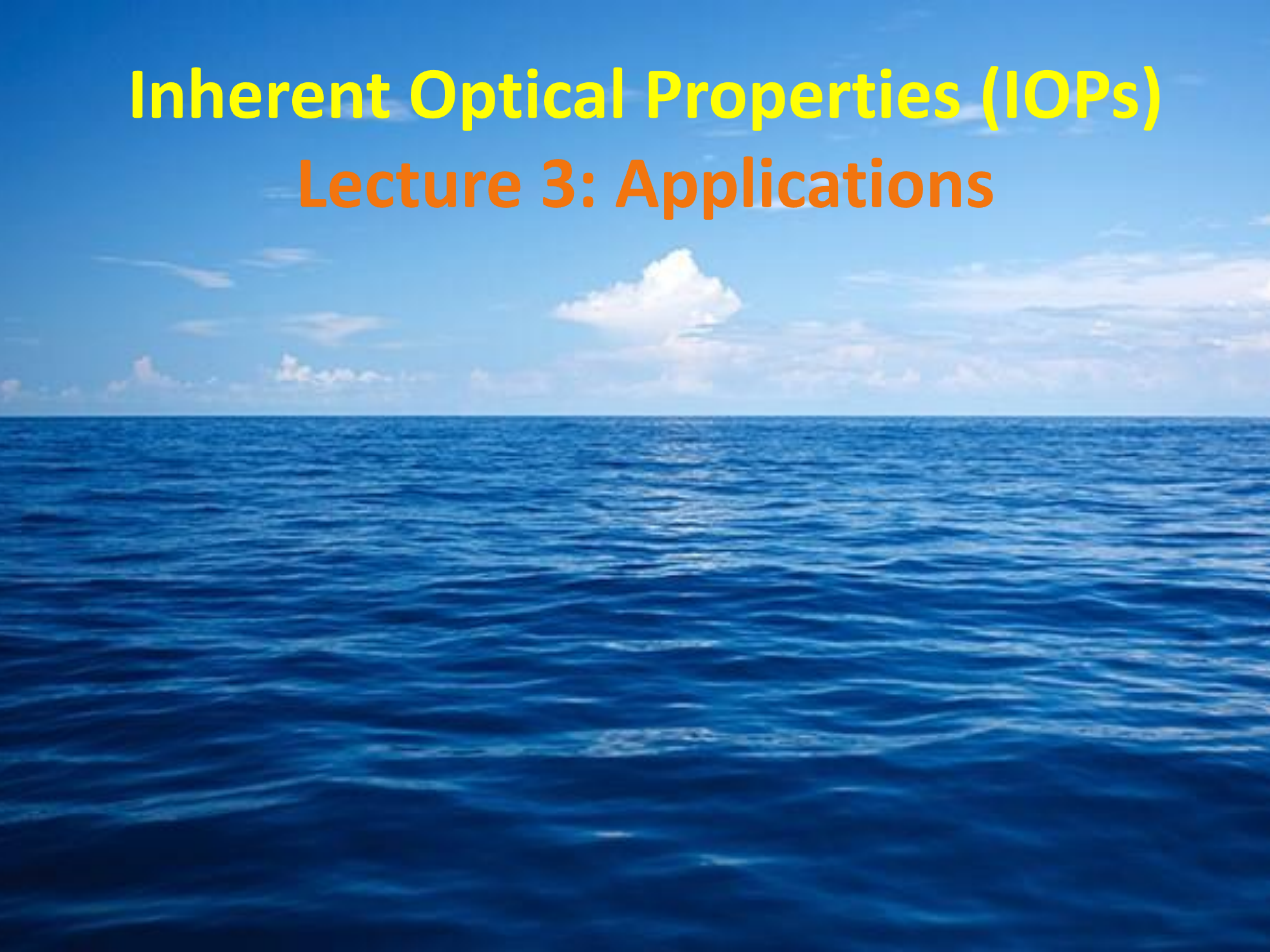
Inherent Optical Properties (IOPs)

Lecture 3: Applications



Rrs(λ)

IOPs

What to do with the retrieved IOPs?

→ Chlorophyll concentration: [Chl]

Examples: Carder et al (1999), GSM (2002), GIOP (2013)

$$R_{rs}(\lambda) = F(a(\lambda), b_b(\lambda))$$

$$a(\lambda) = a_w(\lambda) + M_1 \left\langle a_{ph}(\lambda) \right\rangle + M_2 \left\langle a_{dg}(\lambda) \right\rangle$$

$$\Updownarrow$$

$$a_{ph}(\lambda)$$

$\left\langle a_{ph}(\lambda) \right\rangle$: Varies with temperature in Carder et al (1999);
Fixed globally in GSM (Maritorena et al 2002)
Varies with [Chl] in GIOP (Werdell et al 2013)

Why chlorophyll concentration?

Estimate Primary Production (PP)

“One of the principal applications of satellite ocean color data is to derive net **primary production** (NPP).” --- McClain (Annu. Rev. Mar. Sci., 2009)

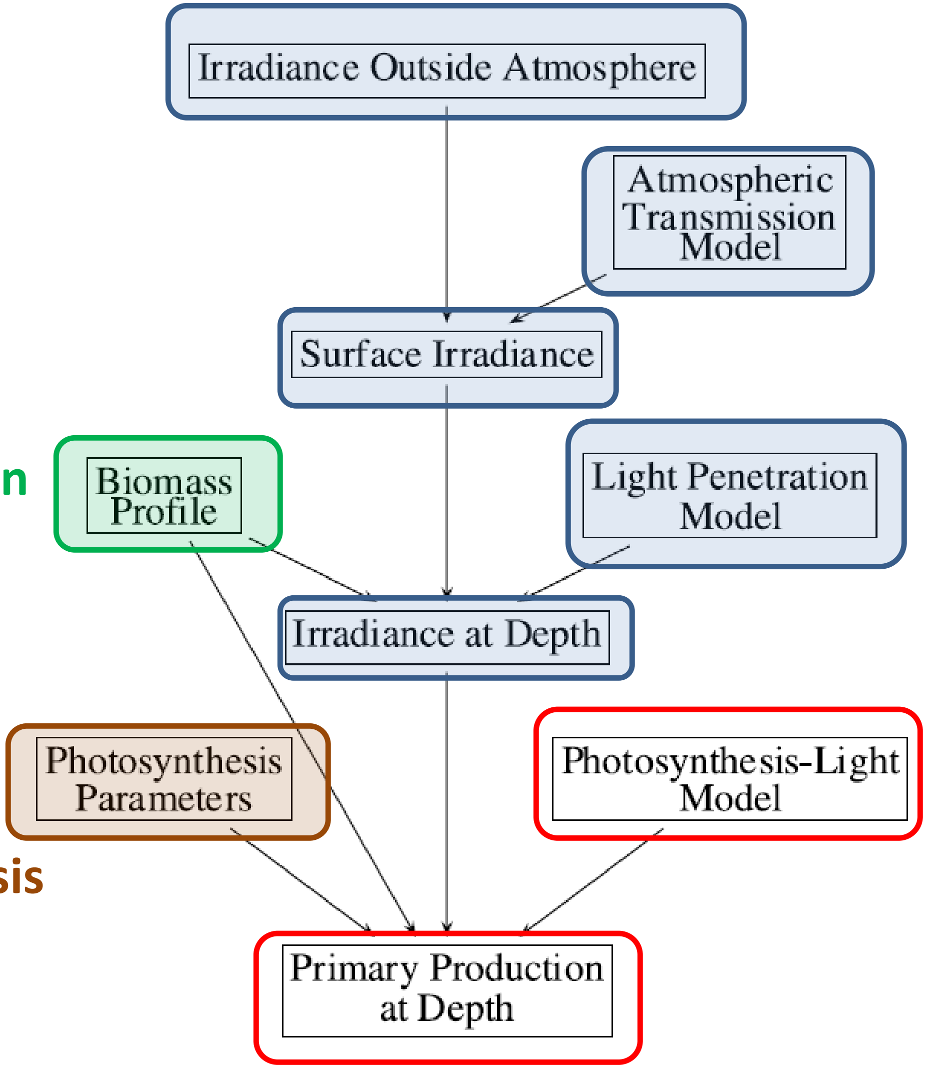
“On a global scale, marine phytoplankton consume fifty thousand million tones of carbon every year in a process referred to as **primary production.**” **-- IOCCG Report #2**

food chain; carbon cycle

Elements of photosynthesis:

CO2 + H2O + phytoplankton + light + nutrient → chemical energy

Components for PP quantification:



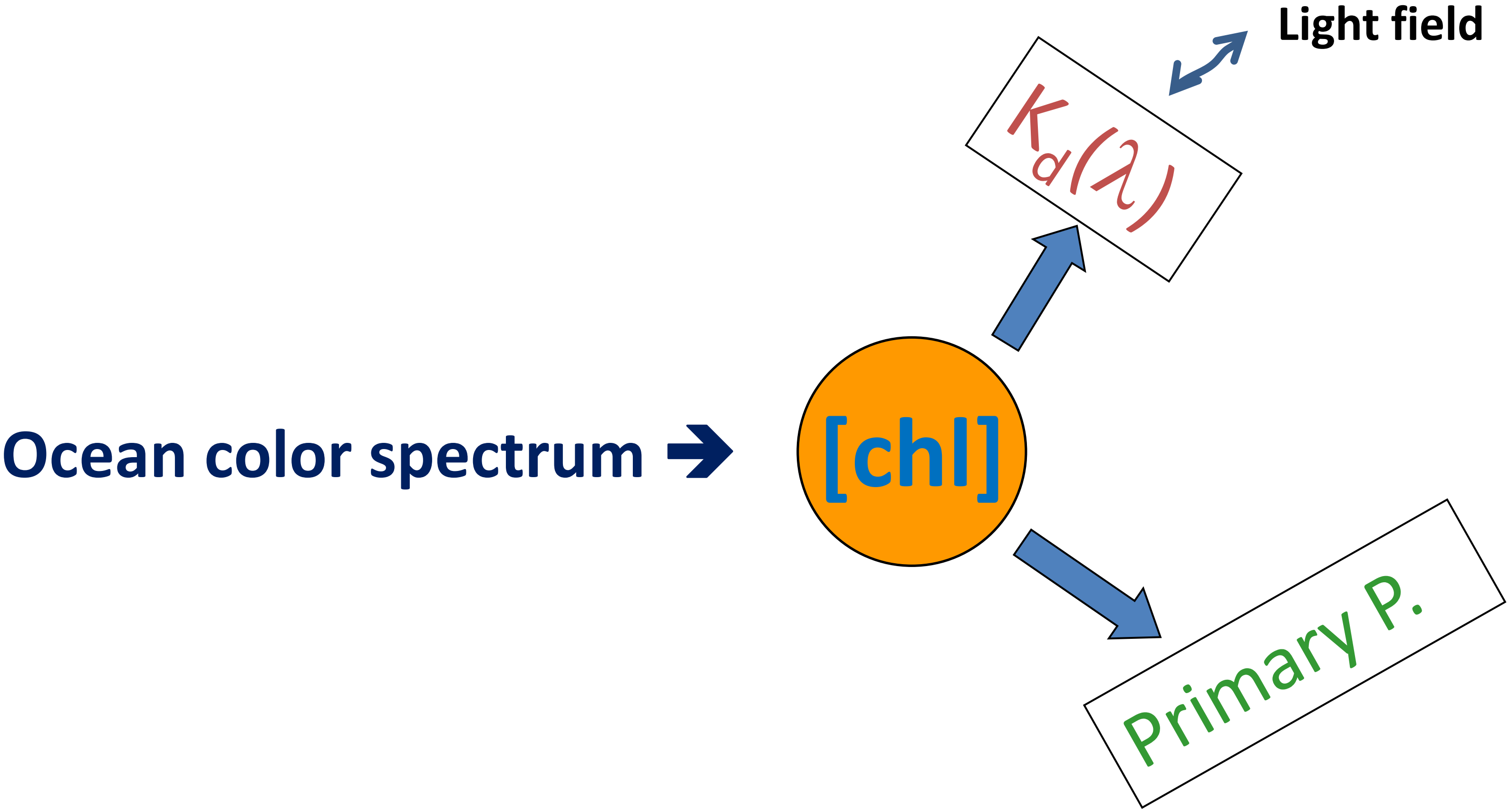
1. Phytoplankton

2. Light penetration

3. Photosynthesis parameters

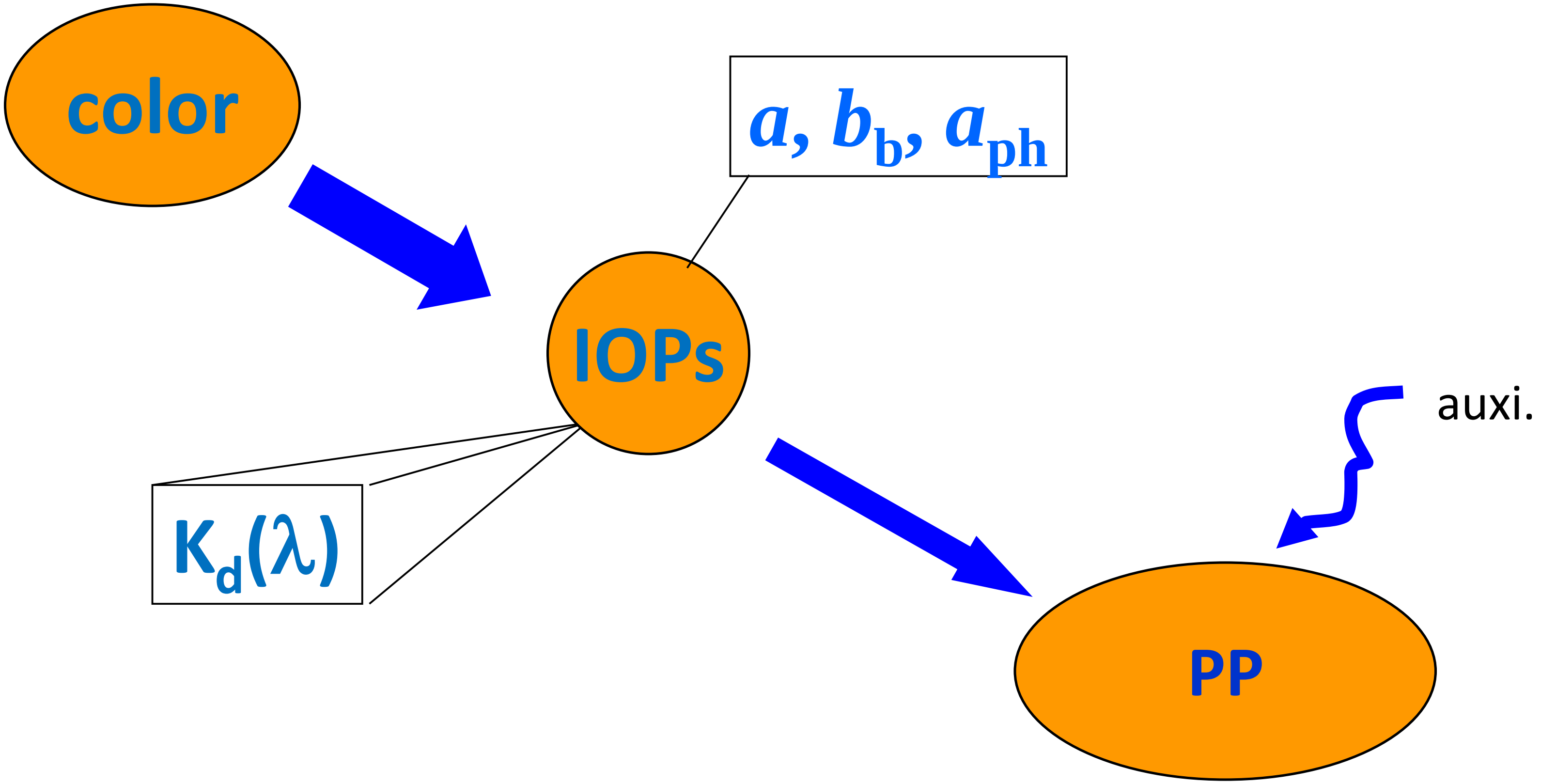
(Platt and Sathyendranath, 2007)

Traditional strategy: [chl] centered system



Works for waters where IOPs co-vary with [Chl].

IOP-based approach:


color

a, b_b, a_{ph}

IOPs

$K_d(\lambda)$

auxi.

PP

Works for most waters.

1. Estimation of diffuse attenuation coefficient (K$_d$) (for light field)

$$K_d(490) \propto \frac{Rrs(\lambda_1)}{Rrs(\lambda_2)}$$

Or,

$$K_d(\lambda) \propto Chl$$

$$K_d(490) = \mathrm{a}_1 + \mathrm{a}_2 X^{-\mathrm{a}_3}$$

or

$$K_d(490) = b_1 X^{-\mathrm{b}_2}$$

with

$$X = \frac{Lwn(488)}{Lwn(551)}$$

or

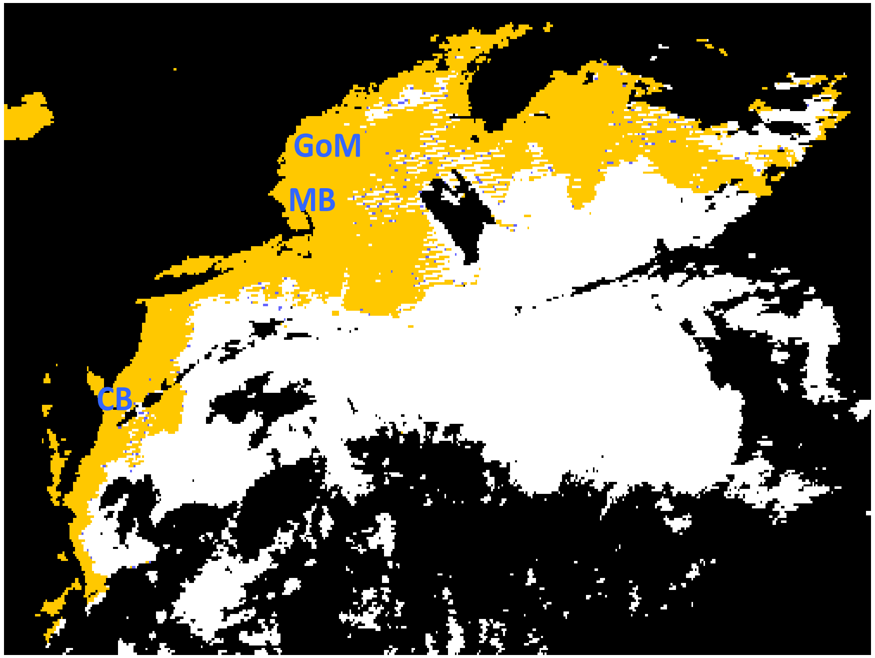
$$X = \frac{Rrs(488)}{Rrs(551)}$$

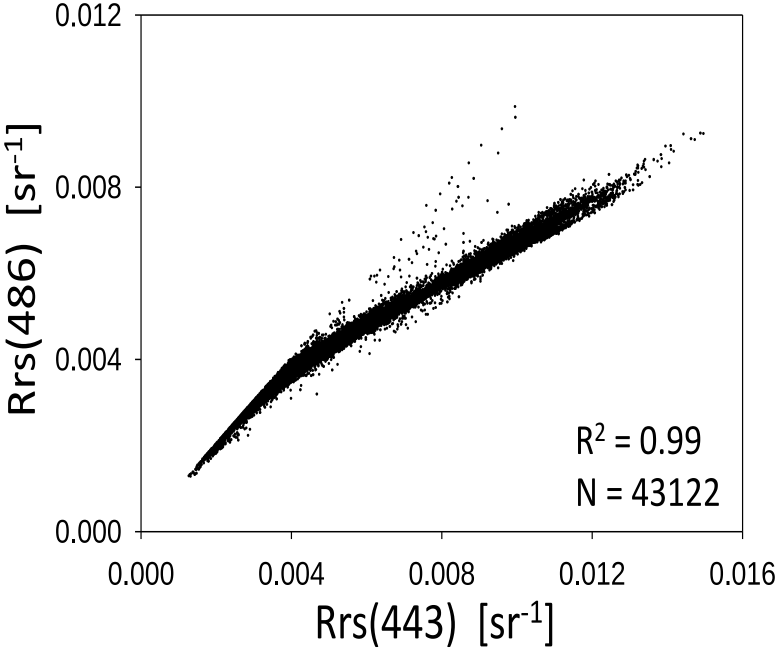
$$\log_{10}(K_{bio}(490)) = a_0 + \sum_{i=1}^{4} a_i \log_{10}\left(\frac{R_{rs}(\lambda_{\text{blue}})}{R_{rs}(\lambda_{\text{green}})}\right)$$

$$Kd\_490 = K_{bio}(490) + 0.0166$$

$$log_{10}(chlor\_a) = a_0 + \sum_{i=1}^{4} a_i log_{10}\left(\frac{R_{rs}(\lambda_{blue})}{R_{rs}(\lambda_{green})}\right)^i,$$

$Rrs(\lambda_{blue})$ = Rrs(443)>Rrs(486)





The standard $K_d(490)$ and Chl products are 100% co-vary in coastal waters; further…

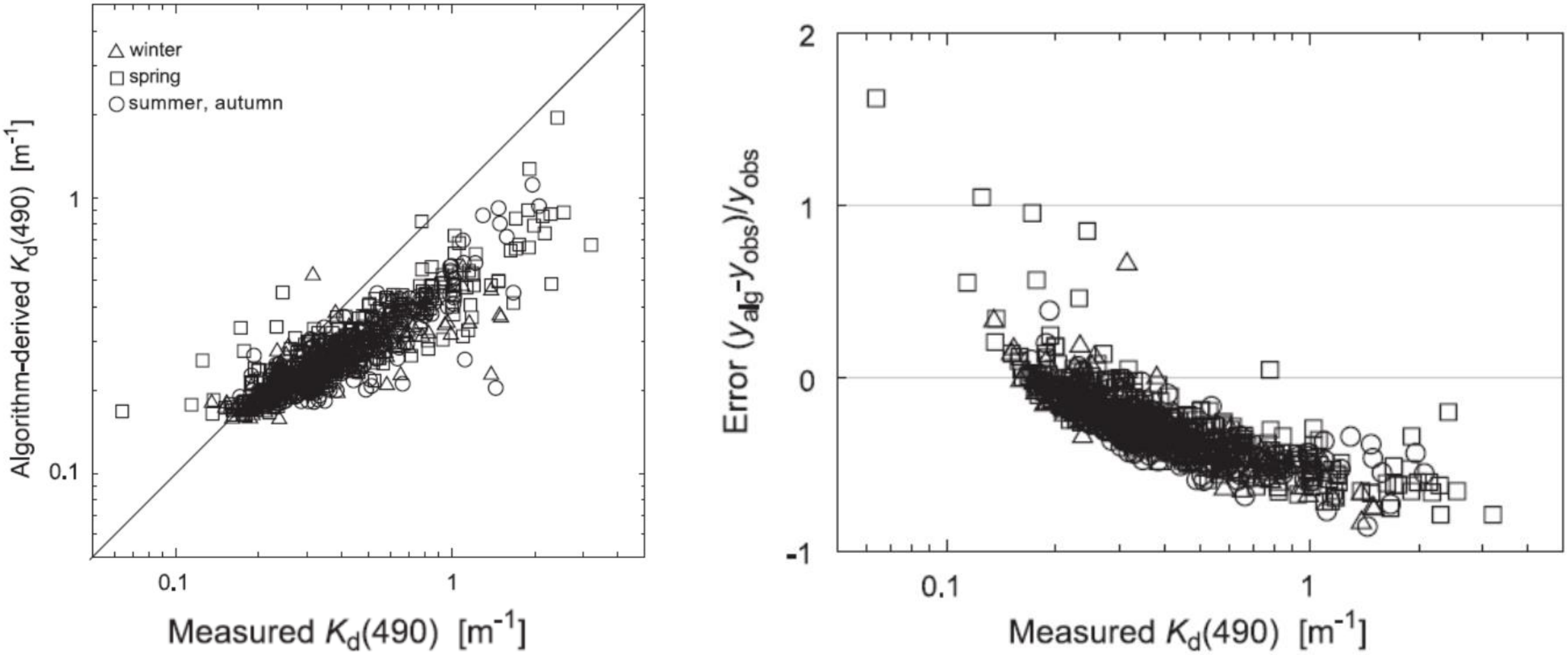
$$K_d(490) = Fun\left(\frac{R_{rs}(490)}{R_{rs}(555)}\right)$$

AOP: sun angle dependent

Nearly independent of sun angle

The two sides do not match in optical nature.

In addition, ratio-derived Kd has large uncertainties



(Darecki and Stramisk, 2004)

Via radiative transfer equation:

$$K_d = \mathrm{m}_0\, a + \mathrm{v}\; b_b$$

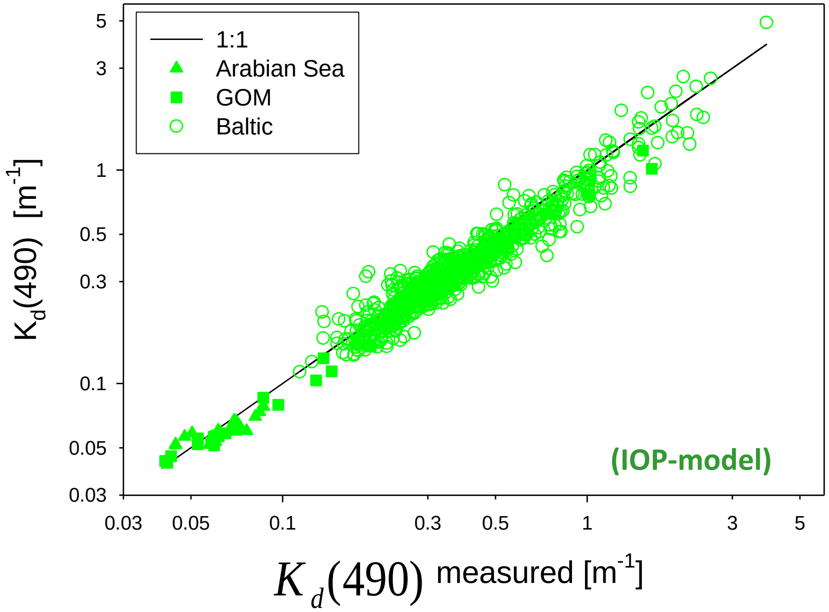
$$\mathrm{v} = m_1(1 - m_2 e^{-m_3 a})$$

(*Lee et al. 2005, 2013*)

($\mathrm{m}_{0,1,2}$ are wavelength independent)

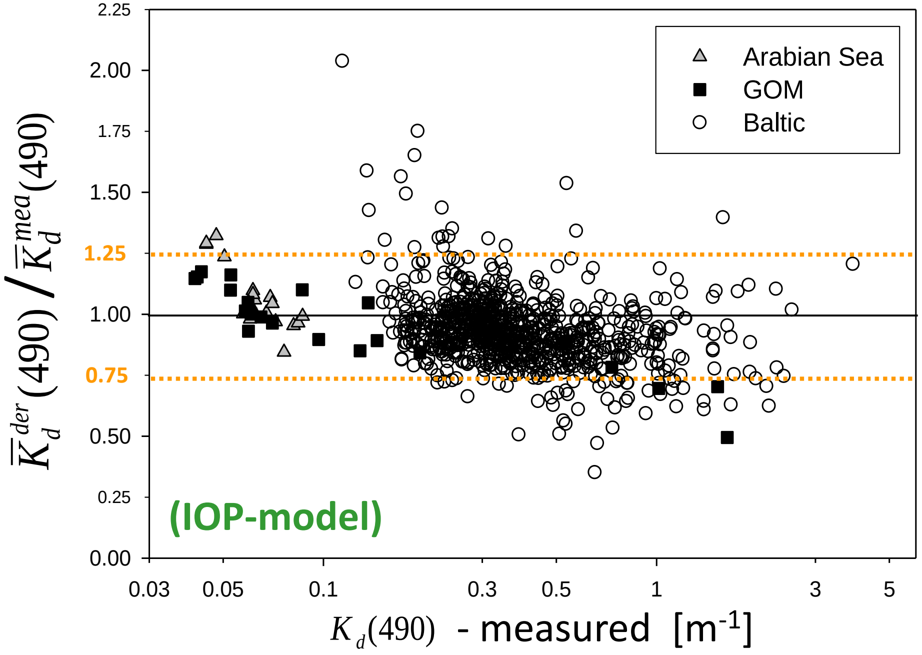
$$\boxed{R_{rs} \Rightarrow a\&b_b \Rightarrow K_d}$$

No division of “Case 1” or “Case 2” waters.

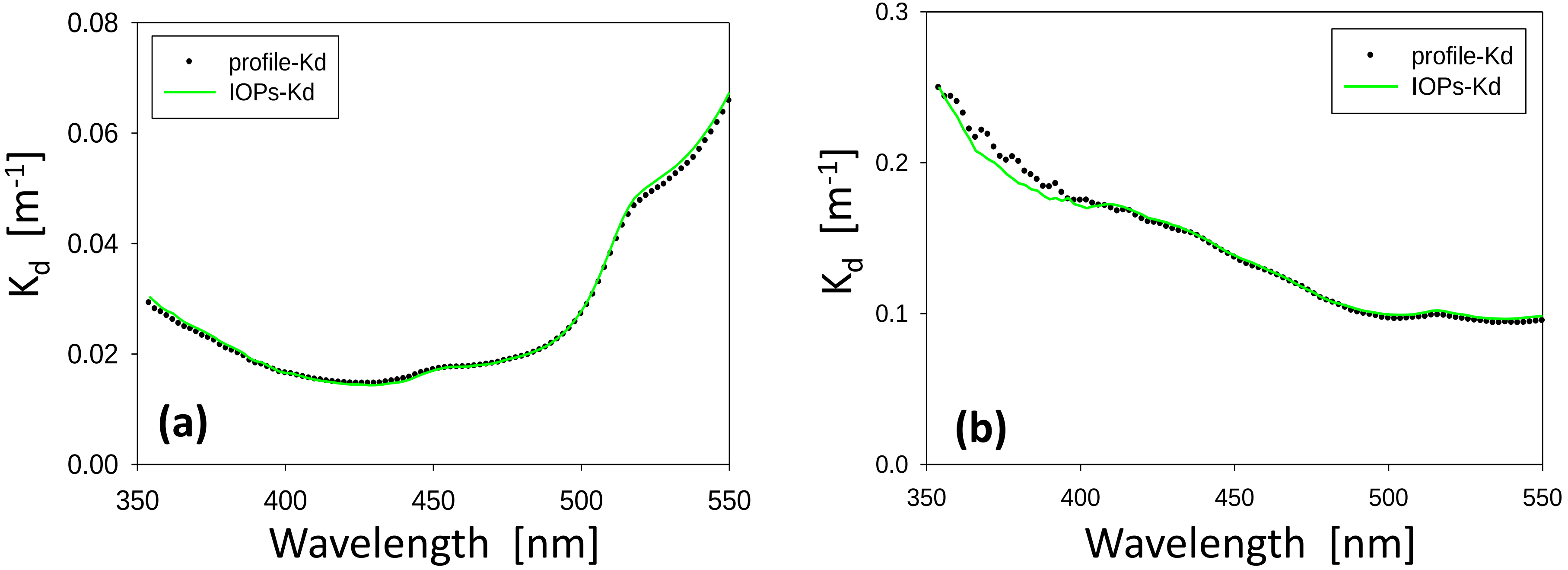


Oceanic & Coastal waters

(*Lee et al. 2005*)



Kd spectrum:



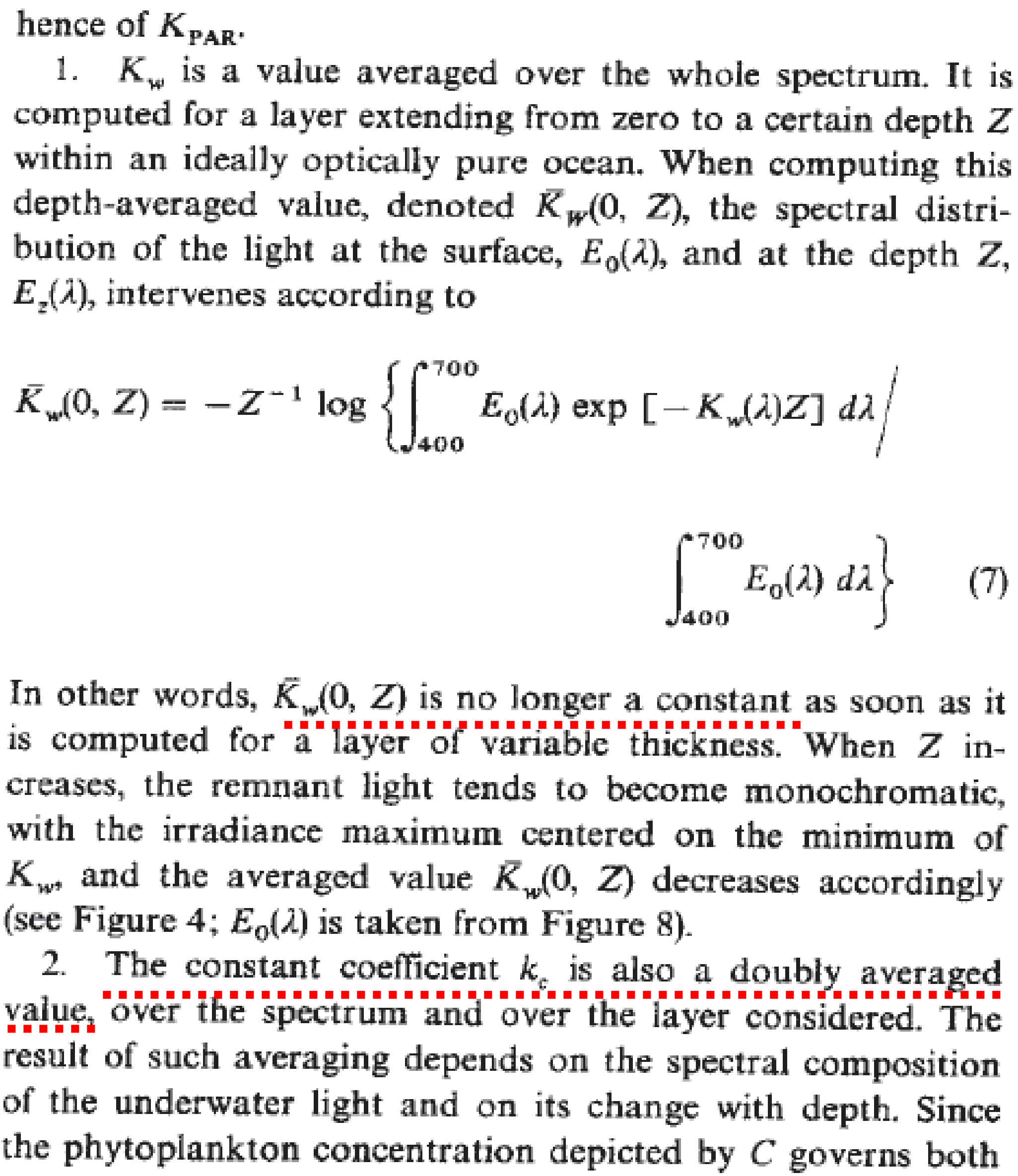
(Lee et al 2013)

2. Attenuation of PAR (K_{PAR}) and euphotic depth

$$PAR(z) = PAR(0)\,e^{-K_{PAR}\,z}$$

$$K_{\mathrm{PAR}} = K_w + k_c C + K_x$$

Good for earlier days,
Not so good for the 21st century



hence of K_{PAR}.

1. K_w is a value averaged over the whole spectrum. It is computed for a layer extending from zero to a certain depth Z within an ideally optically pure ocean. When computing this depth-averaged value, denoted $\bar{K}_w(0, Z)$, the spectral distribution of the light at the surface, $E_0(\lambda)$, and at the depth Z, $E_z(\lambda)$, intervenes according to

$$\bar{K}_w(0, Z) = -Z^{-1} \log \left\{ \int_{400}^{700} E_0(\lambda) \exp\left[-K_w(\lambda) Z\right] d\lambda \Big/ \int_{400}^{700} E_0(\lambda)\, d\lambda \right\} \qquad (7)$$

In other words, $\bar{K}_w(0, Z)$ is no longer a constant as soon as it is computed for a layer of variable thickness. When Z increases, the remnant light tends to become monochromatic, with the irradiance maximum centered on the minimum of K_w, and the averaged value $\bar{K}_w(0, Z)$ decreases accordingly (see Figure 4; $E_0(\lambda)$ is taken from Figure 8).

2. The constant coefficient k_c is also a doubly averaged value, over the spectrum and over the layer considered. The result of such averaging depends on the spectral composition of the underwater light and on its change with depth. Since the phytoplankton concentration depicted by C governs both

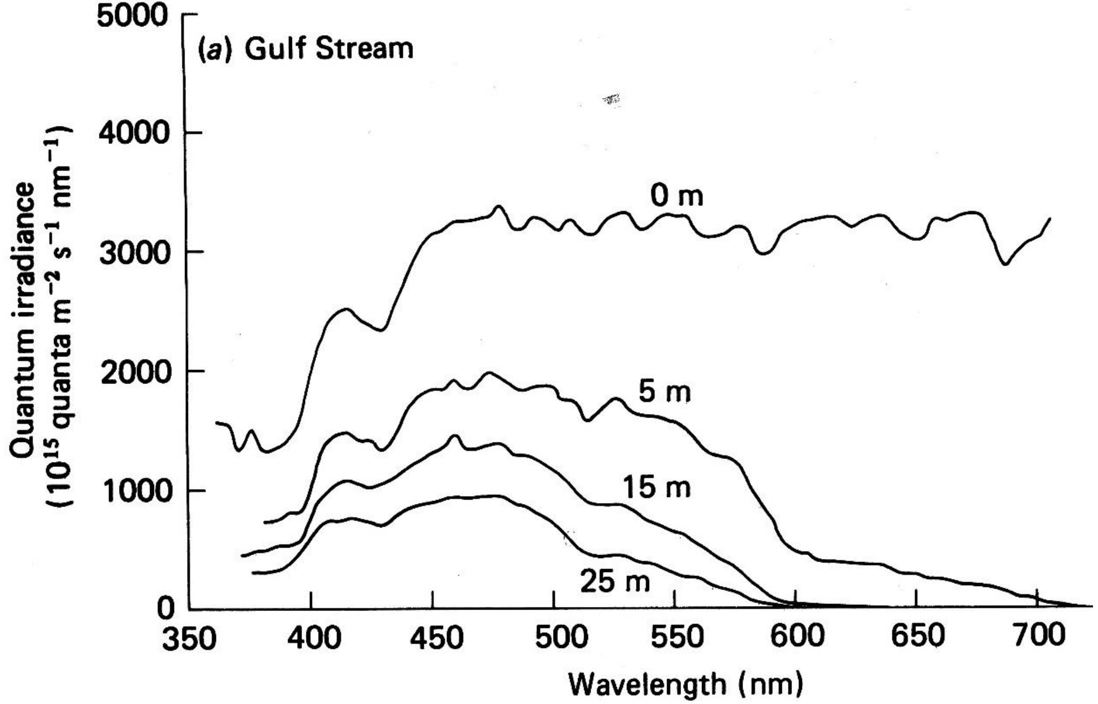
(Morel, 1988, JGR)

$$K_{PAR} = \frac{-1}{z} \ln\left(\frac{PAR(z)}{PAR(0)}\right) \qquad PAR(z) = \int_{400}^{700} E_0(\lambda,0)\, e^{-K(\lambda,z)z}\, d\lambda$$

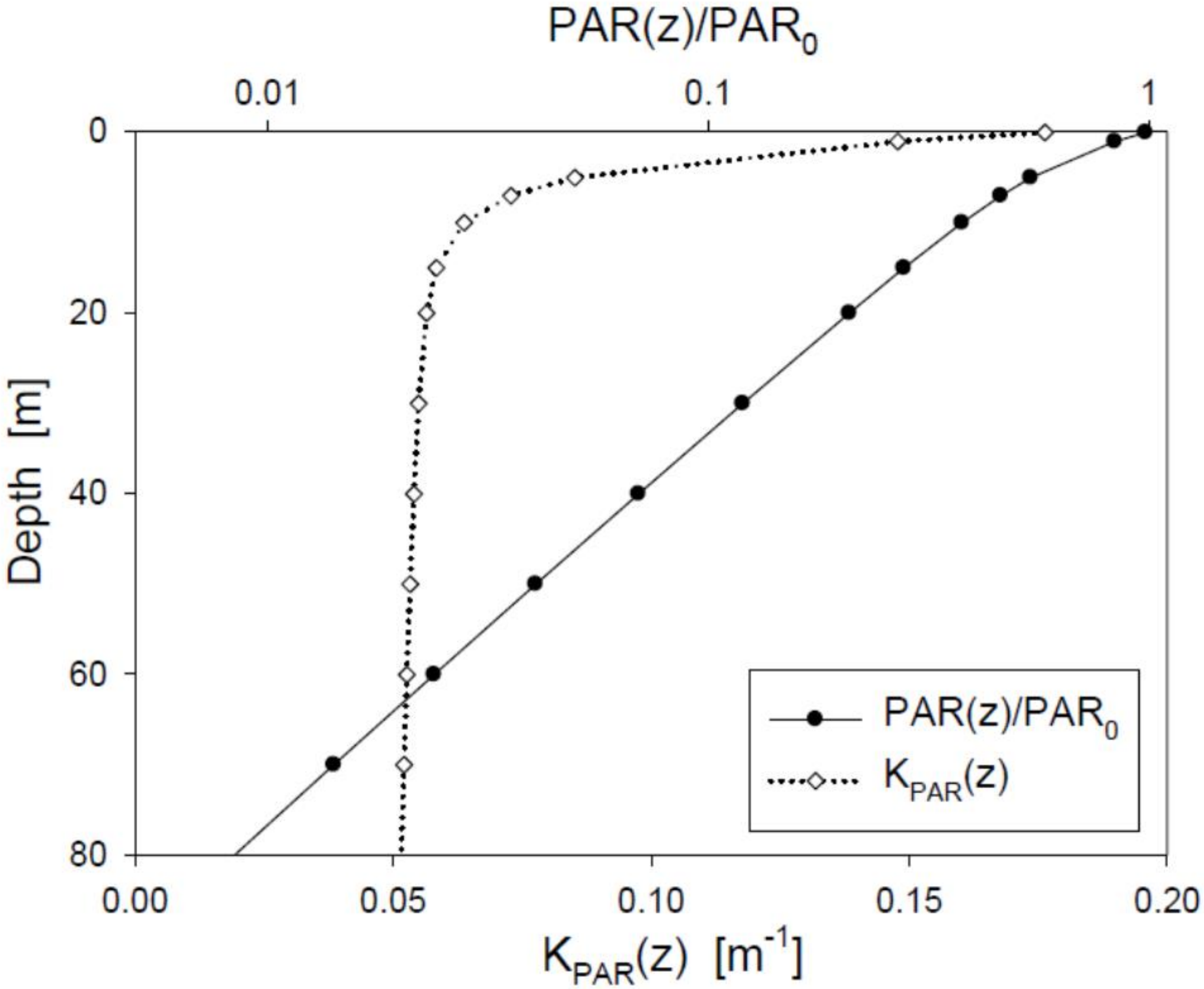
K_{PAR} is light-quality weighted!

Change of light with depth:

(Kirk 1994)



➔ **Light at deeper depth is associated with lower attenuation coefficient**


PAR(z)/PAR0
0.01
0.1
1
0
20
40
60
80
Depth [m]
0.00
0.05
0.10
0.15
0.20
KPAR(z) [m-1]
PAR(z)/PAR0
KPAR(z)

$$PAR(z) = PAR(0)\,e^{-K_{par}(z)\,z}$$

$$K_{PAR}(z) \approx K_1 + \frac{K_2}{(1+z)^{0.5}}$$

$$K_1 = f_1(a(490), b_b(490) \qquad K_2 = f_2(a(490), b_b(490)$$

Key:

K_{PAR} varies with depth, especially in the upper water column!

Transmission estimated from remote sensing:

Transmittance of PAR：

$$\boldsymbol{T = \frac{PAR(z)}{PAR(0)} = \sum a\, e^{-z*b}}$$

The [Chl] based models

MA94

$$T(Chl, z\,) = V_1 e^{-z/z_1} + V_2 e^{-z/z_2}$$

$$V_i, z_i = a_0 + a_1 X + a_2 X^2 + a_3 X^3 + a_4 X^4$$

$$X = \log(Chl)$$

OS00

$$\boldsymbol{T}(Chl, z, \theta_S, CI) = \left(\sum_{i=1}^{4} A_i e^{-K_i z}\right)/0.47$$

$$A_i, K_i = C_1 Chl + C_3 \left(\cos\theta_S\right)^{-1} + C_4$$

MA05

$$T(Chl, z) = 0.5\, e^{-K_{Red} z} + 0.5\, e^{-K_{BG} z}$$

$$K_\lambda = K_W + \chi_\lambda Chl^{e_\lambda}$$

Mu02

$$T(Chl, z) = e^{-z/h_\gamma}$$

$$1/h_\gamma = 0.027 + 0.0518\, Chl^{0.428}$$

↔ Kpar

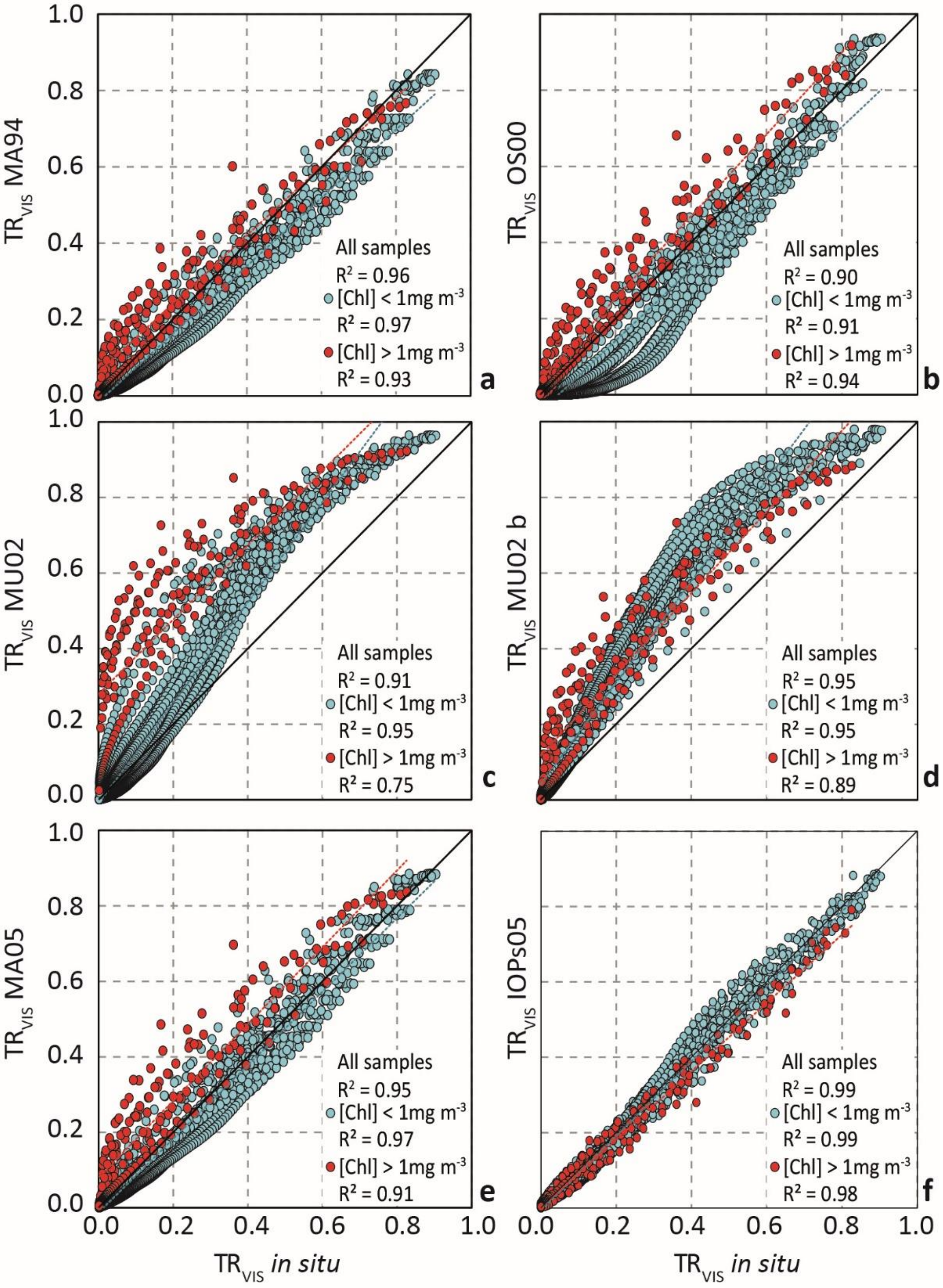
The IOP based model

IOP05

$$T(IOP, z, \theta_S) = e^{-K_{PAR}(IOP,z,\theta_S)z}$$

$$K_{PAR}(IOP, z, \theta_S) = K_1(IOP, \theta_S) + \frac{K_2(IOP, \theta_S)}{(1+z)^{0.5}}$$

Model	R^2	Mean |error|
MA94	0.92	41.2%
OS00	0.88	35.7%
MA05	0.90	49.2%
Mu02	0.87	152.8%
IOPs05	**0.95**	**21.9%**

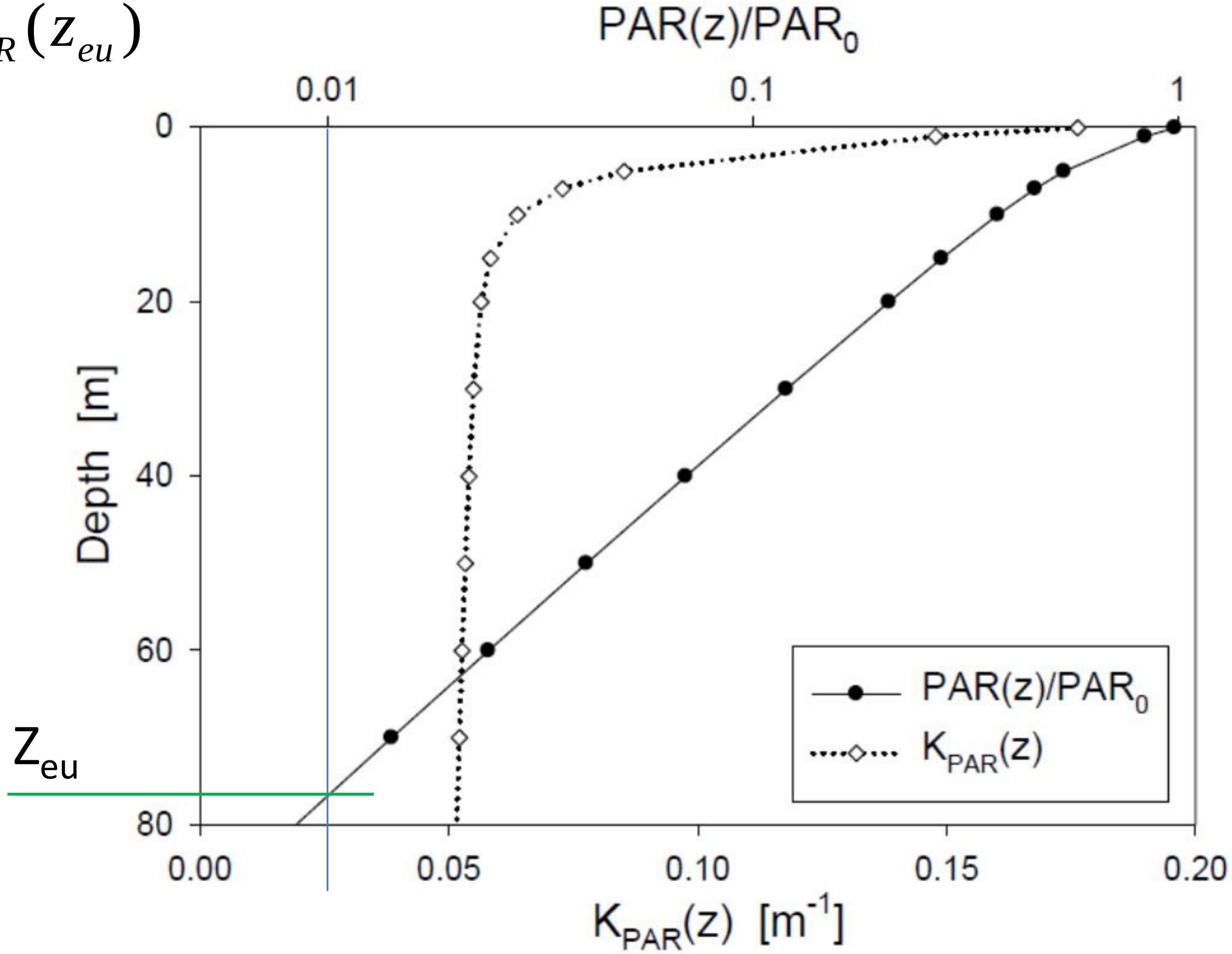


(Zoffoli et al., 2017)

Euphotic depth (z_{eu}):

$$\frac{PAR(z)}{PAR(0)} = 1\%$$

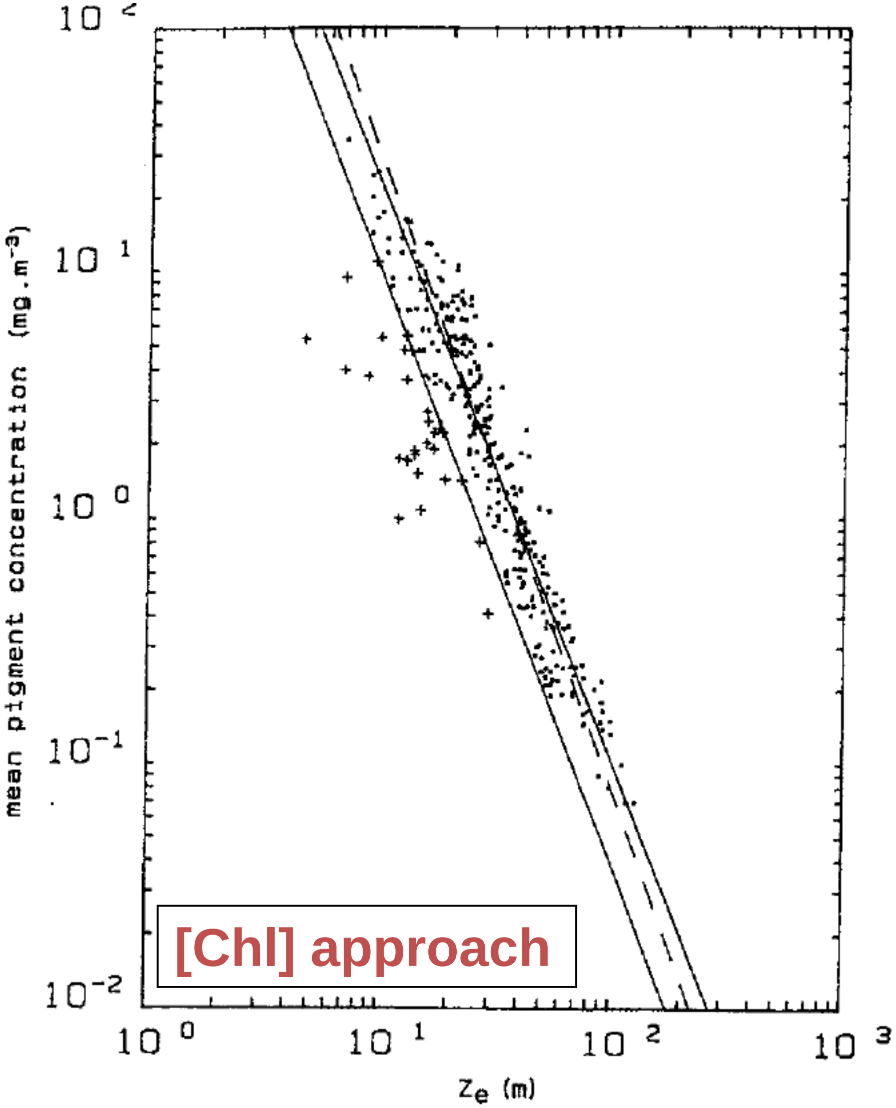
$$z_{eu} = \frac{4.6}{K_{PAR}(z_{eu})}$$



$$\frac{PAR(z_{eu})}{PAR(0)} = 1\%$$

[Chl] (empirical) approach:

$Rrs(\lambda) \rightarrow [Chl] \rightarrow z_{eu}$



[Chl] approach

$$Z_e = 38.0C^{-0.428}$$

(Morel 1988)

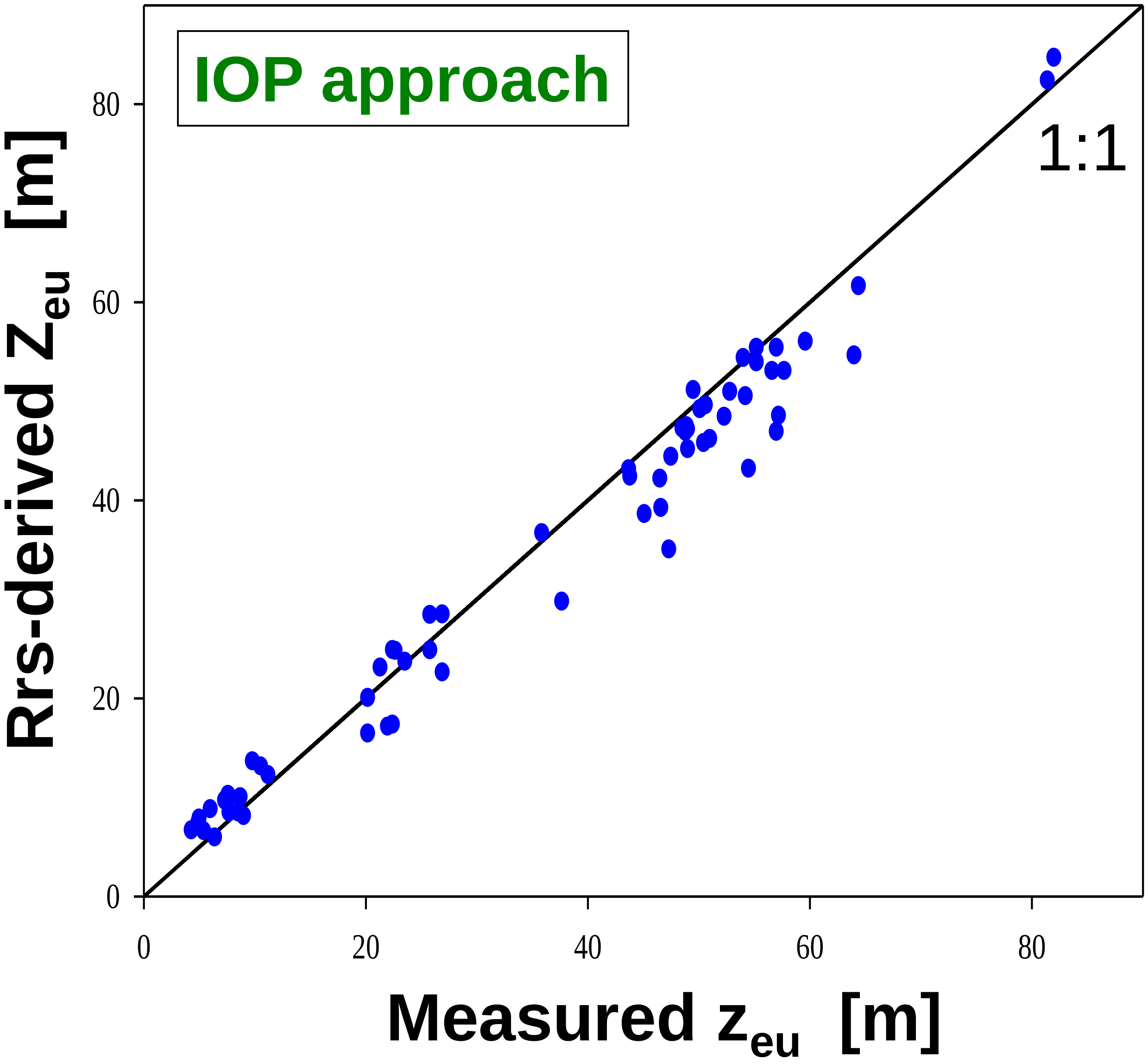
IOP approach:

$$\mathrm{Rrs}(\lambda) \rightarrow a(\lambda)\&b_{\mathrm{b}}(\lambda) \rightarrow K_{\mathrm{PAR}}(\mathrm{z}) \rightarrow z_{\mathrm{eu}}$$

$$\Updownarrow$$

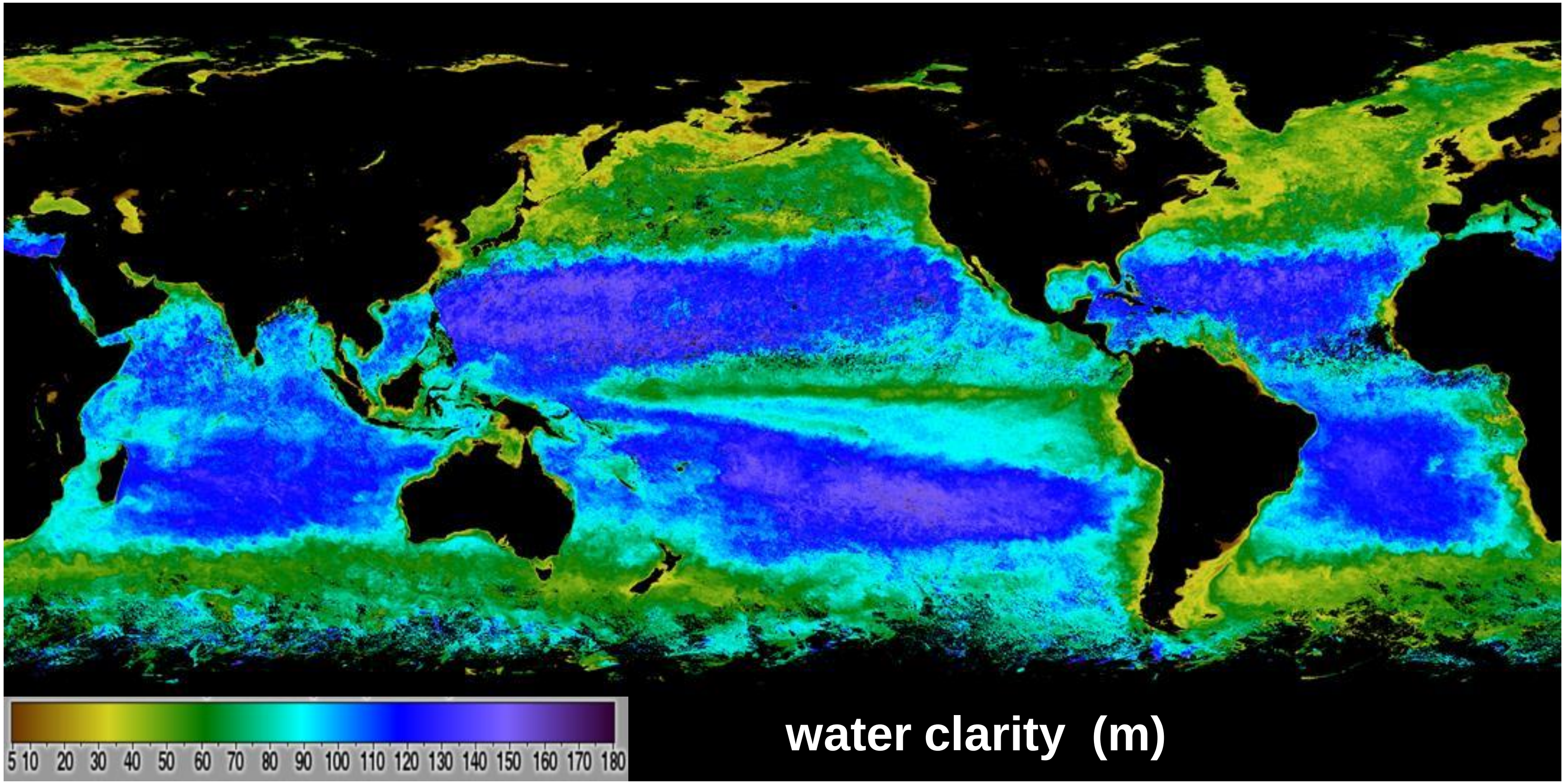
$$K_{PAR}(z) = K_1\left(a(490)\,\&\,b_b(490)\right) + \frac{K_2(a(490)\,\&\,b_b(490))}{(1+z)^{0.5}}$$

Water clarity (z_{eu})



(Lee et al 2005)

Global distribution of Z_{eu}

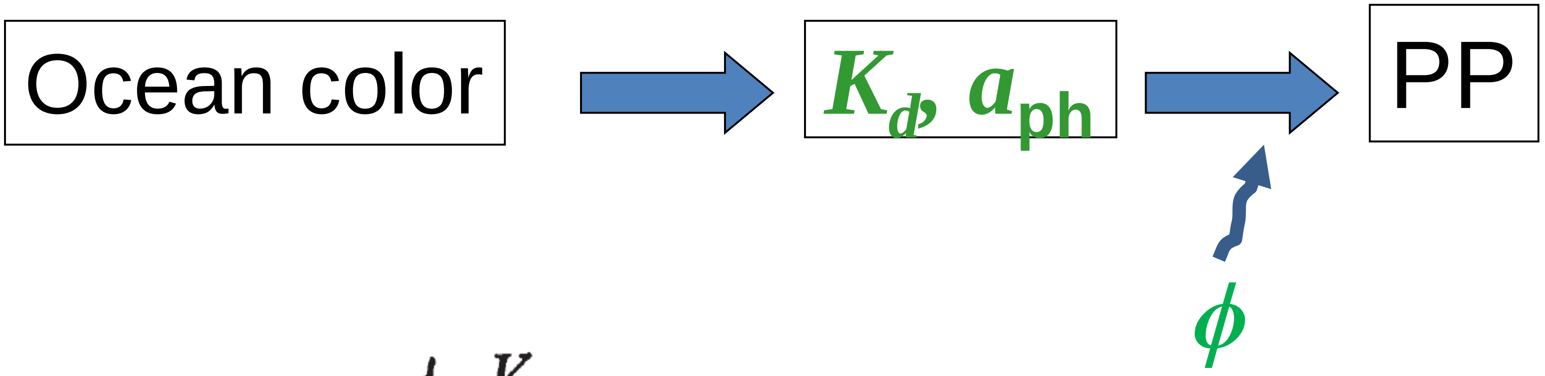


3. PP estimation:

$$PP(z) = \iint \phi \times E_0(\lambda, t, z) \times a_{ph}(\lambda, z)\, d\lambda\, dt$$

represents 'photosynthesis'

ϕ: quantum yield of photosynthesis



Ocean color

K_d, a_{ph}

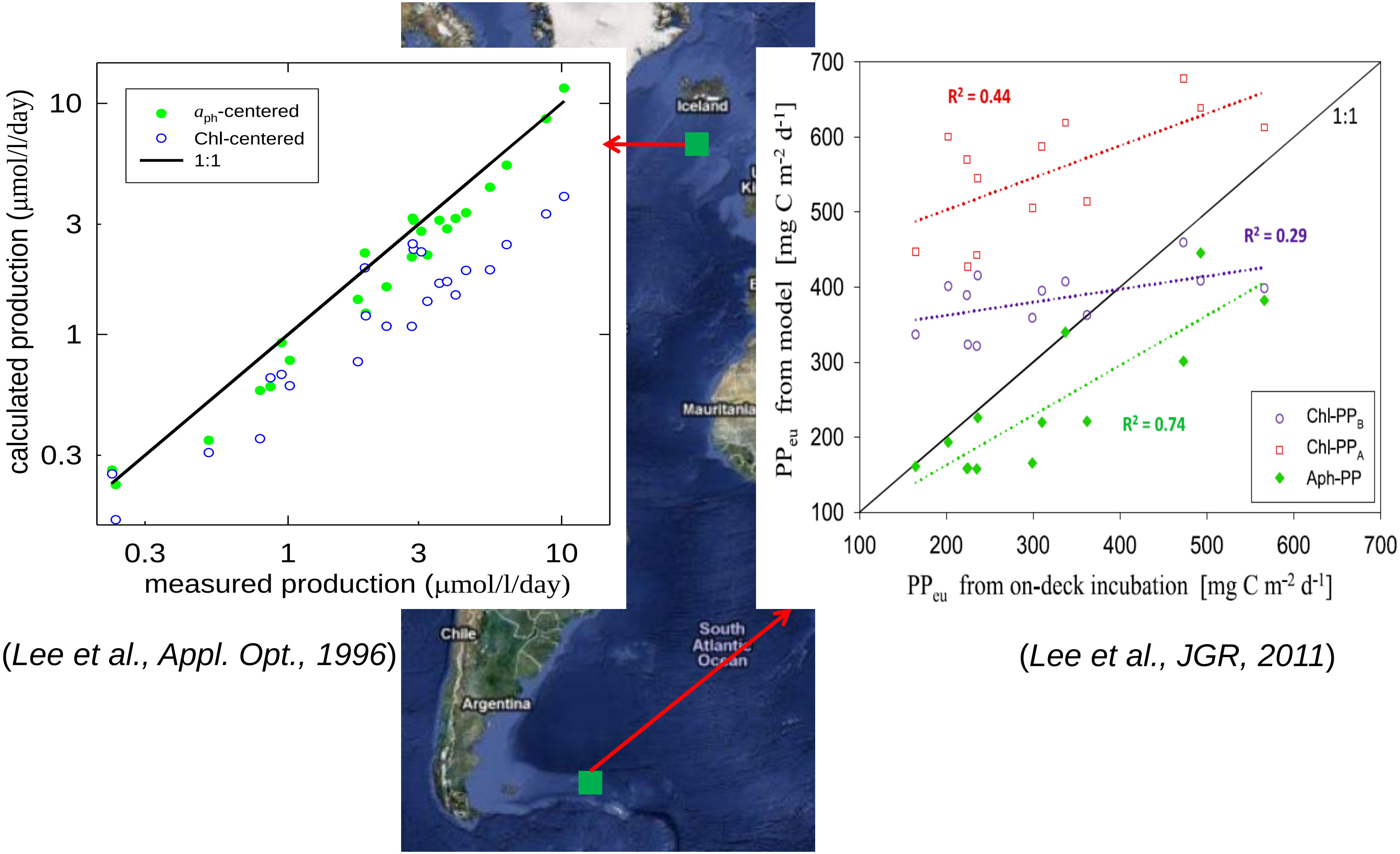
PP

ϕ

$$\phi(E_o) = \frac{\phi_m K_\phi}{K_\phi + E_o} \qquad (2)$$

(Kiefer and Mitchell, 1983, L&O)

Remotely-estimated PP compared with measured PP



(*Lee et al., Appl. Opt., 1996*)

(*Lee et al., JGR, 2011*)

Why a_{ph} based approach is likely better?

Essence of present satellite Chl product:

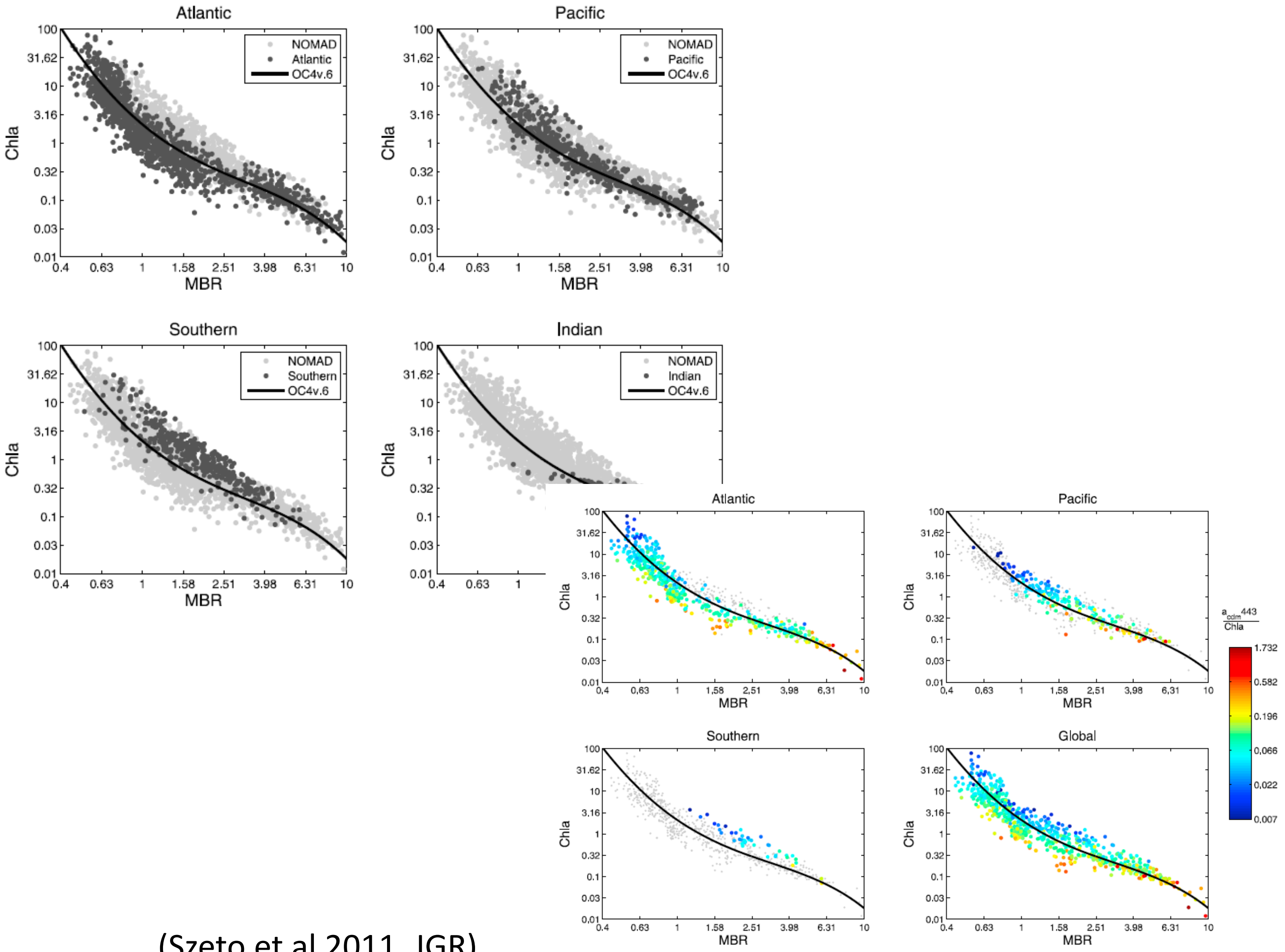
$$Chl = fun\left(\frac{R_{rs}(\lambda_1)}{R_{rs}(\lambda_2)}\right)$$

$$R_{rs} \approx G\frac{b_b}{a + b_b}$$

$$\frac{R_{rs}(440)}{R_{rs}(550)} \approx \frac{a(550)}{a(440)}\frac{b_{bw}(440) + b_{bp}(440)}{b_{bw}(550) + b_{bp}(550)}$$

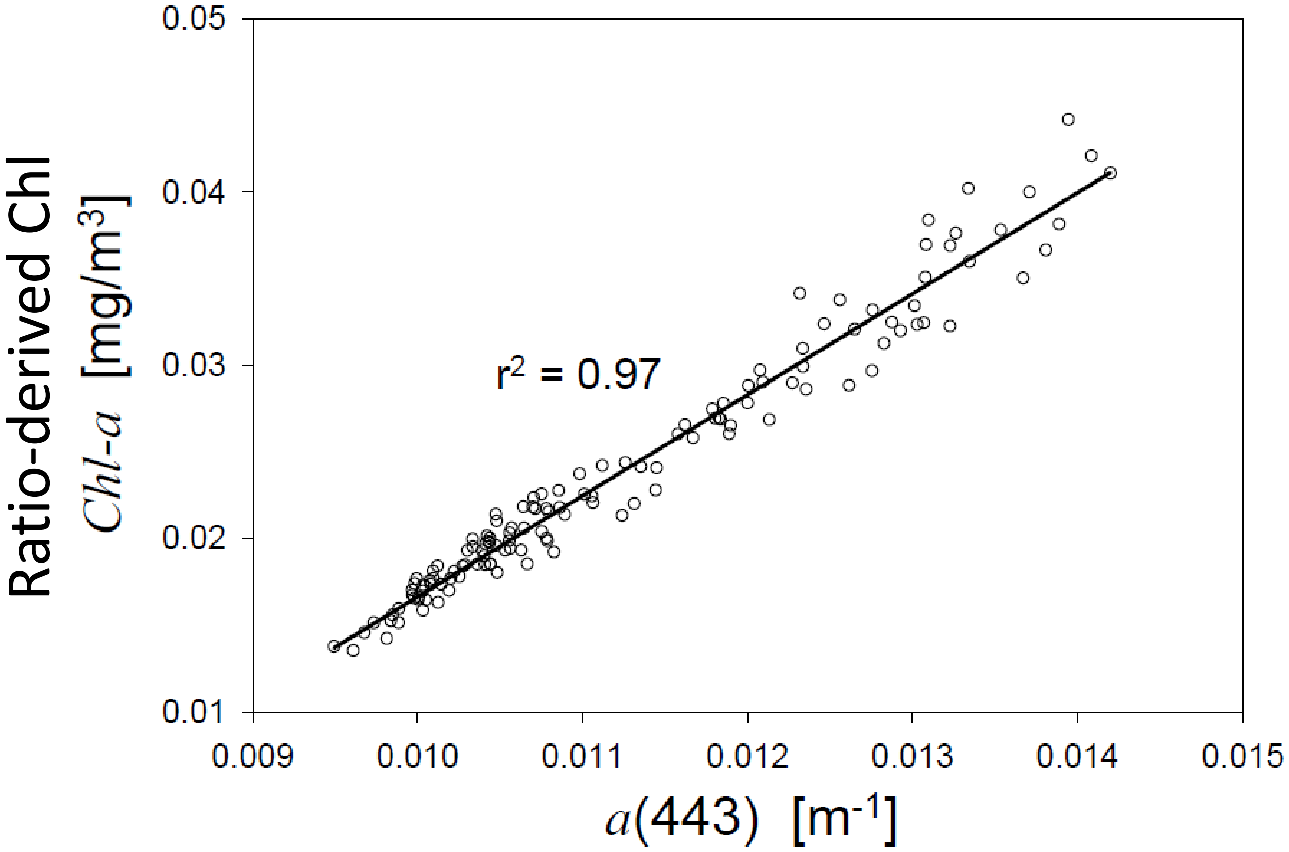
$$\Rightarrow \quad \frac{R_{rs}(440)}{R_{rs}(550)} \propto \frac{a(550)}{a(440)} = \frac{a_w(550) + a_{dg}(550) + a^*_{ph}(550)Chl}{a_w(440) + a_{dg}(440) + a^*_{ph}(440)Chl}$$

Change of R_{rs} band ratio not necessarily represents change of [Chl]!



(Szeto et al 2011, JGR)

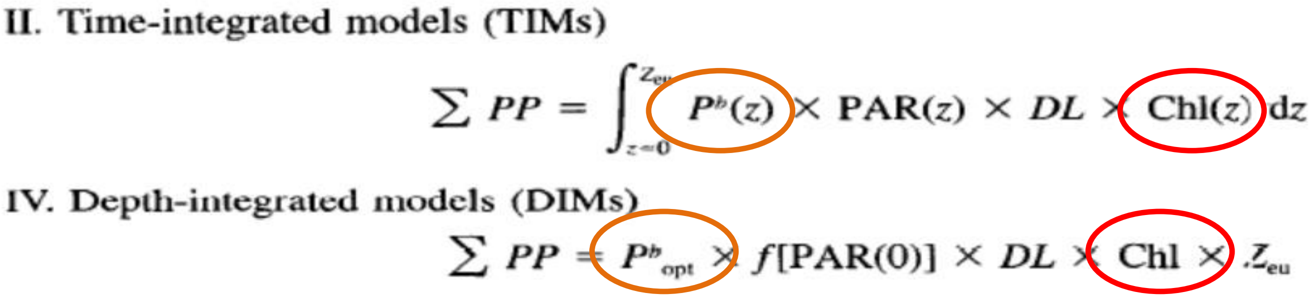
The change of Rrs ratio really reflects change of total absorption!



(Lee et al, 2010, JGR)

p^b:

(Behrenfeld and Falkowski, 1997)



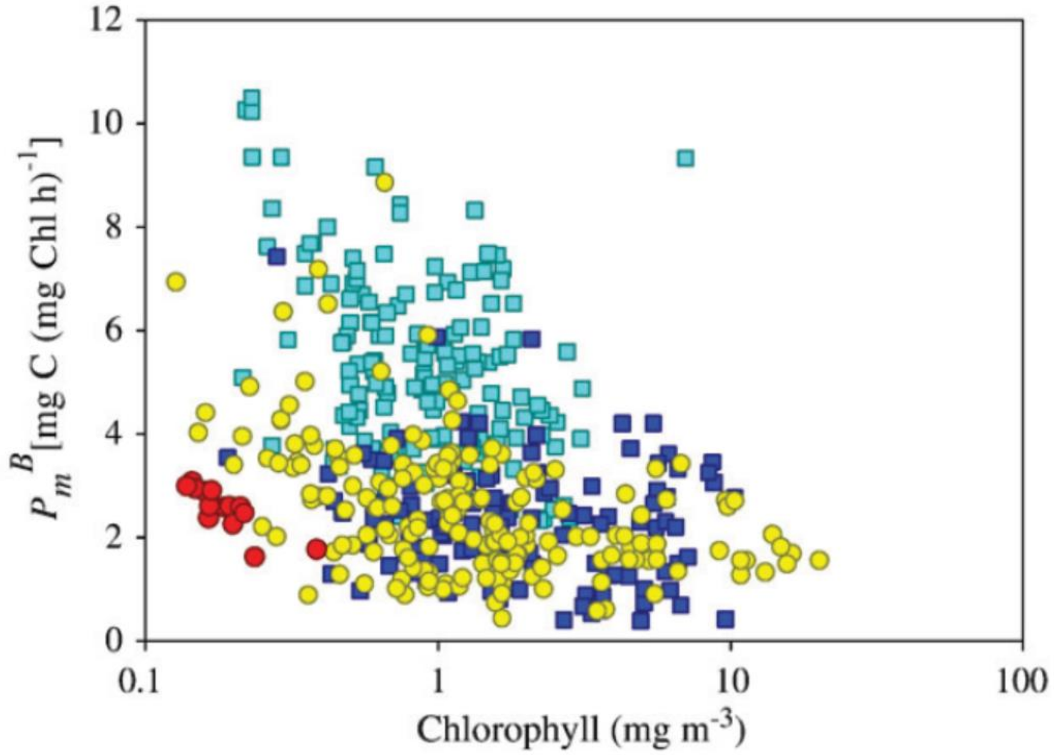
II. Time-integrated models (TIMs)

$$\sum PP = \int_{z=0}^{Z_{eu}} P^b(z) \times \mathrm{PAR}(z) \times DL \times \mathrm{Chl}(z)\, dz$$

IV. Depth-integrated models (DIMs)

$$\sum PP = P^b_{opt} \times f[\mathrm{PAR}(0)] \times DL \times \mathrm{Chl} \times Z_{eu}$$

$$p^b \propto \phi_m\, a^*_{ph}$$



P^b is also dependent on a^*_{ph}, which is not a constant either for a given Chl nor for varying Chl!

(Platt et al 2008, RSE)

Another example of using remotely sensed IOPs for PP

“Carbon” based Production Model (CbPM)

$$\mu = 2 \times \mathrm{Chl{:}C_{sat}} / \left[0.022 + (0.045 - 0.022)\,\exp^{-3\mathrm{Ig}}\right] \times \left(1 - \exp^{-3\mathrm{Ig}}\right).$$

(Behrenfeld et al 2005; Westberry et al 2008)

$$\mu = \frac{Chl}{C} / f(I_g)$$

actually a_{ph} from GSM

b_{bp}

‘IOP-based growth rate’

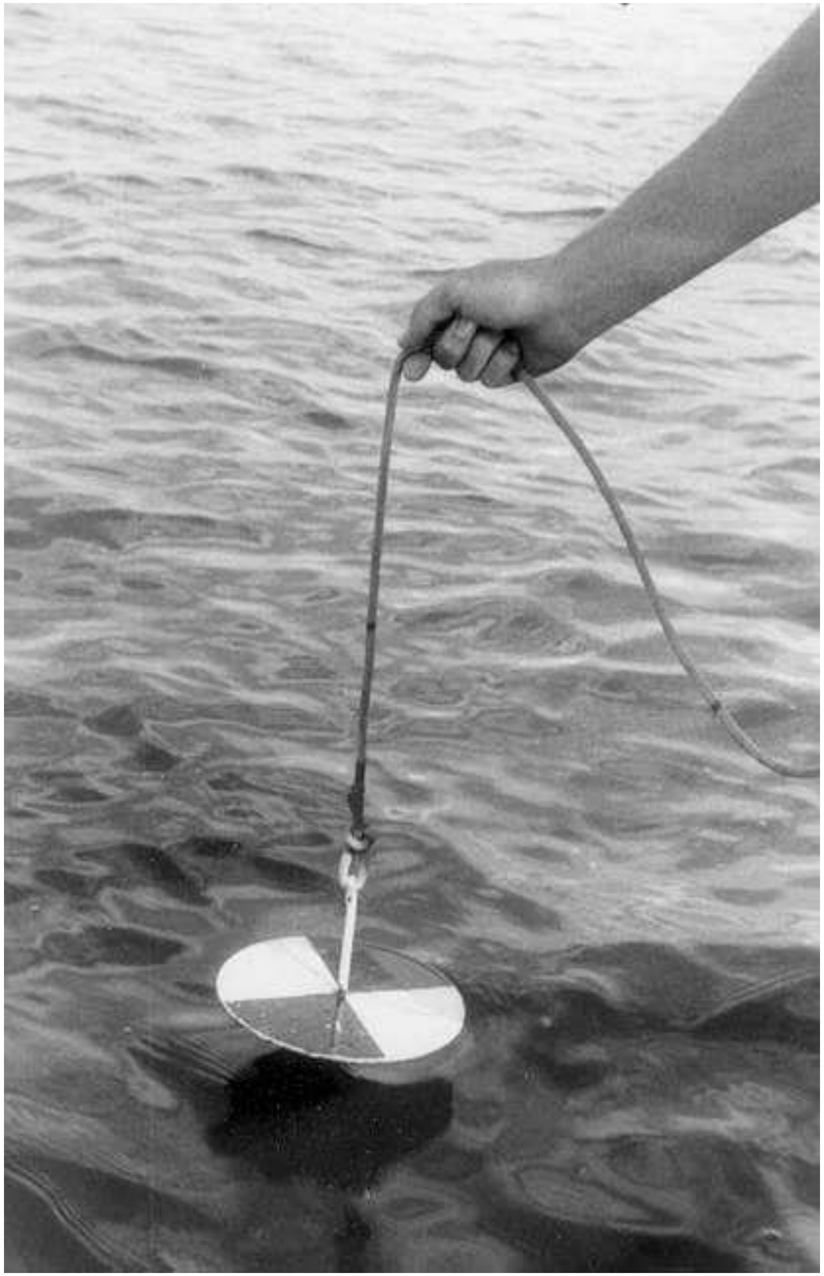
$$\mathrm{NPP} = \mathrm{C} \times \mu \times \mathrm{Z_{eu}} \times h(\mathrm{I_0}),$$

$$NPP \propto Chl \times Z_{eu} \times h(I_0) / f(I_g)$$

a_{ph} from GSM

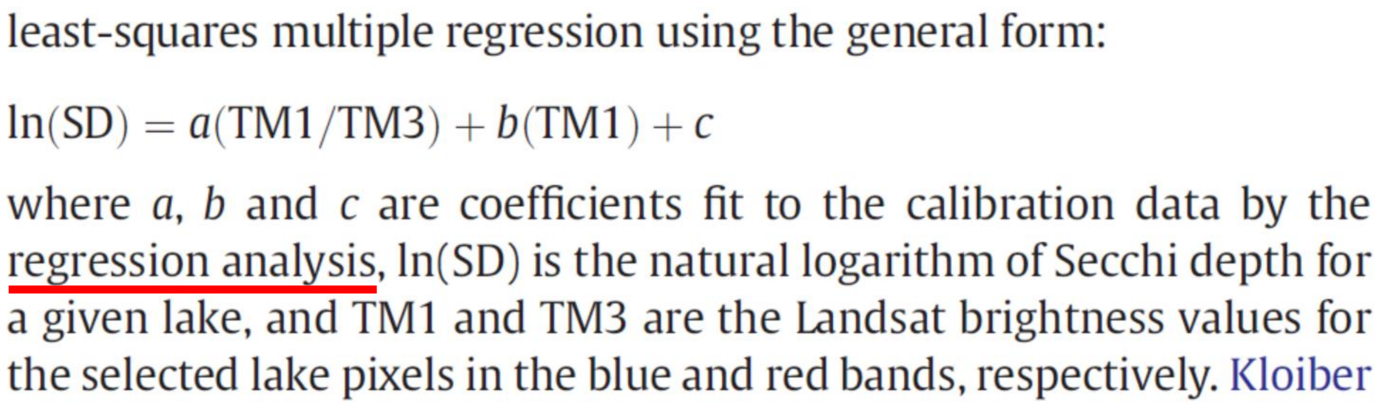
‘IOP-based NPP’

4. Secchi depth (Z_{SD})



Water clarity

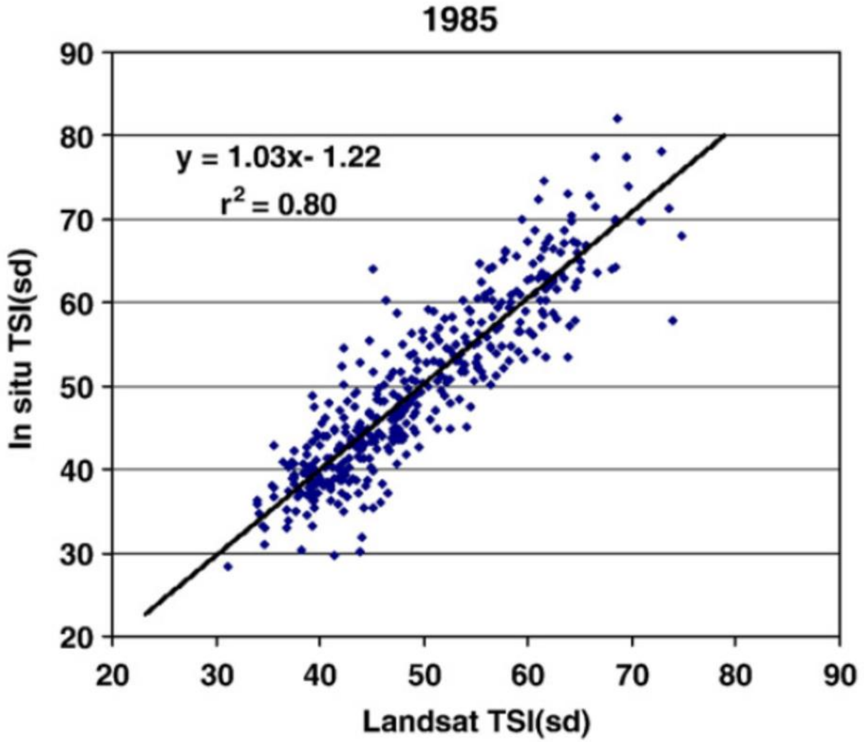
empirical approach:



least-squares multiple regression using the general form:

$$\ln(\mathrm{SD}) = a(\mathrm{TM1}/\mathrm{TM3}) + b(\mathrm{TM1}) + c$$

where a, b and c are coefficients fit to the calibration data by the regression analysis, ln(SD) is the natural logarithm of Secchi depth for a given lake, and TM1 and TM3 are the Landsat brightness values for the selected lake pixels in the blue and red bands, respectively. Kloiber



(Olmanson et al 2008)

New theoretical relationship for Z_{SD}:

$$Z_{SD} \approx \frac{1}{2.5K_d^{tr}} \ln\left(\frac{\left| r_T - r_w^{tr} \right|}{0.013} \right)$$

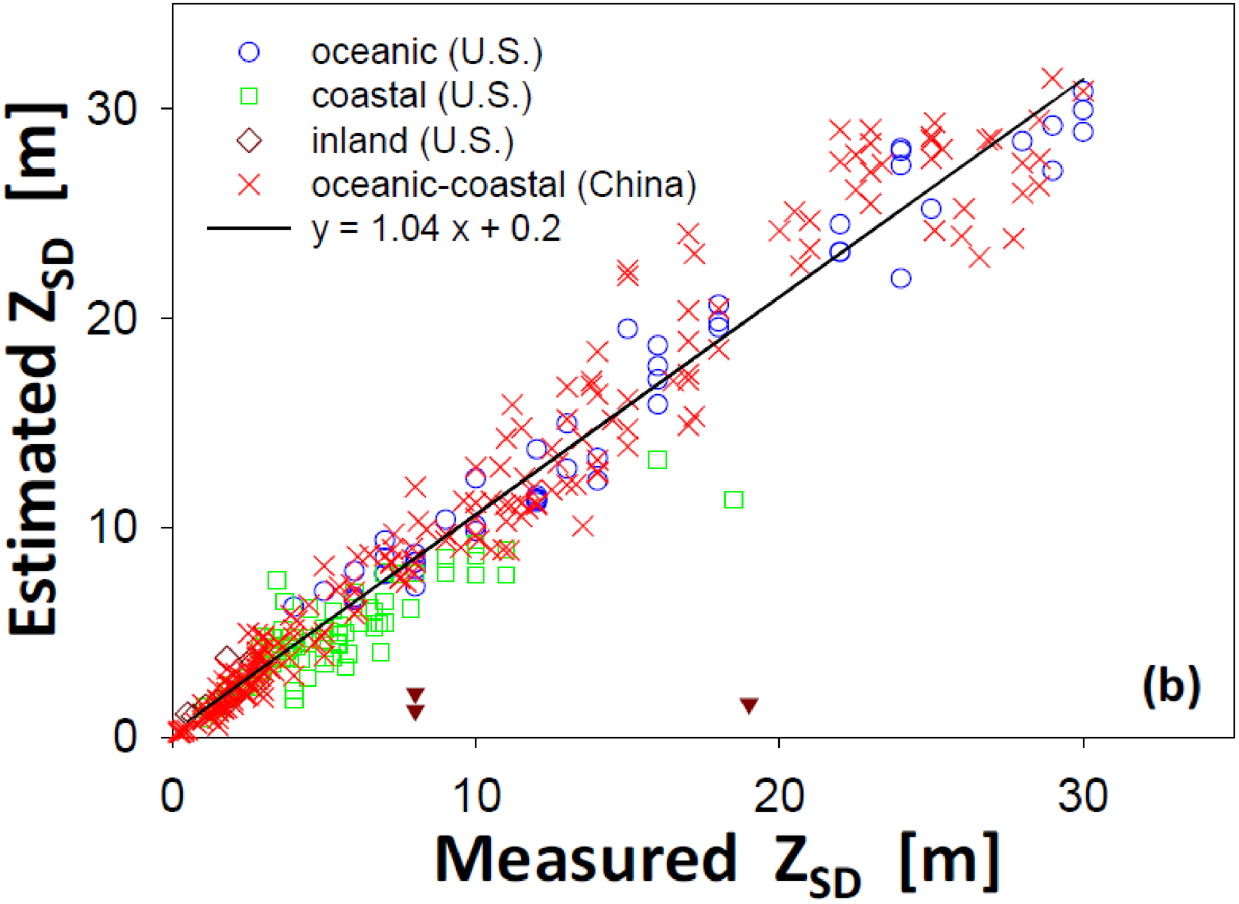
K_d^{tr} **: attenuation coefficient in the transparent window**

(Lee et al 2015)

Verification of the new Secchi disk theory

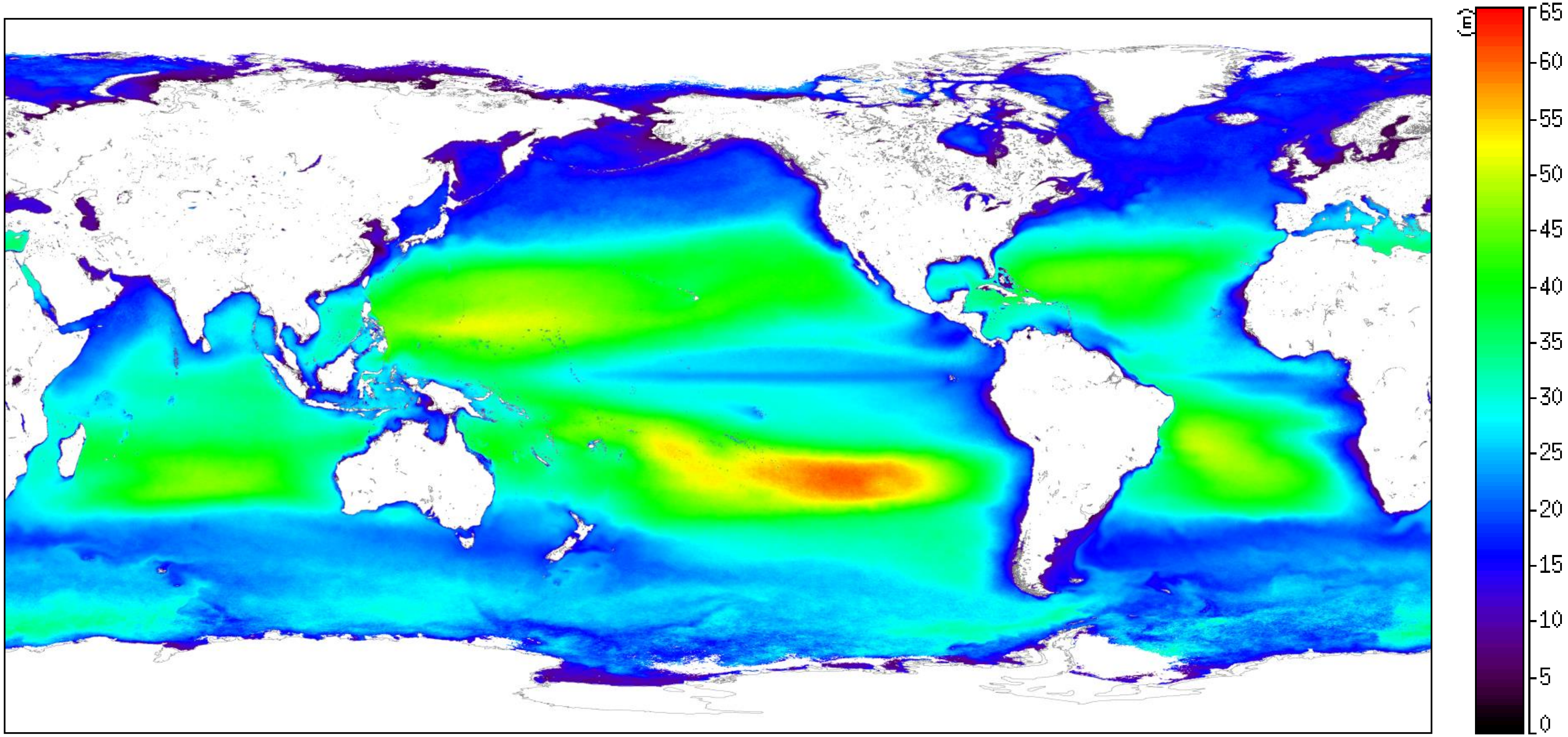
QAA （2002） Lee et al (2013)

$$R_{rs} \rightarrow a\&b_b \rightarrow K_d$$



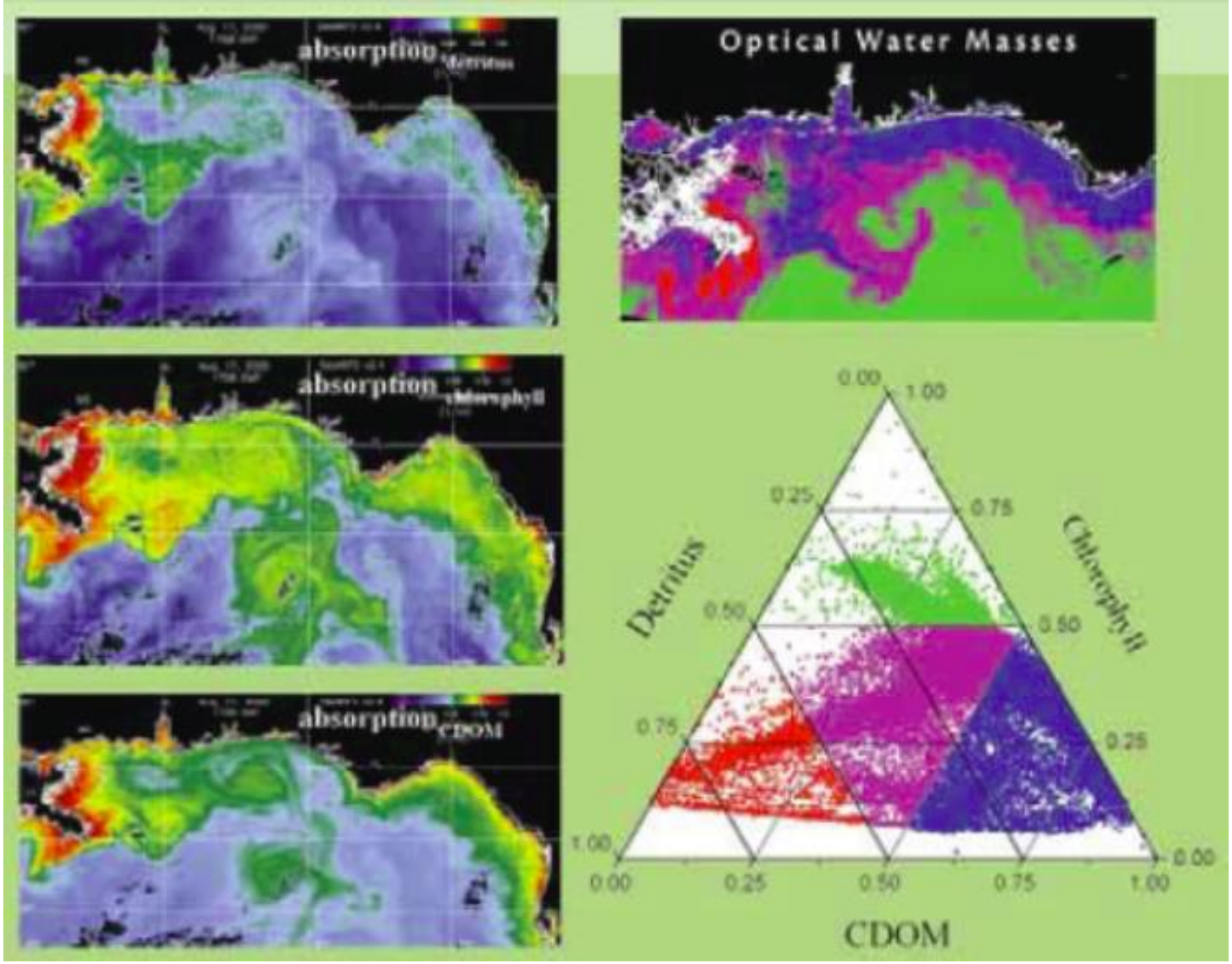
(Lee et al 2015)

Global Z_{SD}



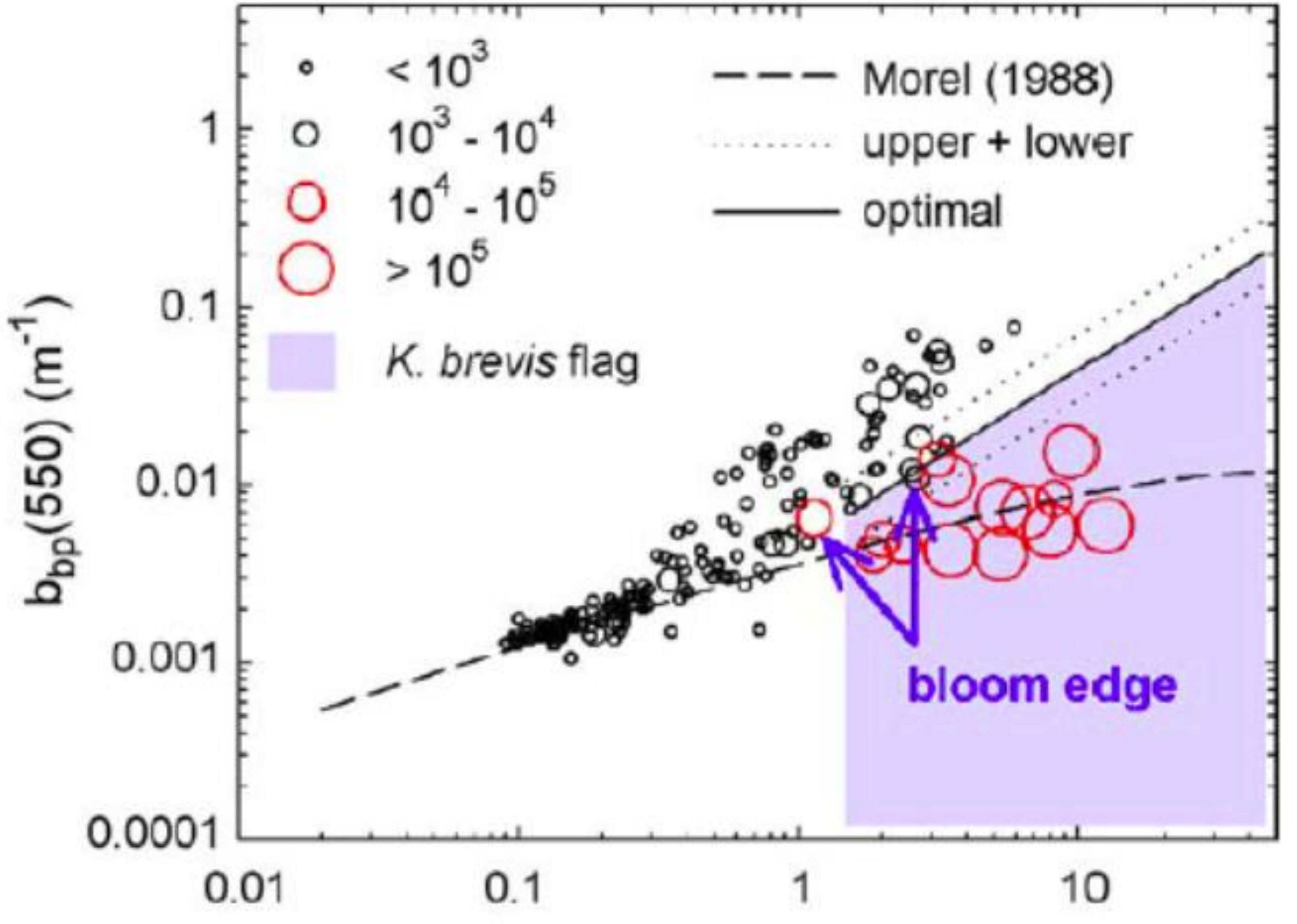
A much more straightforward product for water clarity!

5. Water mass classification



(Arnone et al 2004)

6. HAB identification-1

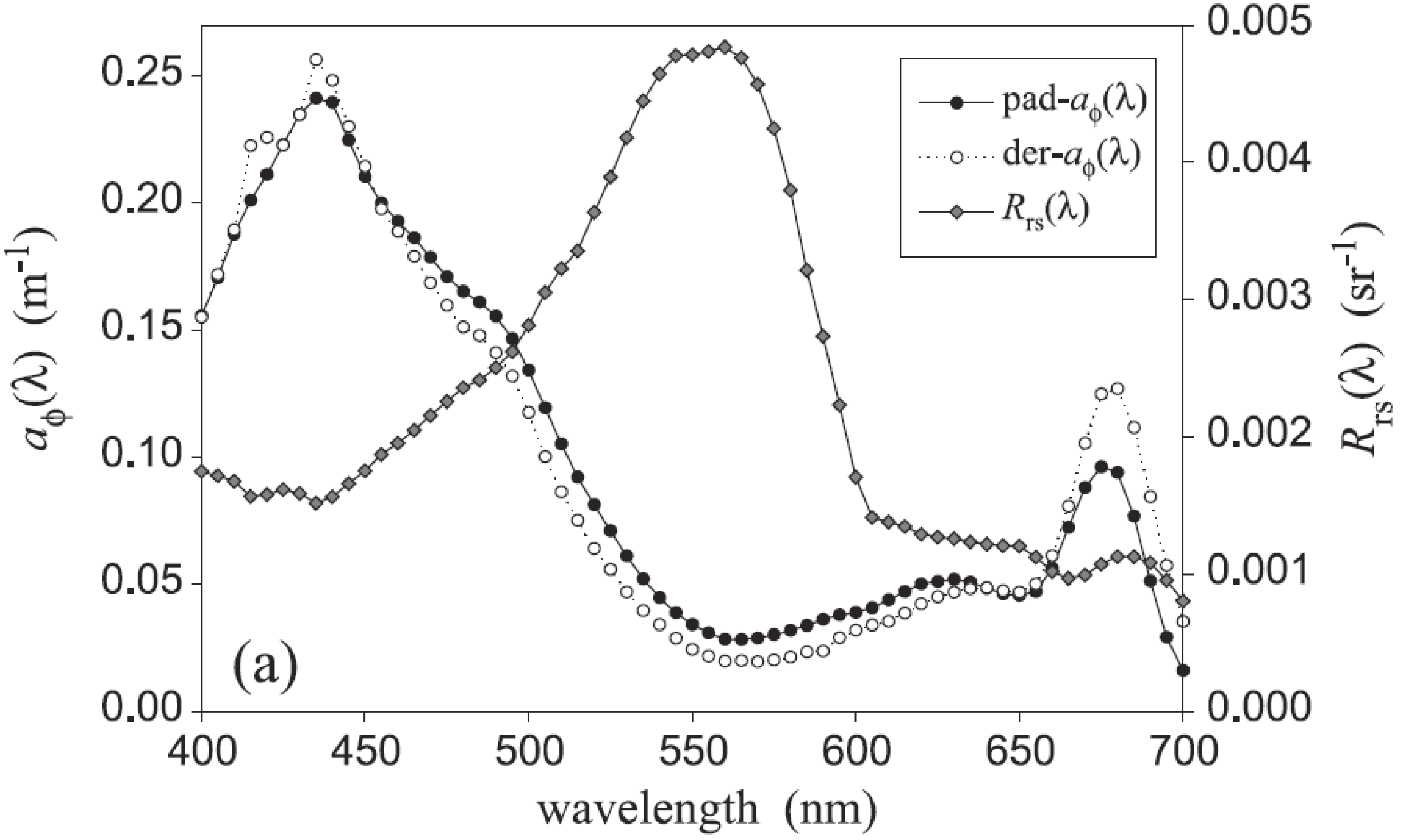


$a_{ph}(675)$ ➔ Chl (mg m^{-3})

(Carnizzaro et al 2006)

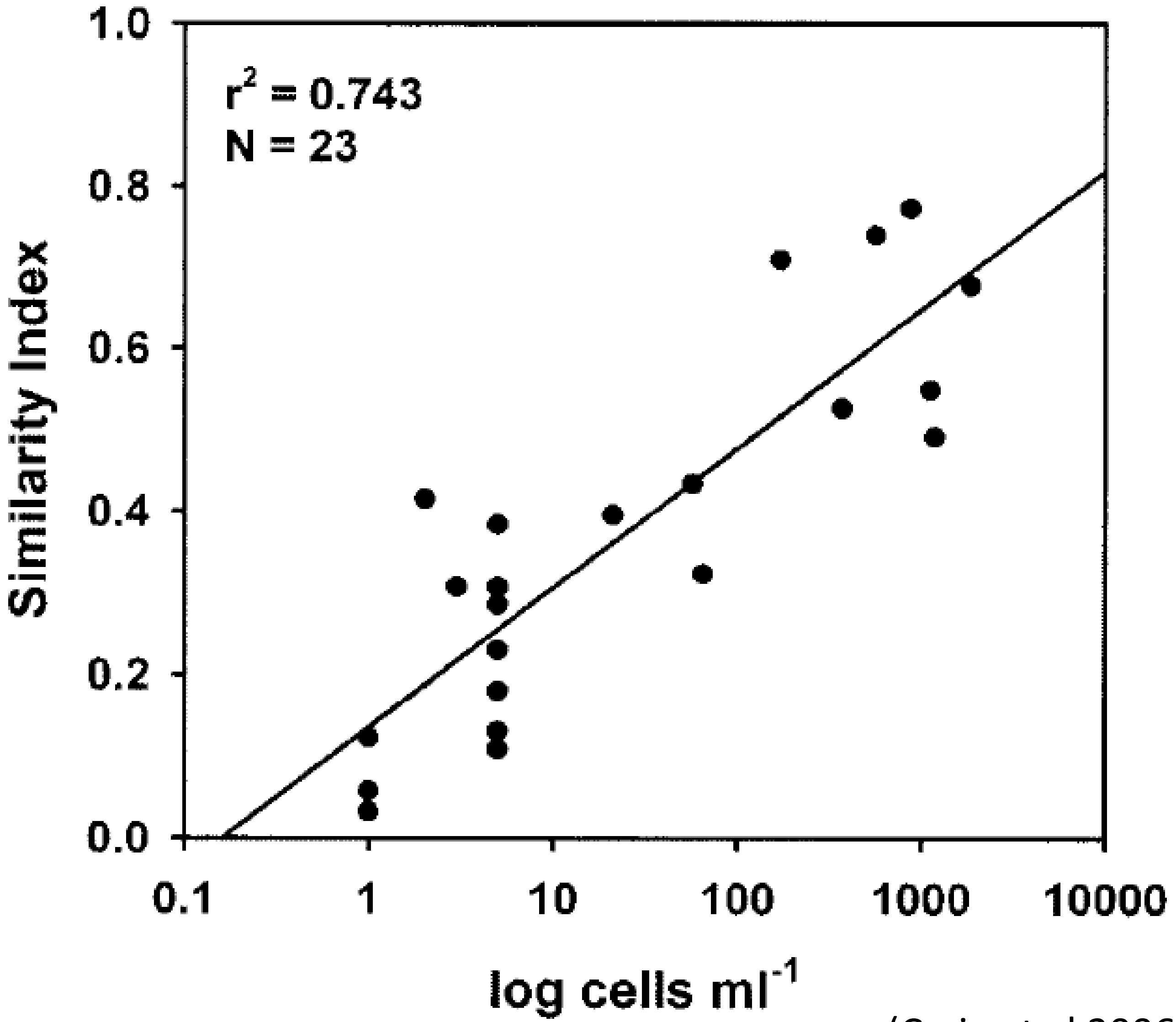
7. HAB identification-2

$$R_{rs}(\lambda) \xrightarrow{\text{QAA}} a_{ph}(\lambda) = a(\lambda) - a_{dg}(\lambda) - a_w(\lambda)$$



(Lee and Carder 2004)

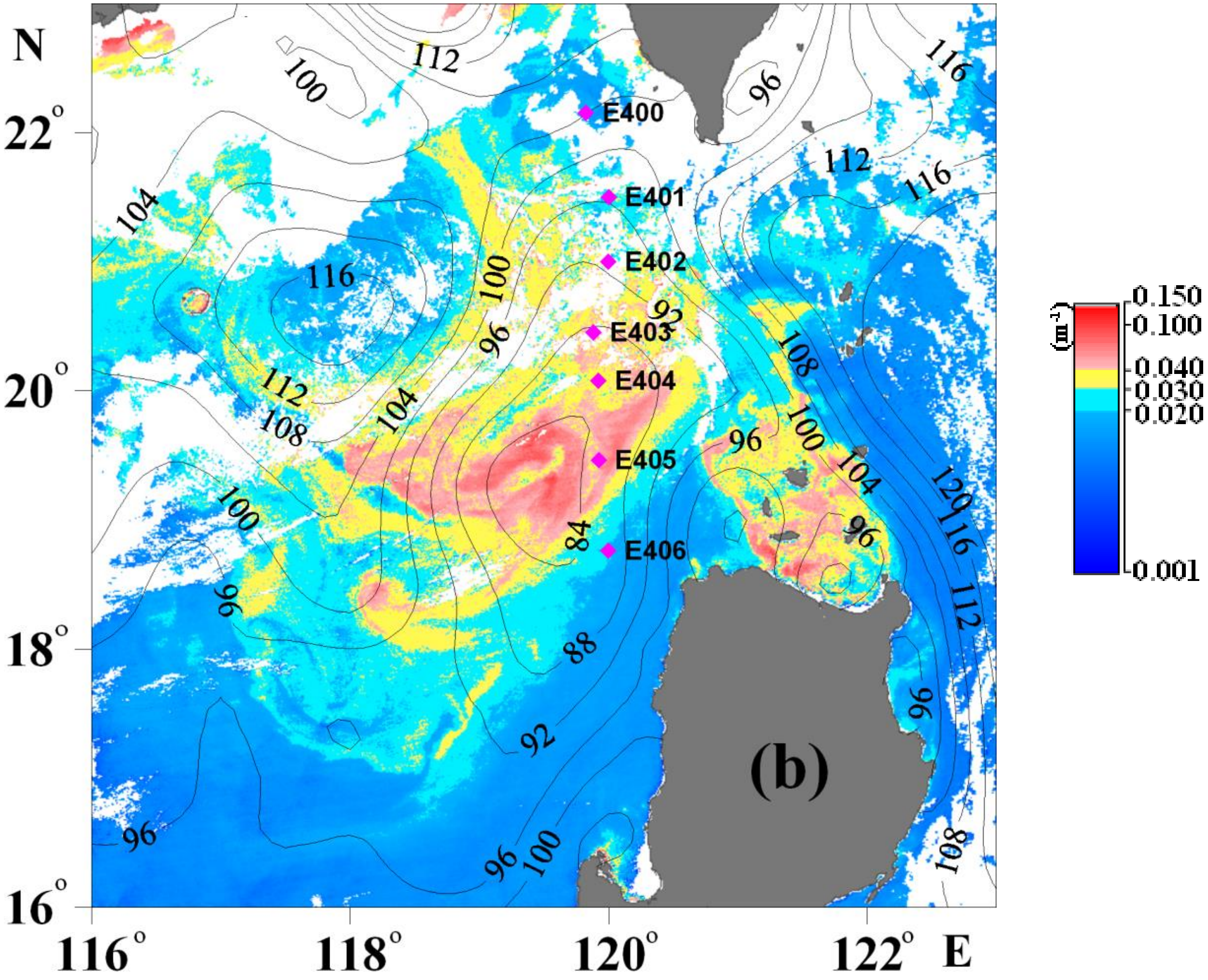
K. brevis



(Craig et al 2006)

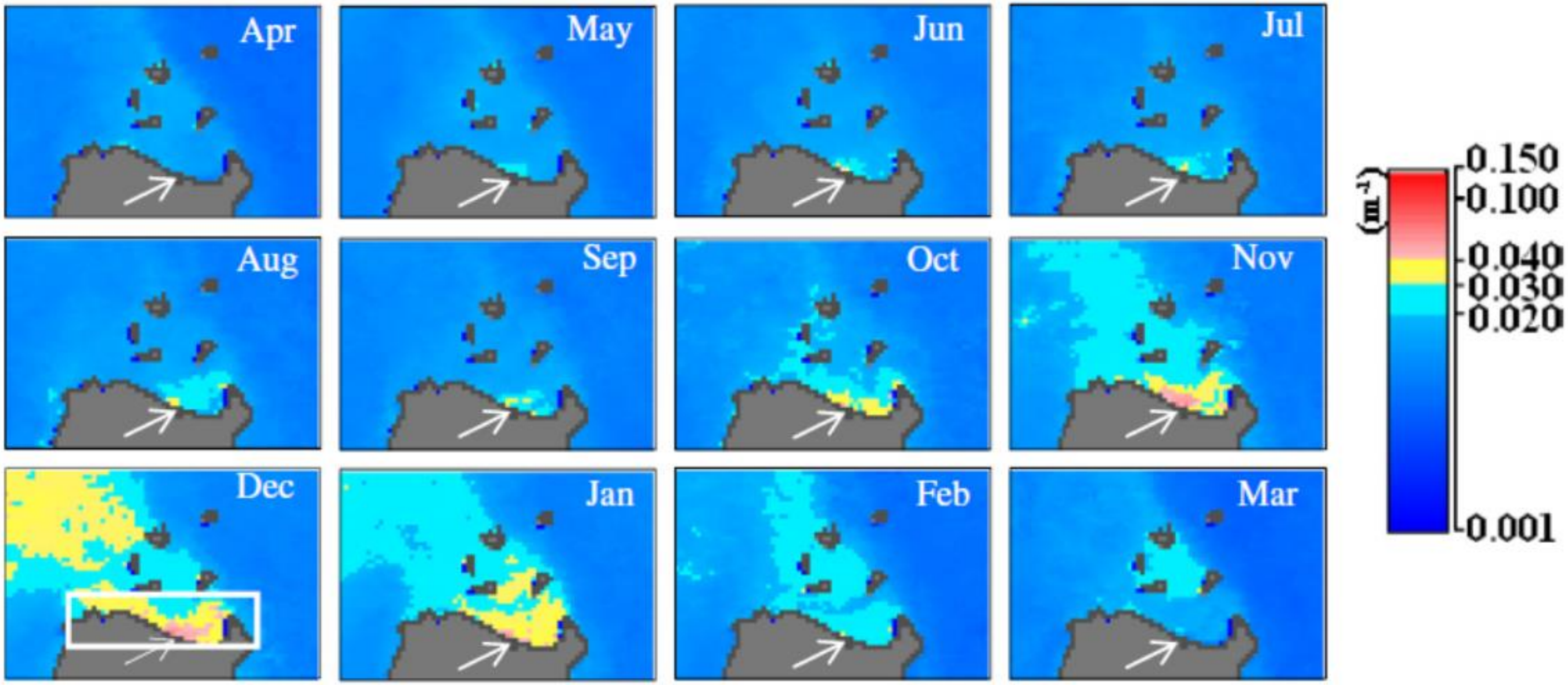
8. Bloom dynamics

Distribution of a_{ph}(440) at Luzon Street



(Shang et al, 2012)

Monthly distribution of $a_{ph}(440)$ at Luzon Street



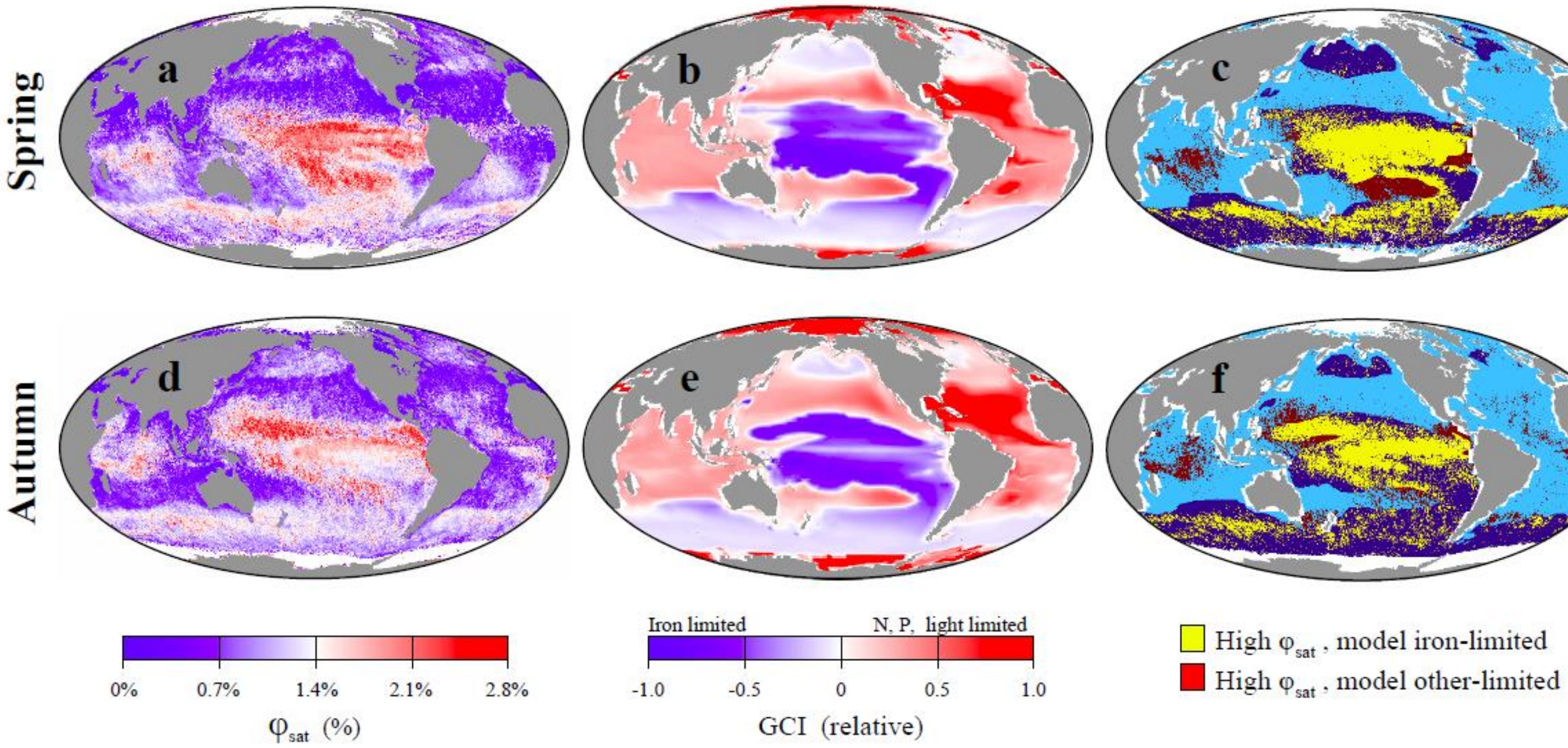
(Shang et al, 2012)

9. Global physiology of ocean phytoplankton

$$F_{\text{sat}} = \overbrace{Chl_{\text{sat}} \times <a^*_{ph}>}^{\textbf{IOP}} \times \phi \times S, \qquad (2)$$

ϕ: quantum yield of fluorescence

(Behrenfeld et al 2009)



(Behrenfeld et al 2009)

10. Salinity estimation

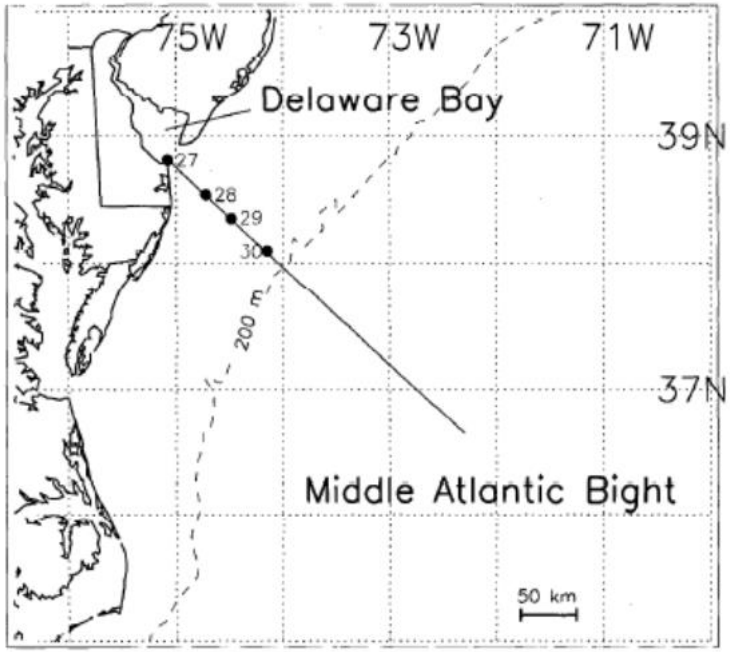
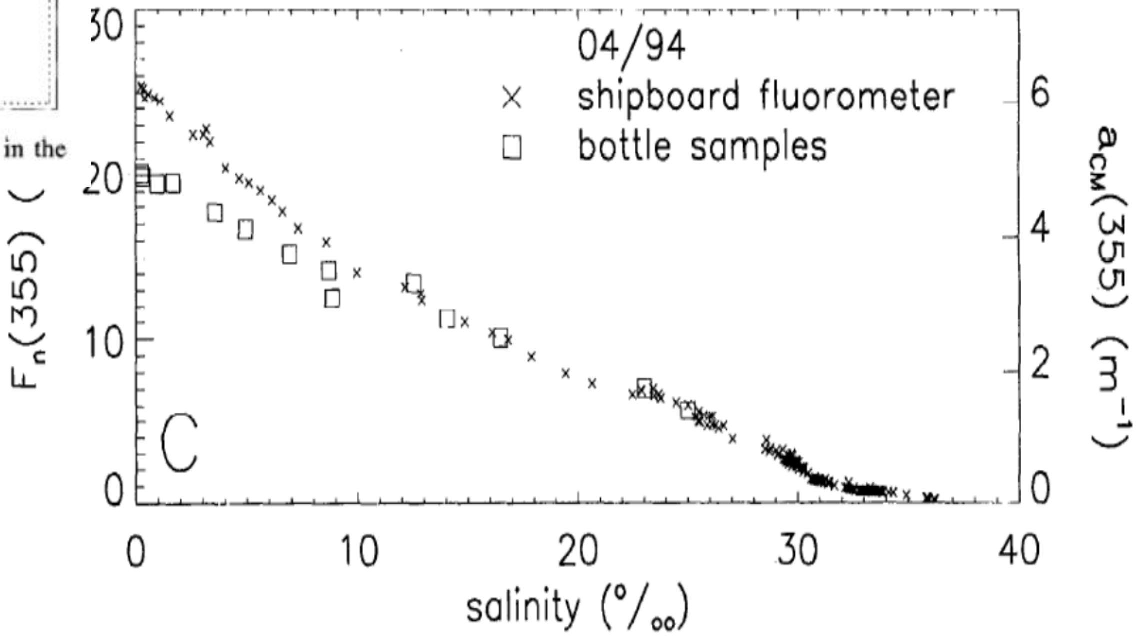
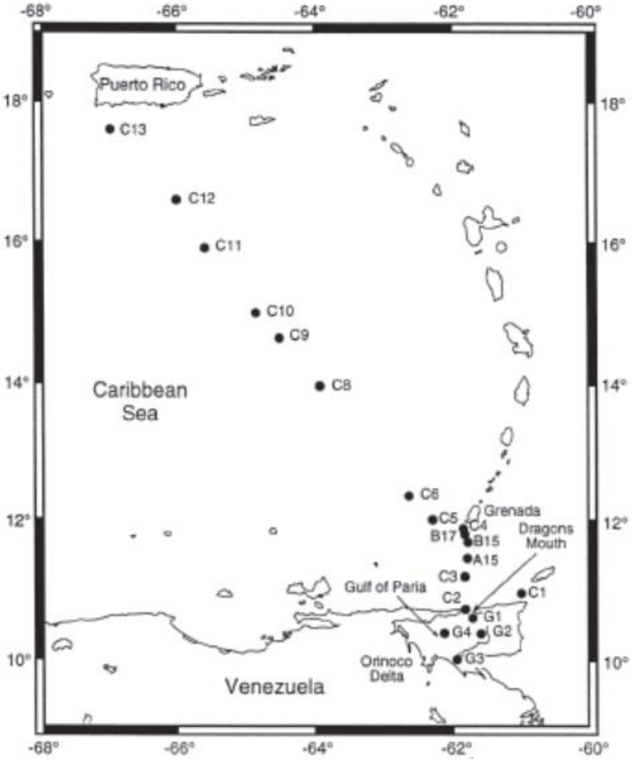
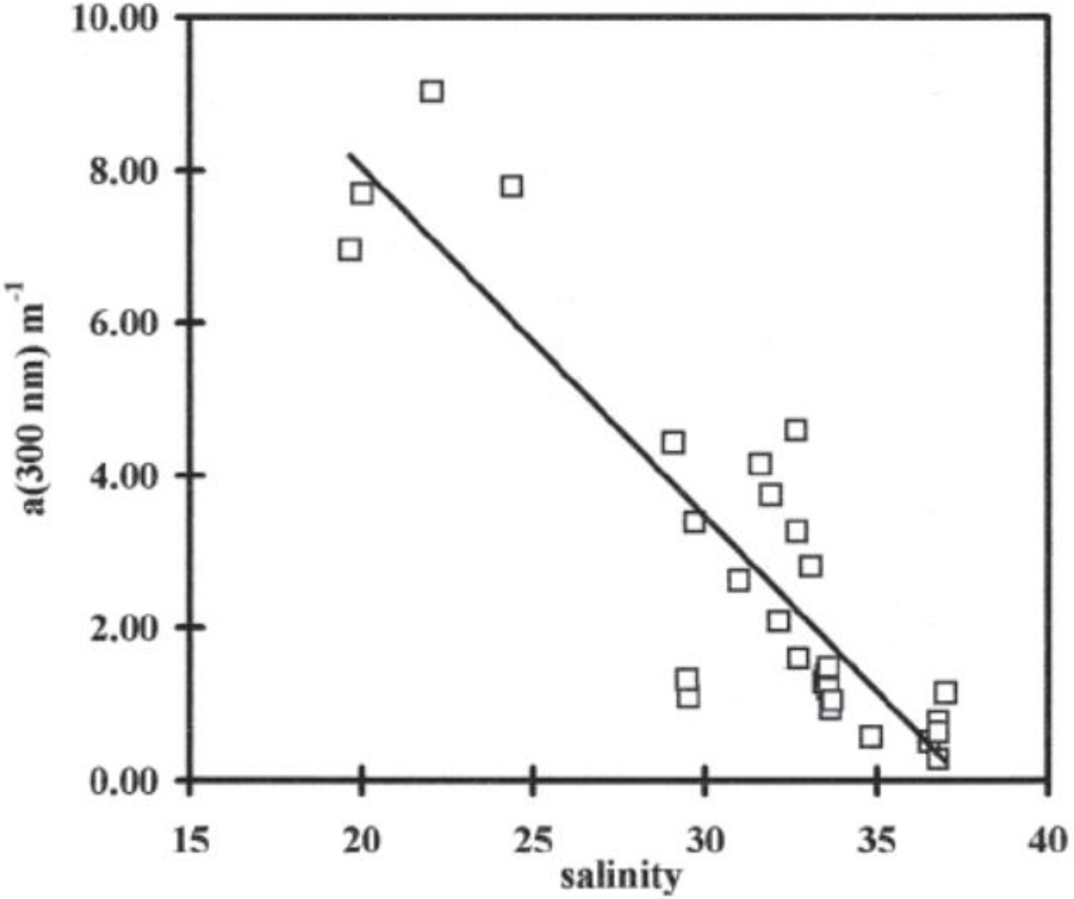


Fig. 1. Locations of the transect and hydrocast stations in the Middle Atlantic Bight.



(Vodacek et al 1997)

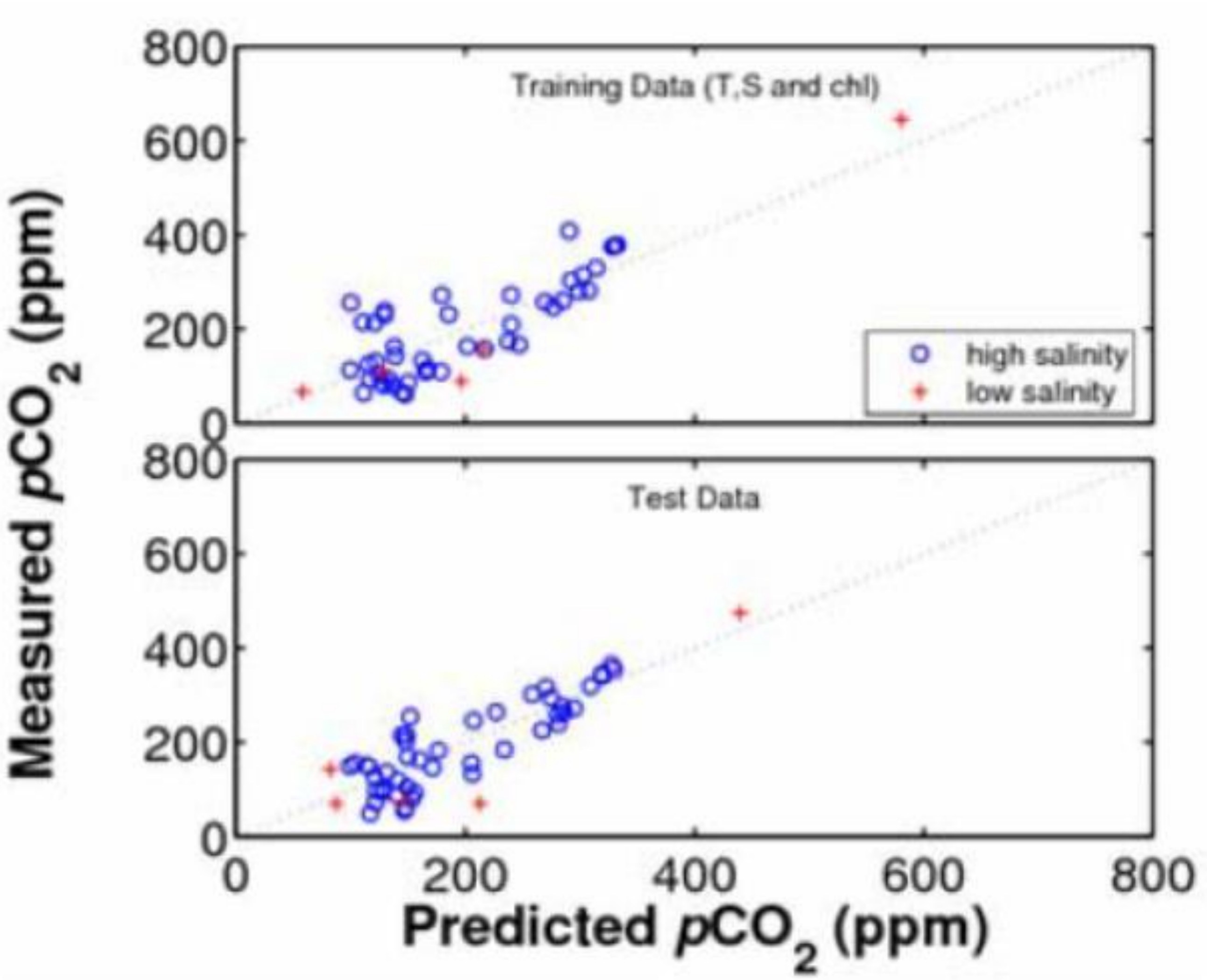




$$\mathrm{SSS} = \mathrm{x}\, a_{\mathrm{CDOM}} + \mathrm{y}$$

(Castellio et al 1999)

11. pCO2 estimation



Lohrenz and Cai (2006)

pCO2 = f(T, S, Chl)

Key Points:

1. Many applications traditionally built around [Chl] can be built around IOPs.

2. Remote sensing and applications centered around IOPs avoided, when necessary, concentration-normalized optical properties.

3. With IOPs as inputs, many products, e.g. Kd, Zeu, Zsd, PP, could be estimated easily and more accurately.

4. When IOPs are known, many other applications could be carried out. **Be creative!!!**

Thank you!