Ven (station 14). Solar elevation: 33°.

phragm, interference filter, PIN diode, and amplification system on deck. Explicitly, the recorded signals can be expressed as

$$\frac{1}{2}[E_o(z_o - \Delta z_A) + E(z_o - \Delta z_A)]_A$$

and

$$\frac{1}{2}[E_o(z_o + \Delta z_B) - E(z_o + \Delta z_B)]_B,$$

where Δz_A and Δz_B are known constants (see Fig. 7).

After one measuring series at wavelength λ_0 the meter is taken up, inverted, and lowered again. This means that the recorded signals now become $\frac{1}{2}(E_o + E)_B$ and $\frac{1}{2}(E_o - E)_A$. Two such lowerings thus give two determinations of the absorption coefficient in absolute values. This measuring technique has the advantages that no calibration of the absorption meter is necessary, calibration factors between the measuring units A and B are readily obtained, and changes in the optical properties of the seawater, the surface irradiance, or both within the measuring period are observed directly by comparing the vertical distribution of $\frac{1}{2}(E_o + E)_A$ and $\frac{1}{2}(E_o + E)_B$.

The two measuring units A and B can also be calibrated in the laboratory so that the same light flux impinging on the collectors for the same angle of incidence results in the same signal reading. This calibration is preferably performed in water to avoid spectral determinations of immersion coefficients.

Such a calibration makes one lowering sufficient for an absolute determination of a (Højerslev 1973) because K_E is obtained from a relative measurement and the expression $E : E_o$ is a ratio (see Eq. 20).

Finally, it should be emphasized that the present a -meter consists of two collectors instead of the three used in other absorption meters of the daylight type (Højerslev 1973; Tyler 1961).

The collecting properties of the collector when in water are shown in Fig. 8. The ideal collector would collect the light flux corresponding to $\frac{1}{2}(1 + \cos \theta)$.

In September 1974 a trial of the light absorption meter was made in the Danish Sound north of Copenhagen, near Ven. It was sunny weather with only a few clouds

Table 3. Daylight measurements in the Danish Sound north of Ven, of the scalar and vector irradiance at 532 nm. The surface light conditions were ideal and the solar elevation about 33°. The absorption coefficients versus depth calculated from these irradiance measurements are presented on the right side of the table.

Depth (m)	$(E_o)_A$	$(E)_A$	$(E_o)_B$	$(E)_B$	$(a)_A$	$(a)_B$
	relative units		relative units		m^{-1}	m^{-1}
1	109	89.2	136	105	0.180	0.141
2	90.4	71.6	114	87.4	0.172	0.138
3	74.3	57.8	96.8	73.2	0.171	0.156
4	59.9	46.1	78.2	57.8	0.164	0.167
5	49.2	37.8	63.4	46.6	0.153	0.158

Table 4. Calculated vertical attenuation coefficients and diffusivity ($E : E_o$) of the underwater light field as a function of depth. These parameters were all obtained from measurements with light collector A.

Depth (m)	$(K_o)_A$	$(K_E)_A$	$(K_1)_A$	$(K_2)_A$	$\left(\frac{E}{E_o}\right)_A$	$\left(\frac{a}{K_E \cos \theta_s}\right)_A$
	m^{-1}					
1	0.187	0.220	0.201	0.130	0.818	1.055
2	0.196	0.214	0.204	0.147	0.792	1.022
3	0.215	0.226	0.220	0.173	0.778	1.004
4	0.197	0.199	0.198	0.174	0.770	0.993
5					0.768	0.991

Table 5. Calculated vertical attenuation coefficients and diffusivity ($E : E_o$) of the underwater light field as a function of depth. These parameters were all obtained from measurements with light collector B.

Depth (m)	$(K_o)_B$	$(K_E)_B$	$(K_1)_B$	$(K_2)_B$	$\left(\frac{E}{E_o}\right)_B$	$\left(\frac{a}{K_E \cos \theta_s}\right)_B$
	m^{-1}					
1	0.176	0.183	0.180	0.119	0.772	0.996
2	0.163	0.177	0.169	0.126	0.767	0.989
3	0.213	0.236	0.223	0.154	0.756	0.976
4	0.210	0.215	0.212	0.194	0.739	0.953
5					0.735	0.948

so that almost stable light conditions prevailed. The sea surface was rather calm but with strong currents in the area. The water belongs to coastal type 3 according to Jerlov's optical classification of water masses (Jerlov 1968). Accordingly, the maximum transmission of daylight lies in the green part of the spectrum and for the same reason the measuring wavelength of 532 nm was selected.

The measurements are presented in Fig. 9. $\frac{1}{2}(E_o + E)_A$ and $\frac{1}{2}(E_o - E)_B$ are measured simultaneously as are $\frac{1}{2}(E_o - E)_A$ and $\frac{1}{2}(E_o + E)_B$. All four parameters display small fluctuations in signal levels when compared to usual downwelling irradiance

measurements. Figure 9 allows calculation of the four parameters $E_o(z)_A$, $E(z)_A$, $E_o(z)_B$, and $E(z)_B$ (Table 3). This is sufficient for calculation of the absorption coefficient and the vertical attenuation coefficients of scalar and vector irradiance (Tables 3-5, Fig. 10). Figure 9 demonstrates that the measuring unit B gives a signal 30% higher than A determined from $\frac{1}{2}(E_o + E)_A$ and $\frac{1}{2}(E_o + E)_B$. The calibration factor between A and B at 532 nm should be derived from these two parameters since shadows from the ship play an insignificant role as long as the direct rays of the sun are not screened out.

The optical properties of the seawater

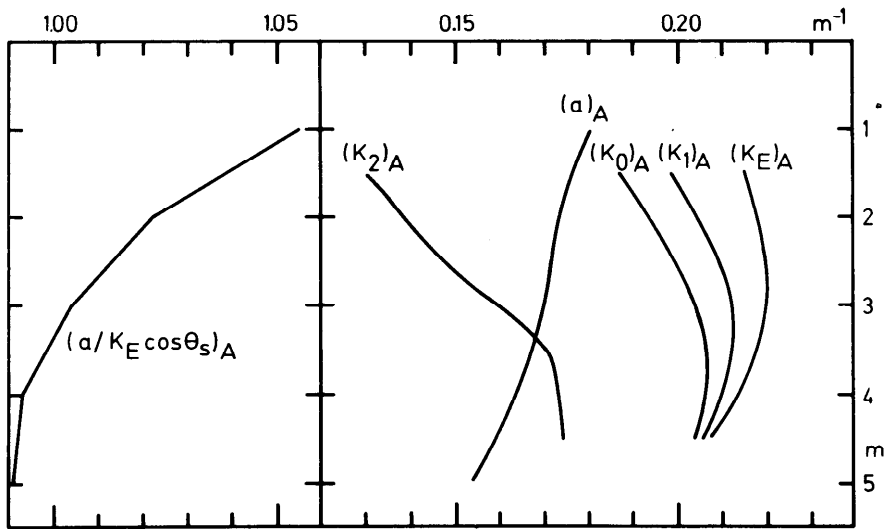


Fig. 10. Selected inherent and apparent optical properties versus depth at station 14.

varied significantly during 5 min due to the strong currents. Similar variation in the light scattering coefficient was observed. The variation in the light absorption coefficient was smaller below 3.5 m.

An approximate determination of the absorption coefficient according to Eq. 23 gives acceptable results. Finally, the ratio $E : E_0 = a : K_E$ decreases with depth, which also should be expected (Jerlov 1968; Jerlov and Nygård 1969; Højerslev 1972, 1973, 1974b).

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References

- BAUER, D., J. C. BRUN-COTTAN, AND A. SALLIOT. 1971. Principe d'une mesure directe dans l'eau de mer du coefficient d'absorption de la lumière. *Cah. Oceanogr.* **23**: 841-858.
- GERSHUN, A. 1936. Fundamental ideas of the theory of a light field (vector methods of photometric calculations) [in Russian]. *Izv. Akad. Nauk SSSR, Otd. Mat. Estestv. Nauk*, p. 417-430.
- GILBERT, G. D., R. C. HONEY, R. E. MYERS, AND G. P. SORENSEN. 1969. Optical absorption meter. *Stanford Res. Inst. Contrib.* N60530-68-C-1423. 60 p.
- HØJERSLEV, N. K. 1972. A theoretical proof of the existence of a constant vertical radiance attenuation coefficient in a horizontally stratified ocean. *Univ. Copenhagen, Inst. Phys. Oceanogr. Rep.* 20. 8 p.
- . 1973. Inherent and apparent optical properties of the western Mediterranean and the Hardangerfjord. *Univ. Copenhagen, Inst. Phys. Oceanogr. Rep.* 21. 23 p.
- . 1974a. Daylight measurements for photosynthetic studies in the western Mediterranean. *Univ. Copenhagen, Inst. Phys. Oceanogr. Rep.* 26. 10 p.
- . 1974b. Inherent and apparent optical properties of the Baltic. *Univ. Copenhagen, Inst. Phys. Oceanogr. Rep.* 23. 16 p.
- INTERNATIONAL ASSOCIATION FOR THE PHYSICAL SCIENCES OF THE OCEAN. 1964. Standard terminology on the optics of the sea.
- JERLOV, N. G. 1961. Optical measurements in the eastern North Atlantic. *Medd. Oceanogr. Inst. Goetb.* **30**: 1-40.
- . 1968. Optical oceanography. Elsevier.
- , AND K. NYGÅRD. 1969. Influence of solar elevation on attenuation of underwater irradiance. *Univ. Copenhagen, Inst. Phys. Oceanogr. Rep.* 4. 6 p.
- PARKER, A. C. 1968. Photoluminescence of solutions. Elsevier.
- PREISENDORFER, R. W. 1959. Theoretical proof of the existence of characteristic diffuse light in natural waters. *J. Mar. Res.* **18**: 1-9.
- SMITH, R. C. 1969. An underwater spectral irradiance collector. *J. Mar. Res.* **27**: 341-351.
- , AND W. H. WILSON, JR. 1972. Photon scalar irradiance. *Appl. Opt.* **11**: 934-938.
- TRAGANZA, E. D. 1969. Fluorescence excitation and emission spectra of dissolved organic matter in sea water. *Bull. Mar. Sci.* **19**(4).
- TYLER, J. E. 1961. An instrument for the measurement of the volume absorption coefficient of horizontally stratified water. *U.S. Gov. Res. Rep.*, PB-154030, Contract bs-72039, Task Rep. 5-4.
- , R. C. SMITH, AND W. H. WILSON, JR. 1972. Predicted optical properties for clear natural water. *J. Opt. Soc. Am.* **62**: 83-91.
- ZANEVELD, J. R. V., AND H. PAK. 1972. Some aspects of the axially symmetric submarine daylight field. *J. Geophys. Res.* **77**: 2677-2680.

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A freely falling probe for small-scale temperature gradients¹

Abstract—A light, simple, and fairly inexpensive hand-launched freely falling probe for measuring temperature microstructure was built. The circuitry operates under pressure in a Freon environment; low-power inte-

grated circuits are used as components. The probe is assembled from plastic pipe and fittings and recovered with monofilament fishing line. Standard XBTs are adapted for use as the data link. Noise levels of 2×10^{-3} °C cm^{-1} were achieved in signals recorded during casts in the ocean and on inland waters. Portions of the water column can be rapidly resampled. The device can be operated from any size vessel.

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