2021 Summer Course

on Optical Oceanography and Ocean Color Remote Sensing

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Apparent Optical Properties

Schiller Coastal Studies Center

Apparent Optical Properties (AOPs)

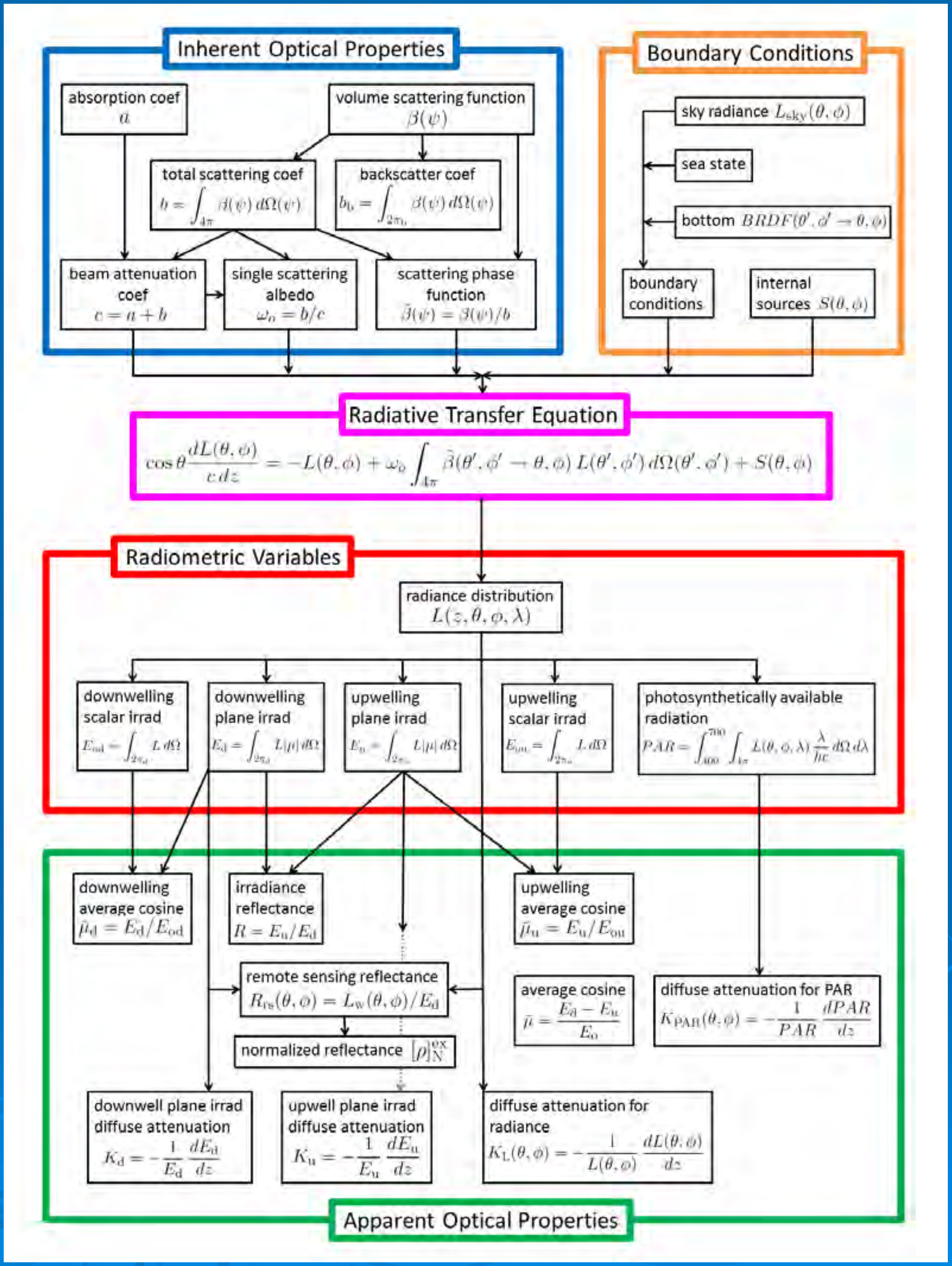
AOPs are quantities that

(1) depend on the IOPs and on the radiance distribution, and

(2) 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 (sun zenith angle, sky condition, surface waves) and strongly on the water IOPs.

AOPs are usually ratios or depth 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 harder to interpret).

In a Perfect World

Light Properties: measure the radiance as a function of location, time, direction, wavelength, $L(x,y,z,t,\theta,\varphi,\lambda)$, 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

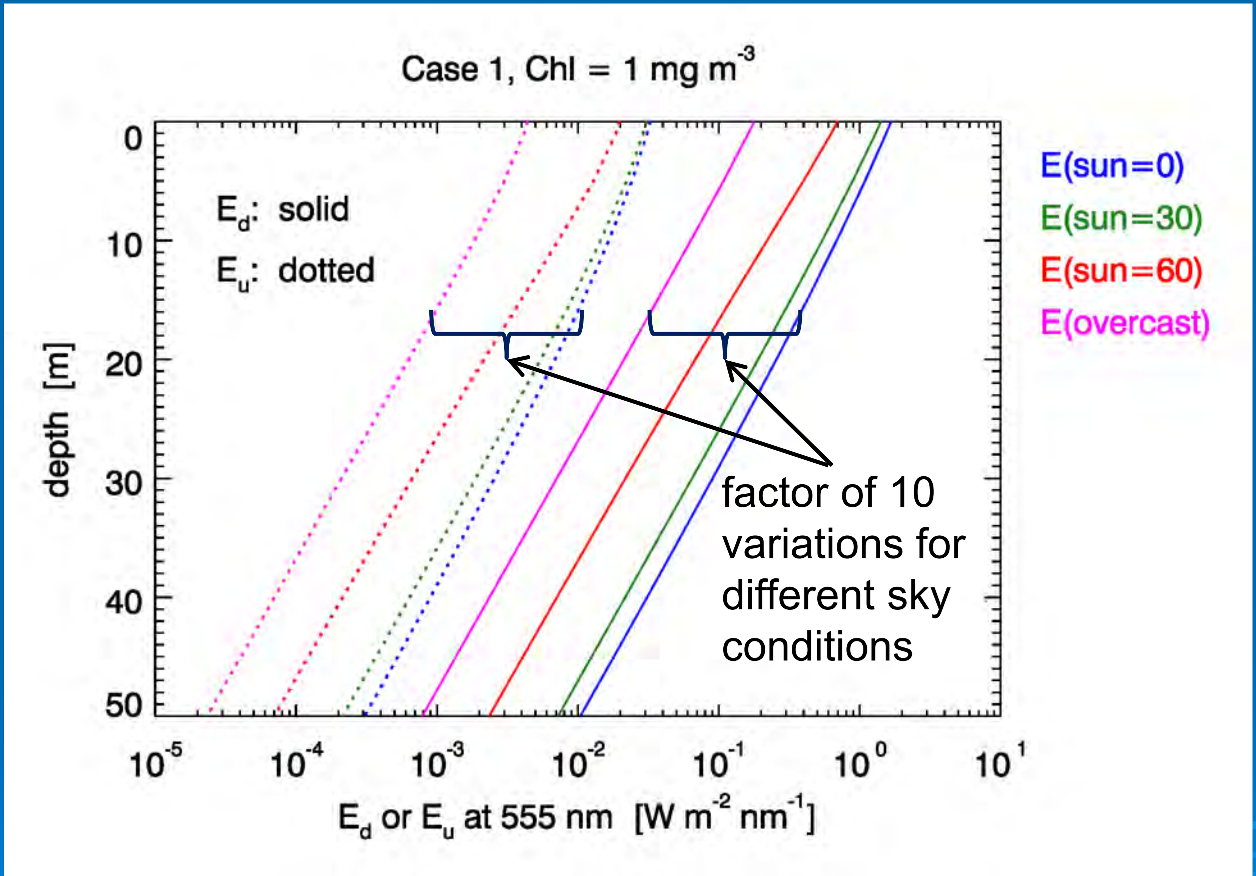
$L(x,y,z,t,\theta,\varphi,\lambda)$ is too difficult and time consuming to measure on a routine basis, and you don't need all of the information contained in L, so therefore measure irradiances, PAR, etc. (ditto for VSF vs b, b_b,….)

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)?

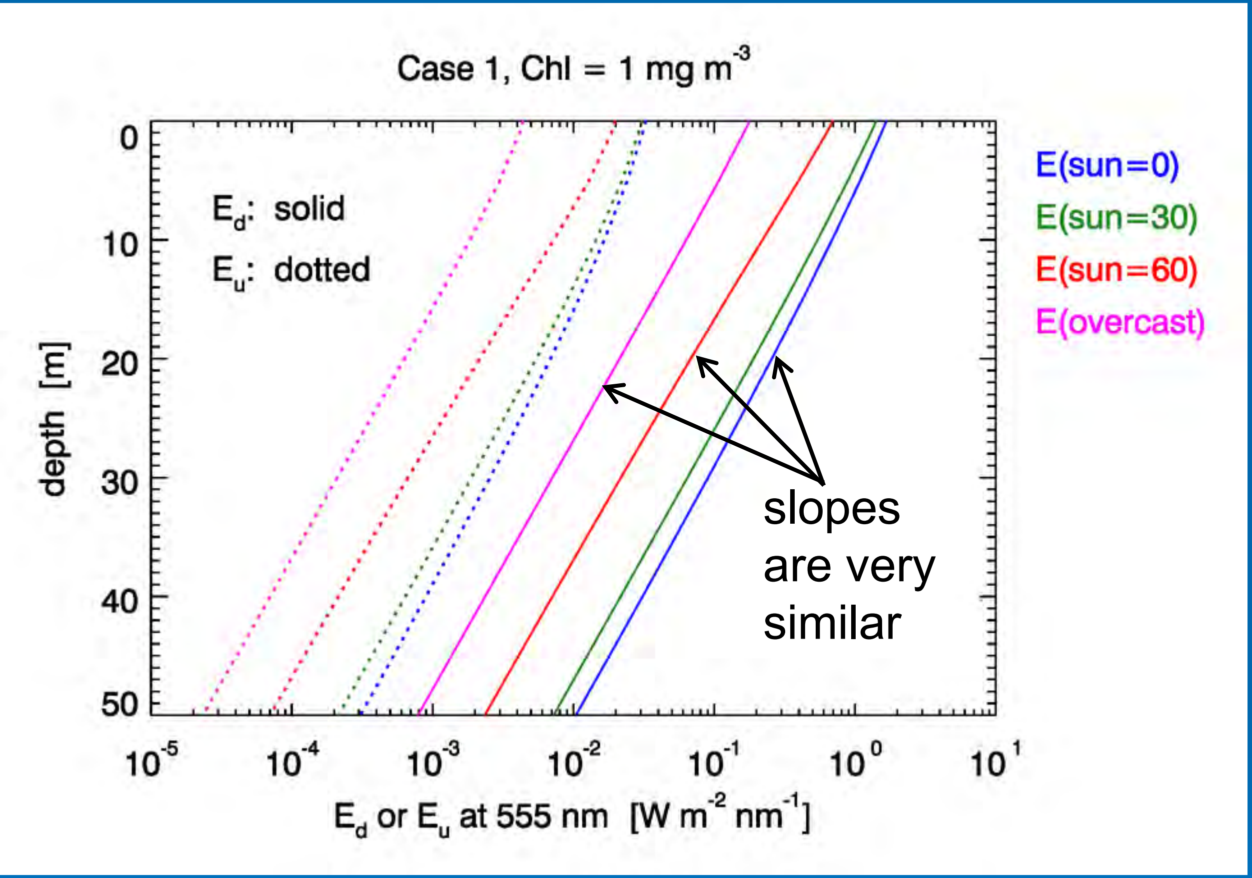
E_d and E_u

HydroLight runs: Case 1 water, Chl = 1.0 mg/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 incident lighting (sun angle and sky condition); 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:

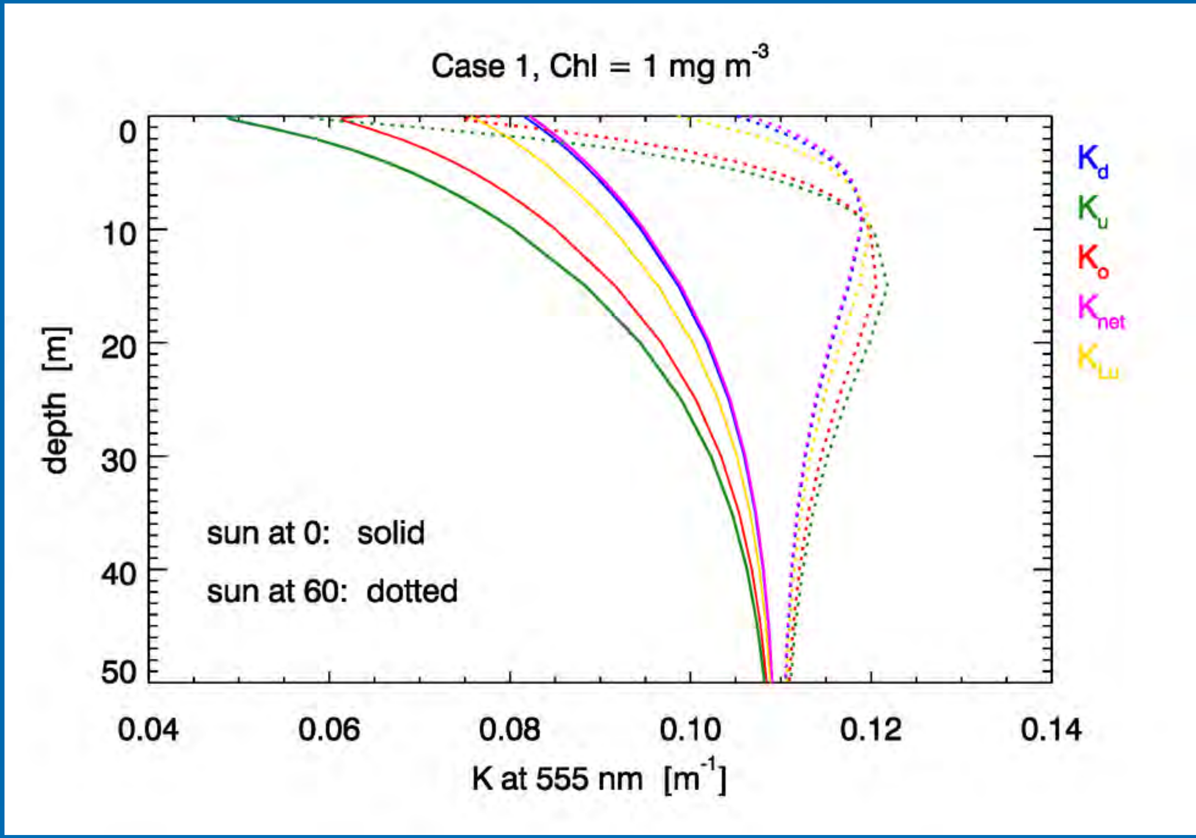
$$K_{\mathrm{d}}(z,\lambda) = -\frac{1}{E_{\mathrm{d}}(z,\lambda)} \frac{dE_{\mathrm{d}}(z,\lambda)}{dz} = -\frac{d \ln E_{\mathrm{d}}(z,\lambda)}{dz} \quad [\mathrm{m}^{-1}]$$

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

Note that the definition of K_d is equivalent to

$$E_d(z,\lambda) = E_d(0,\lambda) \exp\left(-\int_0^z K_d(z,\lambda)\,dz\right)$$

How similar are the different K's?



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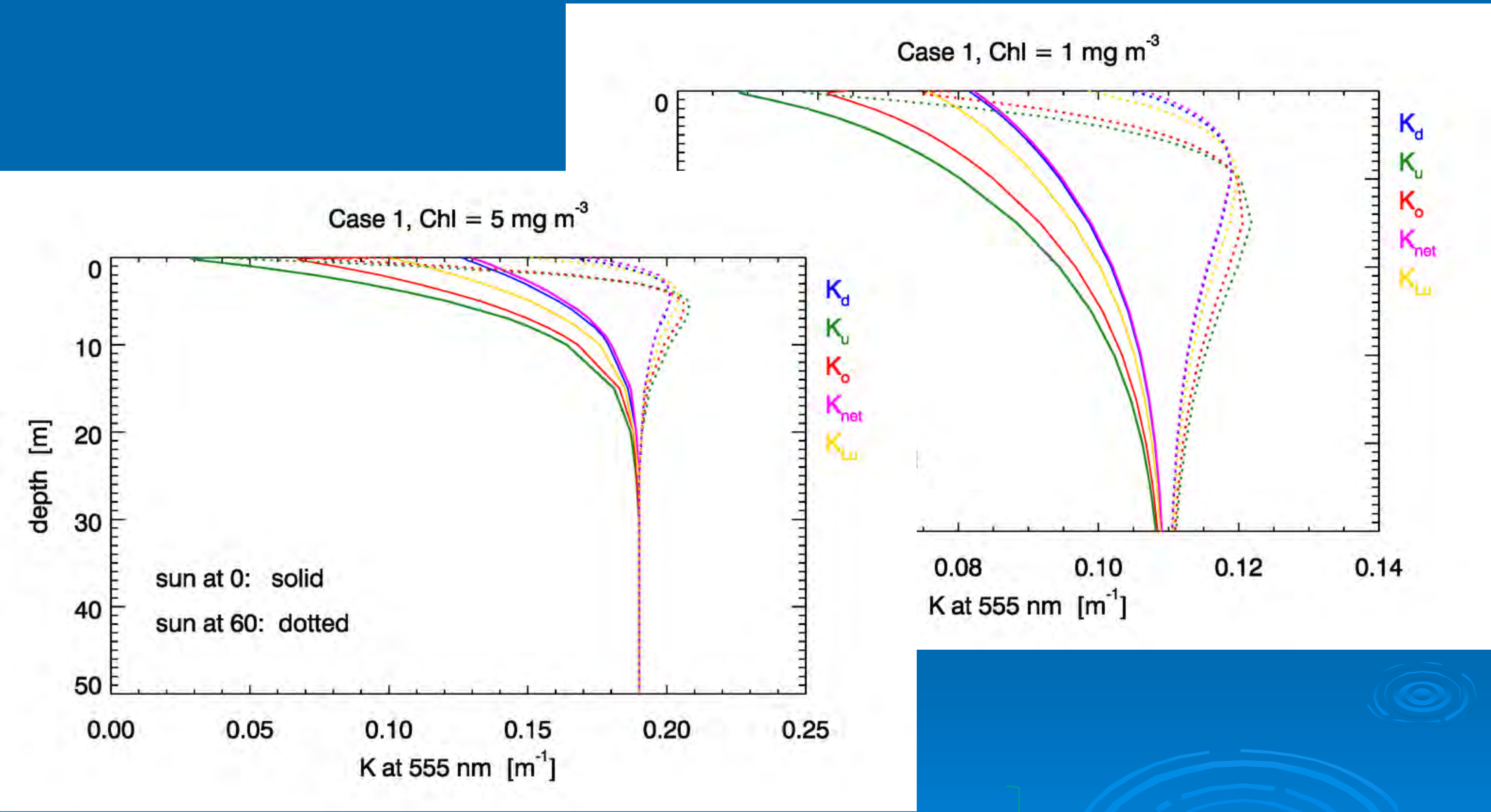
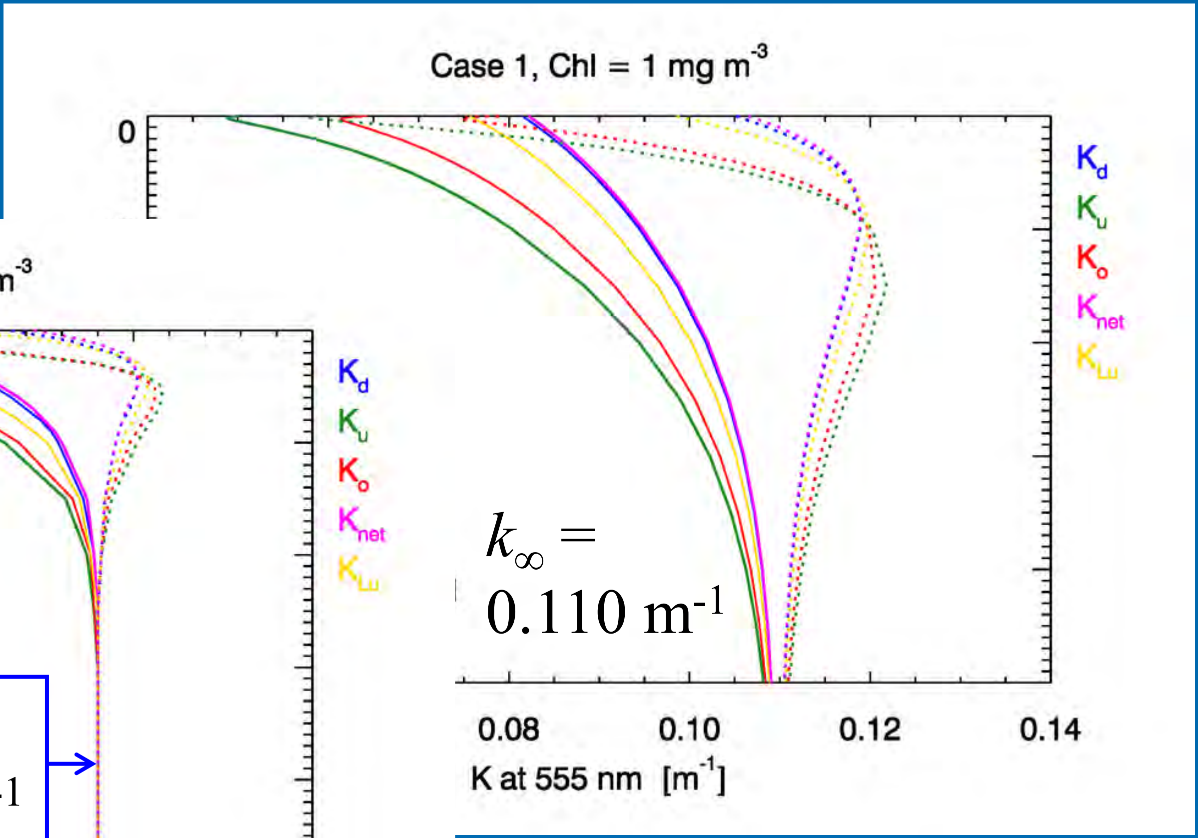
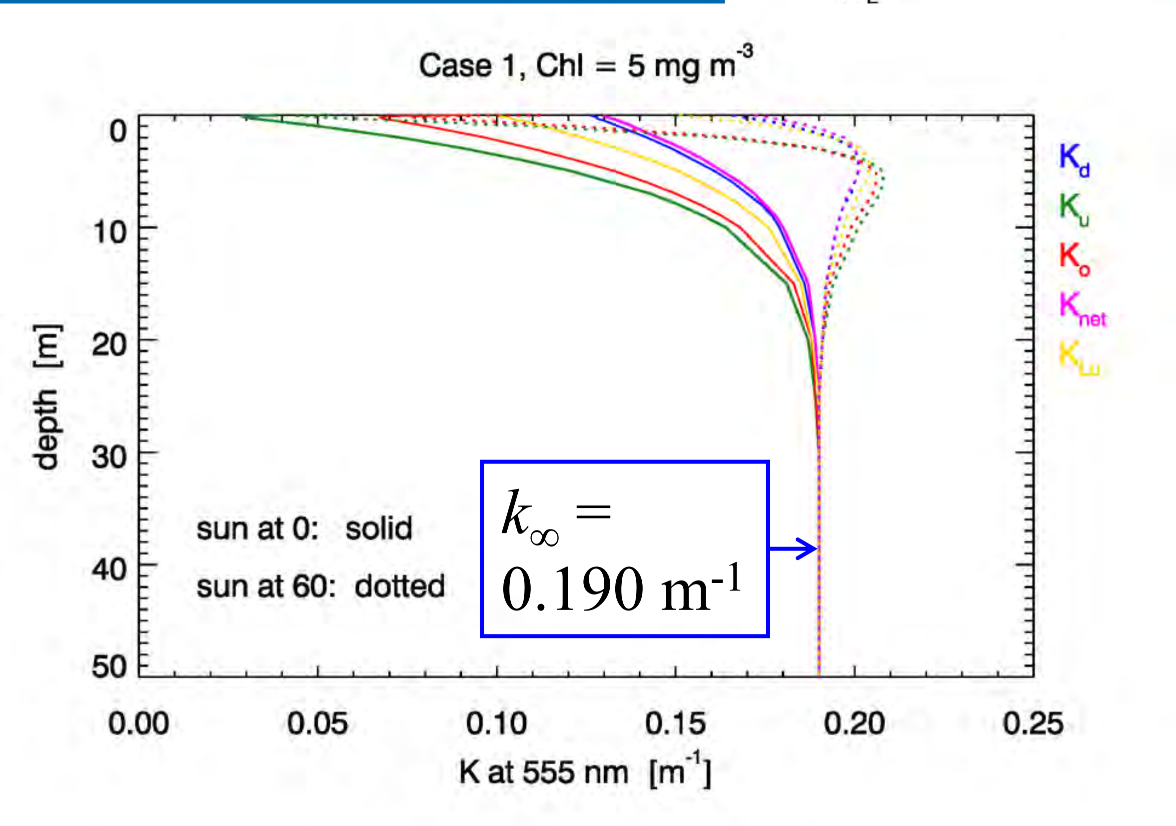


NOTE: The K's depend on depth, even though the water is homogeneous, and they are most different near the surface (where the light field is changing because of boundary effects)

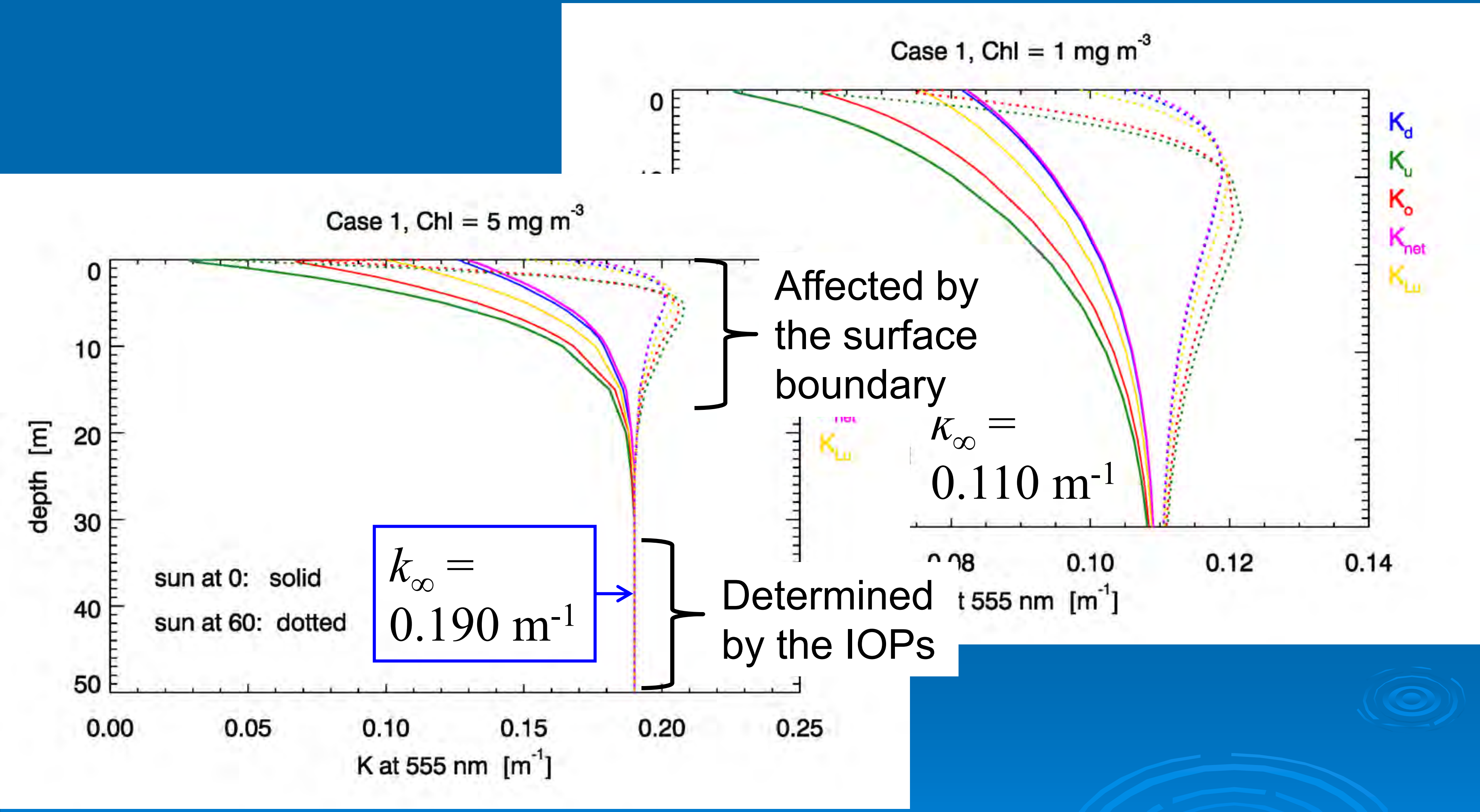
Asymptotic Values





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

Asymptotic Values



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

Models for K_d

$$E_d(z,\lambda) = E_d(0,\lambda)\,\exp\left(-\int_0^z K_d(z,\lambda)\,dz\right) = E_d(0,\lambda)\,e^{-\bar{K}_d(\lambda)\,z}$$

$$K_d(\lambda) = \frac{M(\lambda)}{M(\lambda_o)}\left[K_d(\lambda_o) - K_{dw}(\lambda_o)\right] + K_{dw}(\lambda)$$
$$= M(\lambda)\left[K_d(490) - 0.0224\right] + K_{dw}(\lambda)$$

Austin and Petzold (1986)

K_d(490) is a standard NASA remote-sensing product.

Why 490? Who cares about $K_d(\lambda)$?

Table 3.16. Values of the coefficient $M(\lambda)$ and of the downwelling diffuse attenuation coefficient for pure sea water, $K_{dw}(\lambda)$, for use in Eq. (3.49).[a]

λ (nm)	M (m^{-1})	K_{dw} (m^{-1})	λ (nm)	M (m^{-1})	K_{dw} (m^{-1})	λ (nm)	M (m^{-1})	K_{dw} (m^{-1})
350	2.1442	0.0510	470	1.1982	0.0179	590	0.4840	0.1578
360	2.0504	0.0405	480	1.0955	0.0193	600	0.4903	0.2409
370	1.9610	0.0331	490	1.0000	0.0224	610	0.5090	0.2892
380	1.8772	0.0278	500	0.9118	0.0280	620	0.5380	0.3124
390	1.8009	0.0242	510	0.8310	0.0369	630	0.6231	0.3296
400	1.7383	0.0217	520	0.7578	0.0498	640	0.7001	0.3290
410	1.7591	0.0200	530	0.6924	0.0526	540	0.7300	0.3559
420	1.6974	0.0189	540	0.6350	0.0577	660	0.7301	0.4105
430	1.6108	0.0182	550	0.5860	0.0640	670	0.7008	0.4278
440	1.5169	0.0178	560	0.5457	0.0723	680	0.6245	0.4521
450	1.4158	0.0176	570	0.5146	0.0842	690	0.4901	0.5116
460	1.3077	0.0176	580	0.4935	0.1065	700	0.2891	0.6514

[a] Condensed with permission from Austin and Petzold (1986), who give values every 5 nm.

Light and Water

Models for K_d

$$E_d(z,\lambda) = E_d(0,\lambda)\,\exp\left(-\int_0^z K_d(z,\lambda)\,dz\right) = E_d(0,\lambda)\,e^{-\bar{K}_d(\lambda)\,z}$$

$$\begin{aligned} K_d(\lambda) &= \frac{M(\lambda)}{M(\lambda_o)}\,[K_d(\lambda_o) - K_{dw}(\lambda_o)] + K_{dw}(\lambda) \\ &= M(\lambda)\,[K_d(490) - 0.0224] + K_{dw}(\lambda) \end{aligned}$$

Austin and Petzold (1986)

$K_d(490)$ is a standard NASA remote-sensing product.

Why 490? Who cares about $K_d(\lambda)$?

$$E_d(z,\lambda) = E_d(0,\lambda)\,\exp\left(-\int_0^z K_d(z,\lambda)\,dz\right)$$

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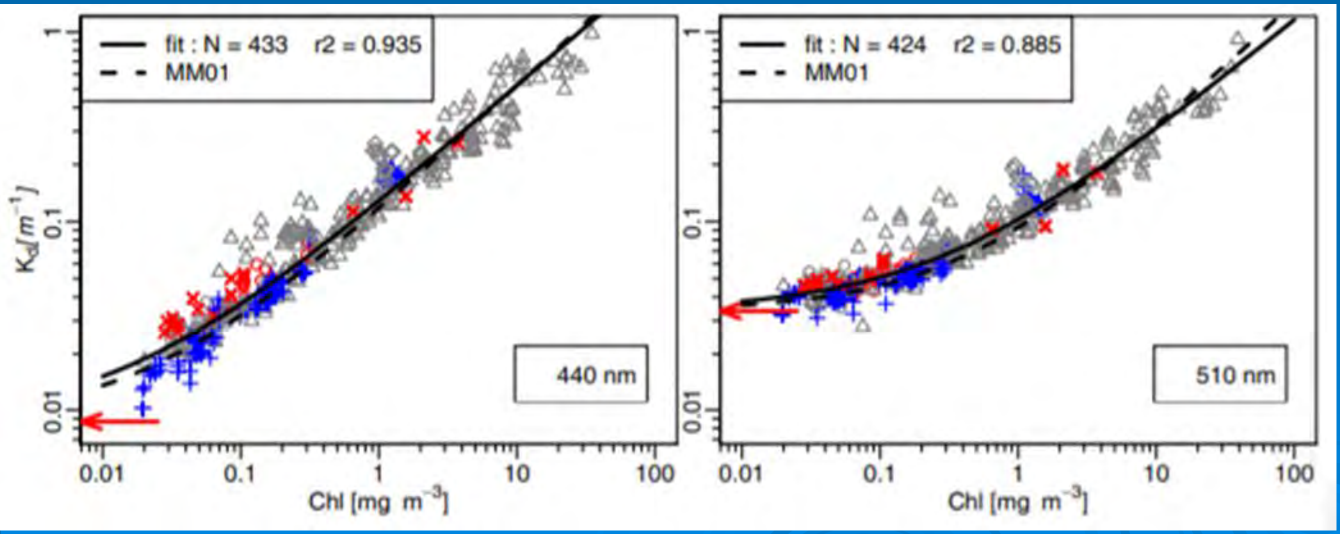
Models for K_d

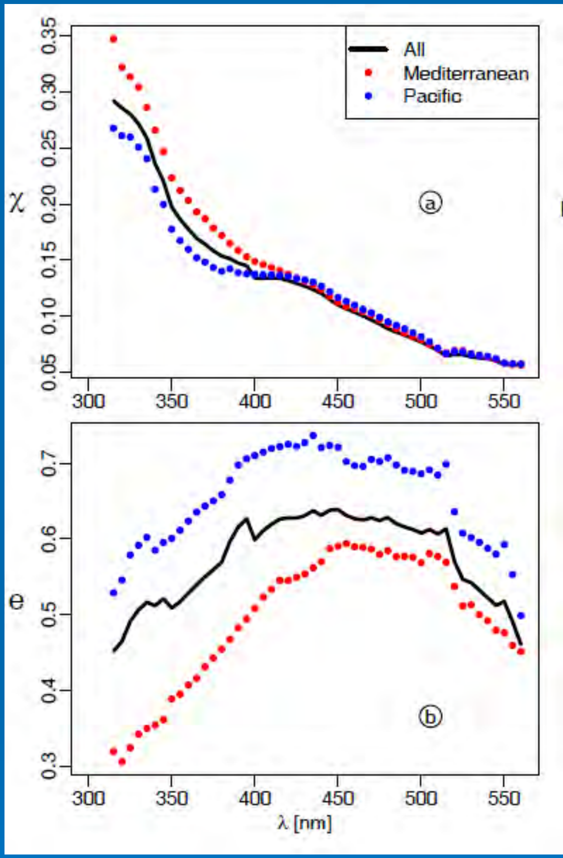
$$\bar{K}_d(\lambda) = K_{dw}(\lambda) + \chi(\lambda)\,[Chl]^{e(\lambda)}$$

Morel (1988)

$\bar{K}_d(\lambda)$ is the depth average of $K_d(z)$ over the euphotic zone

This model is for Case 1 water only!





Morel et al.
BioGeoSci, 2007

Models for K_d

Lee et al., JGR Oceans (2013): A semi-analytic model with 5 geometric parameters fit via HydroLight simulations.

$$\bar{K}_d(\lambda) = (1 + m_o\theta_s)a(\lambda) + m_1\left[1 - \gamma\frac{b_{bw}(\lambda)}{b_b(\lambda)}\right]\left[1 - m_2 e^{-m_3 a(\lambda)}\right] b_b(\lambda)$$
$$= (1 + 0.005\theta_s)a(\lambda) + 4.259\left[1 - 0.265\frac{b_{bw}(\lambda)}{b_b(\lambda)}\right]\left[1 - 0.52 e^{-10.8 a(\lambda)}\right] b_b(\lambda)$$

θ_s is the solar zenith angle in degrees

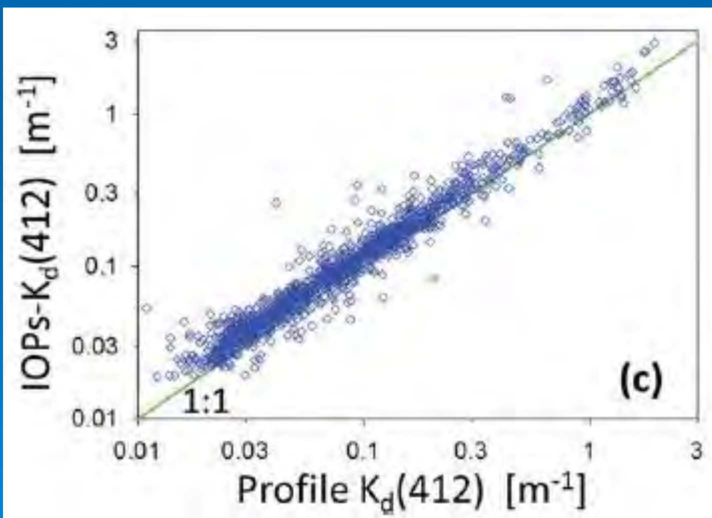
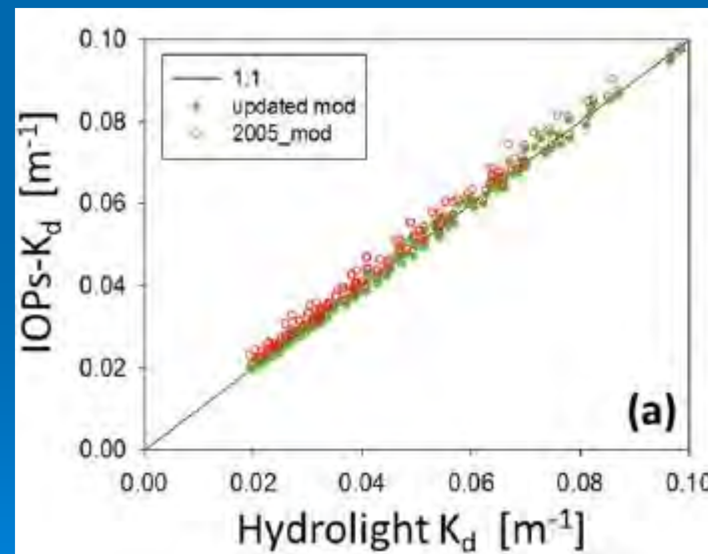


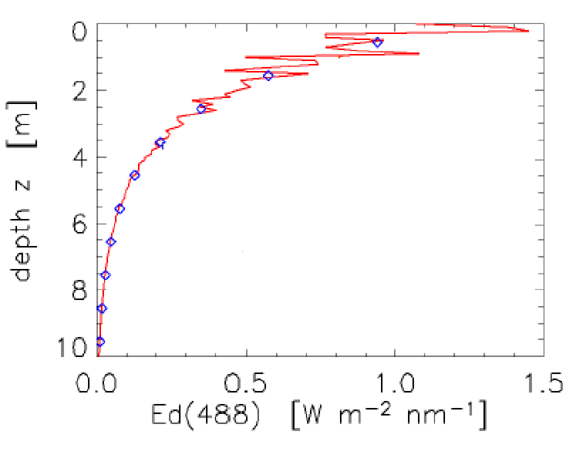
Figure 5. Comparison between retrieved K_d and
(b–d) The NOMAD data set, but limited to the IOP



Something to Think About

- Suppose you measure $E_d(z)$
- 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
- 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 $E_d(0)$ likely to be?



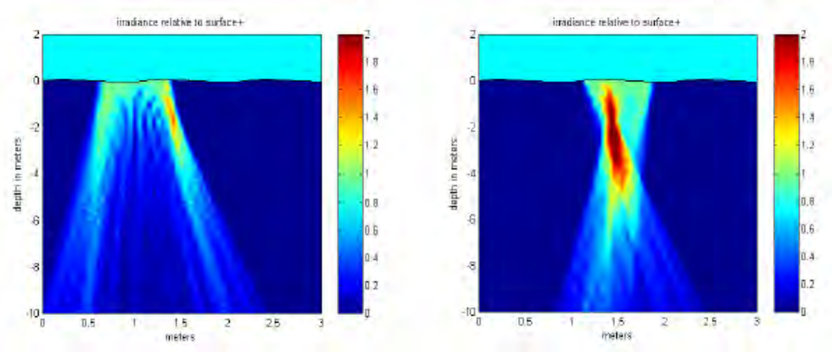
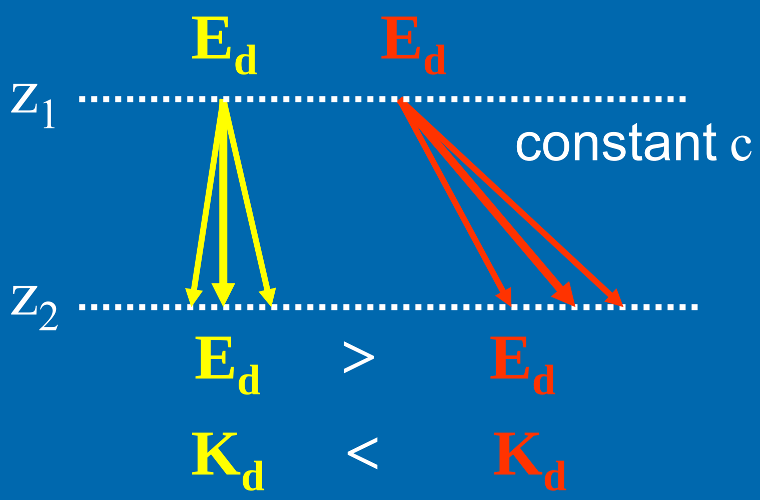
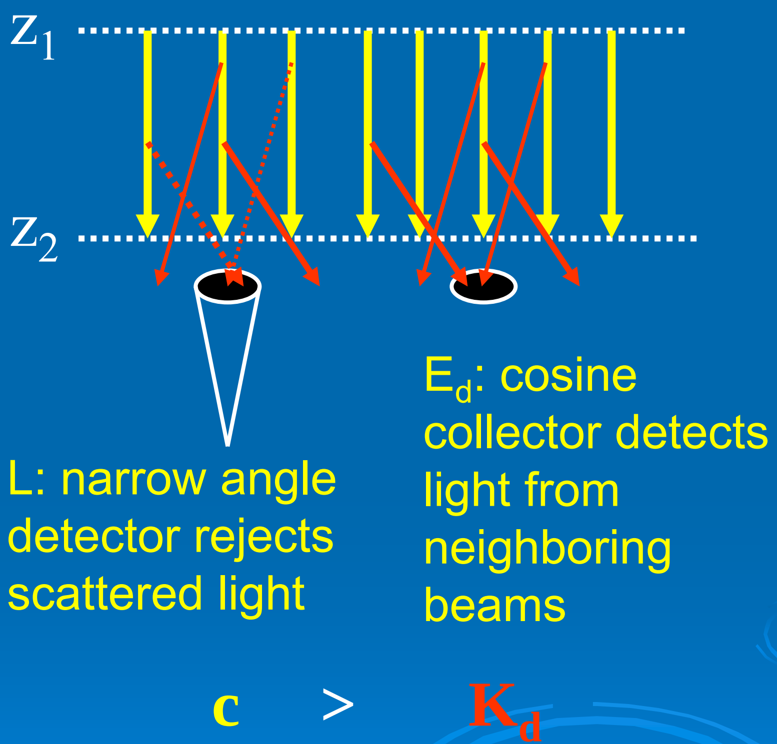


Fig. 2. An example of calculated irradiance relative to the irradiance just above the sea surface. The absorption coefficient is 0.07 m^{-1} and the scattering coefficient is 0.39 m^{-1}; the wave surface is wavelength L=[1.1 0.55 0.05] m and amplitude A= [0.05 0.01 0.0008]m. The beam is 0.8 m wide and 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 $c \neq$ diffuse attenuation K





Virtues and Vices of K's

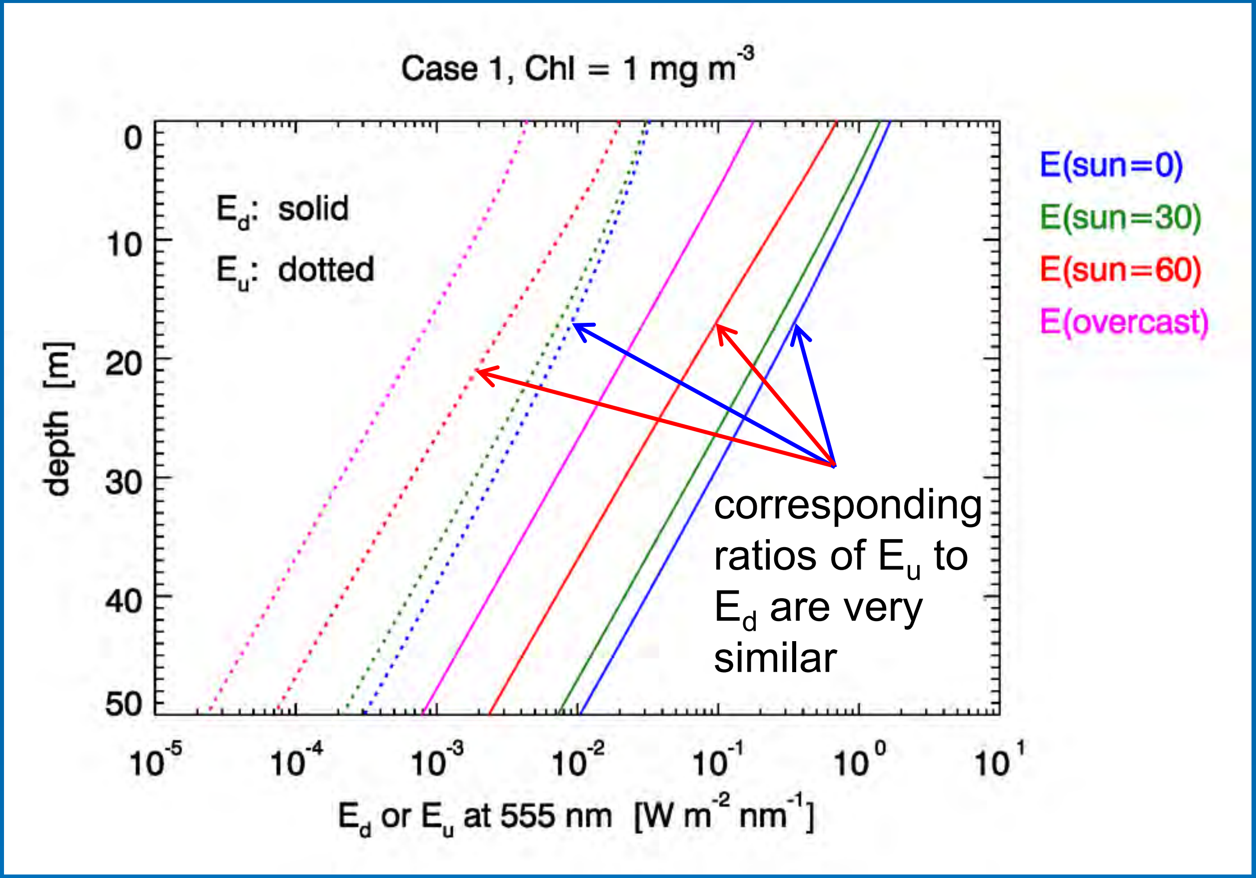
Virtues:

- 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 (e.g., 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



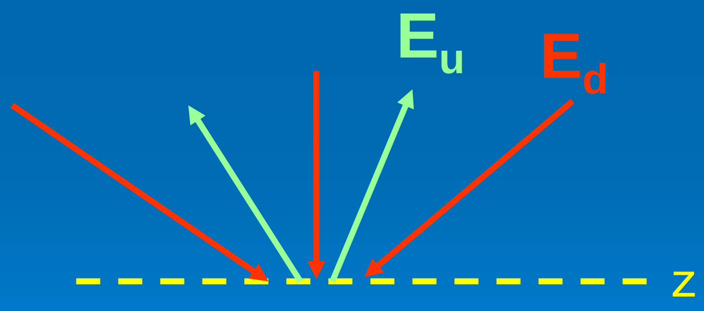
Magnitude changes are due to incident lighting (sun angle and sky condition); ratio of E_u/E_d is determined by water IOPs.

This suggests trying...

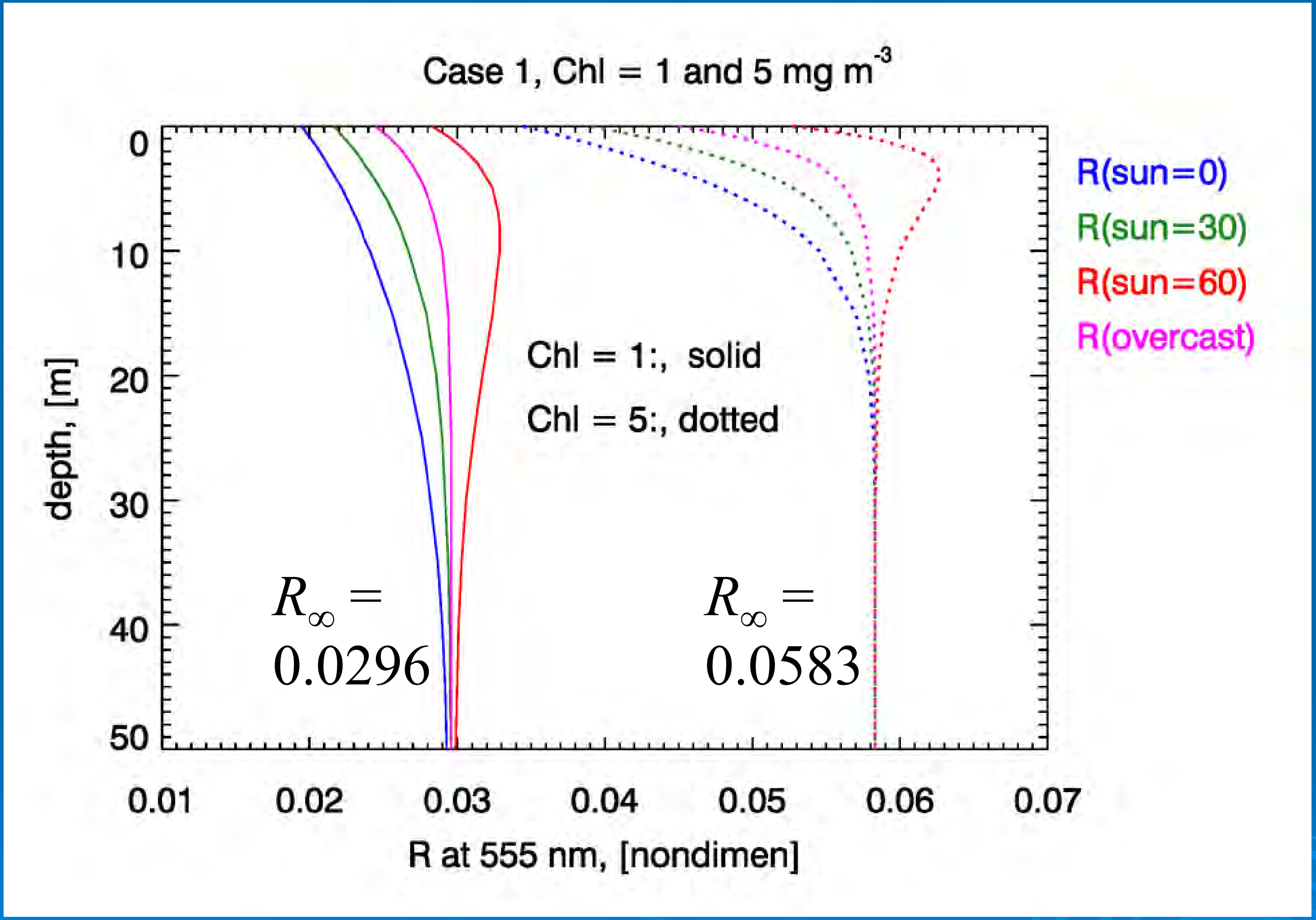
…the ratio of upwelling plane irradiance E_u to downwelling plane irradiance E_d as an AOP.

This is the irradiance reflectance *R*:

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



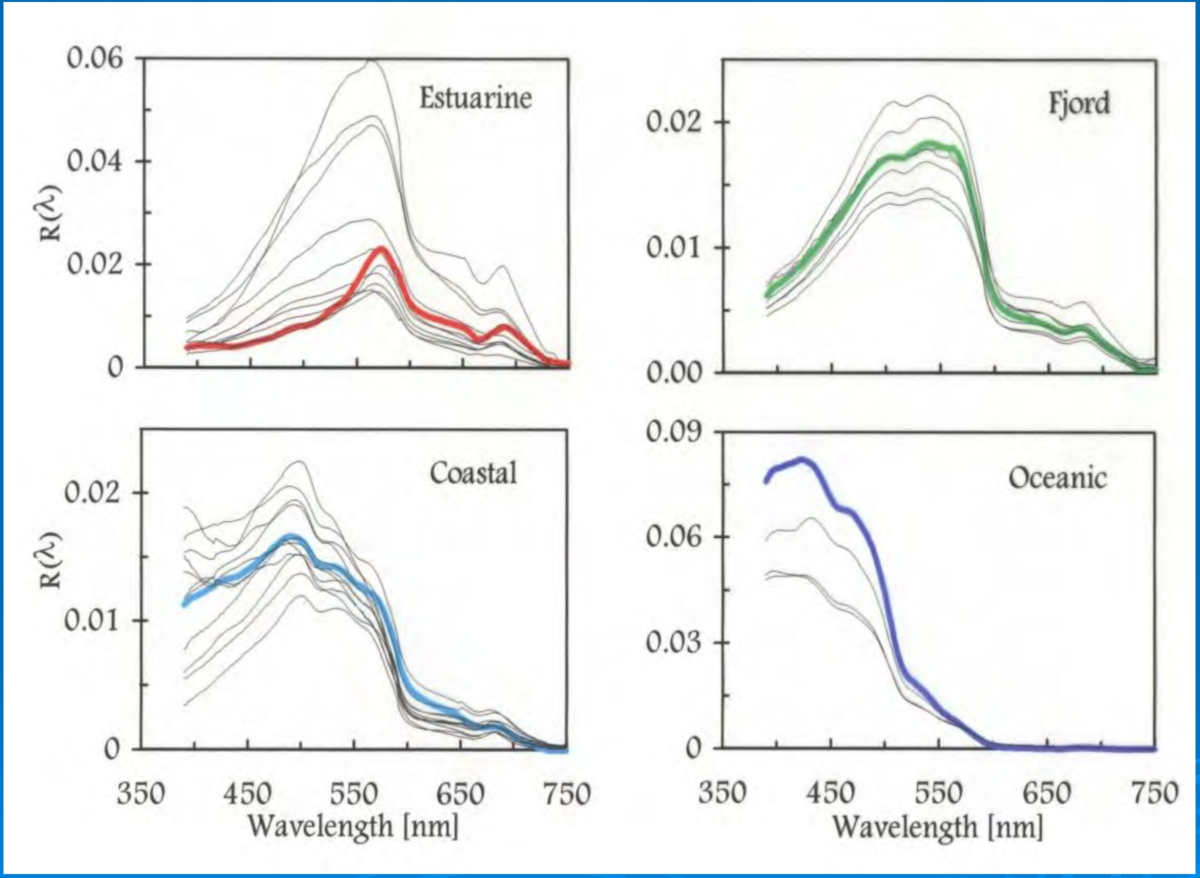
$R = E_u/E_d$



For given IOPs, the *R*'s all approach the same value as you go deeper: the asymptotic reflectance, R_∞, which is an IOP.

Examples of $R = E_u/E_d$

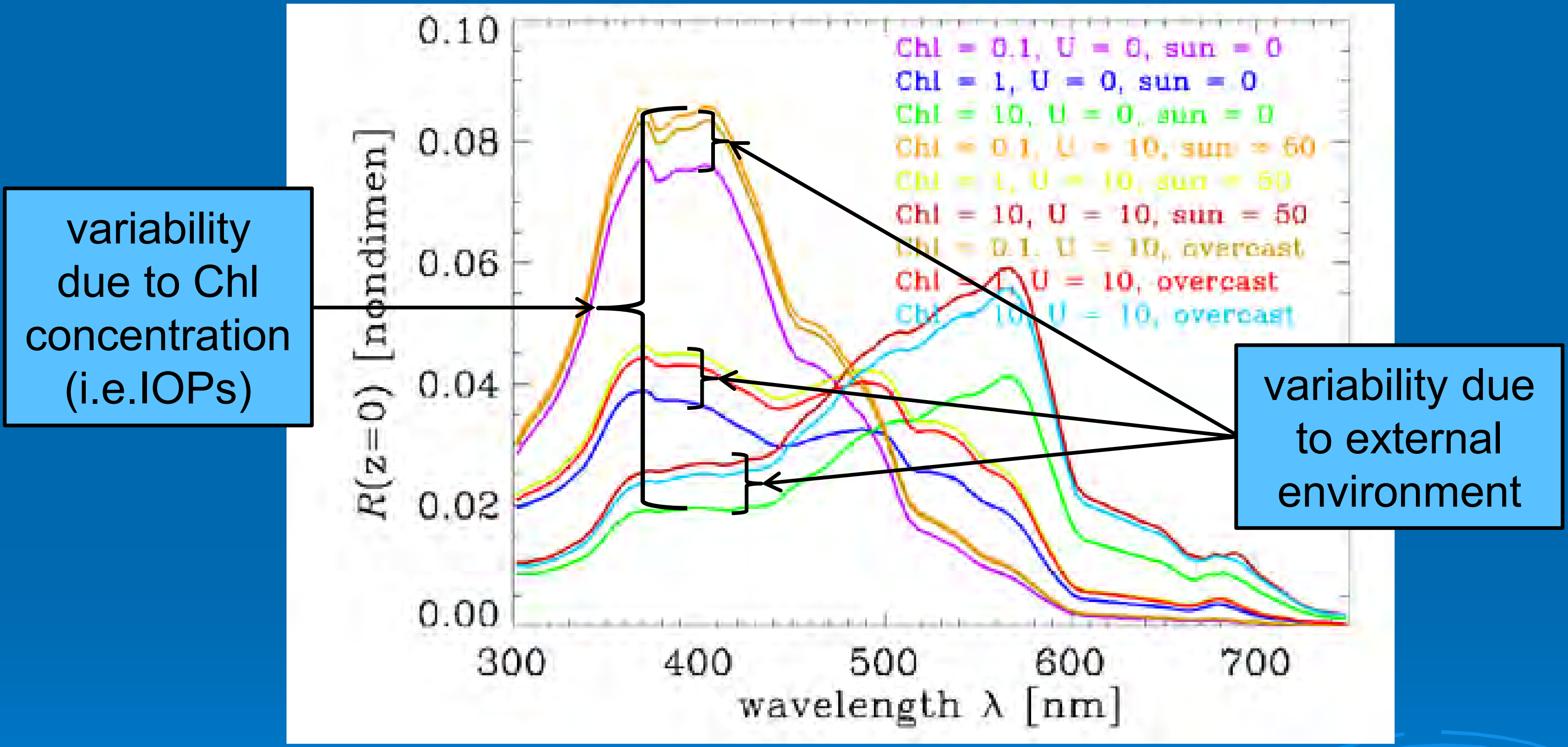
measurements from various ocean waters



Roesler and Perry 1995

$R = E_u/E_d$

HydroLight runs: Chl = 0.1,1, 10 mg/m^3
Sun at 0 and 50 deg in clear sky, and overcast

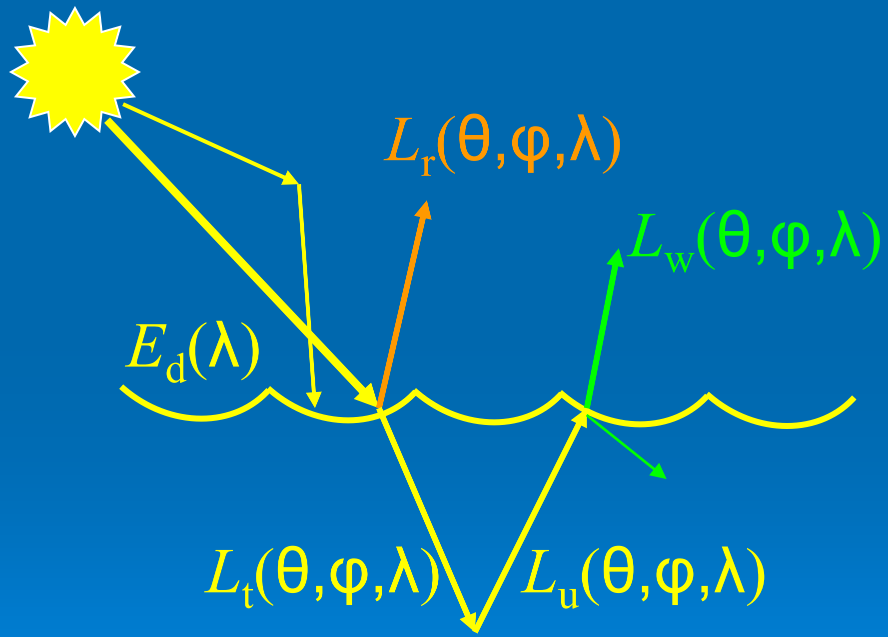


R depends weakly on the external environment and strongly on the water IOPs

The Water-leaving Radiance, L_w

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

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

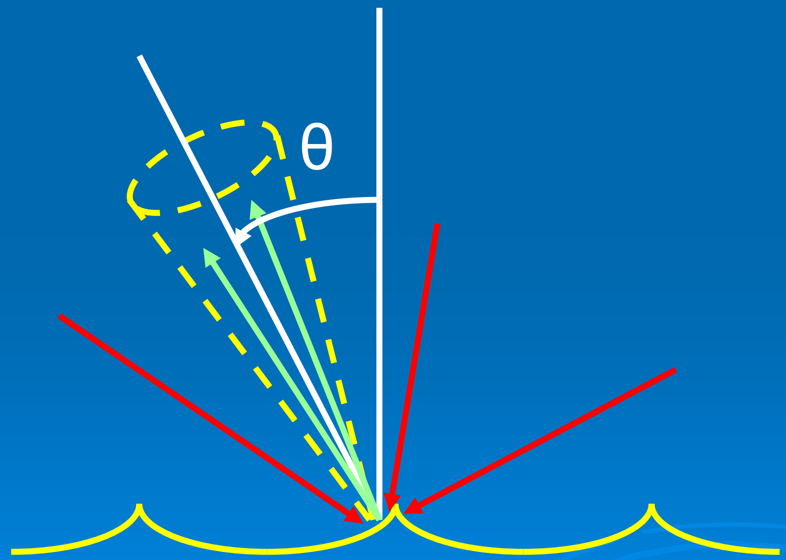


An instrument measures L_u (in air), but L_w is what tells us what is going on in the water. However, it isn't easy to figure out how much of L_u is due to L_w.

Remote-sensing Reflectance R_{rs}

$R_{rs}(\theta,\varphi,\lambda) = \dfrac{\text{upwelling water-leaving radiance}}{\text{downwelling plane irradiance}}$

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



The fundamental quantity used in ocean color remote sensing

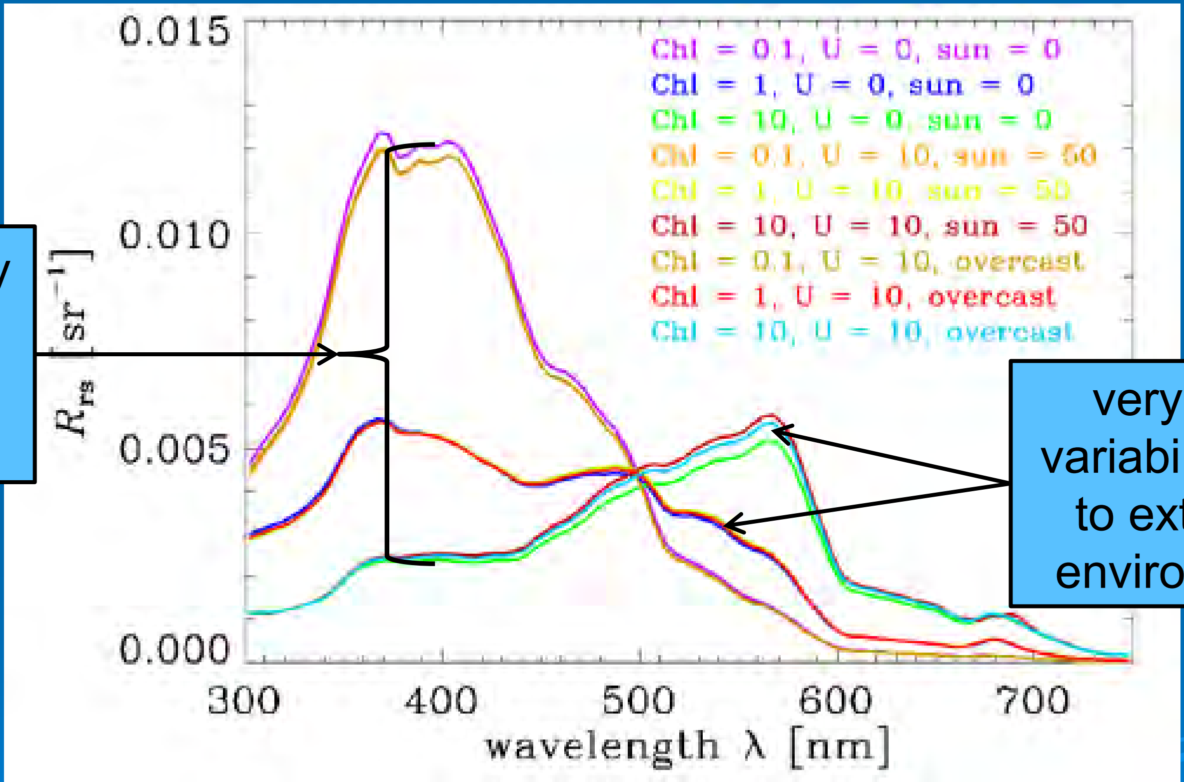
Often work with the nadir-viewing R_{rs}, i.e.,with the radiance that is heading straight up from the sea surface (θ = 0)

sea surface

Example R_{rs}

HydroLight runs: Chl = 0.1,1, 10 mg Chl/m^3
Sun at 0 and 50 deg in clear sky, overcast sky

large variability due to Chl concentration (i.e.IOPs)

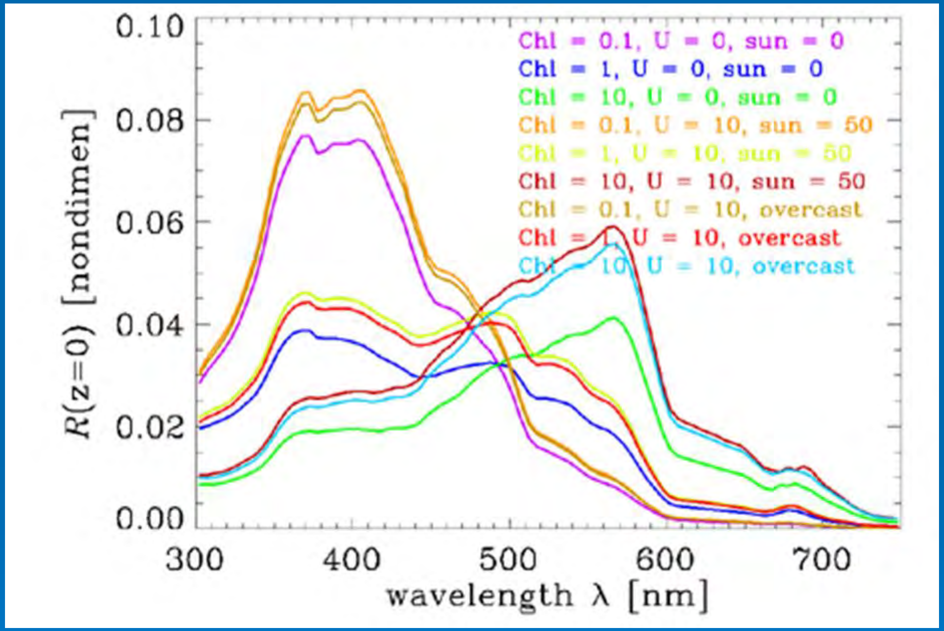


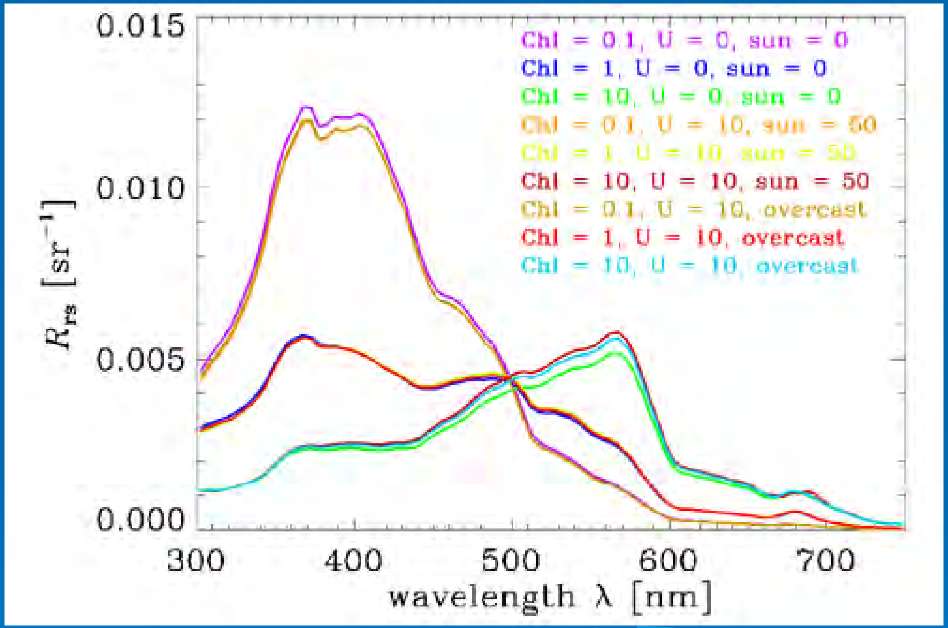
very little variability due to external environment

R_{rs} shows almost no dependence on sky conditions and strong dependence on the water IOPs—a very good AOP

To Reiterate: R vs R_{rs}

R_{rs} shows less dependence on sky conditions that does R, and both show the same dependence on IOPs. R_{rs} is a better AOP than R.

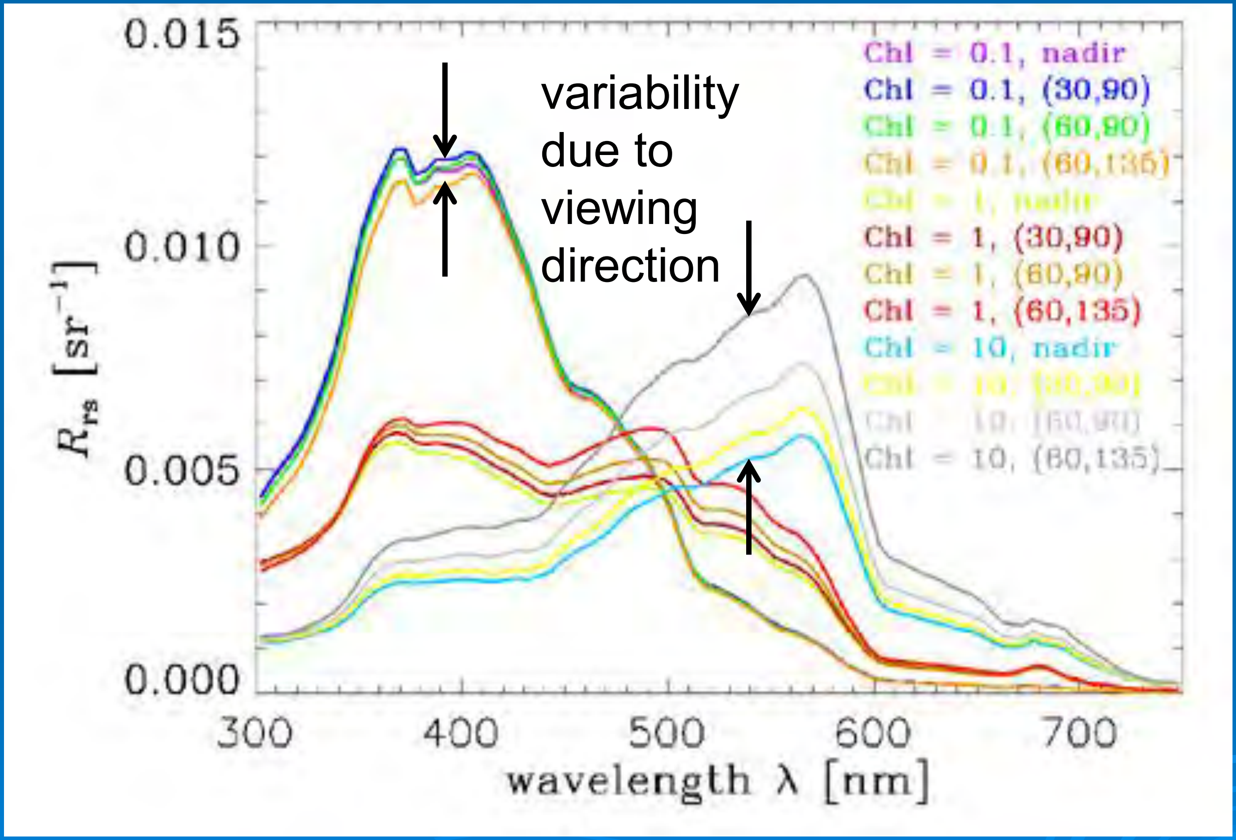




The R_{rs} plots are for nadir-viewing direction.

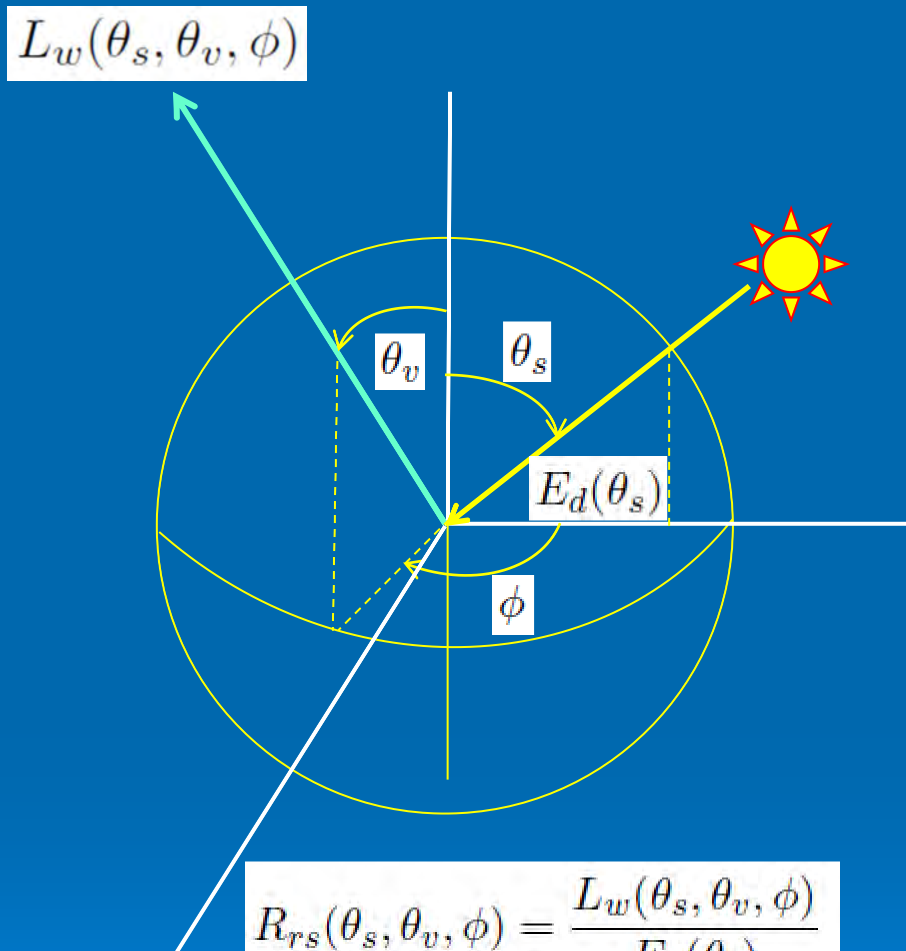
Example R_{rs}

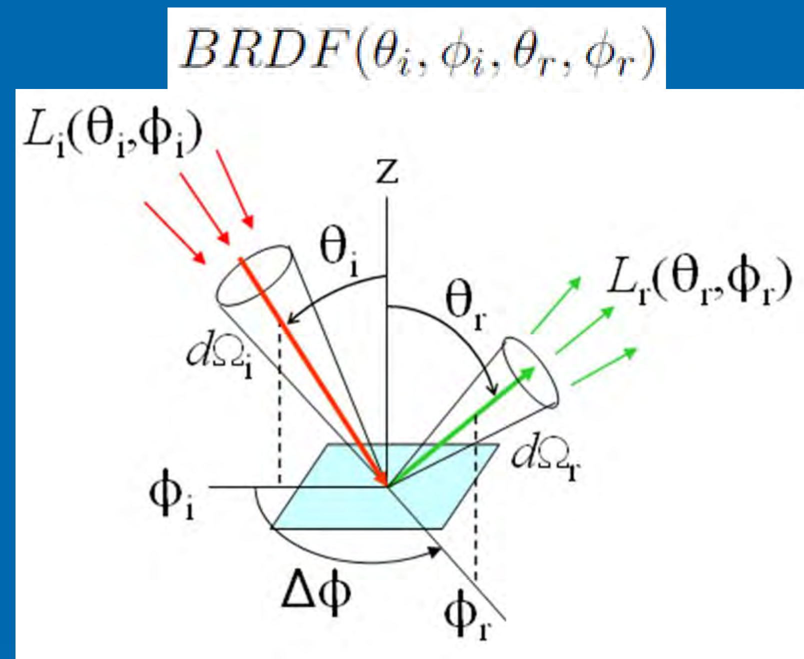
HydroLight runs: Chl = 0.1,1, 10 mg Chl/m^3
Sun at 50 deg in clear sky
R_{rs} for nadir vs off-nadir viewing directions



R_{rs} shows dependence on viewing direction but stronger dependence on the water IOPs—still a good AOP, but could be better…

The "BRDF Effect"





$$R_{rs}(\theta_s, \theta_v, \phi) = \frac{L_w(\theta_s, \theta_v, \phi)}{E_d(\theta_s)}$$

E_d also depends on atmospheric transmittance, distance from Earth to the Sun, and how much gets into the water depends on surface waves (wind speed).

Normalized Reflectance $[\rho]_N$

R_{rs} shows some variability with external environmental conditions and viewing direction. It would be nice to remove those effects.

The *normalized water-leaving radiance* is the water-leaving "*radiance that would be measured by a nadir-viewing instrument, if the Sun were at the zenith in the absence of any atmospheric loss, and when the Earth is at its mean distance from the Sun.*" (Morel et al., 1996, page 4852).

(Note: "absence of any atmospheric loss", not "absence of any atmosphere")

Let $L_w(\theta_s, \theta_v, \phi)$ be the water-leaving radiance for a given sun zenith angle and viewing direction (obtained, perhaps, from atmospheric correction of a TOA radiance). Then the "normalized water-leaving radiance" is

$$[L_w(\theta_v, \phi)]_N \equiv \left(\frac{R}{R_o}\right)^2 \frac{L_w(\theta_s, \theta_v, \phi)}{\cos\theta_s \, t(\theta_s)}$$

$[L_w(\theta_v, \phi)]_N$ still depends on viewing direction

Normalized Reflectance $[\rho]_N$

Morel et al. (2002) developed correction factors that account for surface roughness and the "BRDF effect" of atmospheric conditions, water IOPs, and sun and viewing direction:

$$[L_w]_N^{ex} \equiv [L_w(\theta_v,\phi)]_N \frac{\Re_o(W)}{\Re(\theta_v',W)} \frac{f_o(\mathrm{ATM},W,\mathrm{IOP})}{Q_o(\mathrm{ATM},W,\mathrm{IOP})} \left[\frac{f(\theta_s,\mathrm{ATM},W,\mathrm{IOP})}{Q(\theta_s,\theta_v',\phi,\mathrm{ATM},W,\mathrm{IOP})}\right]^{-1}$$

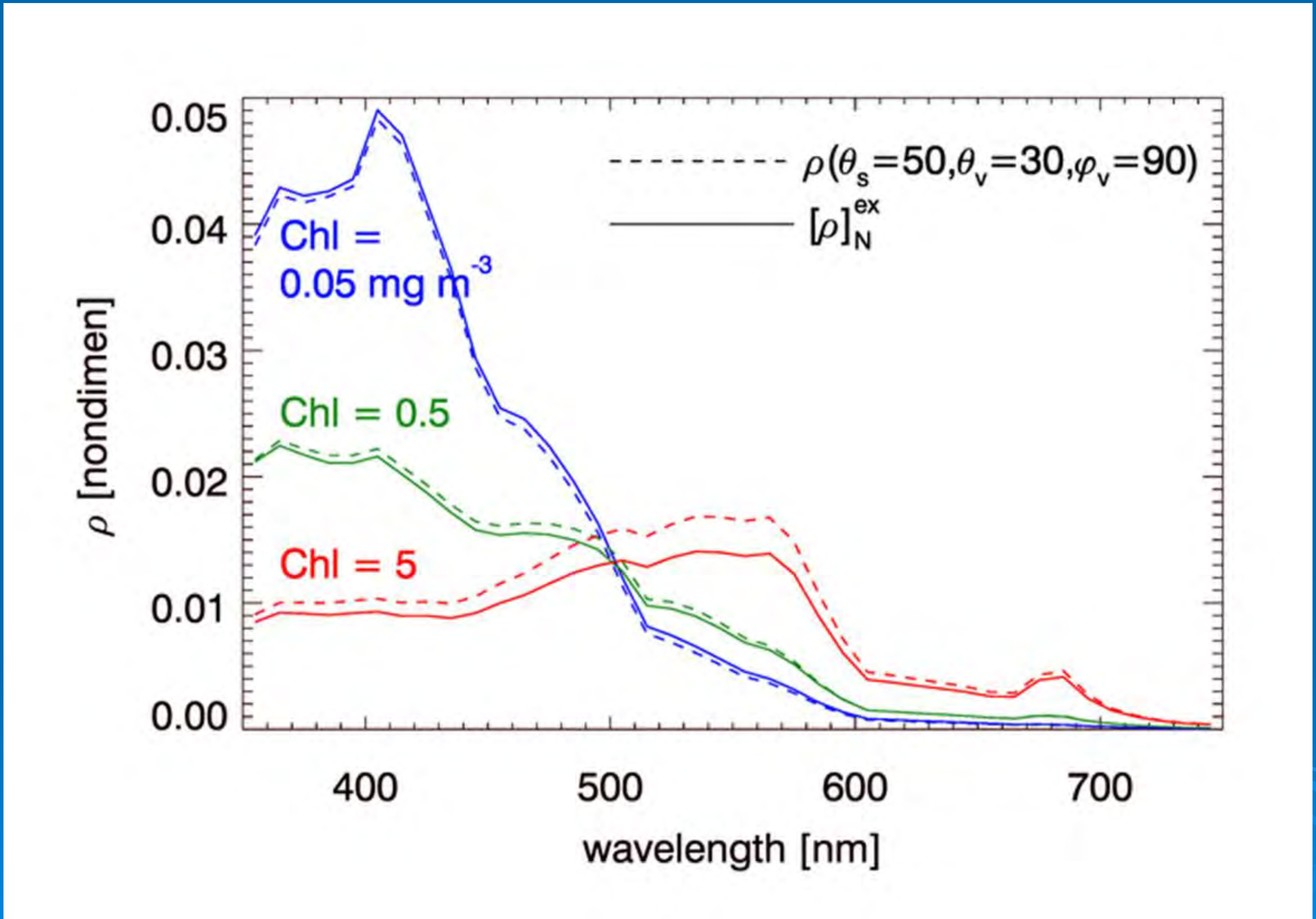
Tabulated factors that depend on atmospheric conditions (ATM), wind speed (W), water IOPs (Chl conc), sun and viewing directions.

$$[\rho_w]_N^{ex} \equiv \frac{\pi}{F_o}[L_w]_N^{ex}$$

is the nondimensional "*exact normalized water-leaving reflectance*". F_o is the extra-terrestrial solar irradiance at the mean Earth-Sun distance.

Note: (1) Everything here depends on wavelength.
(2) The Morel BRDF correction factors were developed using a Case 1 IOP model, so they may not give good results for Case 2 water.
(3) The correction factors require knowing the Chl concentration.
(4) The BRDF correction factors are tabulated only for certain wavelengths as needed for SeaWiFS, MODIS, VIIRS.

Normalized vs Unnormalized Reflectances



Normalized Reflectance $[\rho]_N$

The exact normalized water-leaving reflectance $[\rho_w]_N^{ex}$ is now the standard AOP used for comparisons of measured and remotely sensed radiances.

To compute $[\rho_w]_N^{ex}$ in HydroLight, put the sun at the zenith; then π times the nadir-viewing R_{rs} is $[\rho_w]_N^{ex}$:

$$[\rho_w]_N^{ex} = \pi R_{\rm rs}(\text{HydroLight}; \theta_s = 0, \theta_v = 0)$$

Note: HydroLight works for any IOPs, so HydroLight can give you $[\rho_w]_N^{ex}$ for any IOPs, any bottom conditions, or any wavelength

See the Ocean Optics Web Book page on Normalized Reflectances or OOB Section 10.3 for a full discussion of $[\rho_w]_N^{ex}$

There will be more of this stuff in the Atmospheric Correction lecture.

Average or Mean Cosines

The average or mean cosines give the average of the cos⍰ as weighted by the radiance distribution. This tells you something about the directional pattern of the radiance. For the downwelling radiance we have

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

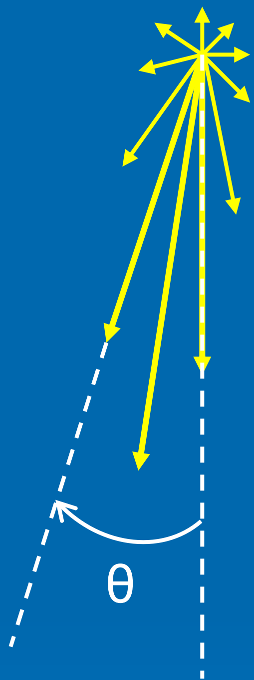
Likewise, for the upwelling radiance, $\bar{\mu}_u = \frac{E_u}{E_{ou}}$

For the entire radiance distribution,

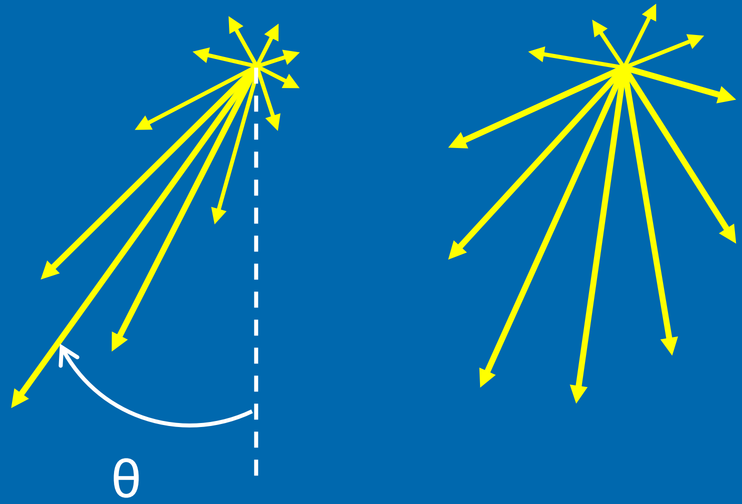
$$\bar{\mu} = \frac{\int_0^{2\pi} \int_0^{\pi} L(\theta,\phi)\ \cos\theta\ \sin\theta\, d\theta\, d\phi}{\int_0^{2\pi} \int_0^{\pi} L(\theta,\phi)\ \sin\theta\, d\theta\, d\phi} = \frac{E_d - E_u}{E_o}$$

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

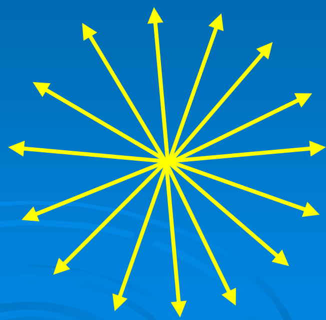
Mean Cosines



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

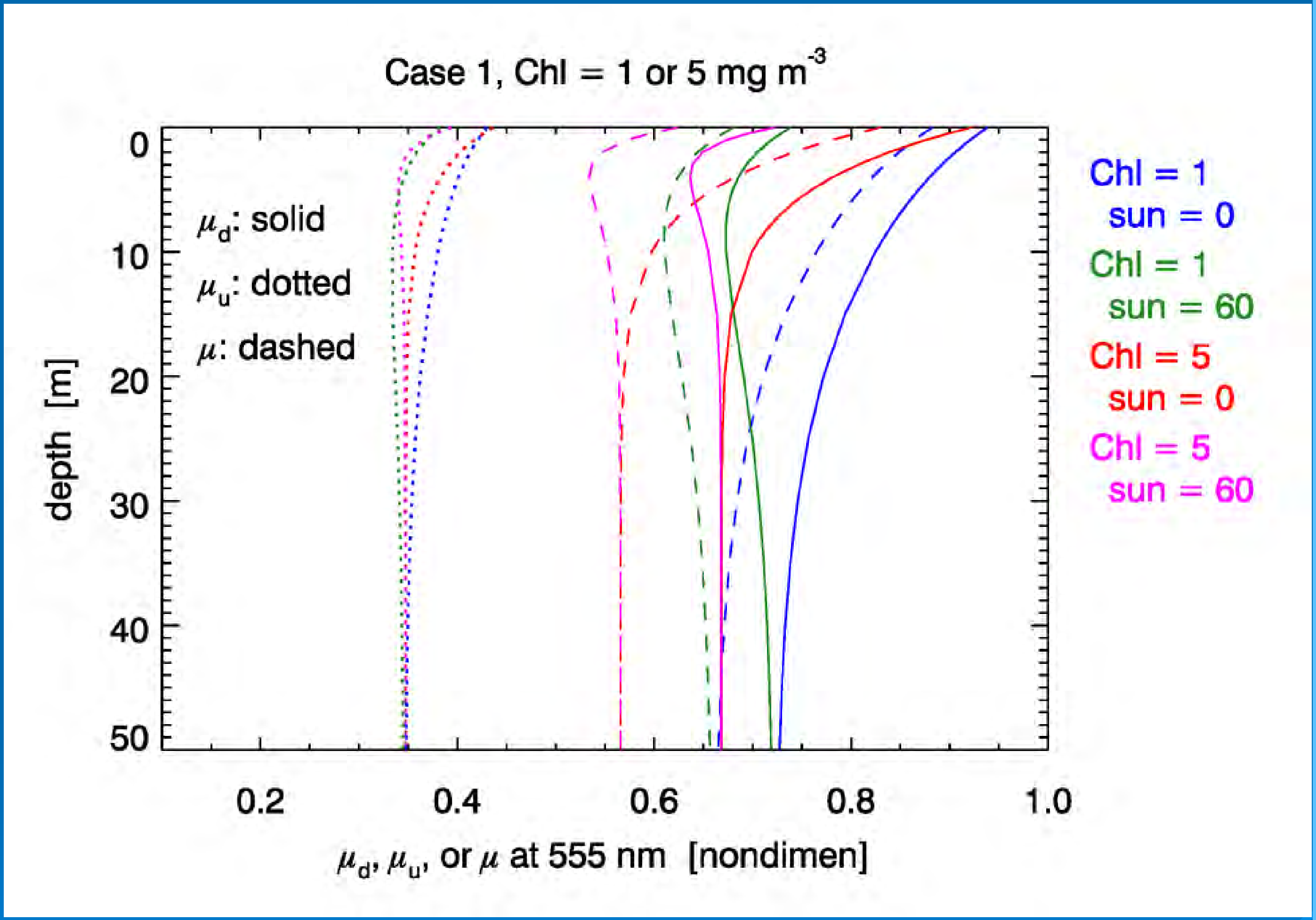


most radiance heading at a large angle, or a diffuse radiance: large average θ, small μ_d



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

Mean Cosines



values at 555 nm:

Albedo of single scattering $\omega_o = b/c$:

$\omega_o(\text{Chl}=1) = 0.85$
$\omega_o(\text{Chl}=5) = 0.93$

Asymptotic values:

Chl = 1:
$\mu_d(\infty) = 0.7222$
$\mu_u(\infty) = 0.3436$
$\mu(\infty) = 0.6600$

Chl = 5:
$\mu_d(\infty) = 0.6682$
$\mu_u(\infty) = 0.3473$
$\mu(\infty) = 0.5658$

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

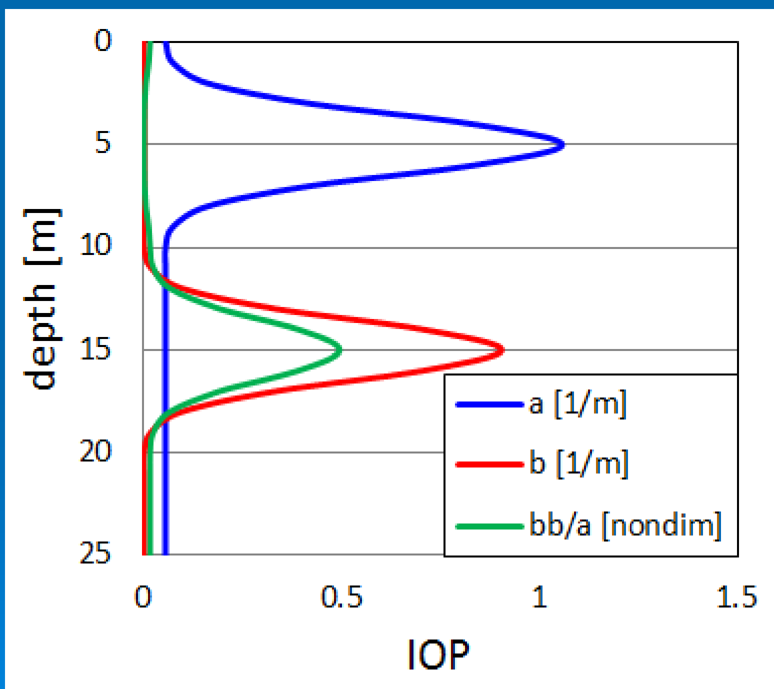
Kd and R Dependence on IOPs

To first order

$$K_d \propto \frac{a + b_b}{\cos \theta_{sw}}$$

$$R \propto \frac{b_b}{a}$$

HydroLight simulation with
pure sea water +
CDOM layer at 5m +
non-absorbing scattering layer at 15 m



What will K_d and
$R = E_u/E_d$ look like?

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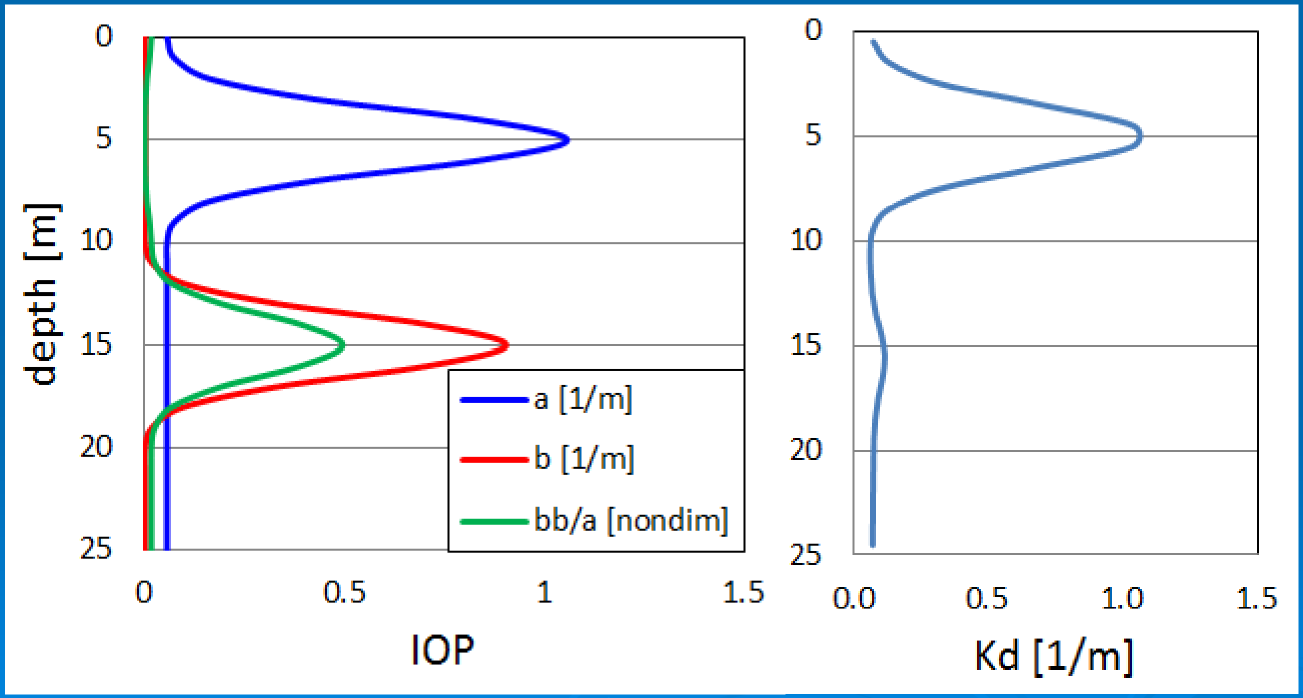

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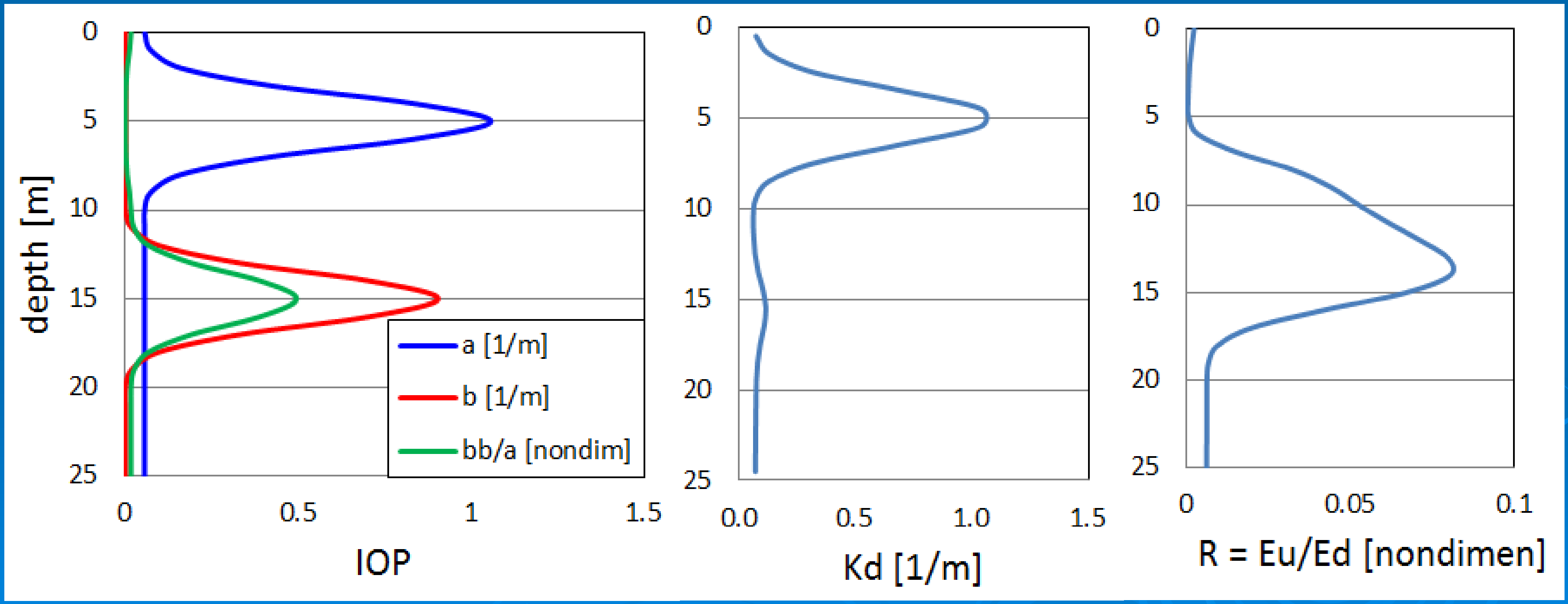
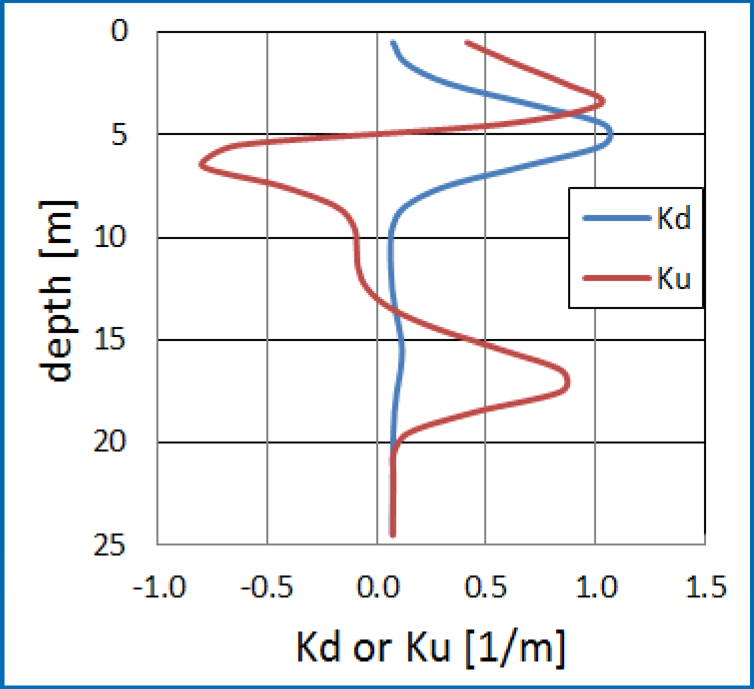
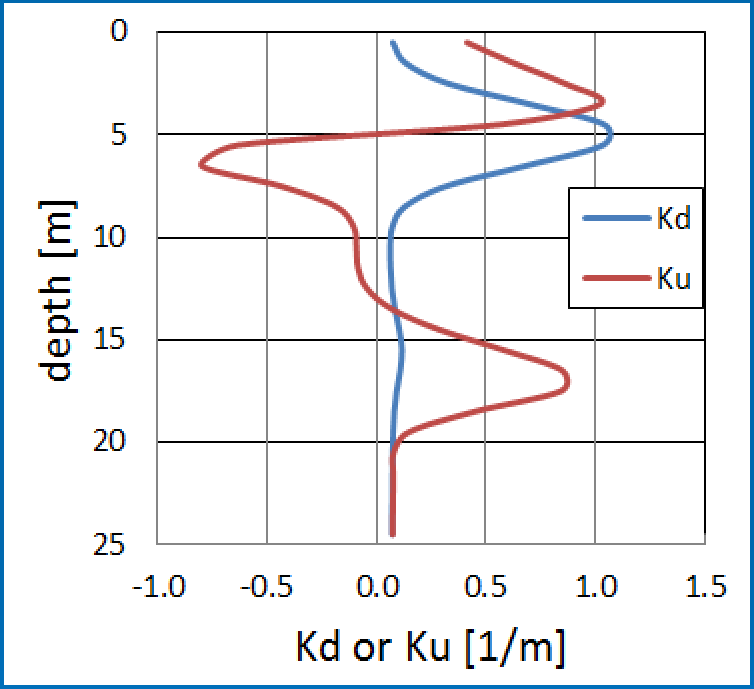
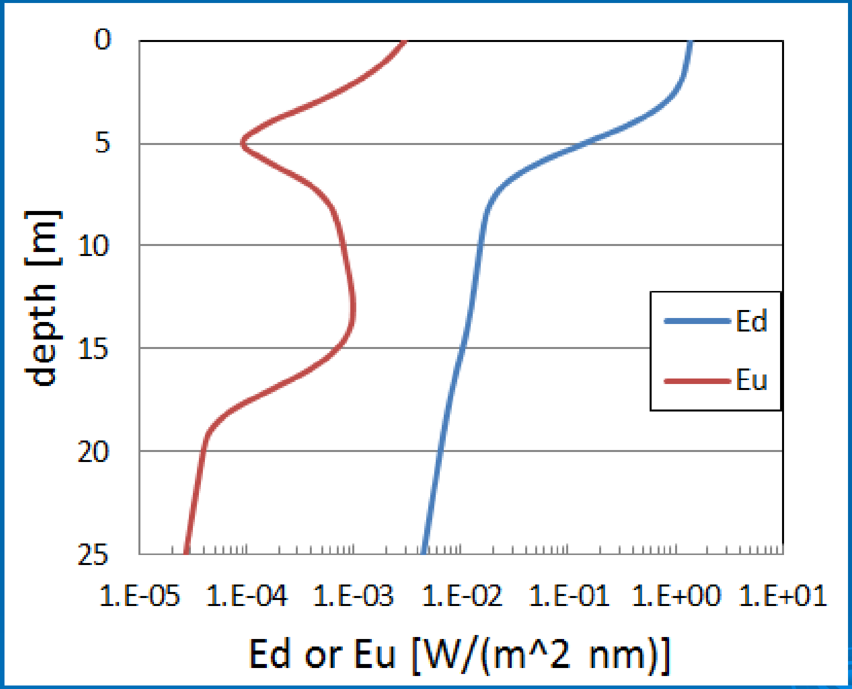


What Does a Negative K_u Mean?



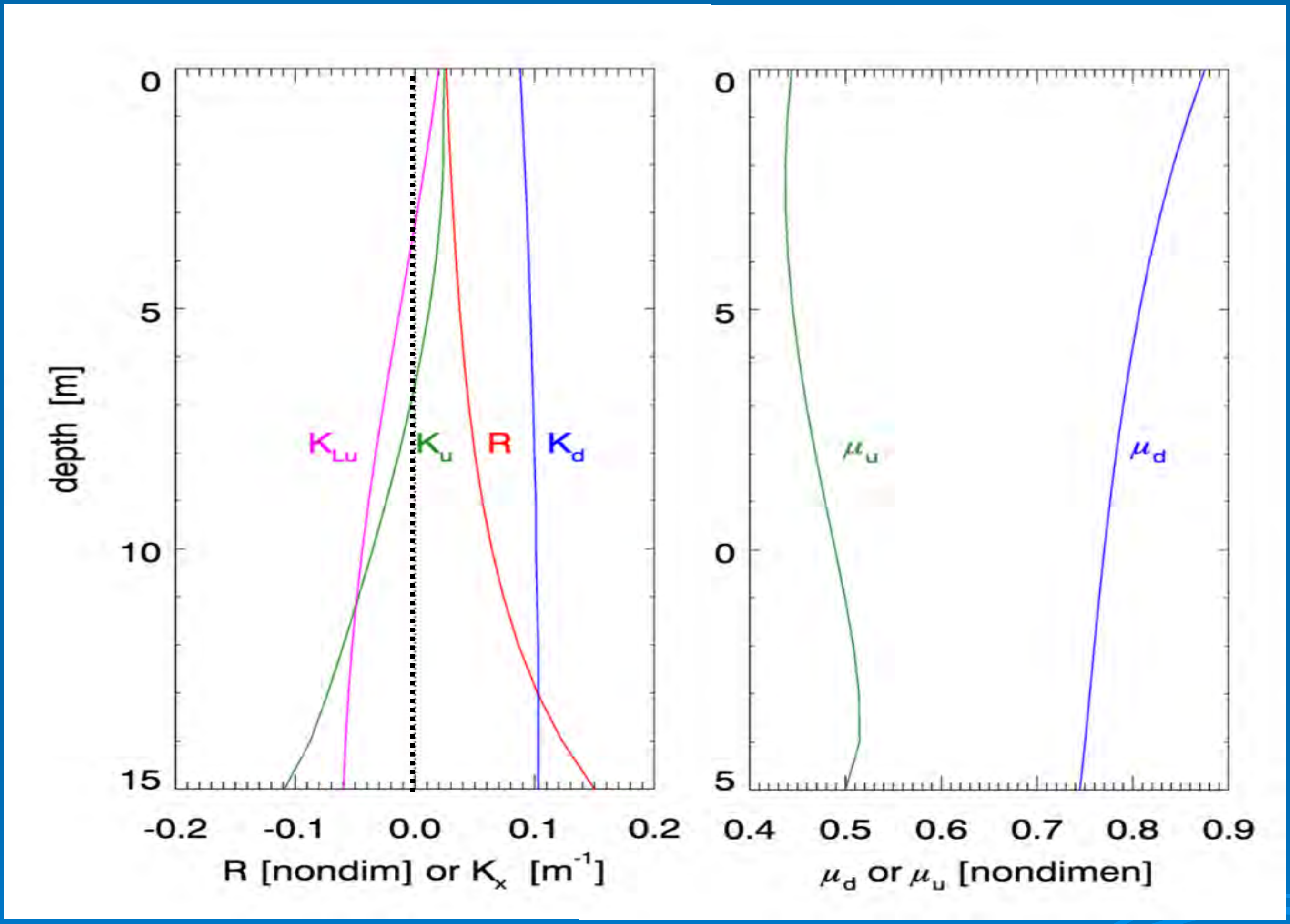
What Does a Negative K_u Mean?





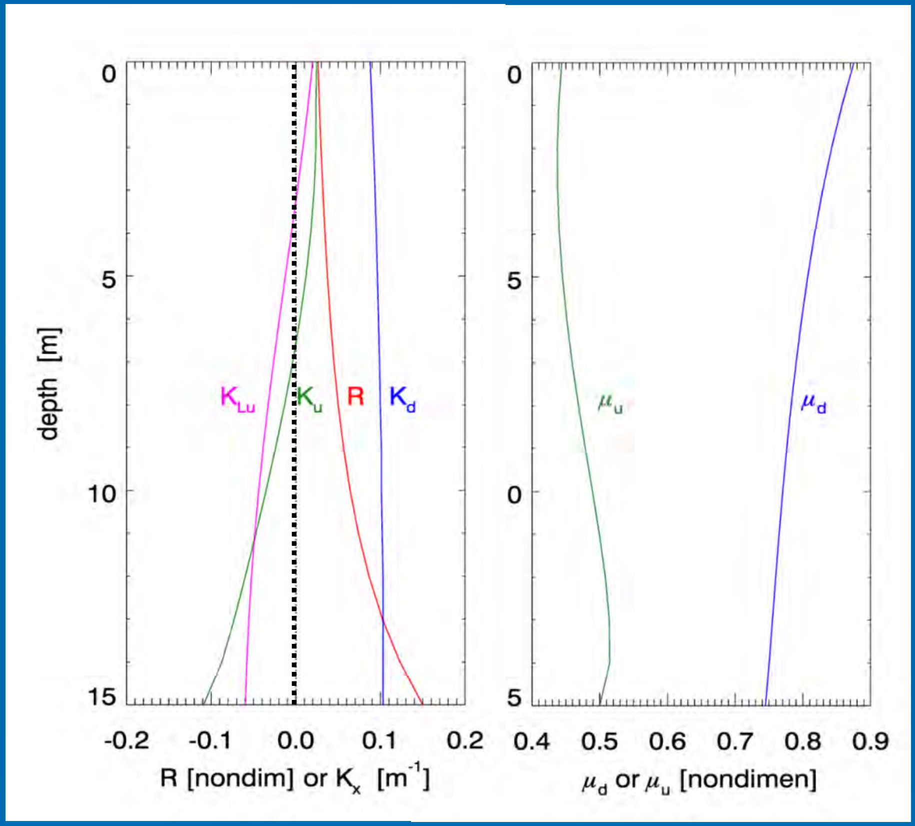
Why is E_u increasing from 5 to 14 m?

Explain These AOPs

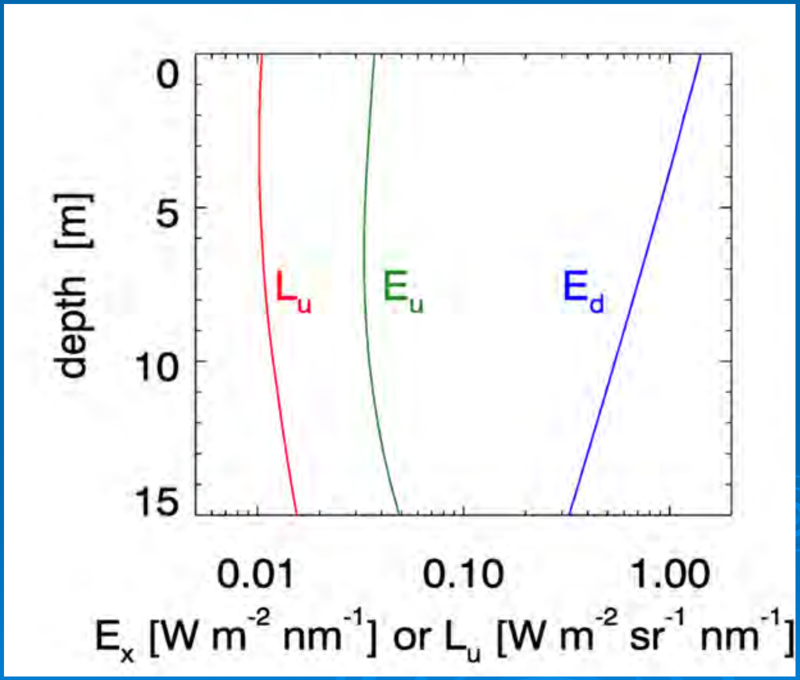


What does $\mu_u = 0.5$ say about the upwelling radiance distribution at 15 m?

Explain These AOPs



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

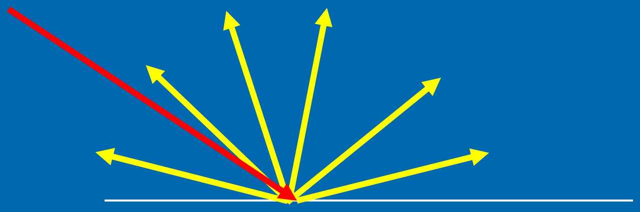


E_d decreases throughout the water column, but E_u and L_u increase with depth close to the 15% reflective bottom.

The Answer

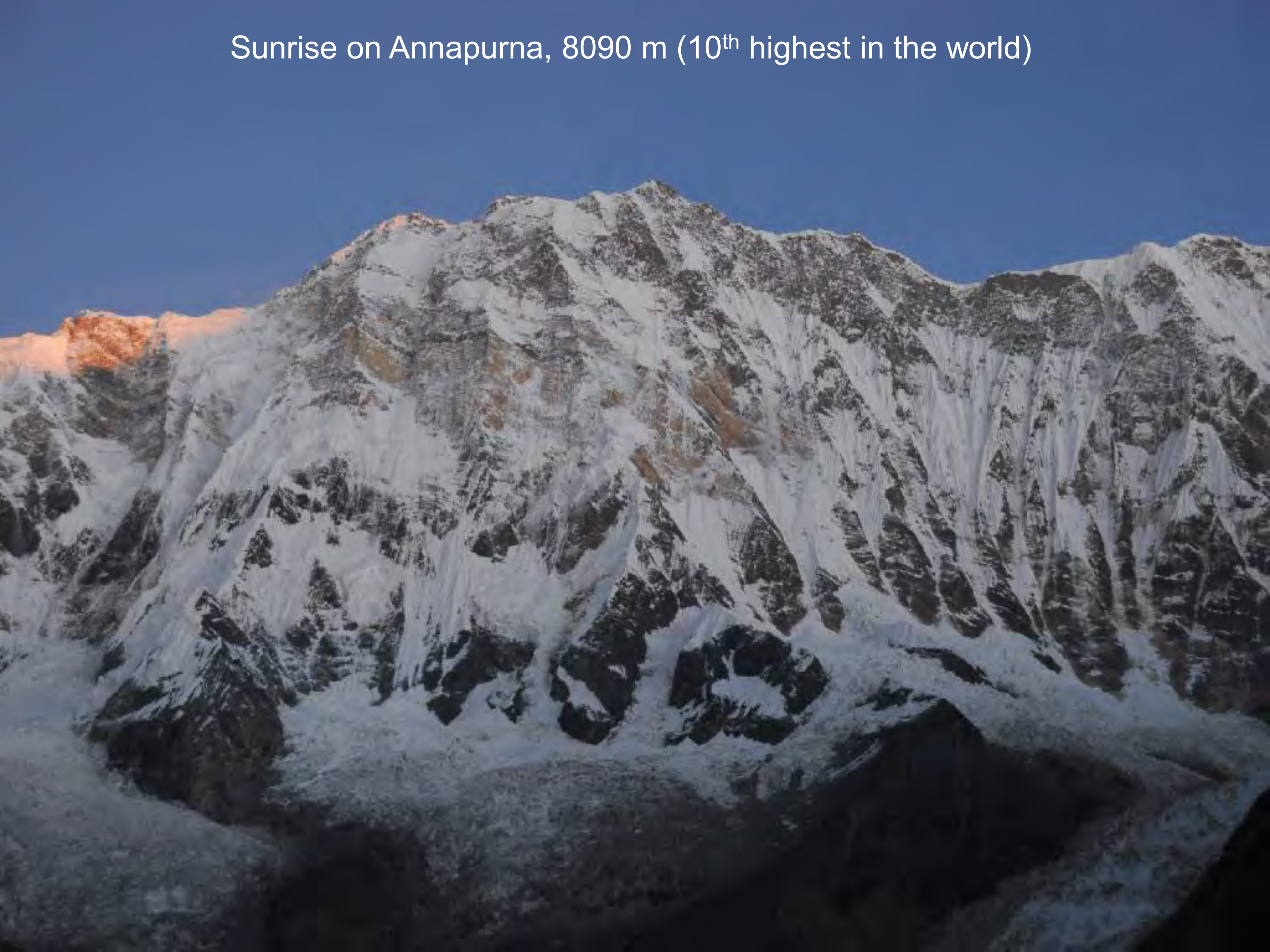
The water was homogeneous (Case 1, Chl = 1 mg/m^3), but there was a Lambertian bottom at 15 m, which had a reflectance of R_b = 0.15

Lambertian means the reflected radiance is the same in all directions
(L_u is isotropic; more on this in week 4)



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

Sunrise on Annapurna, 8090 m (10th highest in the world)



Rhino

Chitwan National Park, Nepal

2011