Ocean Optics Summer Class

Calibration and Validation for
Ocean Color Remote Sensing

Apparent Optical Properties
and
Introduction to Remote Sensing

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Delivered at the Darling Marine Center
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SEQUOIA

HYDROLIGHT

Apparent Optical Properties (AOPs)

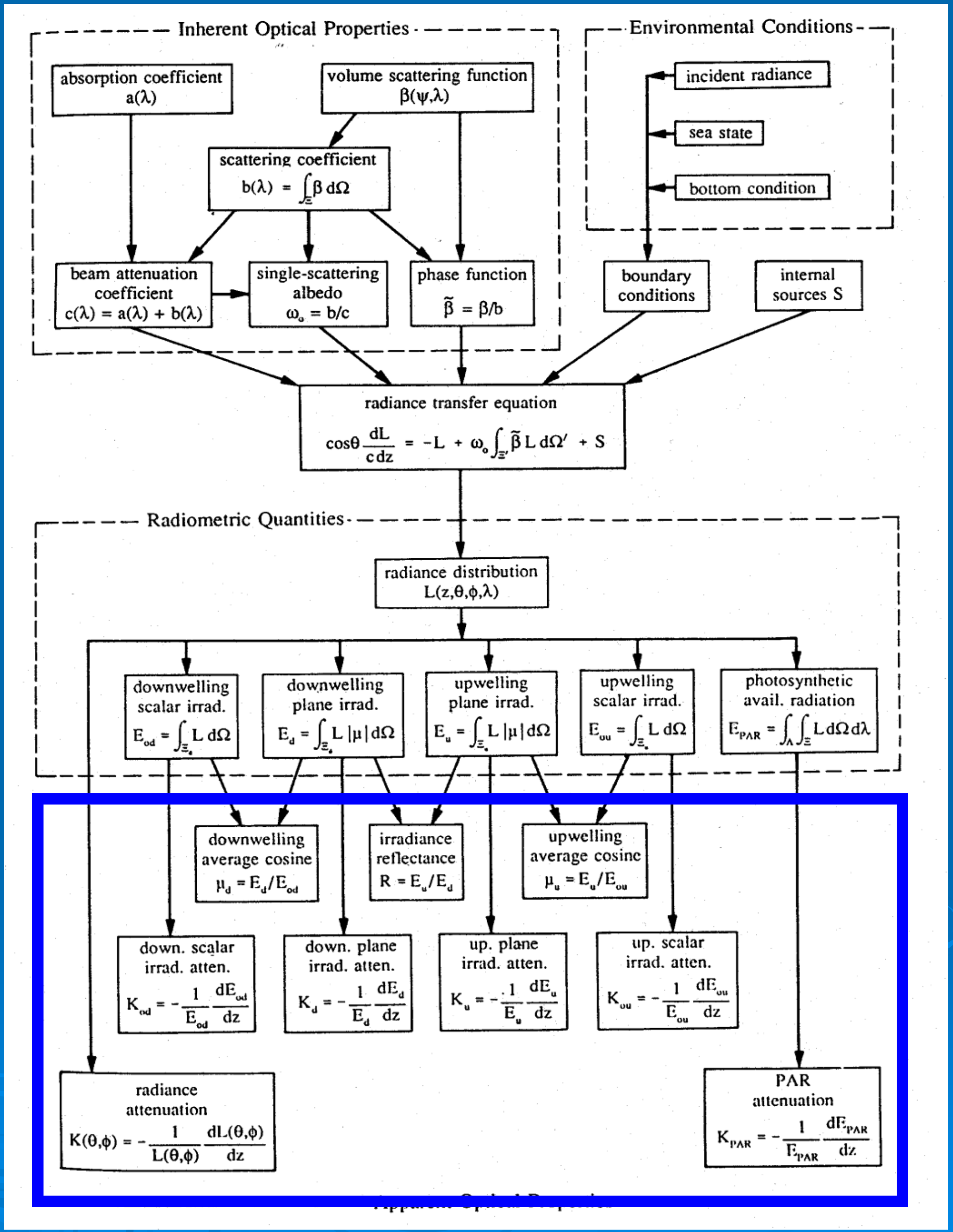
AOPs are quantities that

• depend on the IOPs and on the radiance distribution, and

• they display enough stability to be useful for approximately describing the optical properties of the water body

AOPs can NOT be measured in the lab or on water sample; they must be measured in situ

Radiance and irradiances are NOT AOPs—they don't have stability



Apparent Optical Properties

A good AOP depends weakly on the external environment (sky condition, surface waves) and strongly on the water IOPs

AOPs are ratios or derivatives of radiometric variables

Historically, IOPs were hard to measure (but easy to interpret). This is less true today because of advances in instrumentation.

AOPs were easier to measure (but are often hard to interpret).

In a Perfect World

Light Properties: measure the radiance as a function of location, time, direction, wavelength, $L(x,y,z,t,\theta,\phi,\lambda)$, and you know everything there is to know about the light field. You don't need to measure irradiances, PAR, etc.

Material Properties: measure the absorption coefficient $a(x,y,z,t,\lambda)$ and the volume scattering function $\beta(x,y,z,t,\psi,\lambda)$, and you know everything there is to know about how the material affects light. You don't need to measure b, b_b, etc.

Nothing else (AOPs in particular) is needed.

Reality

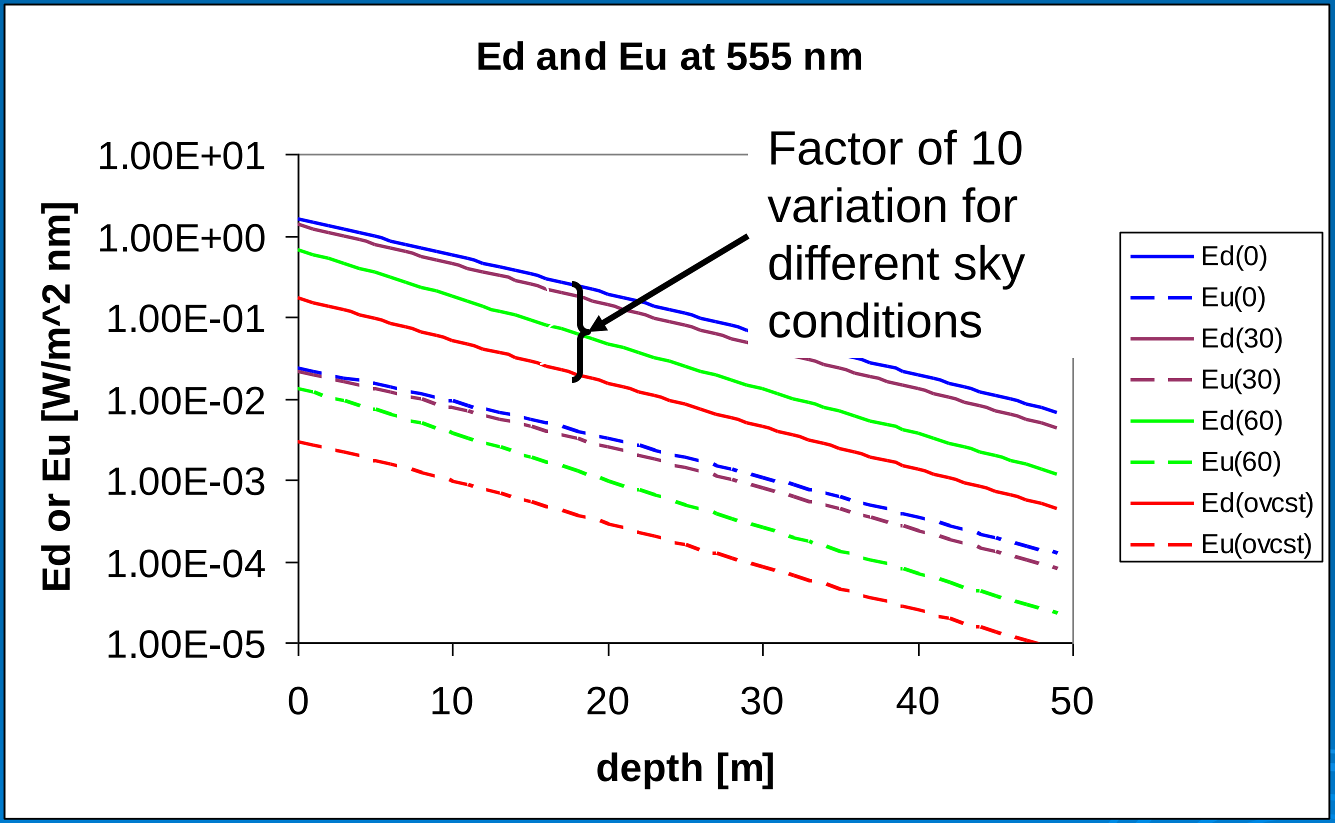
$L(x,y,z,t,\theta,\phi,\lambda)$ is too difficult and time consuming to measure on a routine basis, and you don't need all of the information contained in *L*, so therefore measure irradiances, PAR, etc. (ditto for VSF….)

Idea

Can we find simpler measures of the light field than the radiance, which are also useful for describing the optical characteristics of a water body (i.e., what is in the water)?

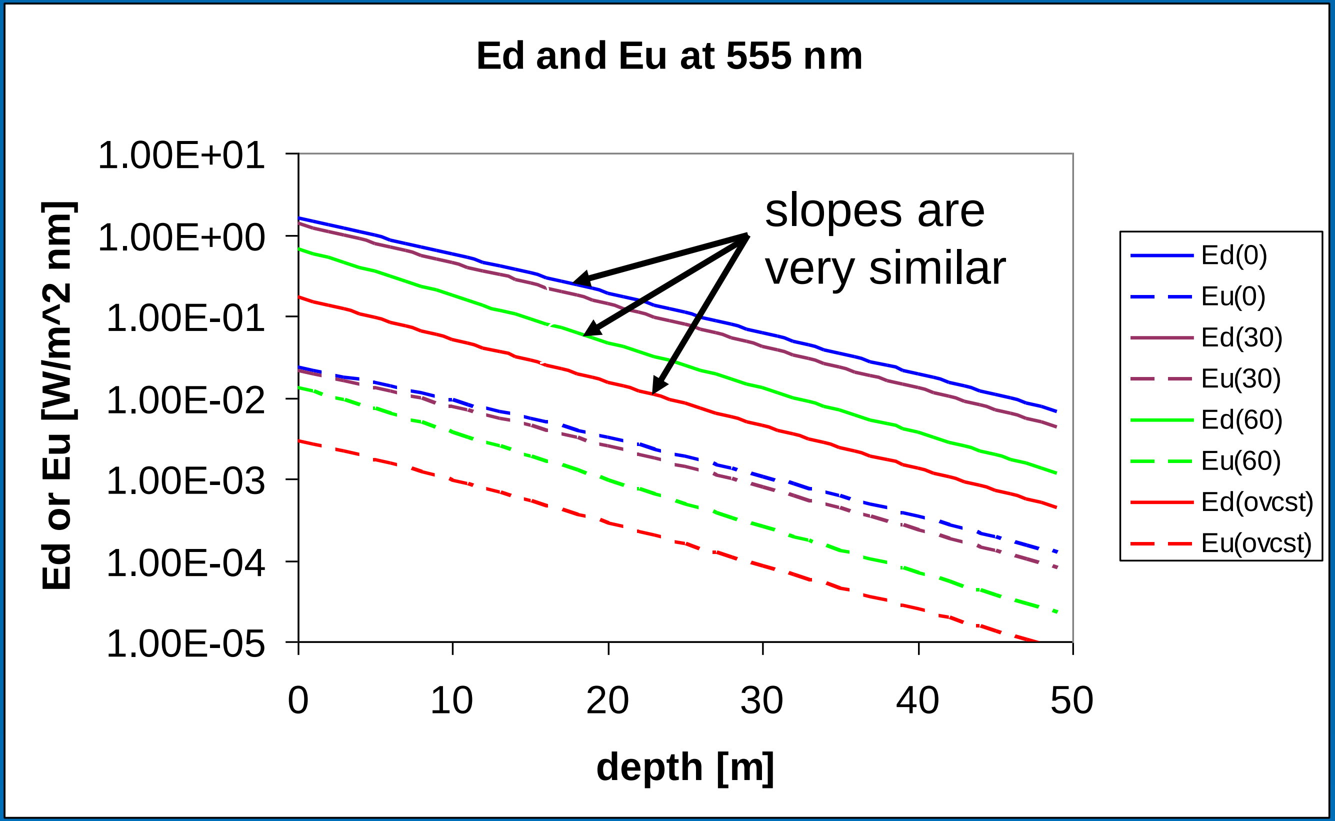
E_d and E_u

HydroLight runs: Chl = 1.0 mg Chl/m^3, etc
Sun at 0, 30, 60 deg in clear sky, and solid overcast



Note: E_d and E_u depend on the radiance and on the abs and scat properties of the water, but they also depend strongly on incident lighting, so not useful for characterizing a water body. Again: irradiances are NOT AOPs!

E_d and E_u



Magnitude changes are due to sun angle; slope is determined by water IOPs.

This suggests trying...

The depth derivative (slope) on a log-linear plot as an AOP. This leads to the diffuse attenuation coefficient for downwelling plane irradiance:

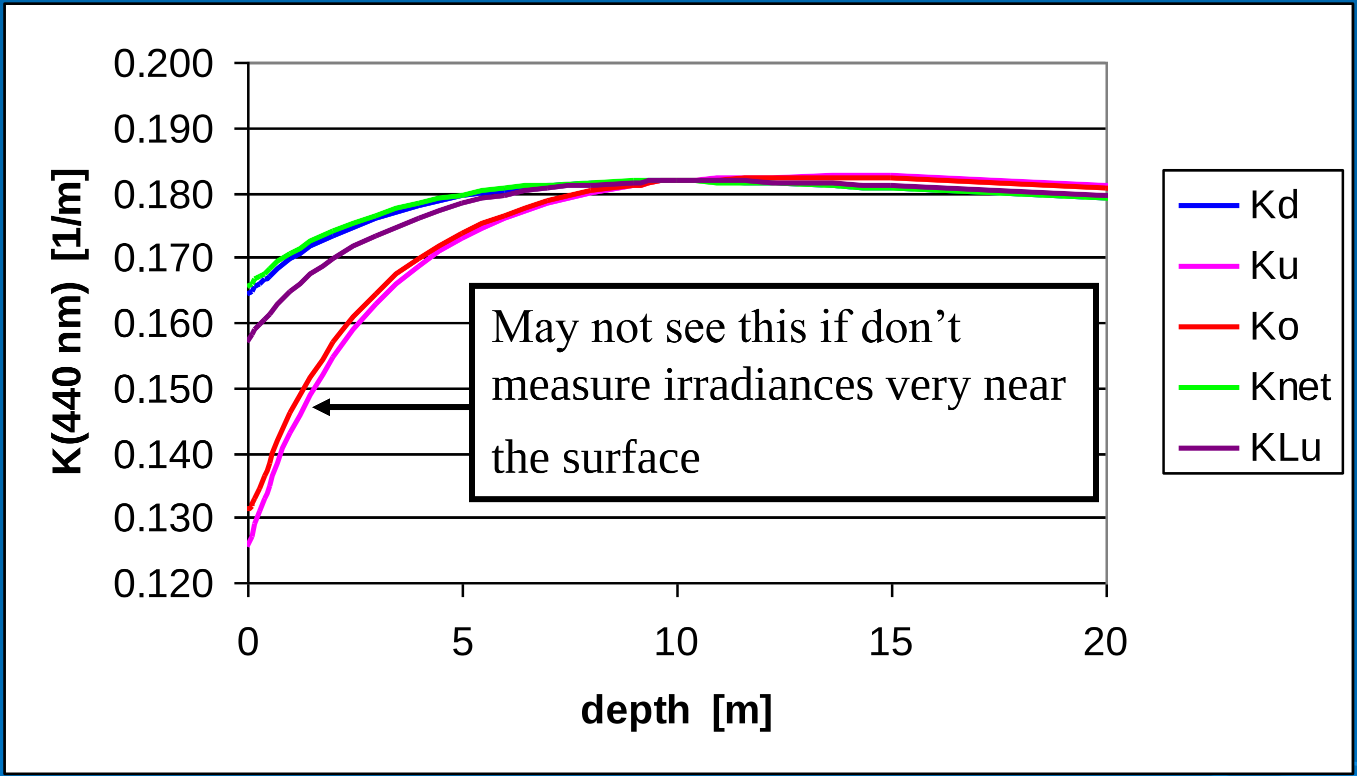
$$K_d(z,\lambda) = -\frac{d \ln E_d(z,\lambda)}{dz}$$

$$= -\frac{1}{E_d(z,\lambda)} \frac{d E_d(z,\lambda)}{dz}$$

We can do the same for E_u, E_o, $L(\theta,\phi)$, etc, and define lots of different K functions: K_u, K_o, $K_L(\theta,\phi)$, etc.

How similar are the different K's?

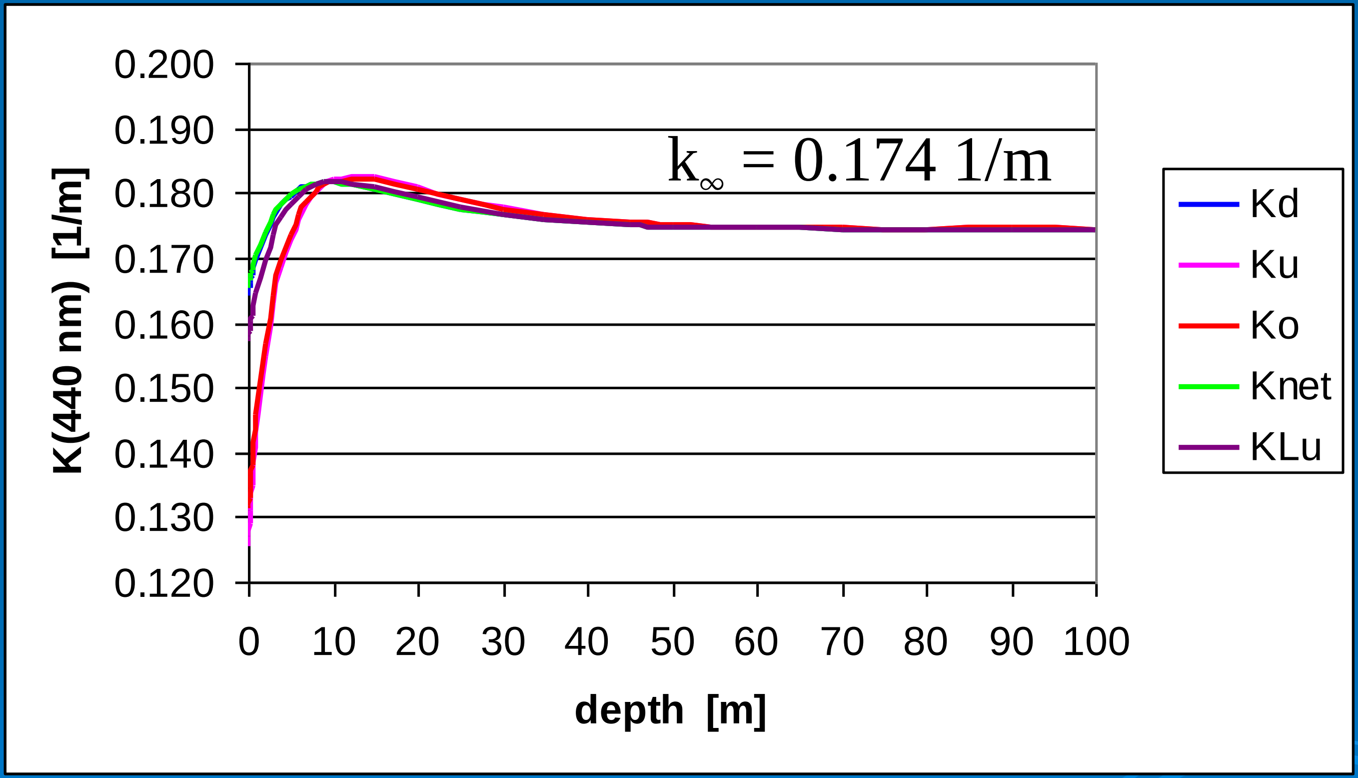
HydroLight: homogeneous water with Chl = 2 mg/m^3, sun at 45 deg, 440 nm, etc.



NOTE: The K's depend on depth, even though the water is homogeneous, and they are most different near the surface (where the light field is changing because of boundary effects)

How similar are the different K's?

HydroLight: homogeneous water with Chl = 2 mg/m^3, sun at 45 deg, 440 nm, etc.

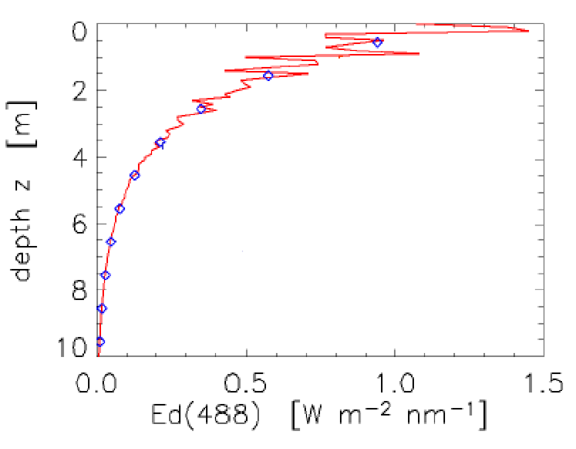


NOTE: the K's all approach the same value as you go deeper: the asymptotic diffuse attenuation coefficient, k_∞, which is an IOP.

Something to Think About

- Suppose you measure $E_d(z)$ [or $L_u(z)$]
- but the data are very noisy in the first few meters because of wave focusing, or bubbles, or…
- so you discard the data from the upper 5 meters
- You then compute K_d from 5 m downward, and get a fairly constant K_d value below 5 m [or K_{Lu} if measuring L_u]
- You then use $E_d(z) = E_d(0)\exp(-K_d\, z)$ and the computed K_d from 5 m downward to extrapolate E_d(5 m) back to the surface

How accurate is this $E_d(0)$ [or $L_u(0)$] likely to be?



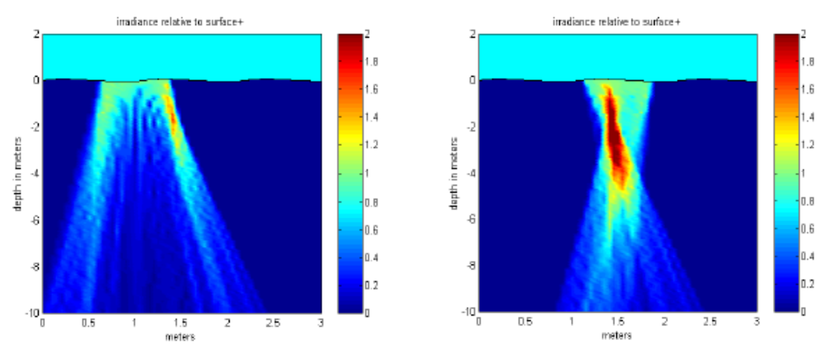
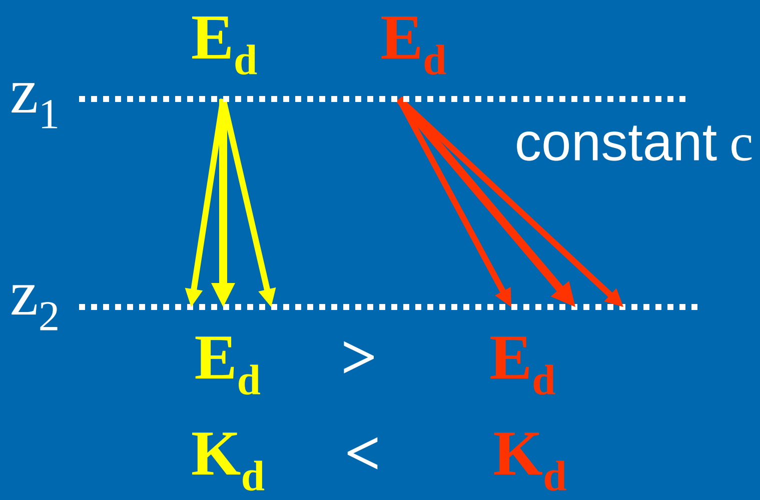
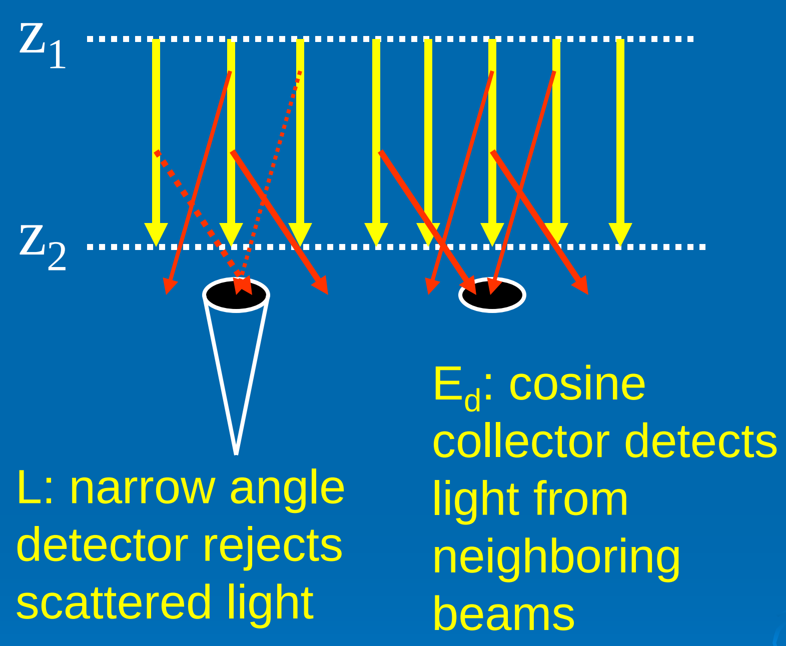


Fig. 2. An example of calculated irradiance relative to the irradiance just above the sea surface. The absorption coefficient is 0.07 m^{-1} and the scattering coefficient is 0.39 m^{-1}; the wave surface is wavelength L=[1.1 0.55 0.05] m and amplitude A= [0.05 0.01 0.0008]m. The beam is 0.8 m wide and enters vertically. Fig.3a shows the beam entering near the trough of a wave, generating a divergent beam. On Fig.3b the light enters near the crest of a wave, generating a convergent beam with a focal area.

wave focusing, from Zaneveld et al, 2001, Optics Express

Beam attenuation *c* ≠ diffuse attenuation *K*





$c > K_d$

Virtues and Vices of K's

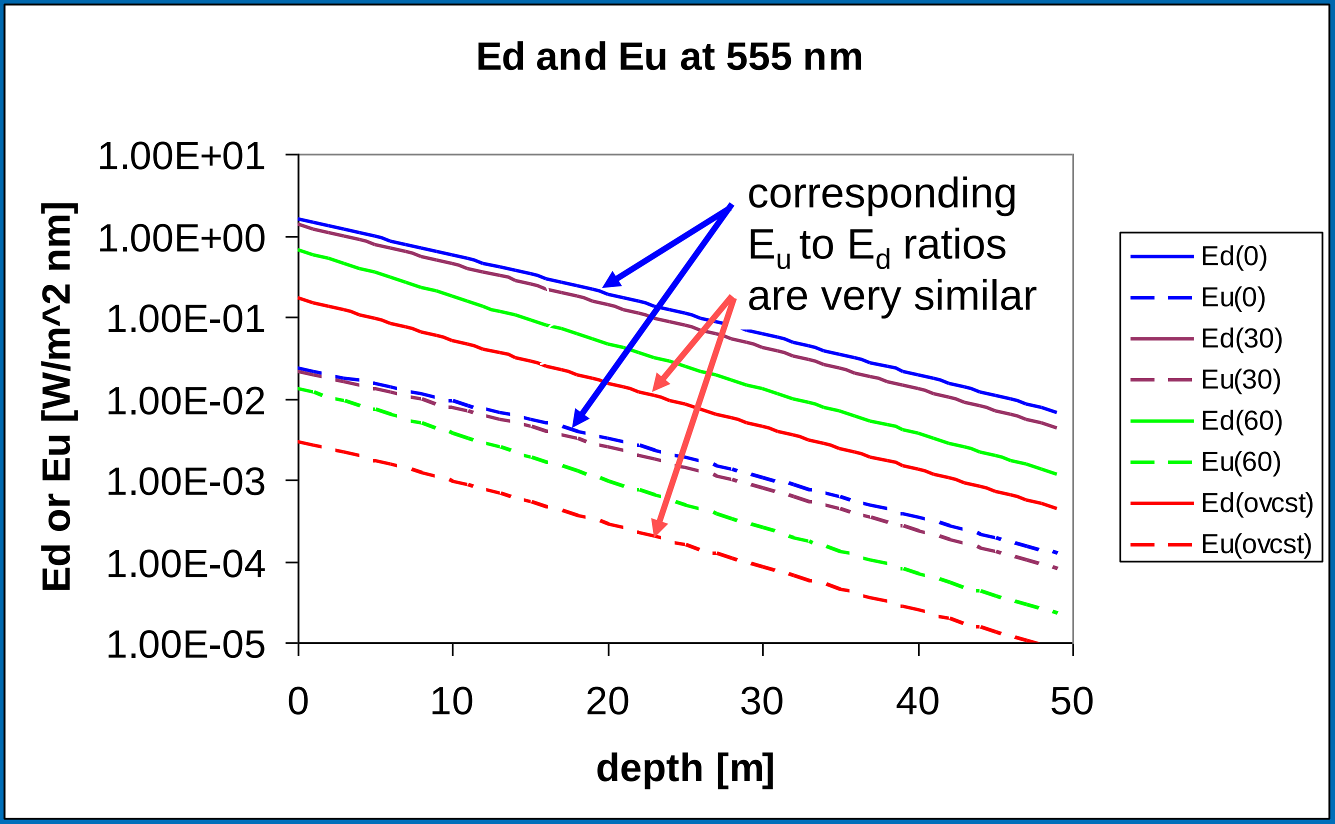
Virtues:

- K's are defined as rates of change with depth, so don't need absolutely calibrated instruments
- K_d is very strongly influenced by absorption, so correlates with chlorophyll concentration (in Case 1 water)
- about 90% of water-leaving radiance comes from a depth of $1/K_d$ (called the penetration depth by Gordon)
- radiative transfer theory provides connections between K's and IOPs and other AOPs (recall Gershun's equation: $a = K_{net}\,\mu$)

Vices:

- not constant with depth, even in homogeneous water
- greatest variation is near the surface
- difficult to compute derivatives with noisy data

E_d and E_u

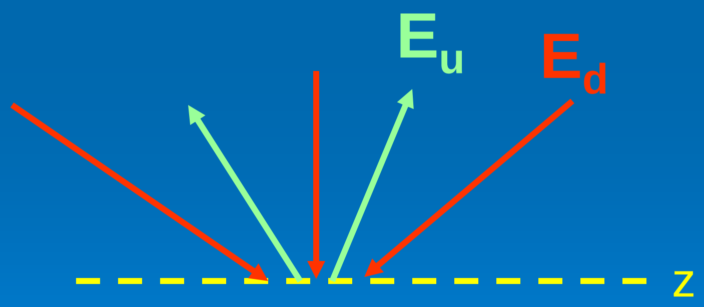


Magnitude changes are due to sun angle; E_u to E_d ratio is determined by water IOPs.

This suggests trying...

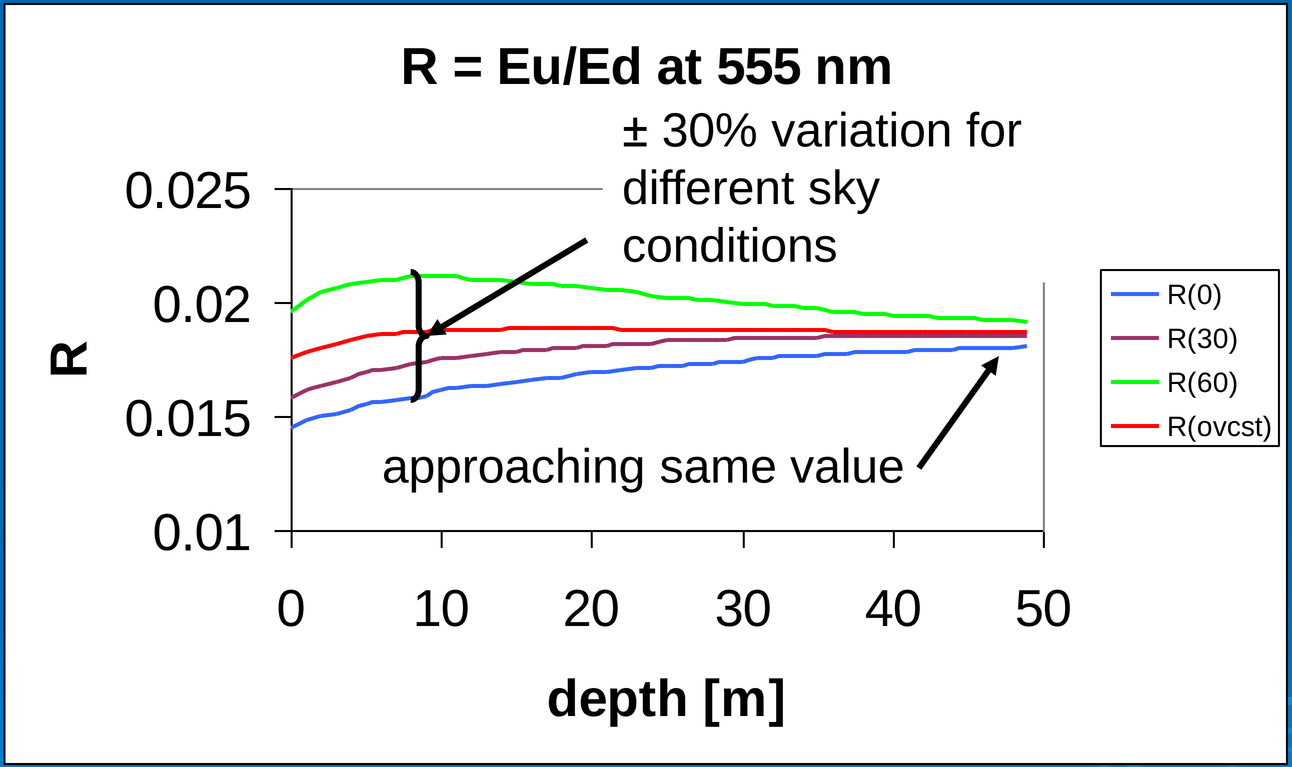
The ratio of upwelling plane irradiance E_u to downwelling plane irradiance E_d. This is the irradiance reflectance R:

$$R(z,\lambda) \equiv \frac{E_u(z,\lambda)}{E_d(z,\lambda)}$$



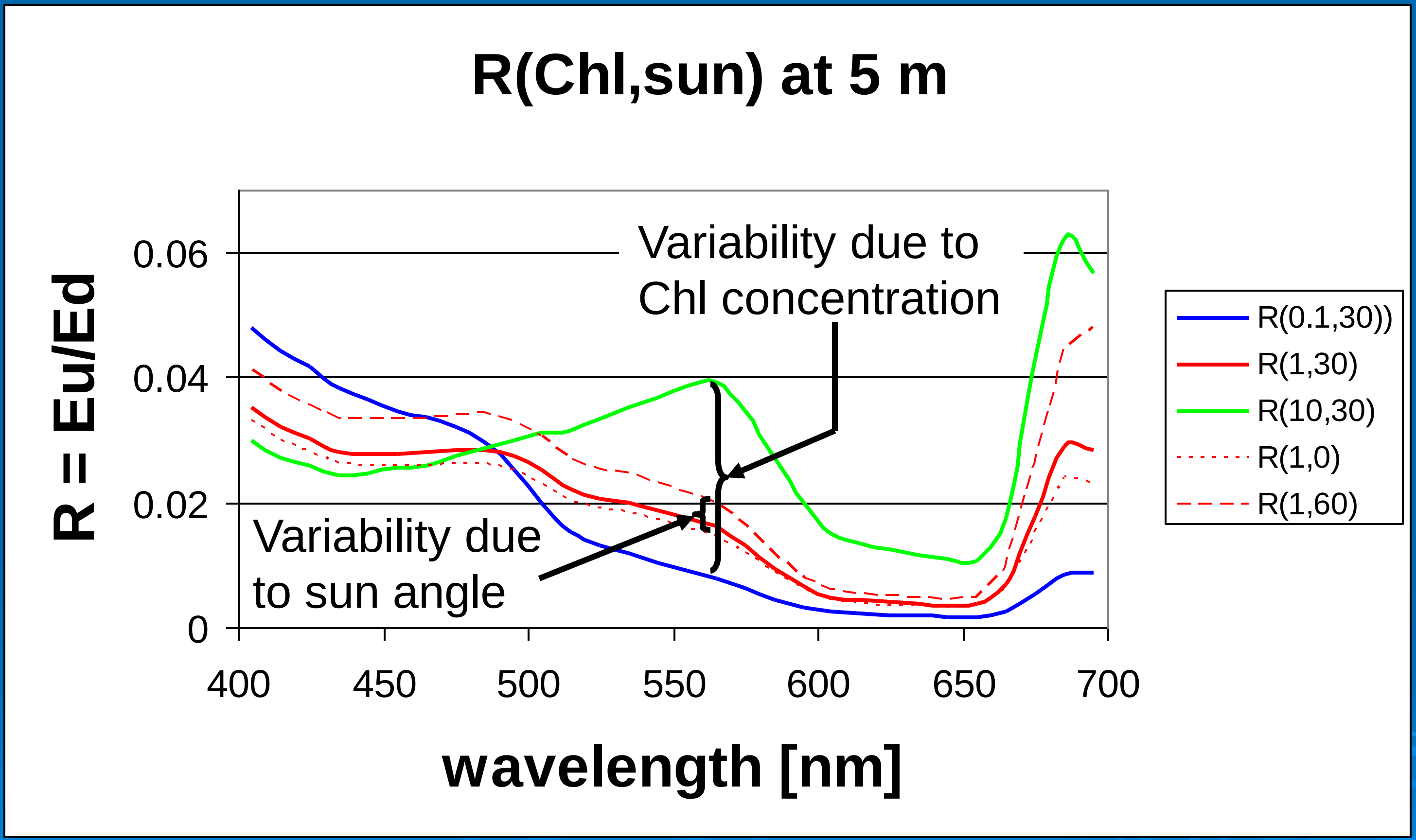
$R = E_u/E_d$

HydroLight runs: Chl = 1.0 mg Chl/m^3, etc
Sun at 0, 30, 60 deg in clear sky, and solid overcast



$R = E_u/E_d$

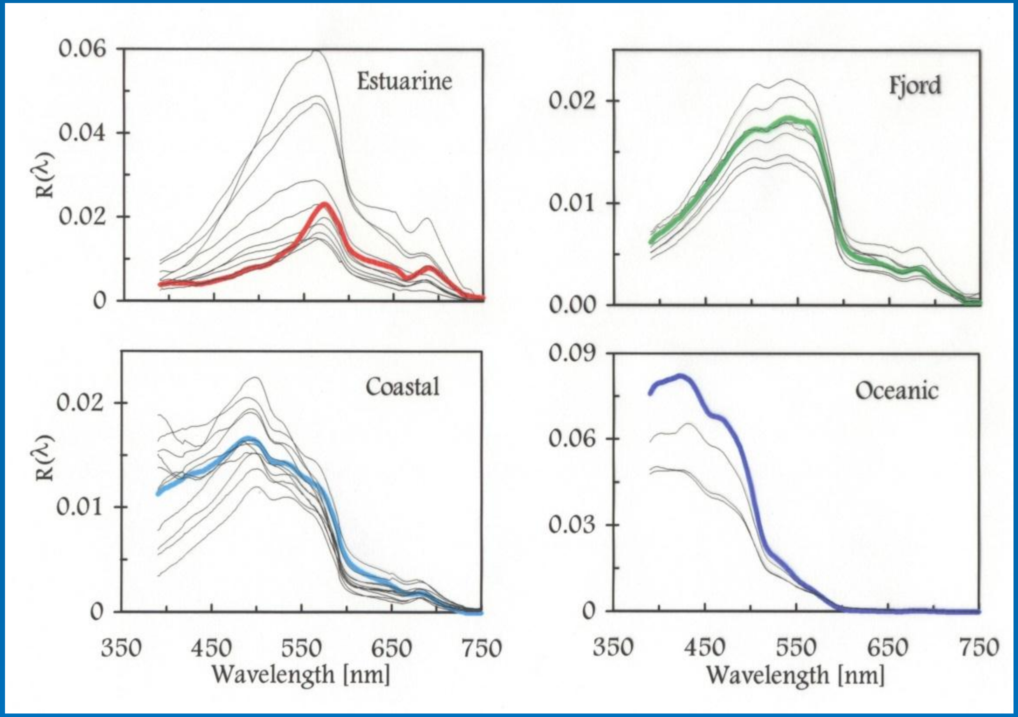
HydroLight runs: Chl = 0.1,1, 10 mg Chl/m^3
Sun at 0, 30, 60 deg in clear sky



R depends weakly on the external environment and strongly on the water IOPs

Examples of $R = E_u/E_d$

measurements from various ocean waters

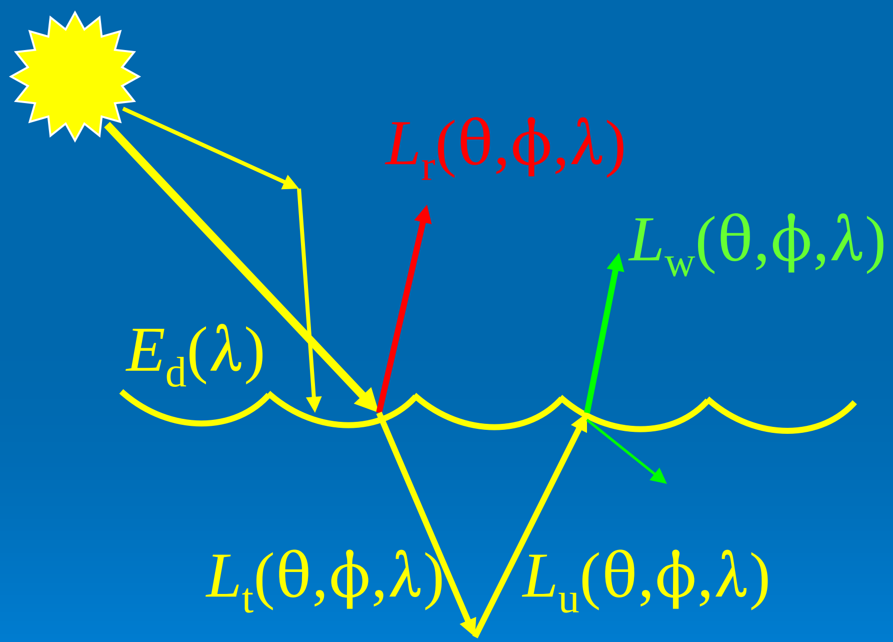


Roesler and Perry 1995

Water-leaving Radiance, L_w

total upwelling radiance in air (above the surface) =
water-leaving radiance + surface-reflected radiance

$$L_u(\theta,\phi,\lambda) = L_w(\theta,\phi,\lambda) + L_r(\theta,\phi,\lambda)$$



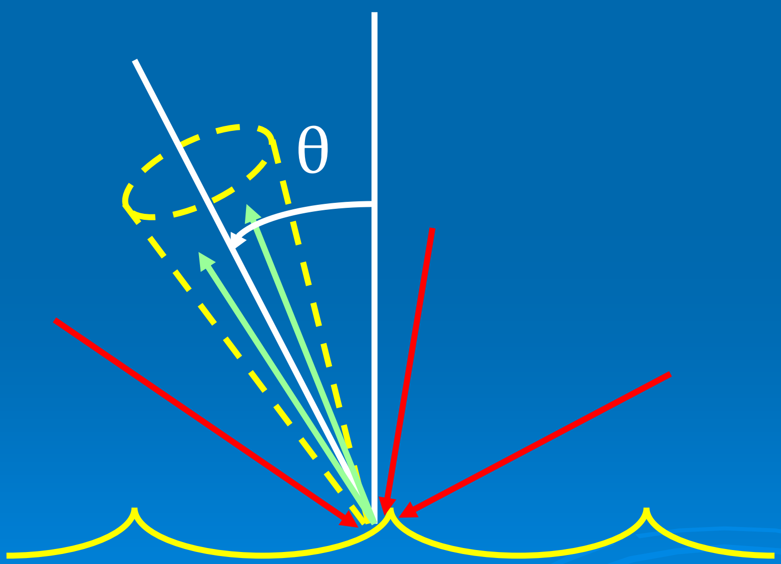
An instrument measures L_u (in air), but L_w is what tells us what is going on in the water. It isn't easy to figure out how much of L_u is due to L_w.

Remote-sensing Reflectance R_{rs}

$R_{rs}(\theta,\varphi,\lambda)$ =

$$\frac{\text{upwelling water-leaving radiance}}{\text{downwelling plane irradiance}}$$

$$R_{rs}(\text{in air}, \theta, \varphi, \lambda) \equiv \frac{L_w(\text{in air}, \theta, \varphi, \lambda)}{E_d(\text{in air}, \lambda)} \qquad [\text{sr}^{-1}]$$



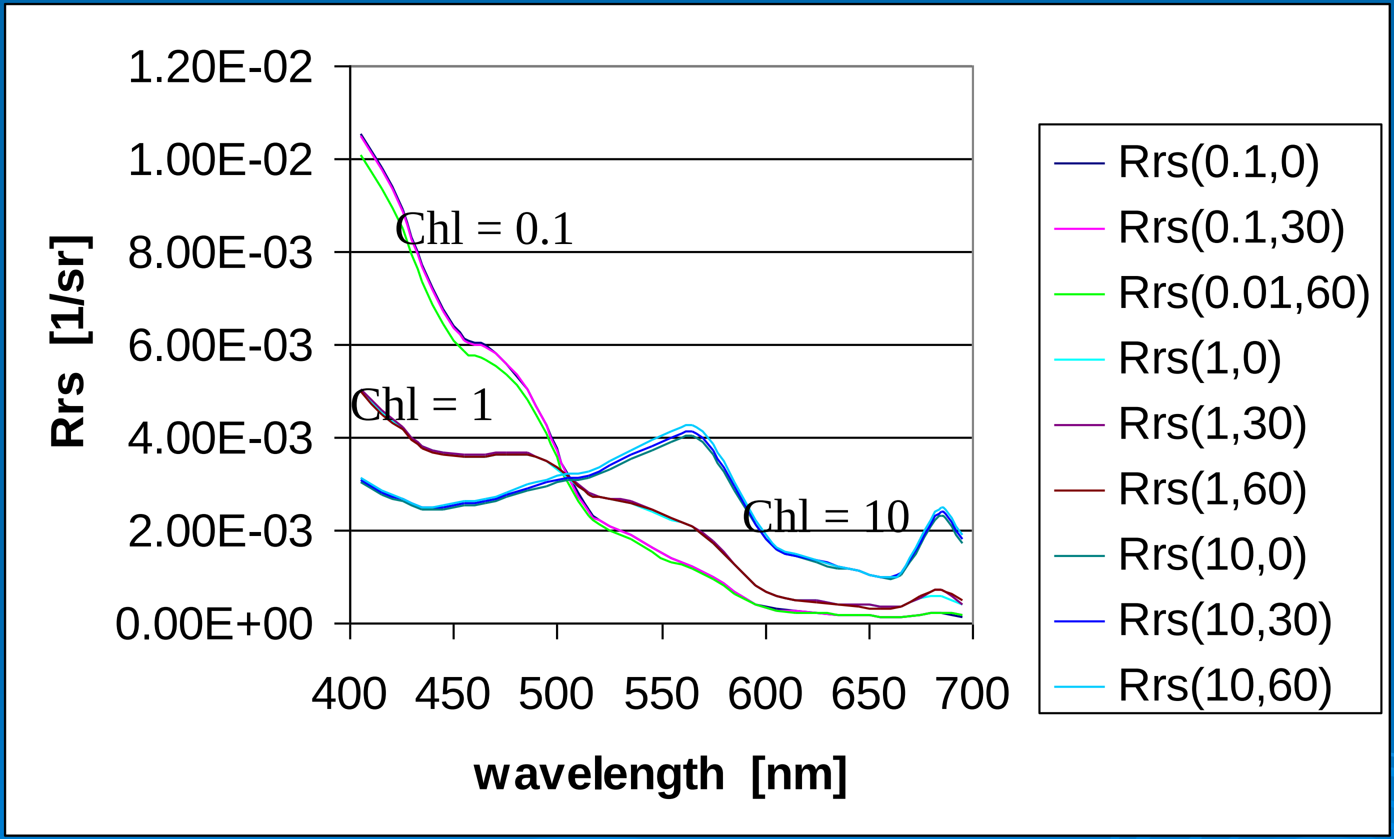
The fundamental quantity used in ocean color remote sensing

Often work with the nadir-viewing R_{rs}, i.e.,with the radiance that is heading straight up from the sea surface (θ = 0)

sea surface

Example R_{rs}

HydroLight runs: Chl = 0.1,1, 10 mg Chl/m^3
Sun at 0, 30, 60 deg in clear sky



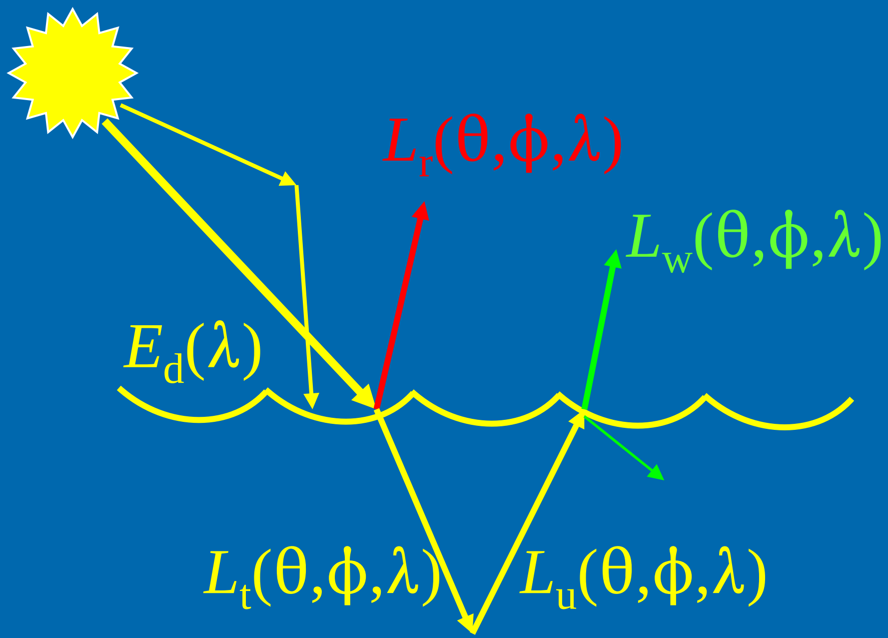
R_{rs} shows very little dependence on sun angle and strong dependence on the water IOPs—a very good AOP

Water-leaving Radiance, L_w

We cannot measure L_w (or R_{rs}) directly. We must estimate them from L_u measurements.

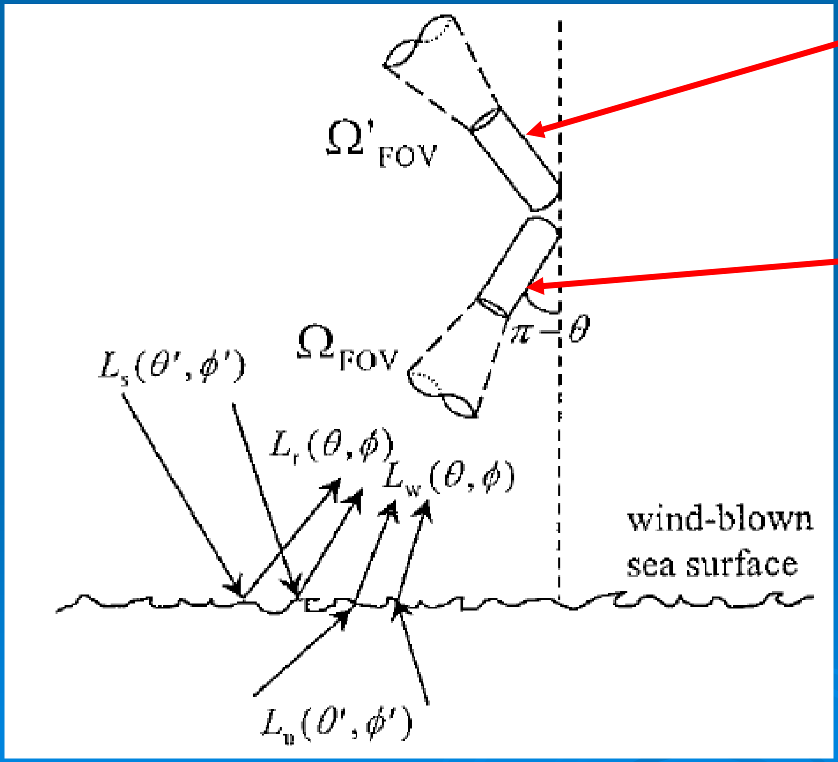
If we have in-water profiles of L_u, we can extrapolate from below the sea surface to get L_w above the surface.

If we have above-surface measurements of L_u, we must remove the contribution of the surface-reflected radiance to get $L_w = L_u – L_r$. This is most often done using the "Carder method."



Estimating L_w and R_{rs} from Above-surface Measurements: The Carder Method

First measure the downwelling (sky) radiance and upwelling (sea surface) radiance at the direction corresponding to specular reflection by a level sea surface.

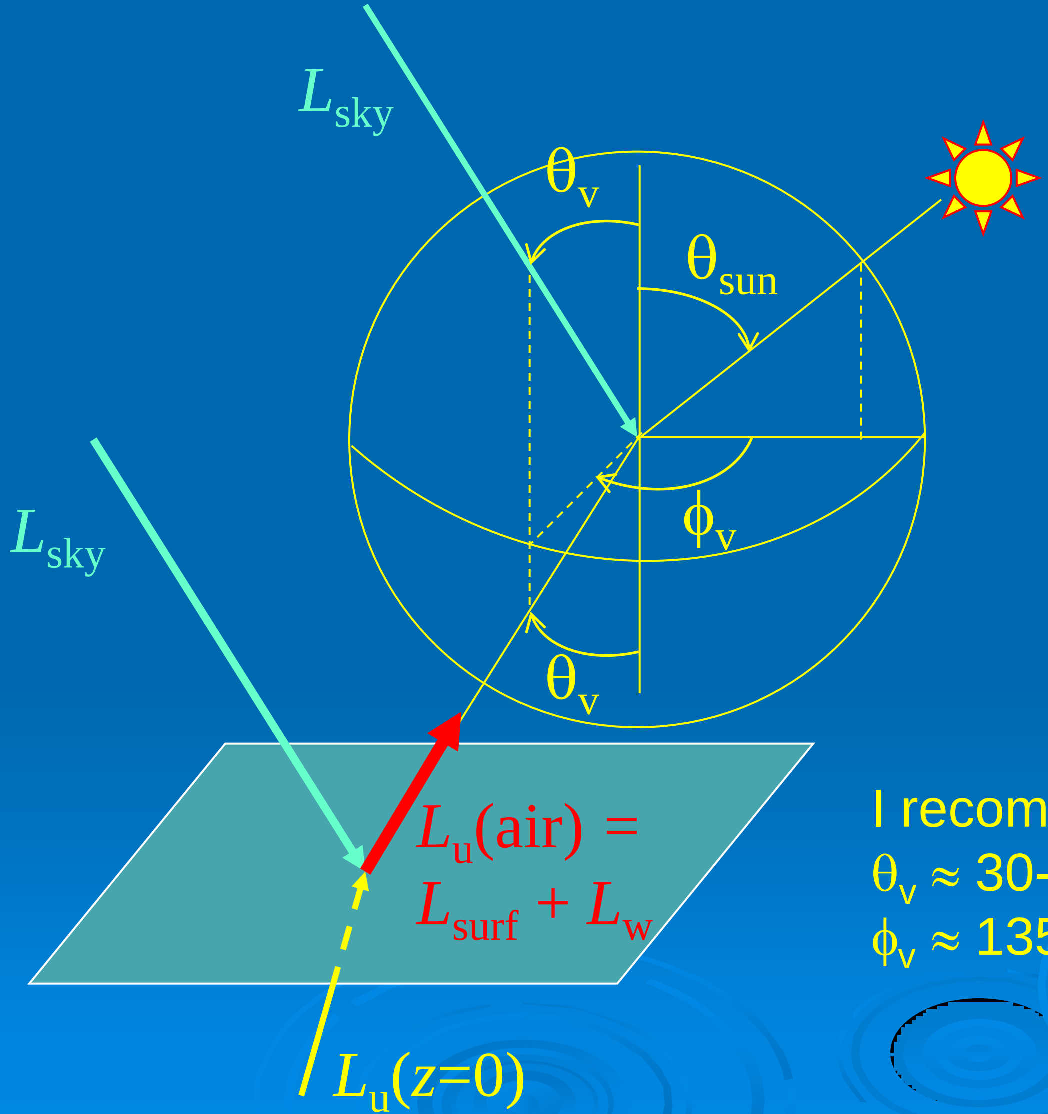


radiometer pointing upward measures $L_{sky}(\theta, \phi, \lambda)$

radiometer pointing downward measures $L_u(\theta, \phi, \lambda)$ = reflected sky radiance + water-leaving radiance

Mobley, AO, 1999

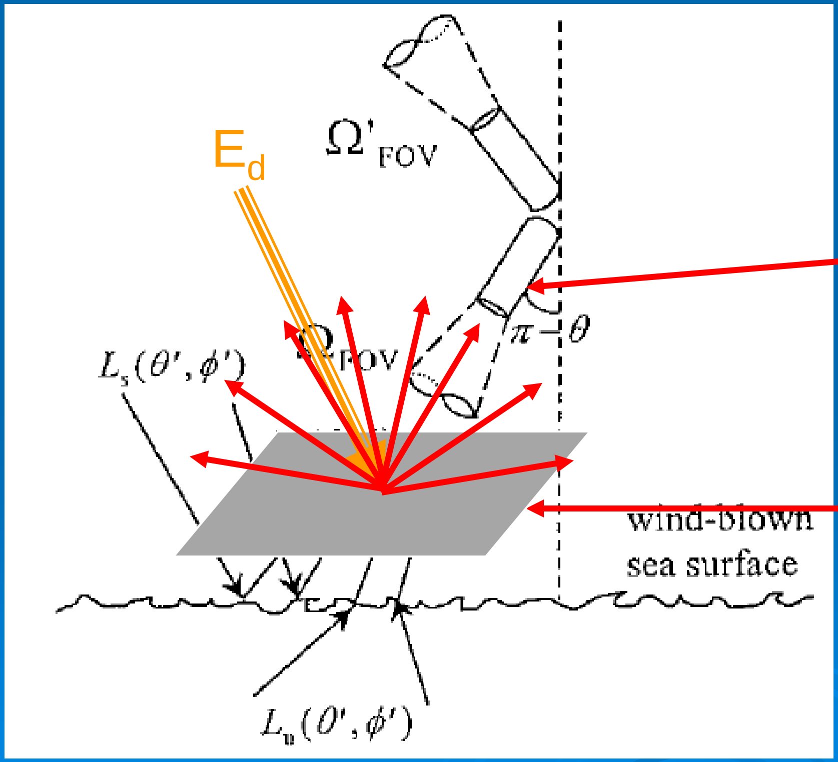
Estimating L_w and R_{rs} from Above-surface Measurements: The Carder Method



I recommend
$\theta_v \approx$ 30-40 deg
$\phi_v \approx$ 135 deg

Estimating L_w and R_{rs} from Above-surface Measurements: The Carder Method

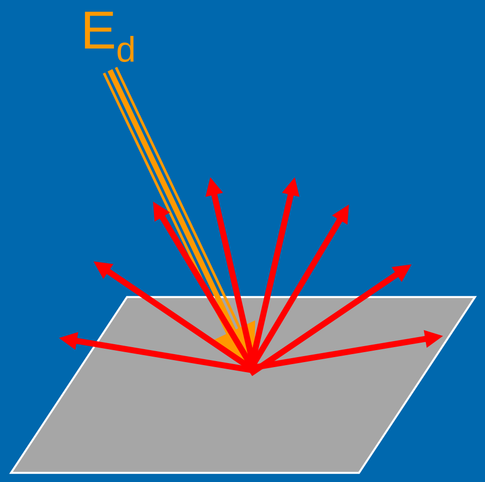
Next measure the radiance reflected by a "gray card" (usually a Spectralon plate) with known irradiance reflectance $R_g(\lambda)$.



radiometer pointing downward now measures $L_g(\theta, \phi, \lambda)$

"gray card"

Estimating L_w and R_{rs} from Above-surface Measurements: The Carder Method



The gray card is assumed to be a Lambertian reflector. Thus the reflected radiance is isotropic and

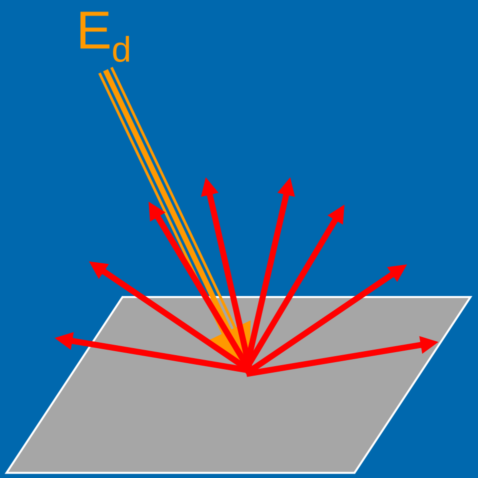
$L_g = (R_g/\pi)E_d$ (can solve for E_d)

A fraction ρ of the measured incident sky radiance L_{sky} is reflected by the sea surface, so $L_{surf} = \rho L_{sky}$

The water-leaving radiance is thus estimated by

$L_w(\theta, \phi, \lambda) = L_u(\theta, \phi, \lambda) - \rho L_{sky}(\theta, \phi, \lambda)$

Estimating L_w and R_{rs} from Above-surface Measurements: The Carder Method



Estimate R_{rs} by

$$R_{rs} = L_w/E_d = \underbrace{(L_u - \rho L_{sky})}_{L_w} / \underbrace{(\pi L_g/R_g)}_{E_d}$$

We could measure E_d with a plane irradiance sensor. However, estimating E_d from the gray-card reflectance means that all measurements are done with the same instrument, and no instrument calibration (other than a dark current correction) is required because any multiplicative calibration factor on L cancels out.

But how do we get the value of ρ?

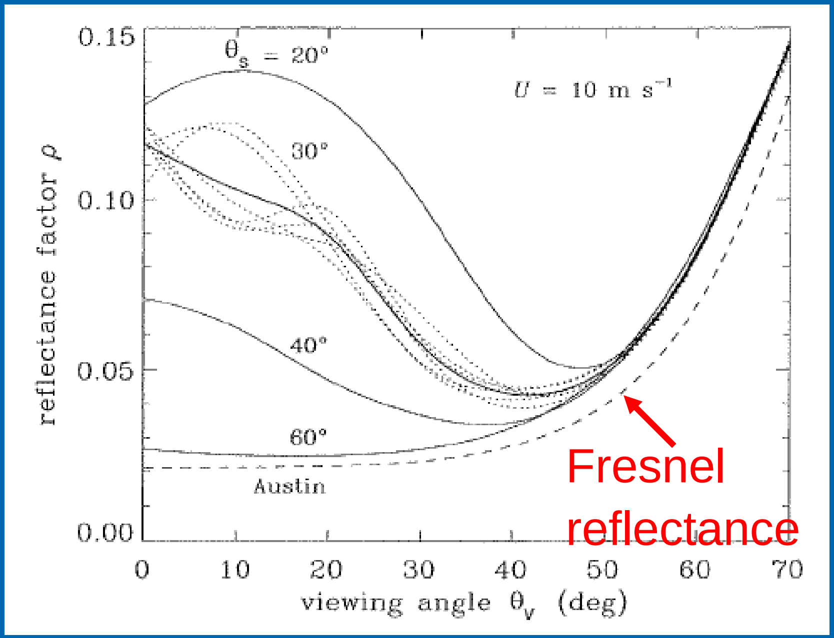
Estimating the Radiance Reflectance ρ

The radiance reflectance ρ depends on viewing direction, sky conditions, and sea-surface wave conditions (and slightly on wavelength). ρ is therefore NOT an IOP, and it is NOT equal to the Fresnel reflectance of the surface, except for a level sea surface.

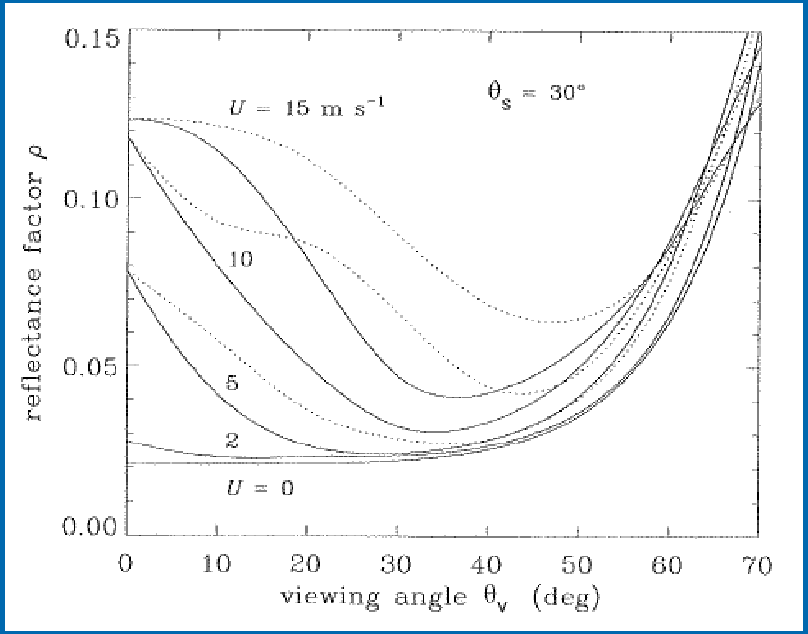
HydroLight computes the surface-reflected radiance for the input sky radiance L_{sky} and surface conditions, so I have used H to compute $\rho = L_{surf} / L_{sky}$ as a function of sun angle, viewing direction, and wind speed. (The ρ value is in the printout for H runs.)

There is a table of HydroLight-computed, clear-sky ρ values in the Papers directory for this course (file rhoTable_AO1999.txt).

Dependence of ρ on Geometry and Wave State

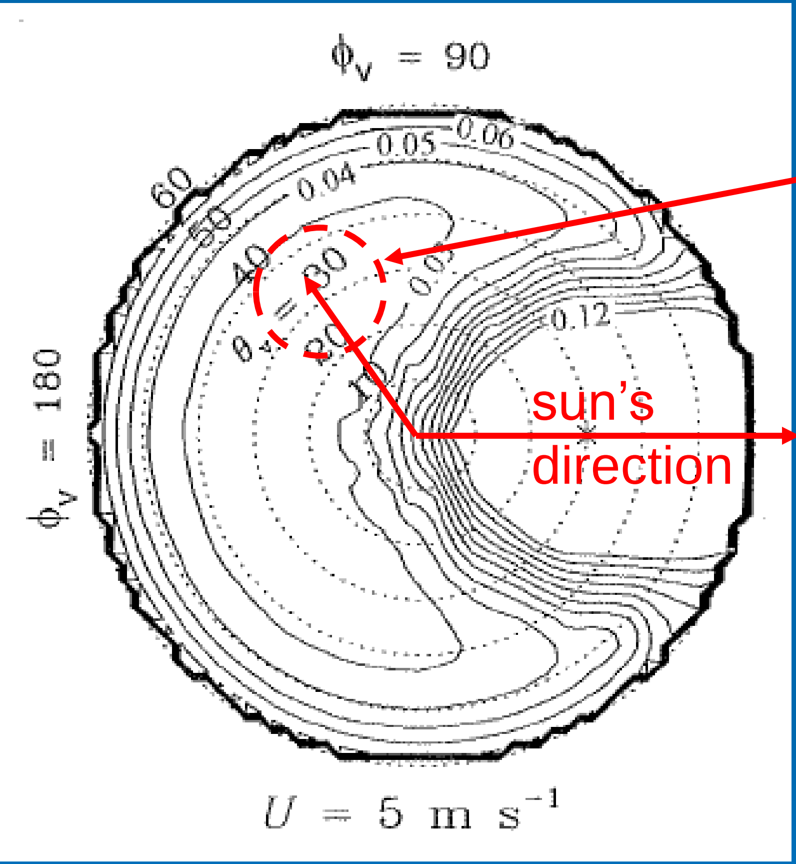


ρ as a function of sun zenith angle for a given wind speed



ρ as a function wind speed for a given sun zenith angle

Dependence of ρ on Geometry and Wave State



ρ as a function of polar and azimuthal viewing angles for a given sun zenith angle and wind speed

an azimuthal viewing direction of roughly 135 degrees from the sun and 30-40 deg from the nadir is optimum for making measurements:

- minimizes sun glitter
- avoids shading by the ship or instrument
- minimum values of ρ and slow variation with viewing direction, so can be reasonably good estimates

See Mobley, *Applied Optics*, 1999 for full details.

Average or Mean Cosines

The average or mean cosines give the average of the cosθ for all of the photons making up the radiance distribution. This tells you something about the directional pattern of the radiance. For the downwelling radiance we have

$$\bar{\mu}_d = \frac{\int_0^{2\pi}\int_0^{\pi/2} L(\theta,\phi)\cos\theta\sin\theta\, d\theta\, d\phi}{\int_0^{2\pi}\int_0^{\pi/2} L(\theta,\phi)\sin\theta\, d\theta\, d\phi} = \frac{E_d}{E_{od}}$$

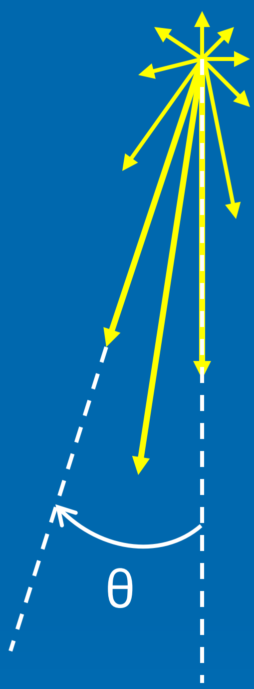
Likewise, for the upwelling radiance, $\bar{\mu}_u = E_u / E_{ou}$

For the entire radiance distribution,

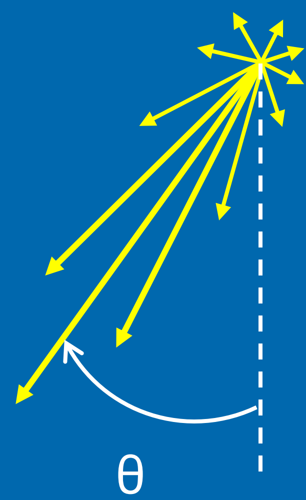
$$\bar{\mu} = \frac{\int_0^{2\pi}\int_0^{\pi} L(\theta,\phi)\cos\theta\sin\theta\, d\theta\, d\phi}{\int_0^{2\pi}\int_0^{\pi} L(\theta,\phi)\sin\theta\, d\theta\, d\phi} = \frac{E_d - E_u}{E_o}$$

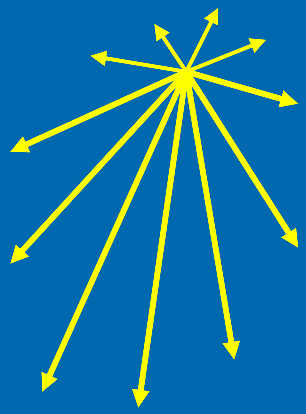
Note: $E_o = E_{od} + E_{ou}$, but $\bar{\mu} \neq \bar{\mu}_d + \bar{\mu}_u$

Mean Cosines

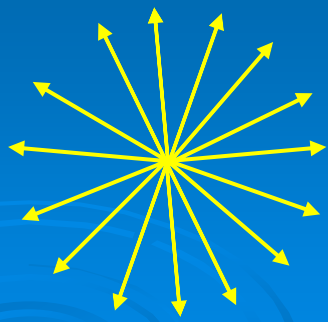


most photons heading
almost straight down:
small average θ, large μ_d





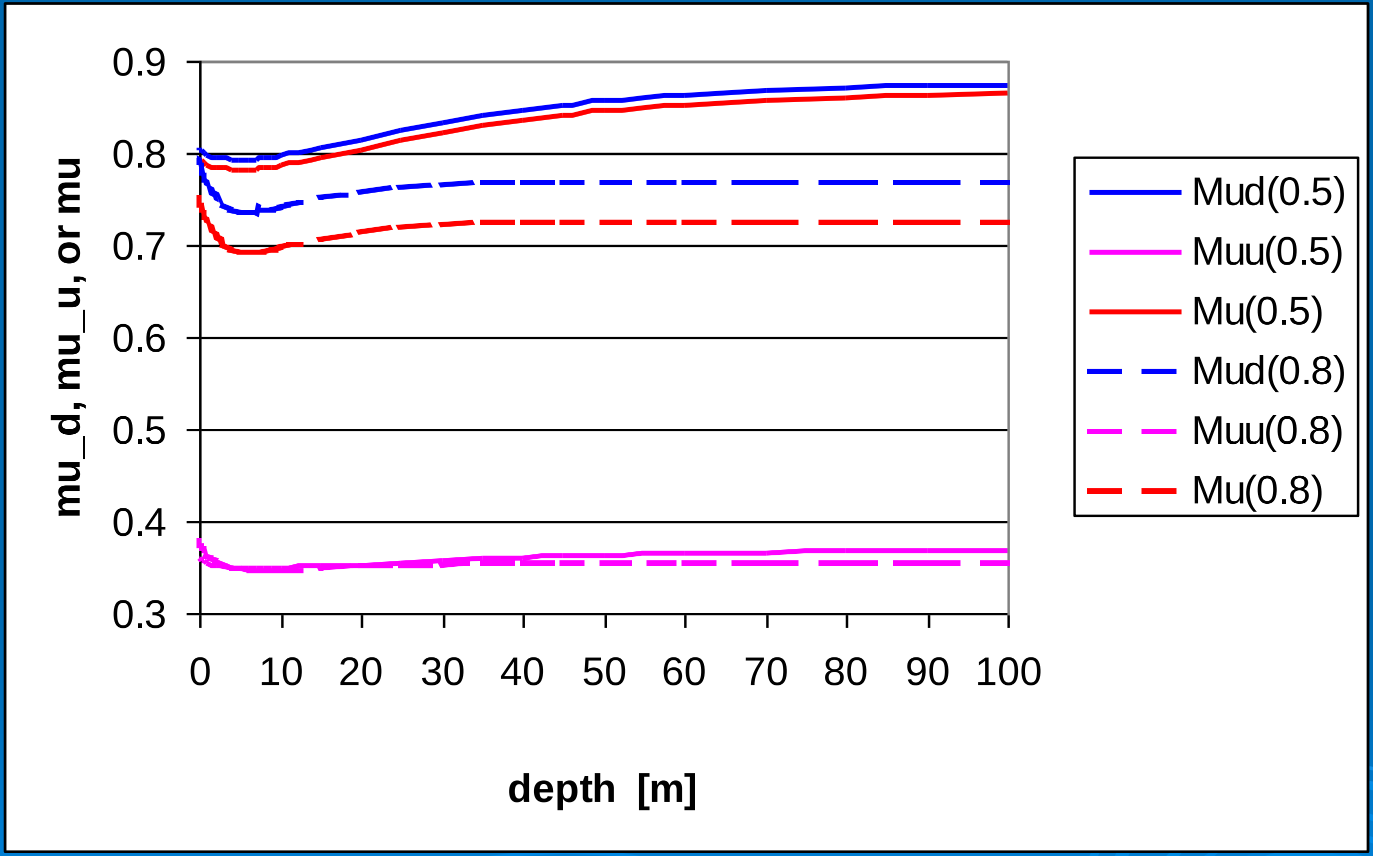
most photons heading at a large angle, or a diffuse radiance: large average θ, small μ_d



isotropic radiance:
$\mu_d = \mu_u = 0.5$
$\mu = 0$

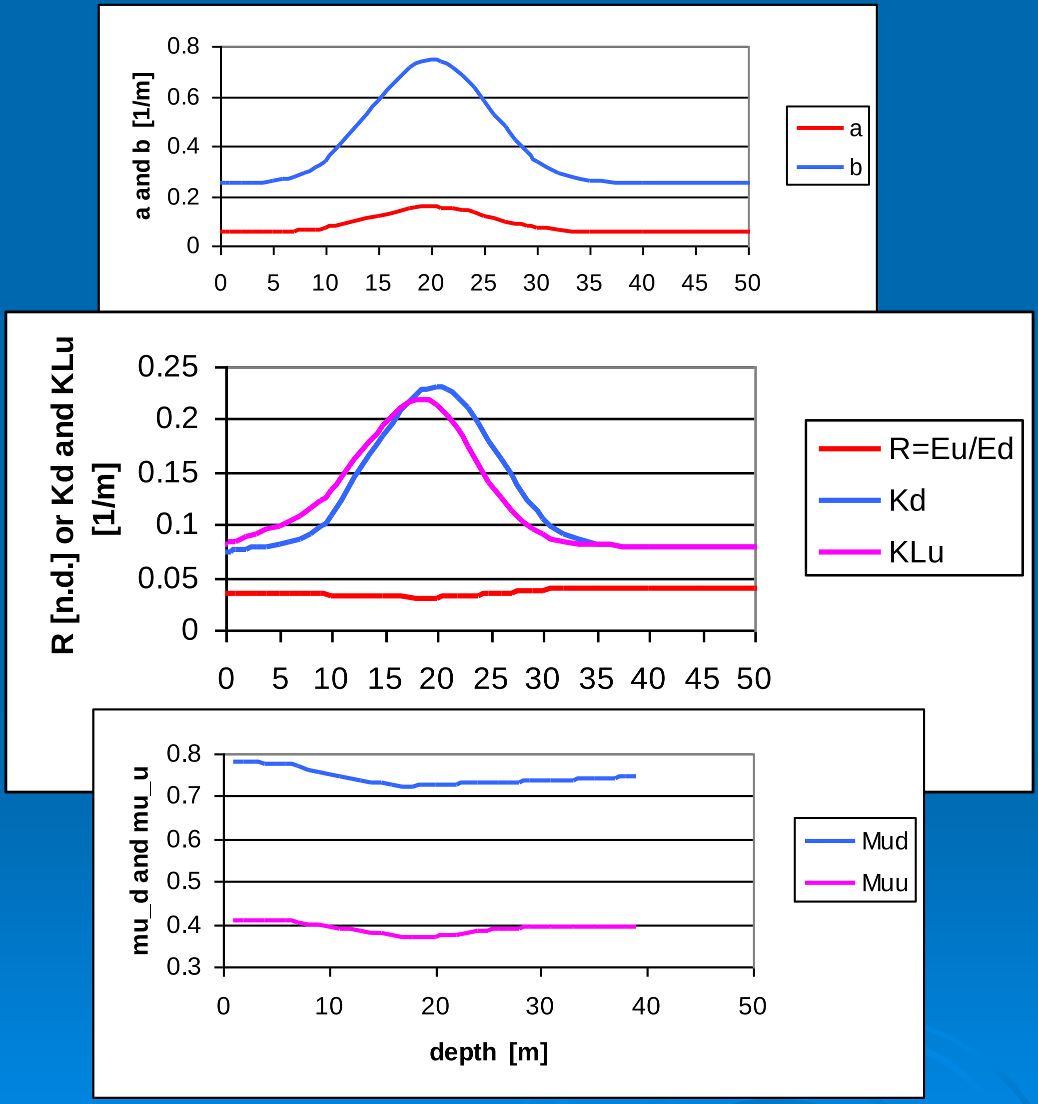
Mean Cosines

highly absorbing water: b = a, so ω_o = 0.5, vs
highly scattering water: b = 4a, so ω_o = 0.8



Note: the highly scattering water approaches asymptotic values quicker than the highly absorbing water.

The Real World: Inhomogeneous Water

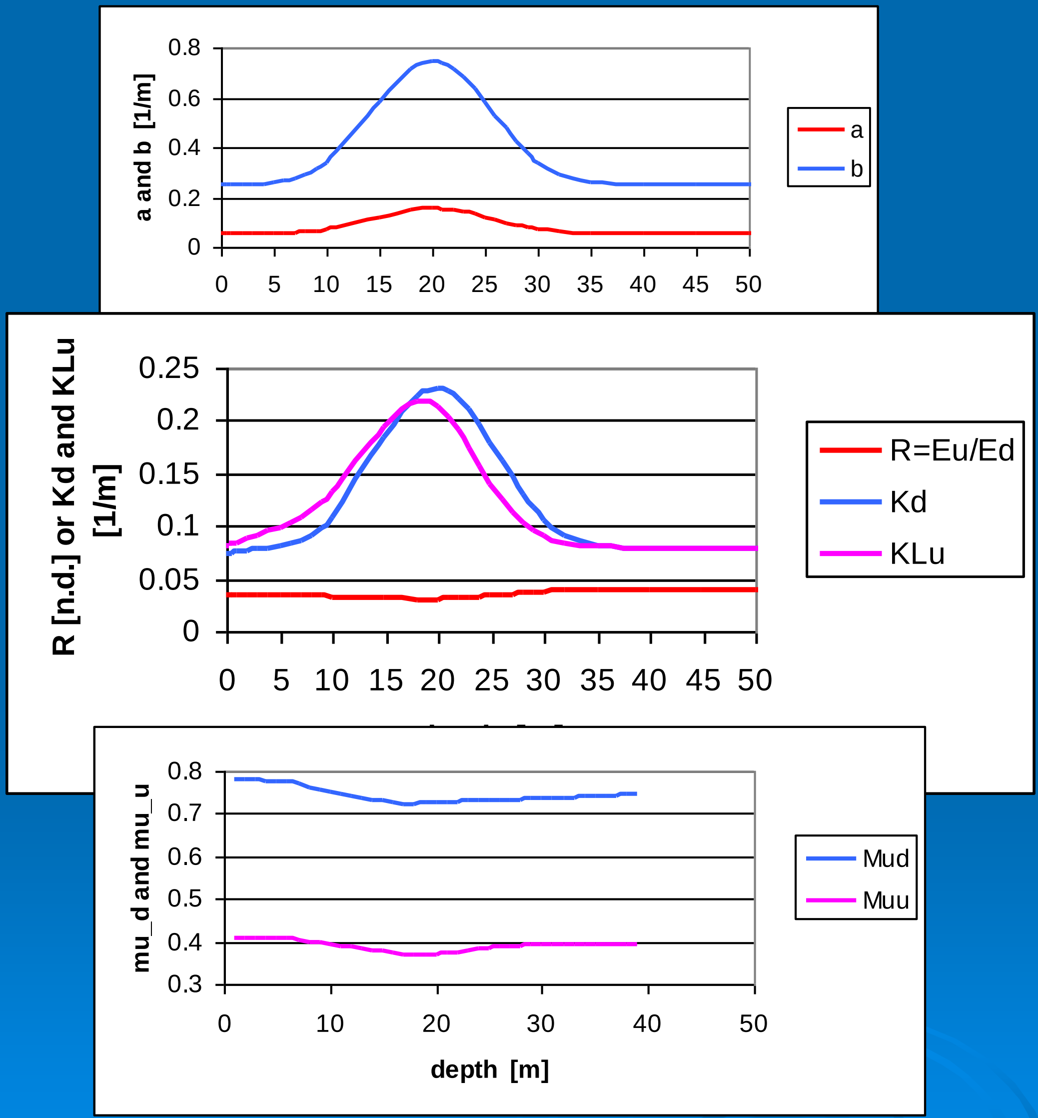


HydroLight run with Chl = 0.5 mg/m^3 background and Chl = 2.5 mg/m^3 max at 20 m

Note how well K_d and K_{Lu} correlate with the IOPs, but R isn't much affected. Why?

μ_d and μ_u are not strongly affected

The Real World: Inhomogeneous Water



HydroLight run with Chl = 0.5 mg/m^3 background and Chl = 2.5 mg/m^3 max at 20 m

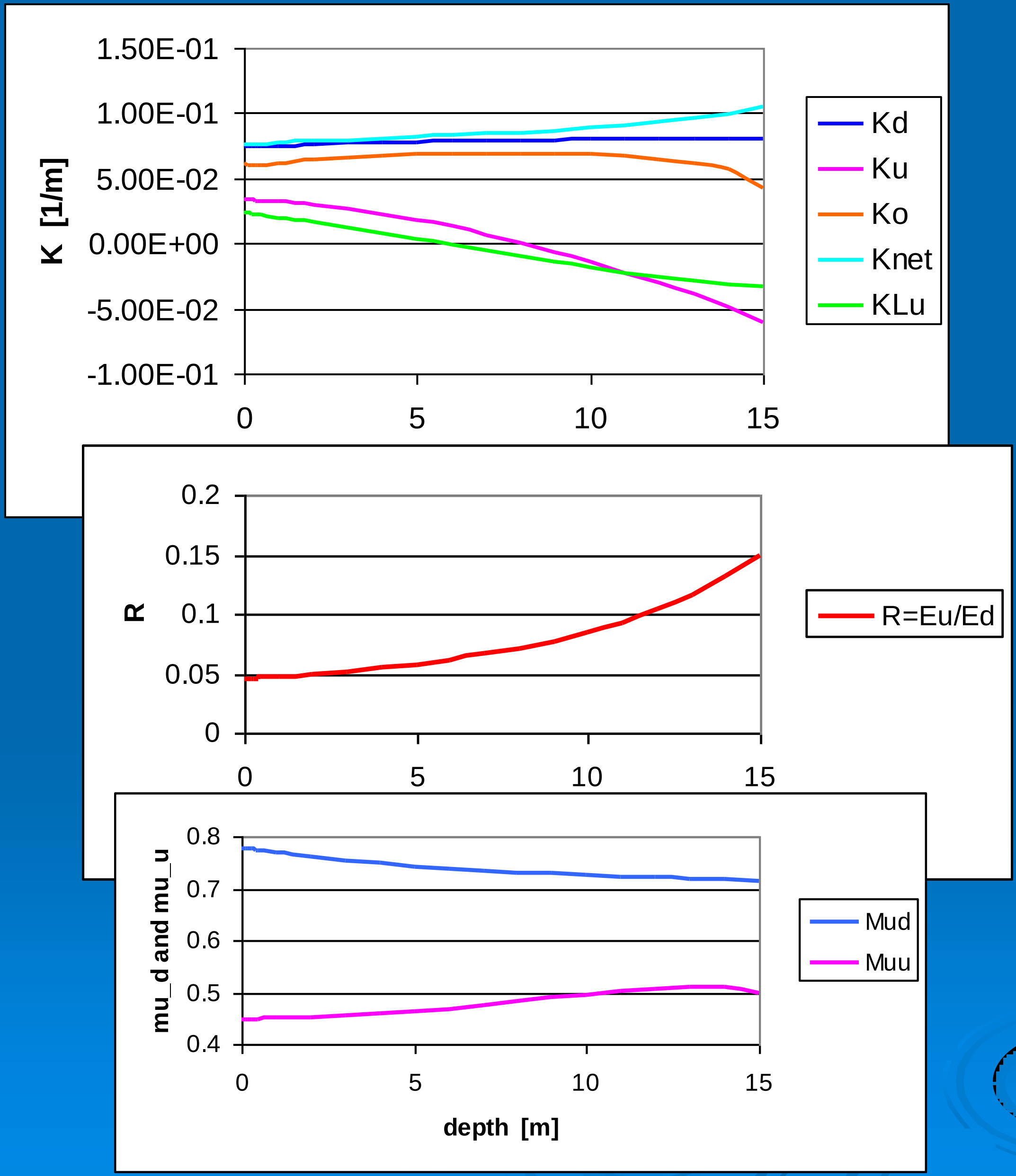
Note how well K_d and K_{Lu} correlate with the IOPs, but R isn't much affected. Why?

$K \propto a$

$R \propto b_b/a$

What would happen to K and R if there were a layer of highly scattering but non-absorbing particles in the water?

Explain These AOPs



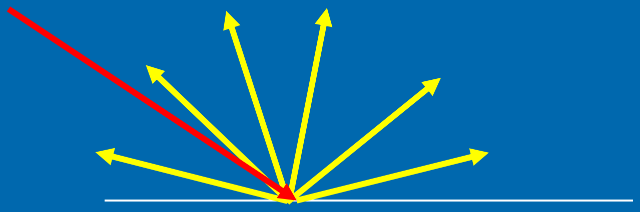
What does it mean for K_u and K_{Lu} to become negative?

What does $\mu_u = 0.5$ say about the upwelling radiance distribution at 15 m?

The Answer

The water IOPs were homogeneous, but there was a Lambertian bottom at 15 m, which had a reflectance of R_b = 0.15

Lambertian means the reflected radiance is the same in all directions
(L is isotropic)



Exercise: compute μ_d, μ_u, and μ for an isotropic radiance distribution: $L(\theta,\phi) = L_o$ = a constant

Relations Among IOPs and AOPs

Manipulating the radiative transfer equation (see *Light and Water*, Section 5.12) gives various relations among IOPs and AOPs, which are exact if there are no internal sources, e.g.

$$a \le K_d \bar{\mu}_d \le c$$

$$\bar{\mu} = \frac{a}{K_{net}} = \frac{a(1 - R)}{K_d - R K_u}$$

Rewritten forms of Gershun's eq.

$$\frac{d}{dz}(E_d - E_u) = -a E_o ,$$

$$R = \frac{K_d - a/\bar{\mu}_d}{K_u + a/\bar{\mu}_u}$$

These relations sometimes provide useful sanity checks on data.

Orchids in the Singapore National Botanic Garden

Orchids in the Singapore National Botanic Garden




all photos by Curtis Mobley