Lectures on
Optical Oceanography and
Ocean Color Remote Sensing

Introduction to Remote Sensing

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This Lecture

- Basic terminology used in ocean color remote sensing
- Forward and inverse models: Remote-sensing is a radiative transfer inverse problem
- AOPs for remote sensing
- Estimation of remote-sensing reflectance R_{rs}
    - from above-surface measurements; Carder method
    - from above-surface measurements; Lee method
    - from in-water measurements
- Introduce the Atmospheric Correction Problem (ways to solve the problem tomorrow and in week 4)

Data Resolution

The quality of remote sensing data is determined by the spatial, spectral, radiometric, and temporal resolutions.

- Spatial resolution: The “ground” size of a pixel, typically ~1 m for airborne to ~1000 meters for satellite systems
- Spectral resolution: The number and width of the different wavelength bands recorded.
- Radiometric resolution: The number of discrete levels of radiation the sensor is able to distinguish. Typically ranges from 8 to 14 bits, corresponding to $2^8 = 256$ to $2^{14} = 16,384$ levels or "shades" of color in each band. Useable resolution depends on the instrument noise.
- Temporal resolution: The frequency of flyovers by the sensor. Relevant for time-series studies, or if cloud cover over a given area makes it necessary to repeat the data collection.
- 4H = high spatial, high spectral, high SNR, high temporal

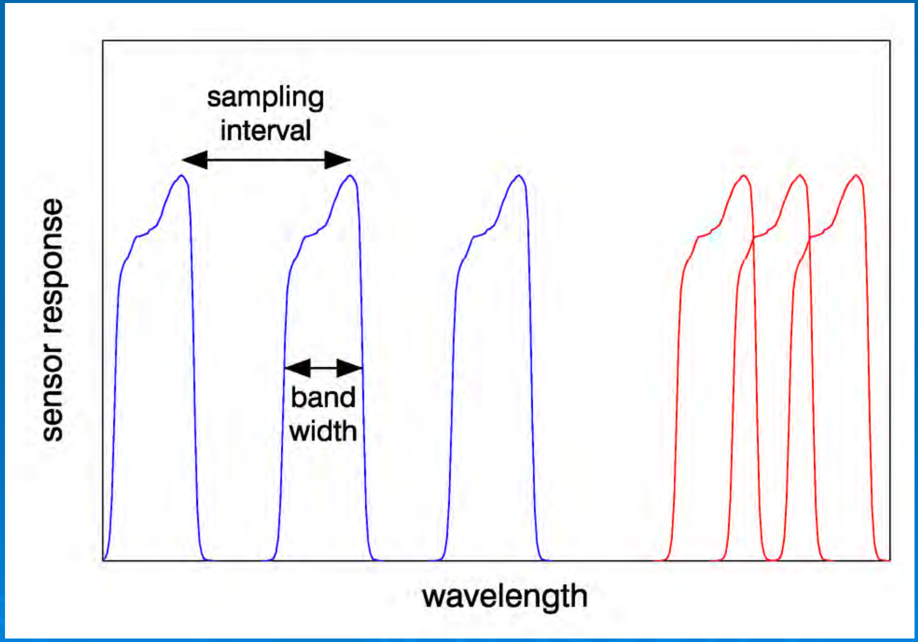
Spectral Resolution

Monochromatic:
1 very narrow wavelength band, e.g. at a laser wavelength

Panchromatic:
1 very broad wavelength band, usually over the visible range (e.g., a black and white photograph)

Multispectral:
Several (typically 5-10) wavelength bands, typically 10-20 nm wide

Hyperspectral:
30 or more bands with 10 nm or better resolution
Typically have >100 bands with ~5 nm resolution



NASA Data Processing Levels

- Level 0: Unprocessed instrument data at full resolution (volts, digital counts)

- Level 1a: Unprocessed instrument data at full resolution, but with radiometric and geometric calibration coefficients and georeferencing parameters appended, but not yet applied, to the Level 0 data.

- Level 1b: Level 1a data that have been processed to sensor units (e.g., radiance units) by application of the calibration coefficients. Level 0 data are not recoverable from level 1b data. Science starts with Level 1b data.

  Atmospheric correction converts Level 1b TOA radiance to normalized reflectance $[\rho]_N$ at the start of Level 2

- Level 2: Derived geophysical variables (e.g., normalized water-leaving reflectance, chlorophyll concentration, CDOM absorption, bottom depth) at the same resolution and location as Level 1 data.

- Level 3: Variables mapped onto uniform space-time grids, usually with missing points interpolated, complete regions mosaiced together from multiple orbits, etc.

- Level 4: Model output or results from analyses of lower level data (i.e., variables not measured by the instruments but instead derived from these measurements).

Modeling

A *model* is a representation of the real world, which tries to retain the essential features of nature while discarding the less important details.

Predictive: Predict something we don't know from something we do know, e.g., predict the radiance from the IOPs and boundary conditions (HydroLight)

vs.

Diagnostic: Analyze or transform known information, e.g., curve fitting to data to show that the data fit a given theory

Direct or Forward: E.g., solve the RTE to compute the radiance given the IOPs

vs.

Inverse: E.g., deduce the IOPs given the radiance

Approximate Analytical: E.g., the single-scattering solution of the RTE

vs.

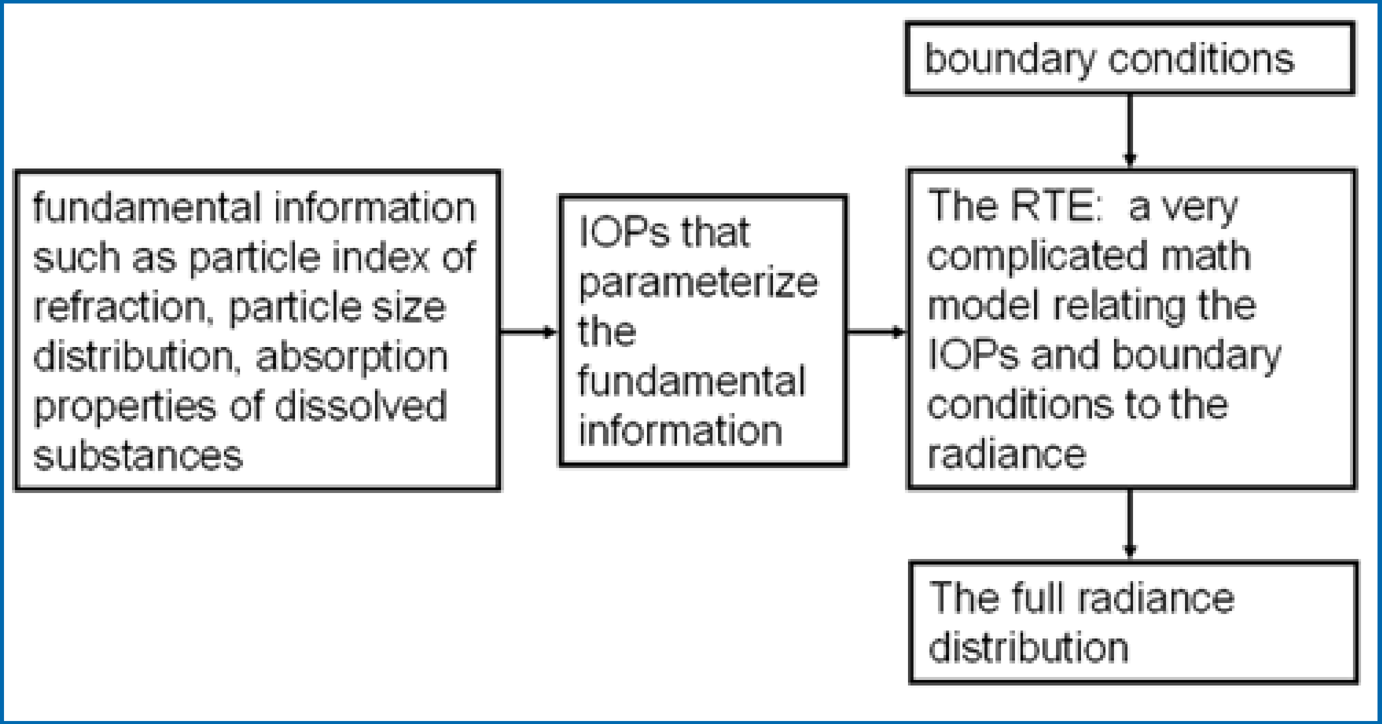
Exact Numerical: E.g., HydroLight or Monte Carlo solution of the RTE

Deterministic: No statistical noise; e.g., HydroLight solution of the RTE

vs.

Probabilistic: Has statistical noise; e.g., Monte Carlo solution of the RTE

The Radiative Transfer Forward Problem



This is a solved problem: We know how to solve the RTE. All you need is accurate inputs and computer time.

The RTE is a fixed, predictive, forward model whose variable input parameters are the IOPs and the boundary conditions, and whose output is the radiance.

Remote Sensing is an Inverse Problem

Inverse problems may have a unique solution *in principle* (e.g., if you have complete and noise-free data), but *they seldom have a unique solution in practice* (e.g., if you have incomplete or noisy data). For example, there may be more than one set of IOPs that give the same R_{rs} within the error of the R_{rs} measurement.

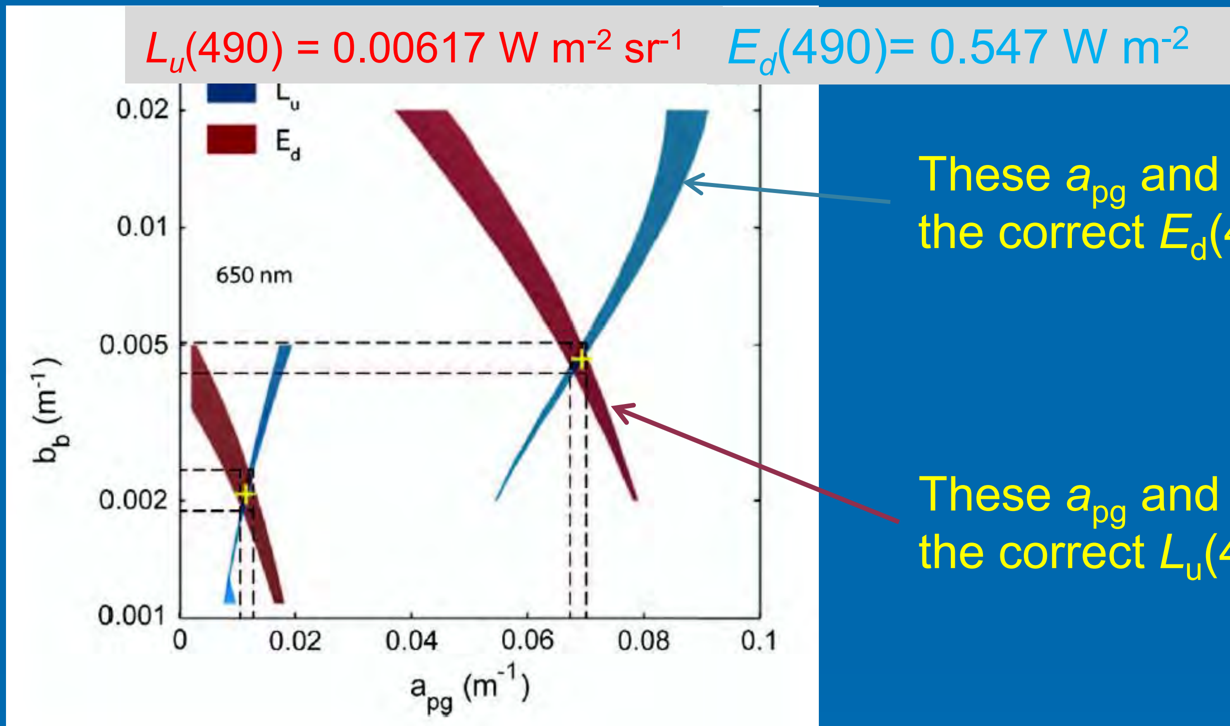
To solve an inverse problem, it is usually necessary to either

(1) add constraints on the solution, to eliminate "wrong" or unphysical mathematical solutions, or

(2) solve for only limited information given the available data (e.g., solve for only b_b/a given R_{rs})

We always have to worry about non-uniqueness when solving inverse problems, including remote sensing.

Example of Non-uniqueness when inverting L_u and E_d to get IOPs a, b, b_b



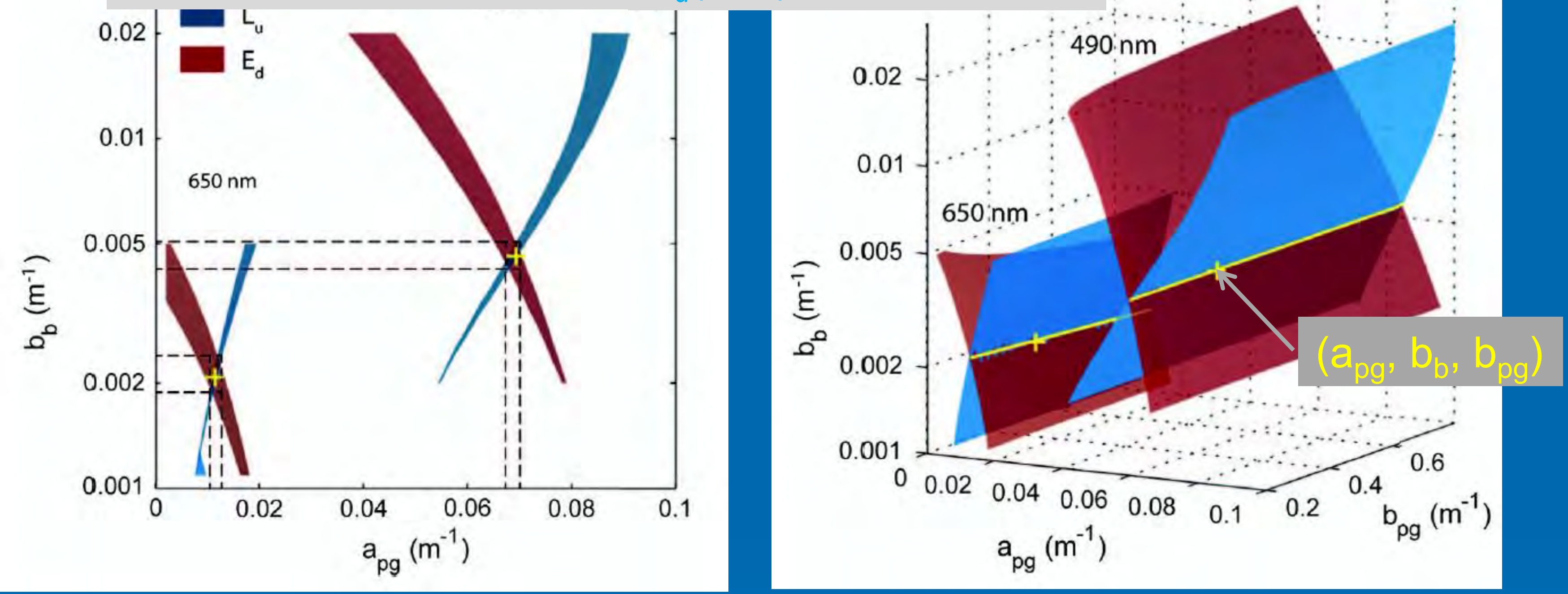
These a_{pg} and b_b values give the correct $E_d(490)$

These a_{pg} and b_b values give the correct $L_u(490)$

Isosurfaces for L_u (blue) and E_d (red) at 490 and 650 nm, representing a subset of the domain of IOPs (a_{pg}, b_b, b_{pg}) that can produce given values of L_u or E_d. The yellow cross identifies the true value of the IOP triplet for the given L_u or E_d. (a) Isosurface intersection (yellow lines) indicates the range of possible inverse solutions when given both L_u and E_d, indicating that b is completely ambiguous. (b) The projected dashed lines show the range of uncertainty in the estimated IOPs a_{pg} and b_b if no value of b is specified. From Rehm and Mobley, AO 2013.

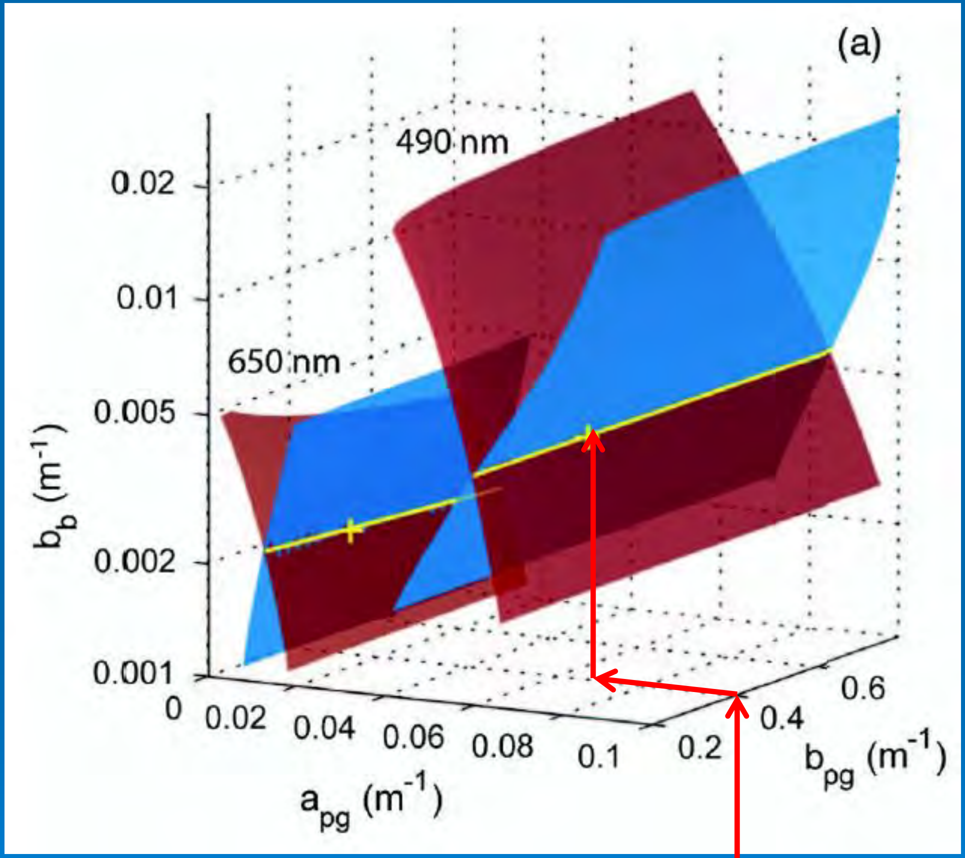
Example of Non-uniqueness when inverting L_u and E_d to get IOPs *a, b,* b_b

$L_u(490) = 0.00617$ W m^{-2} sr^{-1} $E_d(490) = 0.547$ W m^{-2}



Given just L_u and E_d, we can retrieve a_{pg} and b_b, but not also total b_{pg}. To get b_{pg}, we need to add more information, i.e., constrain the range of possible solutions for b_{pg}.

Example of Non-uniqueness when inverting L_u and E_d to get IOPs a, b, b_b



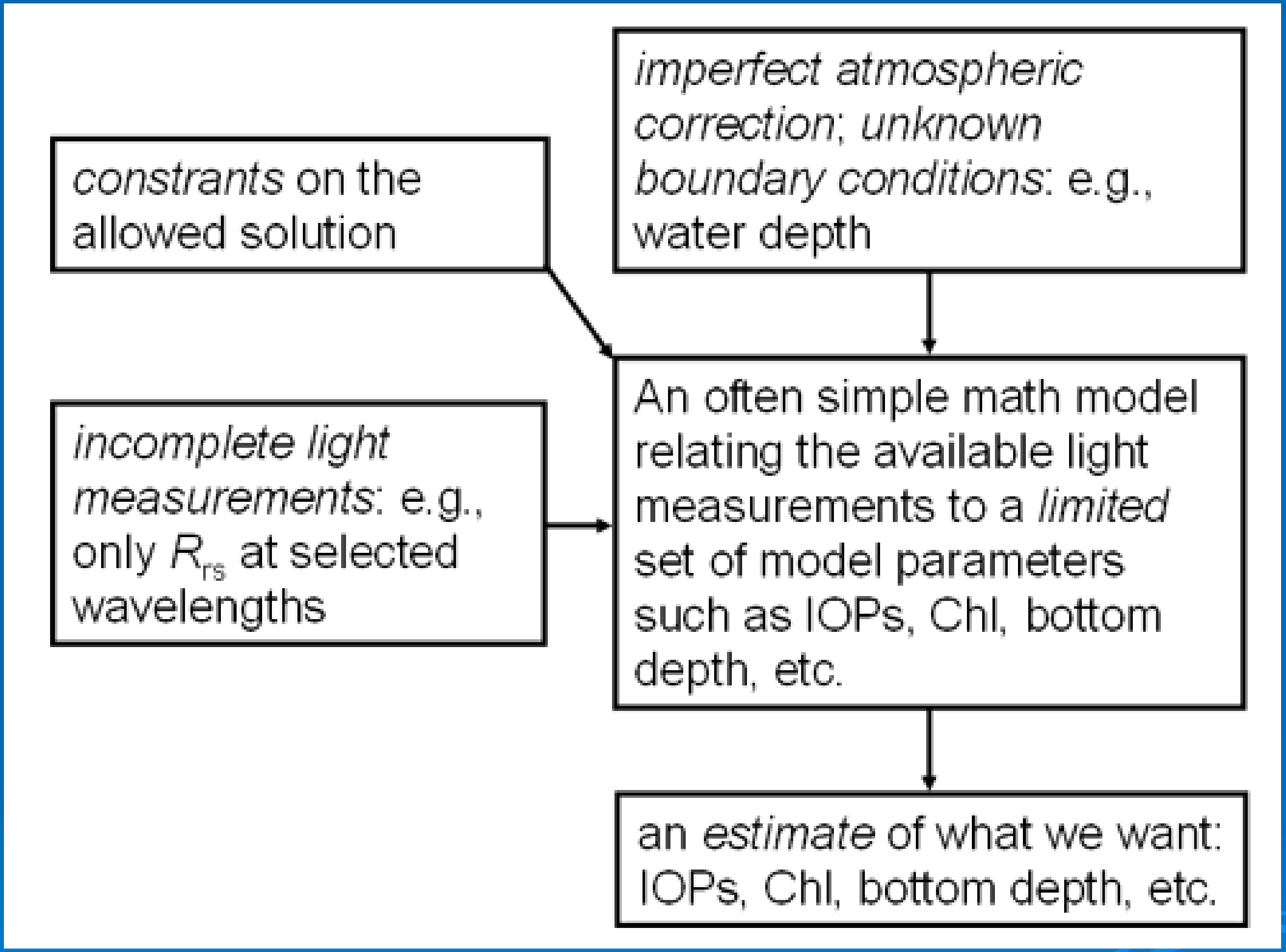
Suppose we use L_u and E_d to retrieve a and b_b, which can be done well.

Then if we know beam c, we can get $b = c – a$.

The value of c is a constraint on the inverse solution. The value of c adds information, and allows us to remove the non-uniqueness of the b value.

The constraint for c pins down the (a, b, b_b) solution point.

The Remote-Sensing Inverse Problem



This is NOT a solved problem. There are many models for retrieval of the same thing (based on different simplifications and data sets), and there are uniqueness problems.

Explicit and Implicit Inverse Problems

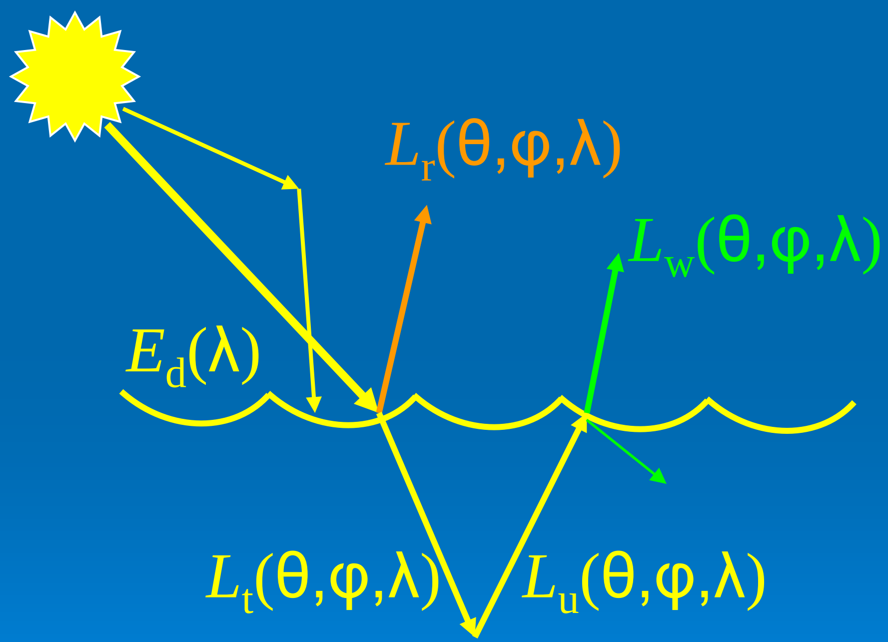
Explicit solutions are formulas that give the desired IOPs as functions of measured radiometric quantities or AOPs. A simple example is Gershun's law, $a = -(1/E_o)\, d(E_d - E_u)/dz$, when solved for the absorption in terms of the irradiances.

Implicit solutions are obtained by *solving a sequence of direct or forward problems*. In crude form, we can imaging having a measured remote-sensing reflectance (or set of underwater radiance or irradiance measurements). We then solve direct problems to predict the reflectance for each of many different sets of IOPs. Each predicted reflectance is compared with the measured value. The IOPs associated with the predicted reflectance that most closely matches the measured reflectance are then taken to be the solution of the inverse problem. Such a plan of attack can be efficient if we have a rational way of changing the IOPs from one direct solution to the next, so that the sequence of direct solutions converges to the measured reflectance or radiance.

Recall 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.

AOPs for Remote Sensing

Radiometric variables such as radiance or irradiance depend not just on the IOPs of the water column, but also on the incident lighting (sun angle, sky conditions). Therefore, it is hard to separate IOP and boundary effects on the (ir)radiance. We therefore usually do not use a radiometric variable such as upwelling radiance for ocean color remote sensing.

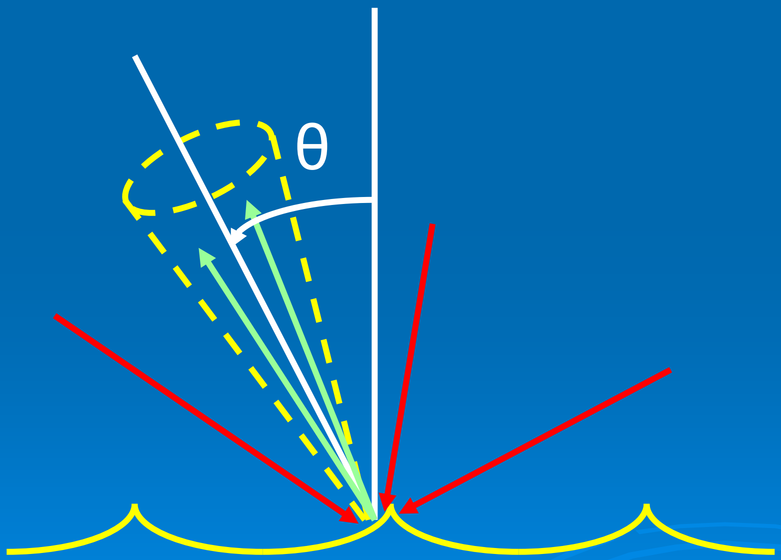
We need to find an apparent optical property (AOP) that is strongly dependent on the IOPs of the ocean water column, but which is only weakly dependent on sun angle, sky conditions, surface waves, etc.

We have already seen that L_w normalized by E_d is a good AOP.

Remote-sensing Reflectance R_{rs}

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

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

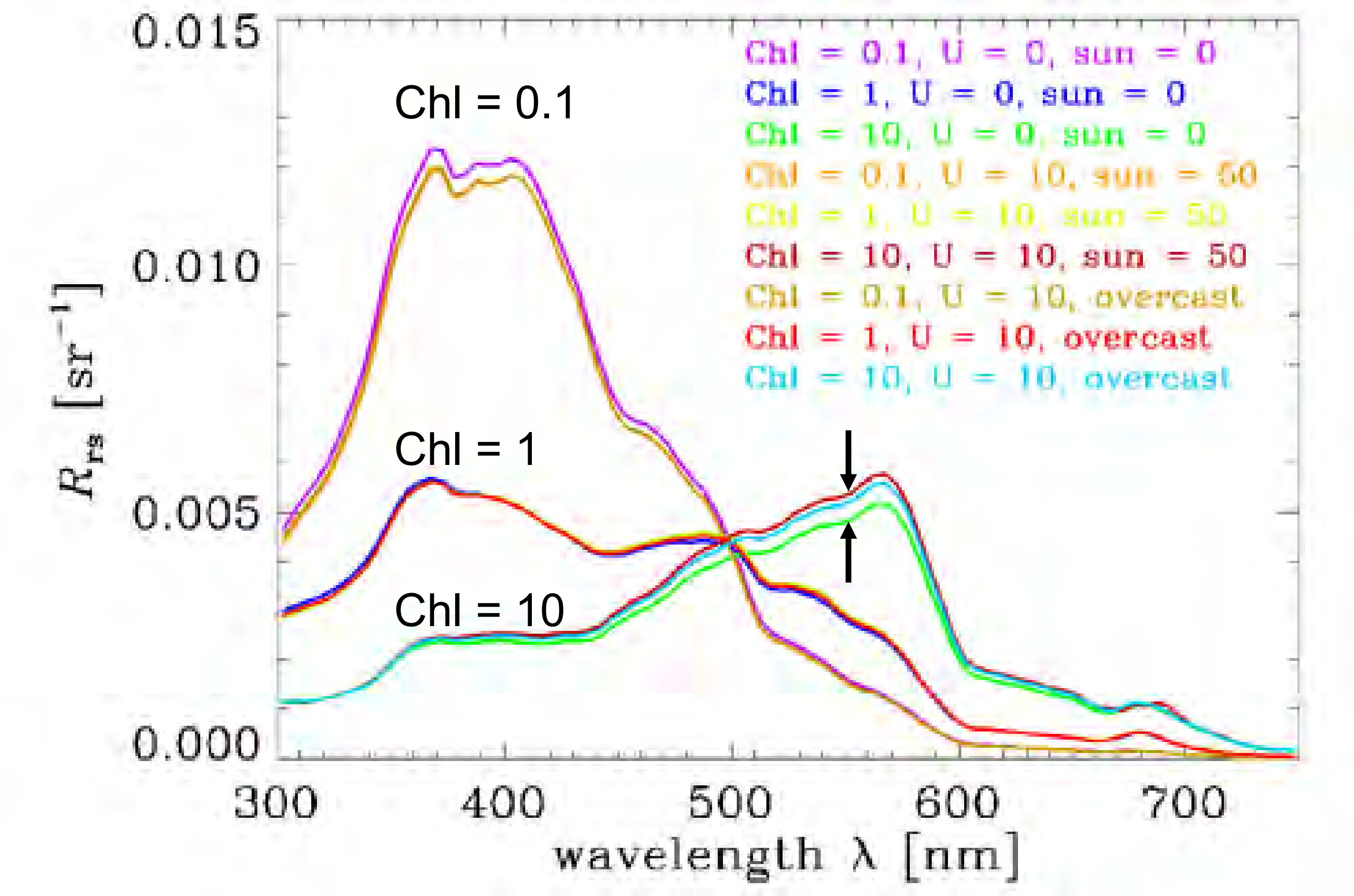


sea surface

The fundamental quantity used today in ocean color remote sensing

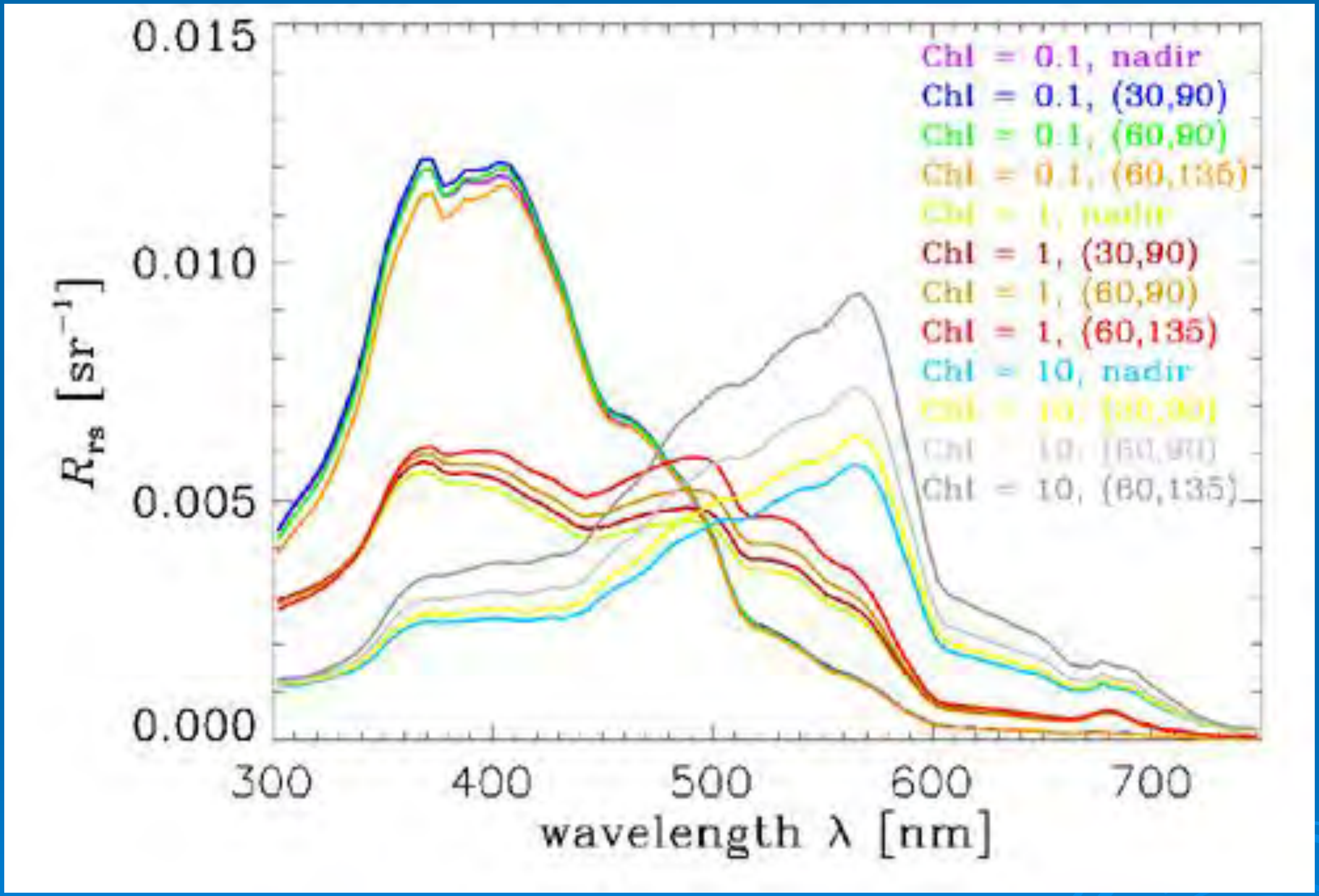
Often use the nadir-viewing R_{rs}, i.e. the radiance that is heading straight up from the sea surface ($\theta = 0$)

Dependence of R_{rs} on IOPs and Environmental Conditions



Curves separate by *Chl* value, and show very little dependence on sky conditions and wind speed: R_{rs} is a much better AOP than R.

Dependence of R_{rs} on IOPs and Viewing Direction



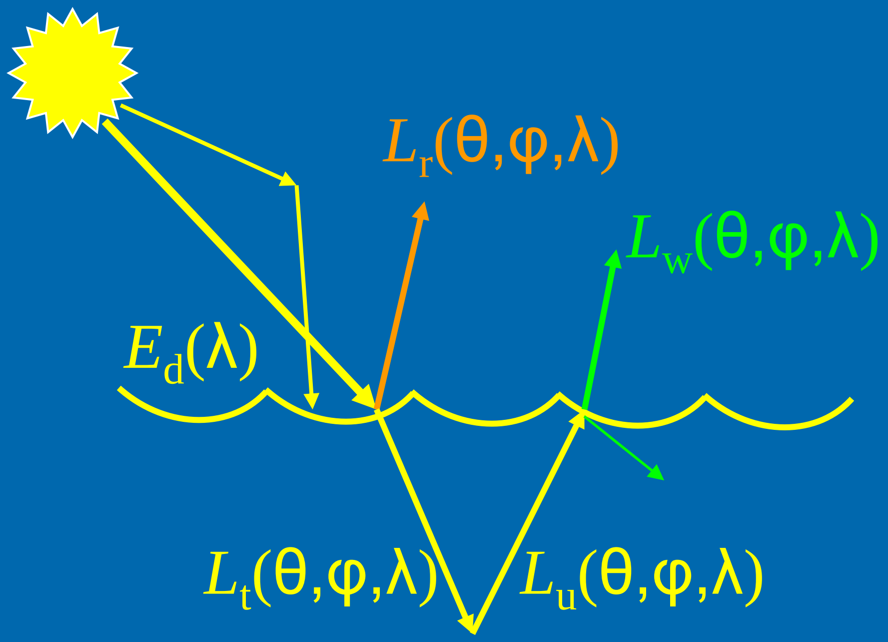
Not much dependence on off-nadir viewing direction until > 30 deg. for high *Chl* values. OK for most remote sensing geometries.

How Do We Get L_w?

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

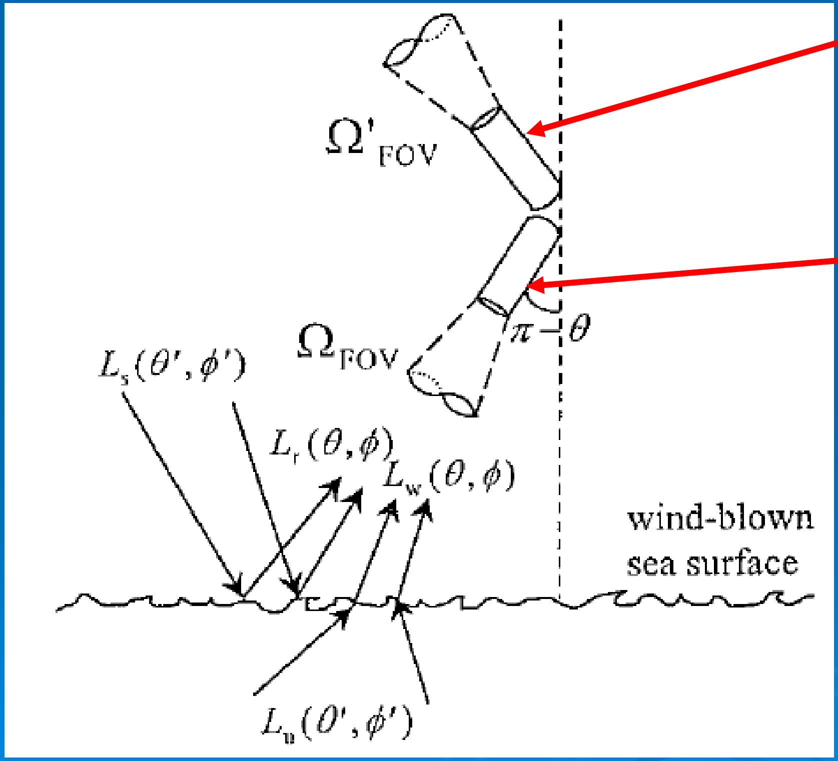
(1) If we have in-water profiles of L_u, we can extrapolate from below the sea surface to get L_w above the surface.

(2) If we have above-surface measurements of L_u, we must remove the contribution of the surface-reflected radiance to get $L_w = L_u - L_r$



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.

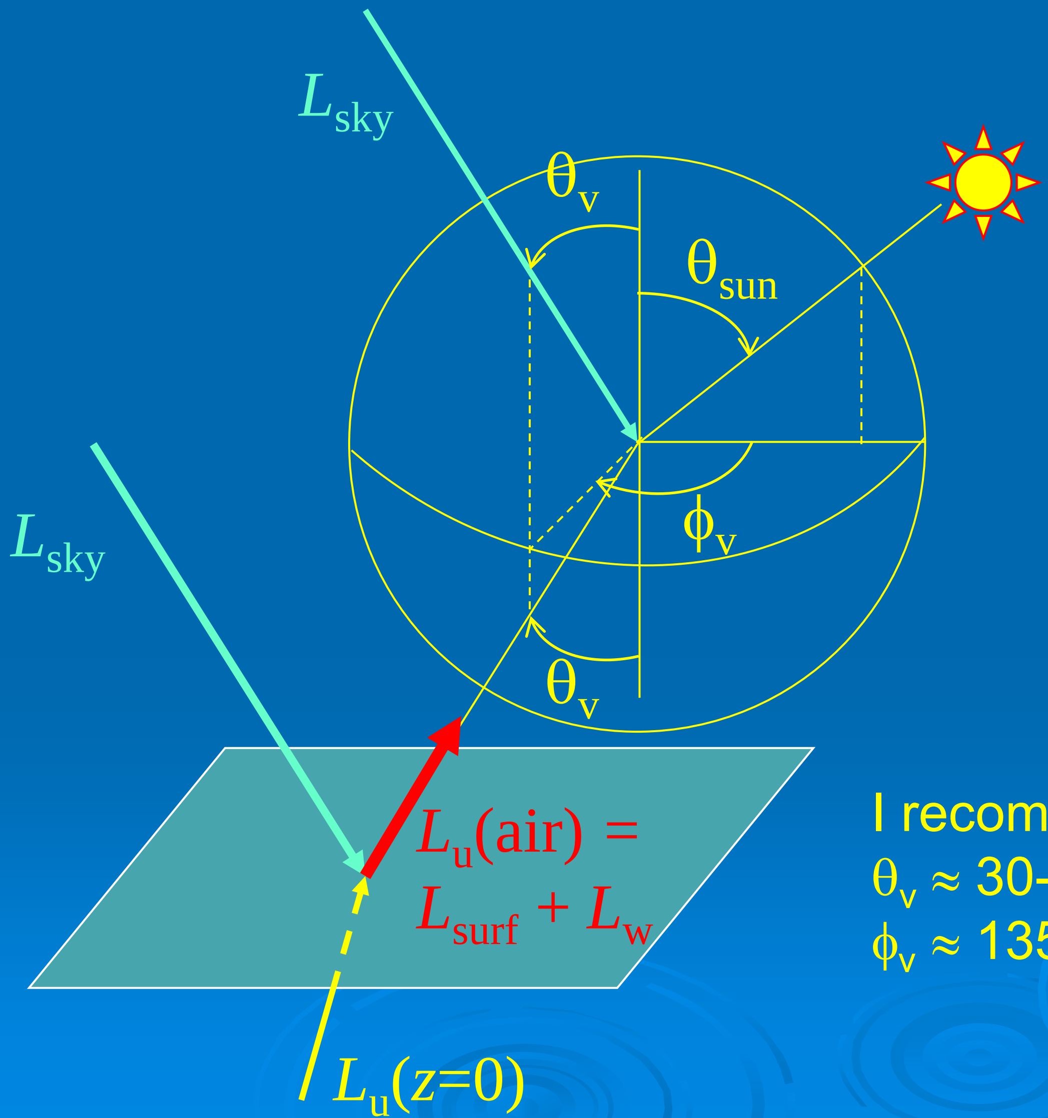


radiometer pointing upward measures $L_{sky}(\theta, \phi, \lambda)$

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

Mobley, AO, 1999

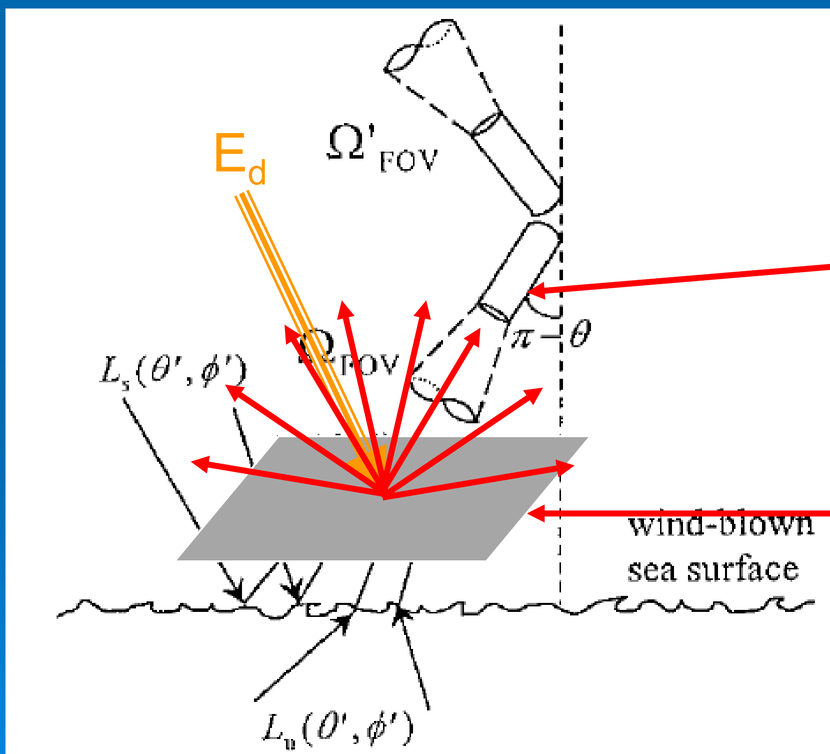
Estimating L_w and R_{rs} from Above-surface Measurements: The Carder Method



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

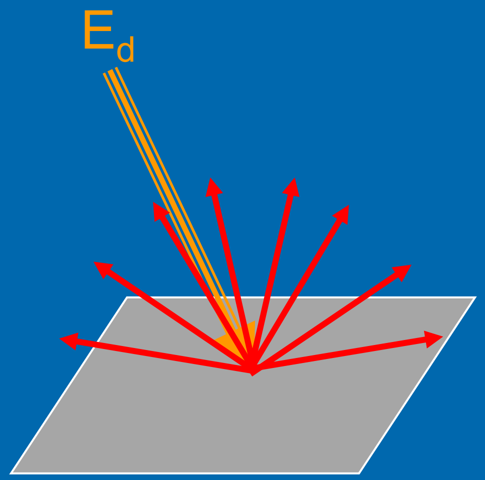
Next measure the radiance reflected by a "gray card" (often a Spectralon plate) with known irradiance reflectance $R_g(\lambda)$.



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

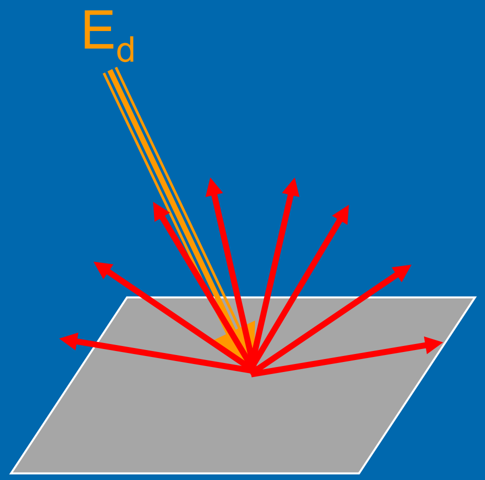
$$L_g = (R_g/\pi)E_d \qquad \text{(can solve for } E_d\text{)}$$

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



Retrieval algorithms usually use R_{rs}, not L_w, as their input. 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?

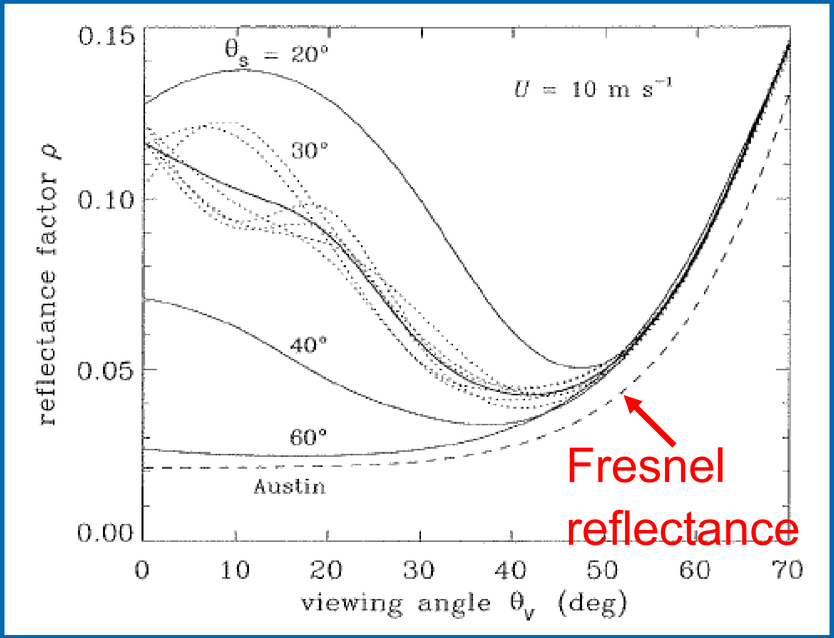
Estimating the Radiance Reflectance ρ

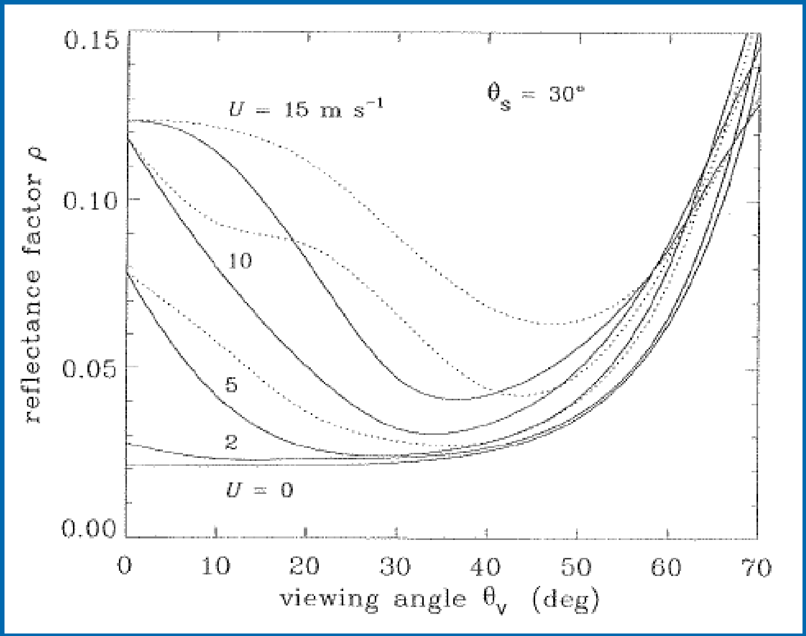
The radiance reflectance ρ depends on viewing direction, sky conditions, sea-surface wave conditions, and wavelength.
ρ is therefore NOT an IOP, and it is NOT equal to the Fresnel reflectance of the surface, except for a level sea surface.

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 Library (file rhoTable_AO1999.txt) (broad-band average values).

Dependence of ρ on Geometry and Wave State

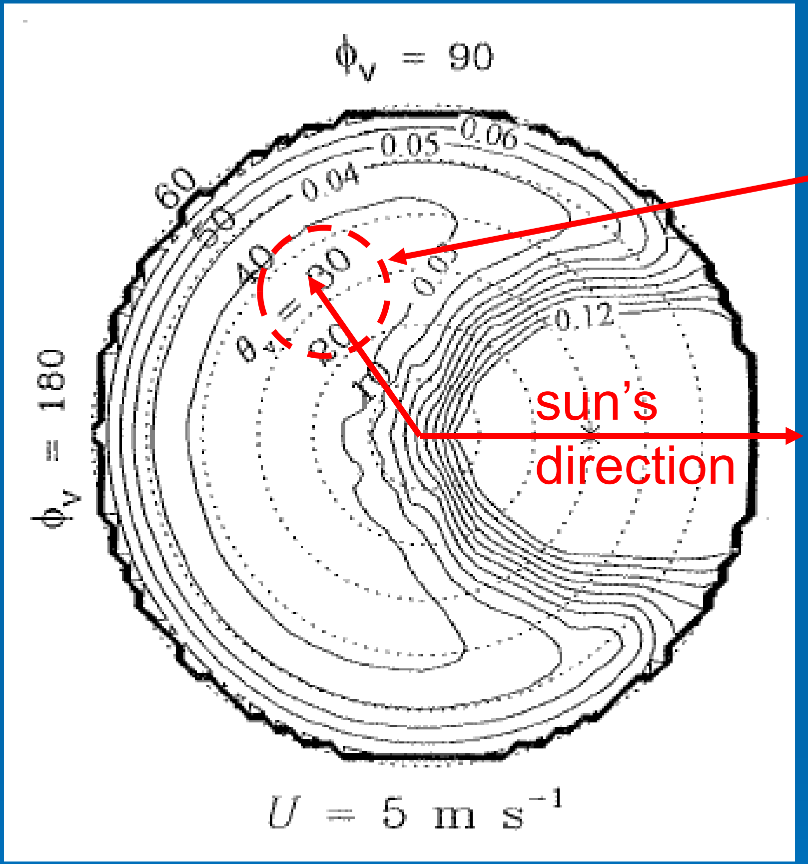




ρ as a function of sun zenith angle for a given wind speed

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.

Dependence of ρ on Wavelength

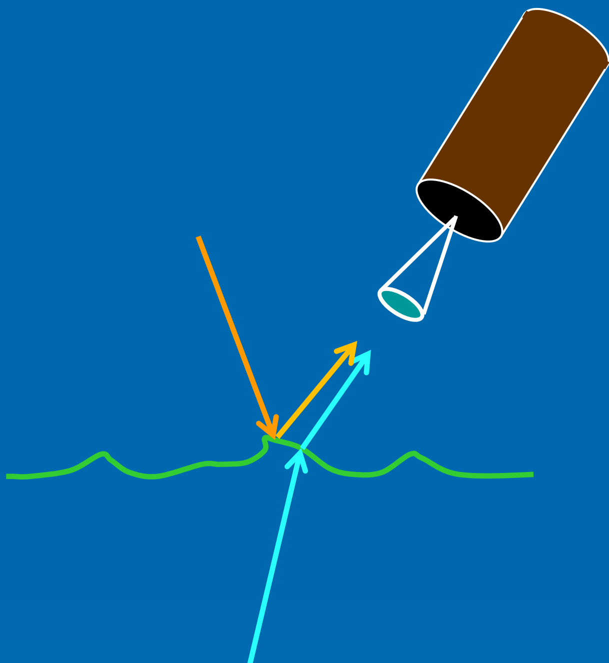
The value of ρ depends on the angular distribution of the sky radiance.

The sky radiance distribution depends on wavelength: more diffuse at blue wavelengths due to Rayleigh scattering; more direct at red wavelengths.

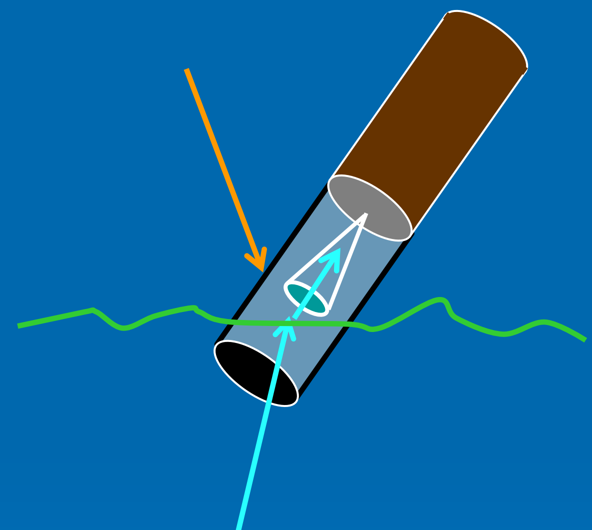
Therefore, ρ also depends on wavelength.

The $\rho(\lambda)$ dependence is significant; see Lee et al. (2010).

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



Carder method: Surface-reflected radiance is detected and must be removed



Lee method: Surface-reflected radiance is blocked by a tube; only L_w is detected

Estimating L_w and R_{rs} from Below-surface Measurements

A second approach to estimating R_{rs} is to make in-water measurements, and then extrapolate those values upward through the sea surface. Various ways have been tried:

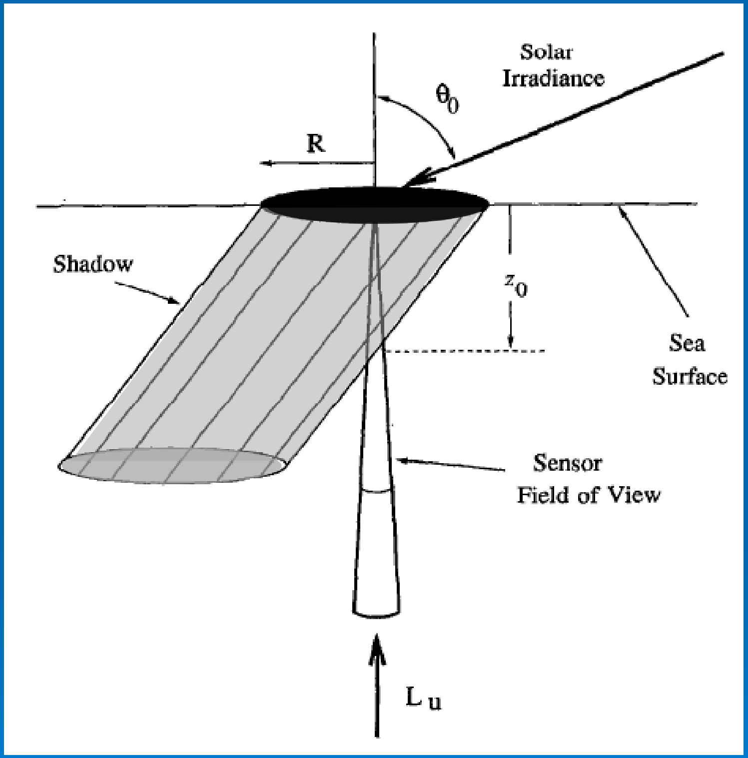
- measure $L_u(z,\lambda)$ at 1 or more depths (either at fixed depths on a mooring or using a profiling instrument). Extrapolate $L_u(z, \lambda)$ upward through the surface to get $L_w(\lambda)$; then $R_{rs} = L_w(\lambda)/E_d(\text{in air}, \lambda)$.

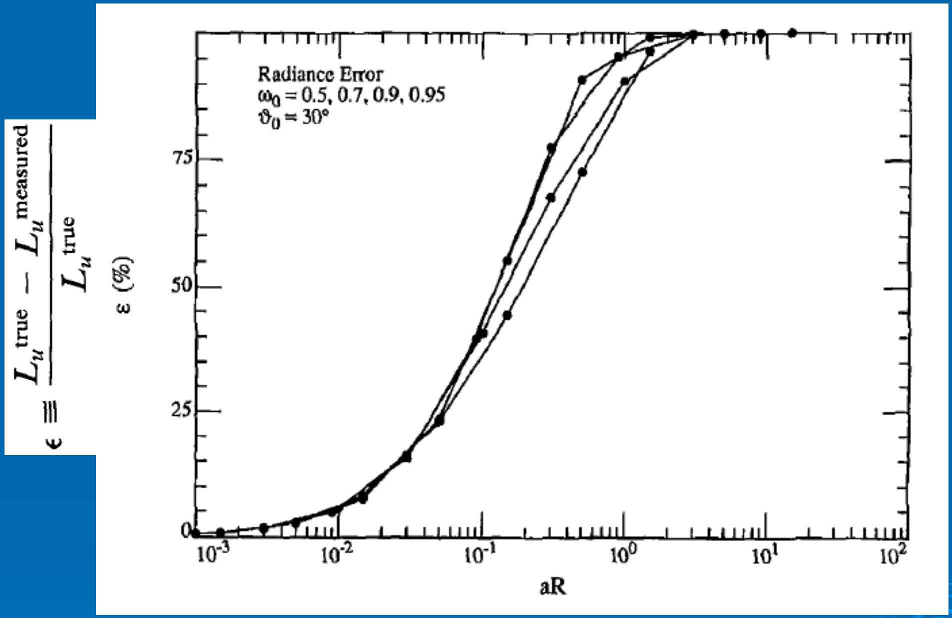
- measure $L_u(z, \lambda)$ and $E_d(z, \lambda)$ in water. Extrapolate the in-water ratio $\text{RSR}(z, \lambda) = L_u(z, \lambda)/E_d(z, \lambda)$ upward to get $R_{rs}(\lambda)$ in air.

Note that in-water measurements of L_u must be corrected for instrument self-shading. The effect of self shading depends on the water IOPs (mostly the absorption coefficient), sun zenith angle, and size and shape of the instrument.

Instrument Self-Shading Effect on L_u

An instrument measuring L_u or E_u is looking at water that is partially shaded by the instrument. The errors in L_u can be large if the water is highly absorbing and the instrument is large.





Gordon and Ding (1992)

Estimating L_w and R_{rs} from Below-surface Measurements

The upward extrapolation of $L_u(z,\lambda)$ (after correction for self shading) requires estimating $K_{Lu}(z,\lambda)$.

$K_{Lu}(z,\lambda)$ can be estimated from a profile of $L_u(z,\lambda)$ values (curve fitting), or by modeling (e.g., with HydroLight).

Extrapolating a measured profile is hard because of noise (especially wave focusing)

Computing $K_{Lu}(z,\lambda)$ requires knowing the IOPs, or using some other model for $K_{Lu}(z,\lambda)$.

Therefore, in-water methods have just as much uncertainly as above-water methods. (an example of Mobley's Law of Conservation of Misery)

Estimating L_w and R_{rs} from Below-surface Measurements

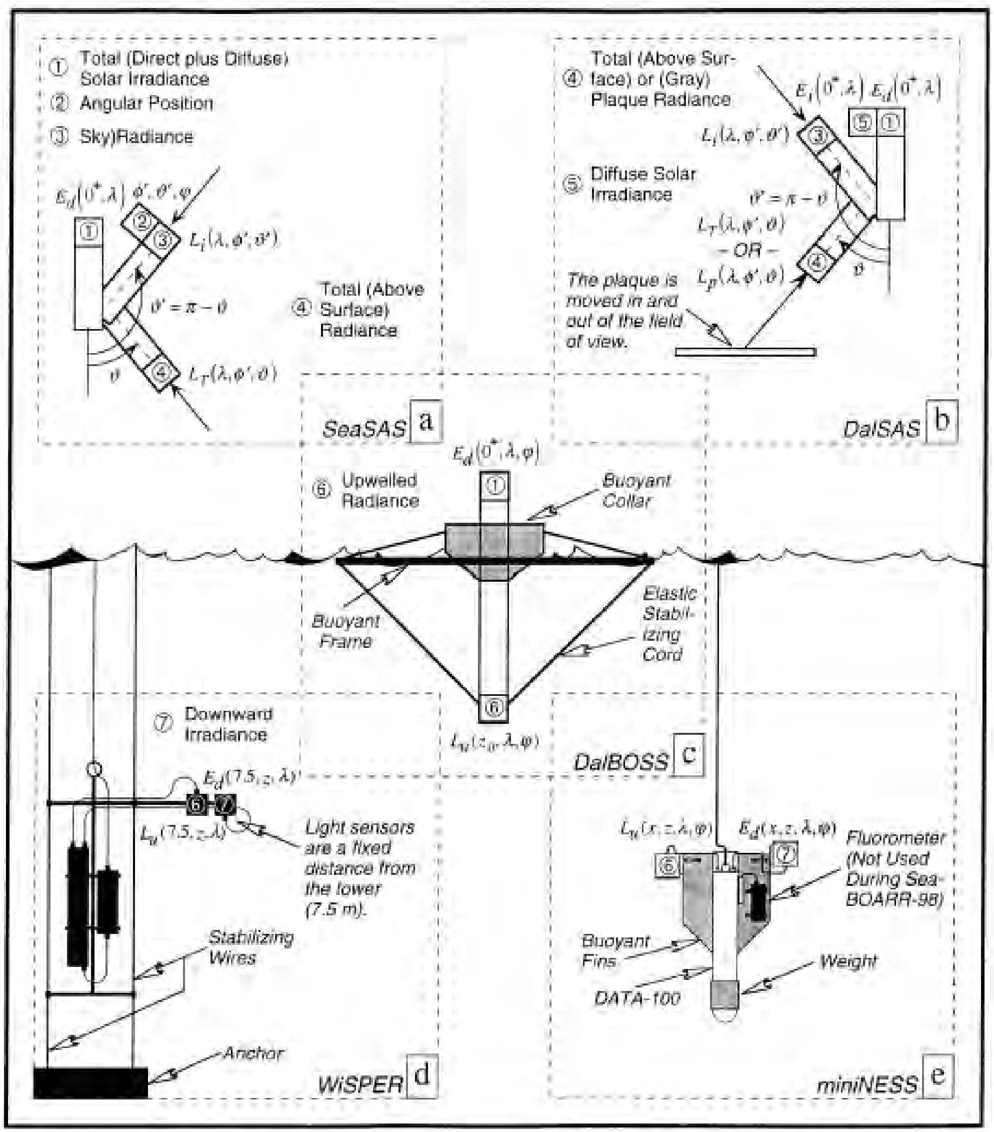


FIG. 2. A generalized schematic of the above- and in-water instruments used du... DalBOSS, (d) WiSPER, and (e) miniNESS.

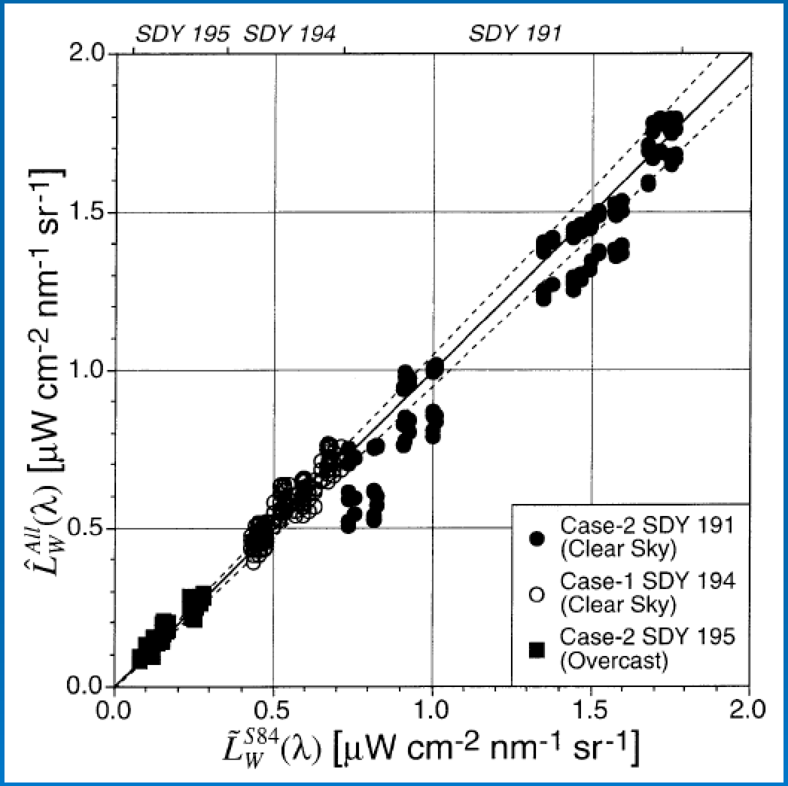
Hooker et al. (2002)

Hooker et al (2002) compared several ways of removing surface glint from above-water measurements, and several ways to extrapolate upward from in-water measurements.

In-water:

- mooring L_u and E_d at fixed depth; extrapolate L_u/E_d upward through the surface
- surface float: L_u(in water), extrapolated upward; then divide by E_d(in air)
- profiling of L_u and E_d in water; extrapolated L_u/E_d upward

Estimating L_w and R_{rs} from Below-surface Measurements



all above-water

in-water method S84

Hooker et al (2002)

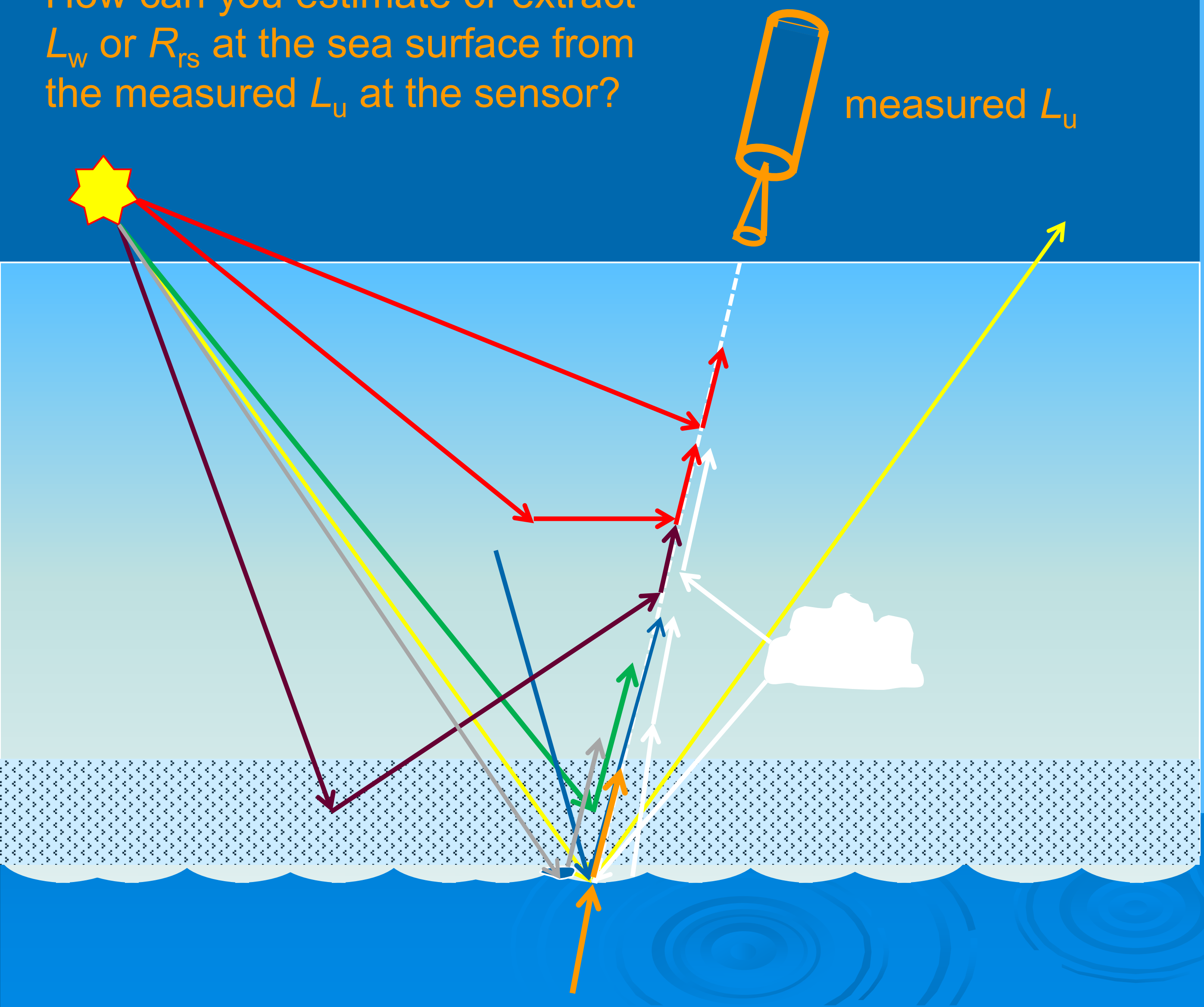
Above-water and in-water techniques should give the same L_w.

Closure can be achieved, but only with great effort and care in making the measurements and in processing the data.

See Hooker et al. (2002) for examples.

The Atmospheric Correction Problem

How can you estimate or extract L_w or R_{rs} at the sea surface from the measured L_u at the sensor?



contributions to at-sensor radiance

specular reflectance

Rayleigh (atmospheric gases)

aerosol (particles)

surface sky relf

Rayleigh-aerosol

cloud refl and adjacency effect

whitecaps

water-leaving L_w

Examples of Atmospheric Effects

The MODTRAN atmospheric model was used to compute the radiance incident onto the sea surface.

HydroLight was used with the MODTRAN incident radiance to compute the water-leaving and surface-reflected radiances.

MODTRAN was then used to propagate the water-leaving and surface-reflected radiances to the sensor, and to compute the atmospheric path radiance contribution to the measured total radiance at the sensor.

(The combined HydroLight-MODTRAN code was developed for Northrop-Grumman for design of the NPP-VIIRS sensor and is proprietary to N-G. I can show you results but not the code or algorithms. (NPP-VIIRS = National Polar-orbiting Partnership, Visible and Infrared Imaging Radiometer Suite, launched in 2011 and now operational.)

Simulated Atmosphere and Water

MODTRAN was run with typical atmospheric conditions for mid-latitude summer, marine aerosols, sun at 50 deg, cloudless sky, 6 m/s wind speed, etc. These conditions were typical of excellent remote-sensing conditions (63 km horizontal visibility at sea level)

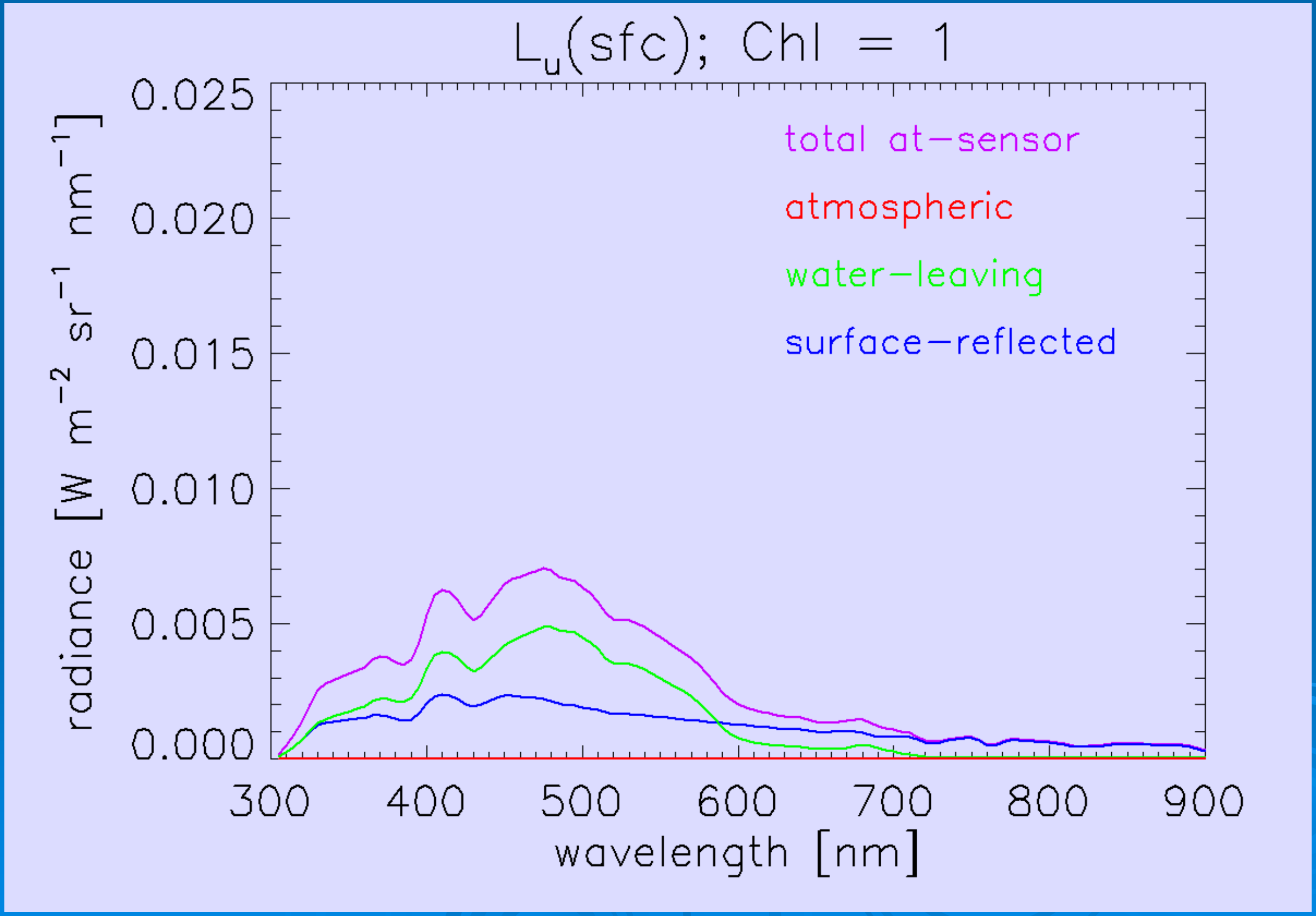
HydroLight was run for homogeneous, infinitely deep, Case 1 or 2 water.

--Case 1: used the "new" Case 1 IOP model with *Chl* = 1

--Case 2: used *Chl* = 1 plus calcareous sand mineral particles with a concentration of 2 gm/m^3

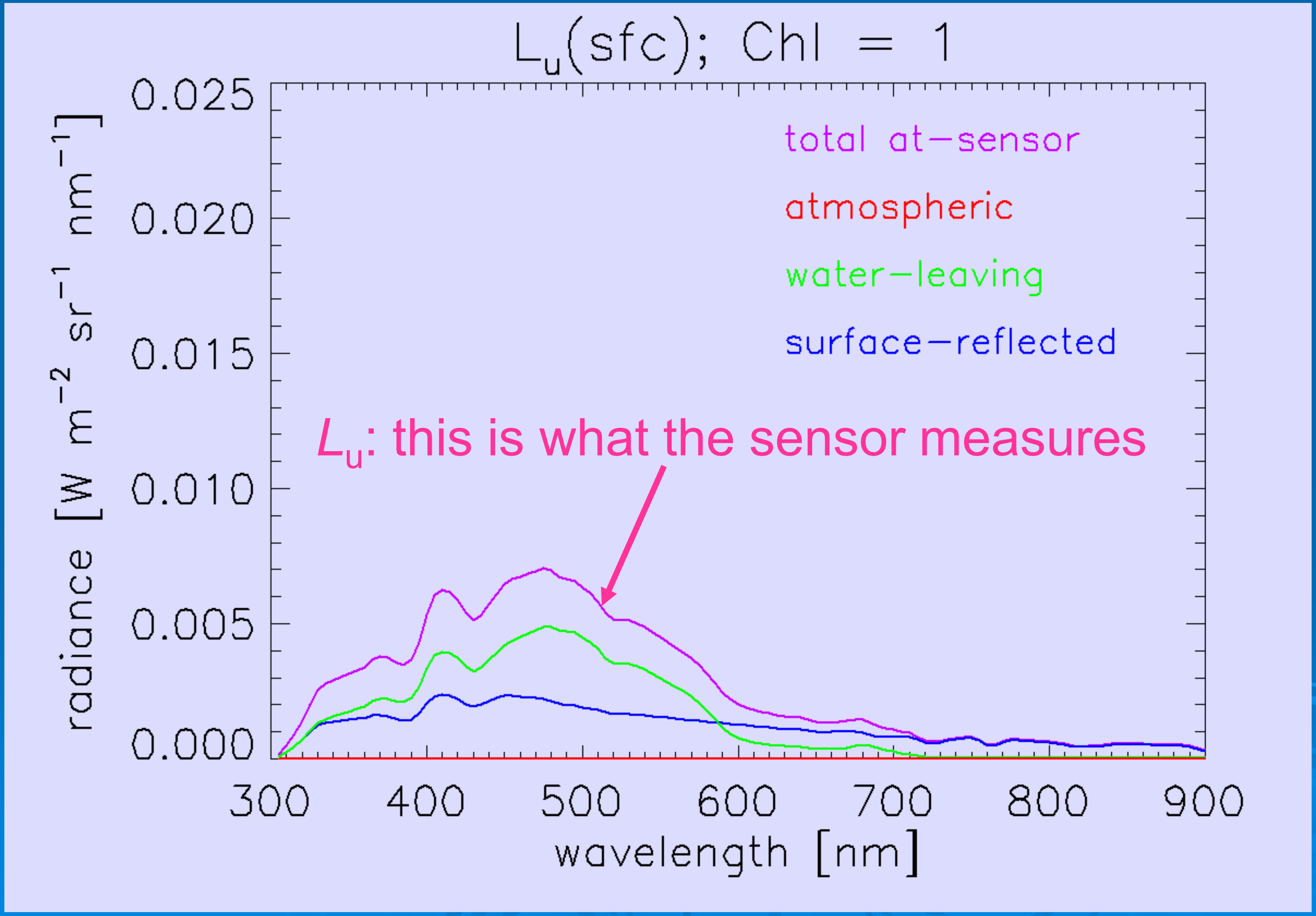
Example for Case 1 Water

Upwelling (nadir-viewing) radiance measured just above the sea surface.



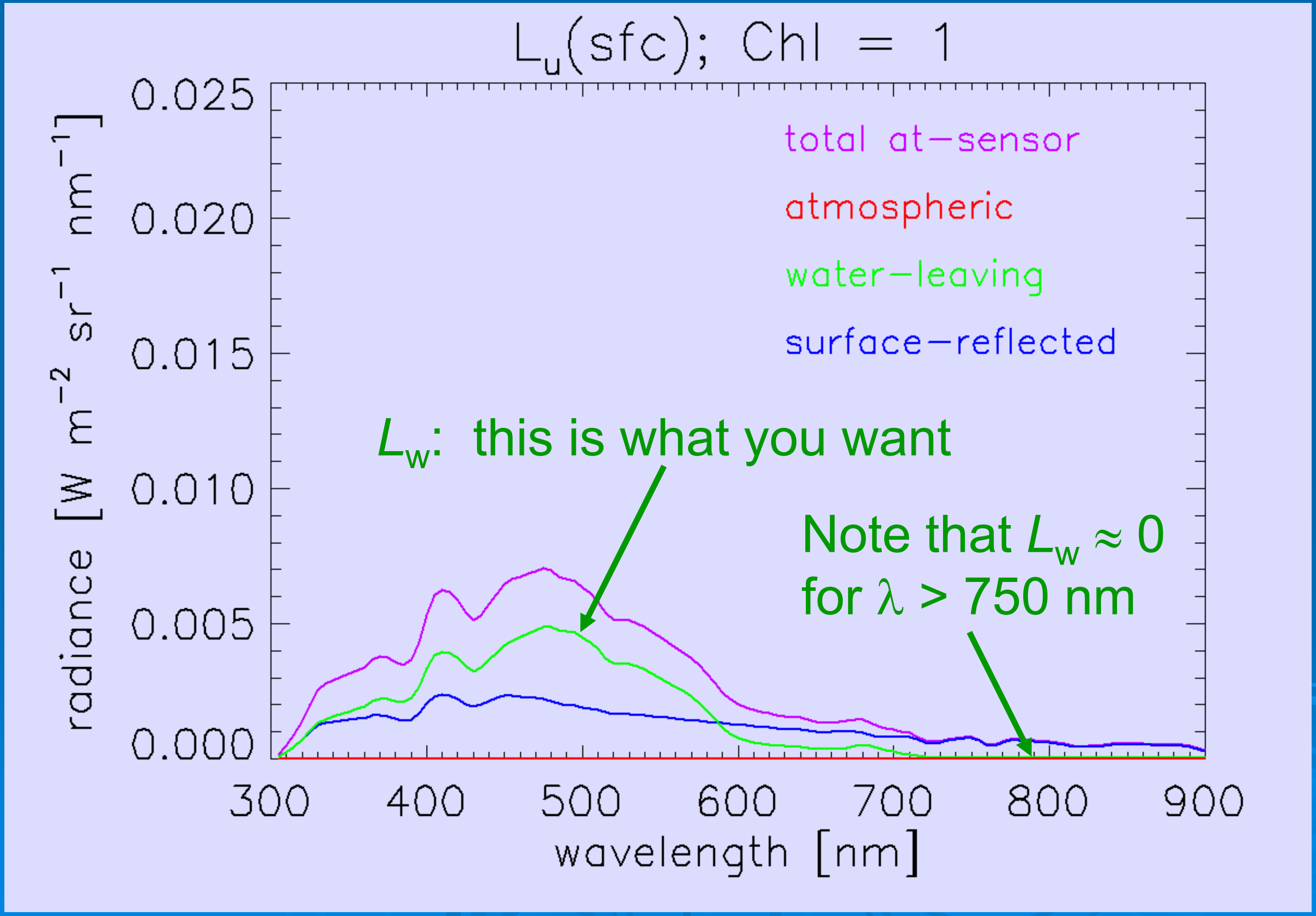
Example for Case 1 Water

Upwelling (nadir-viewing) radiances just above the sea surface



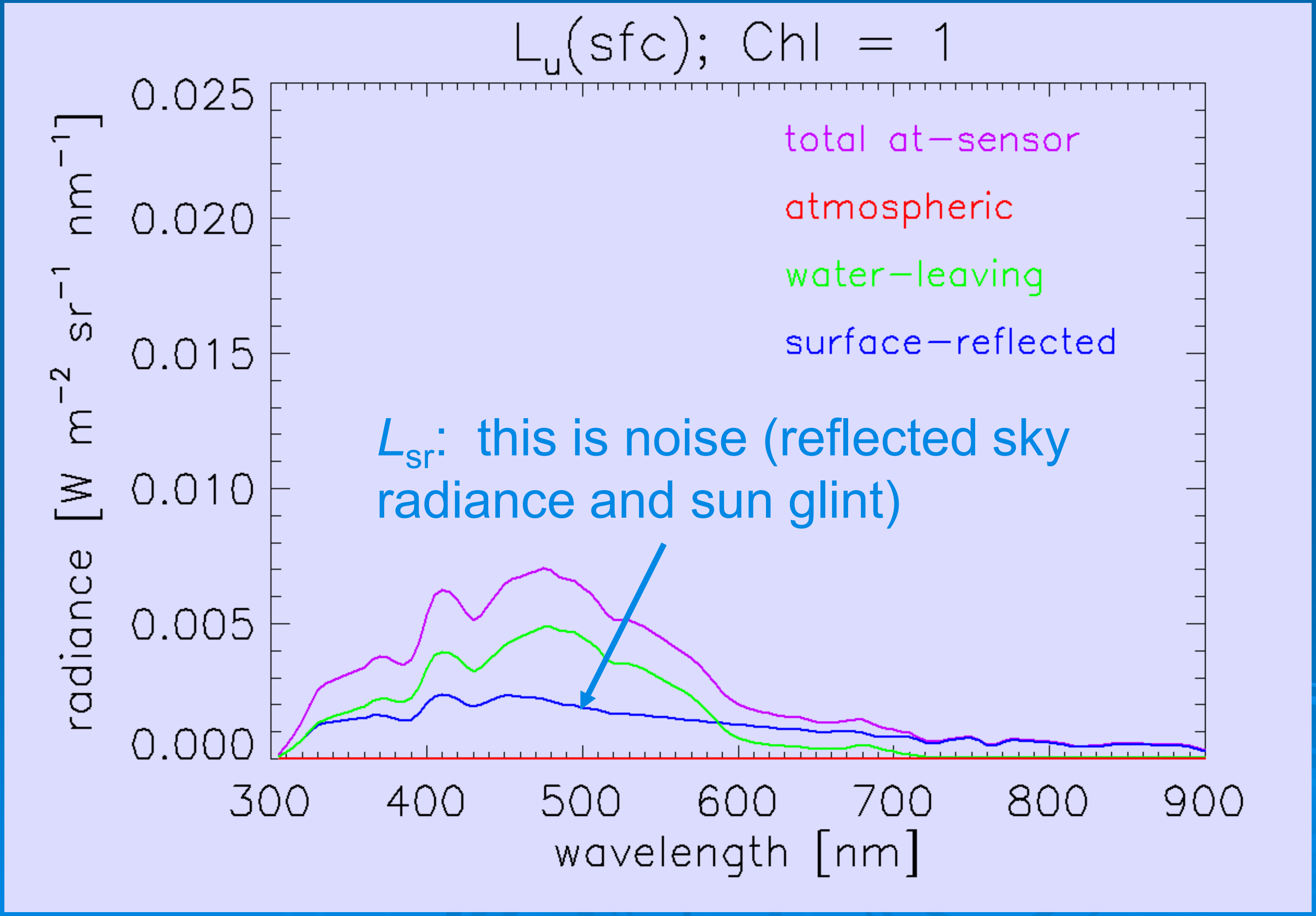
Example for Case 1 Water

Upwelling (nadir-viewing) radiances just above the sea surface



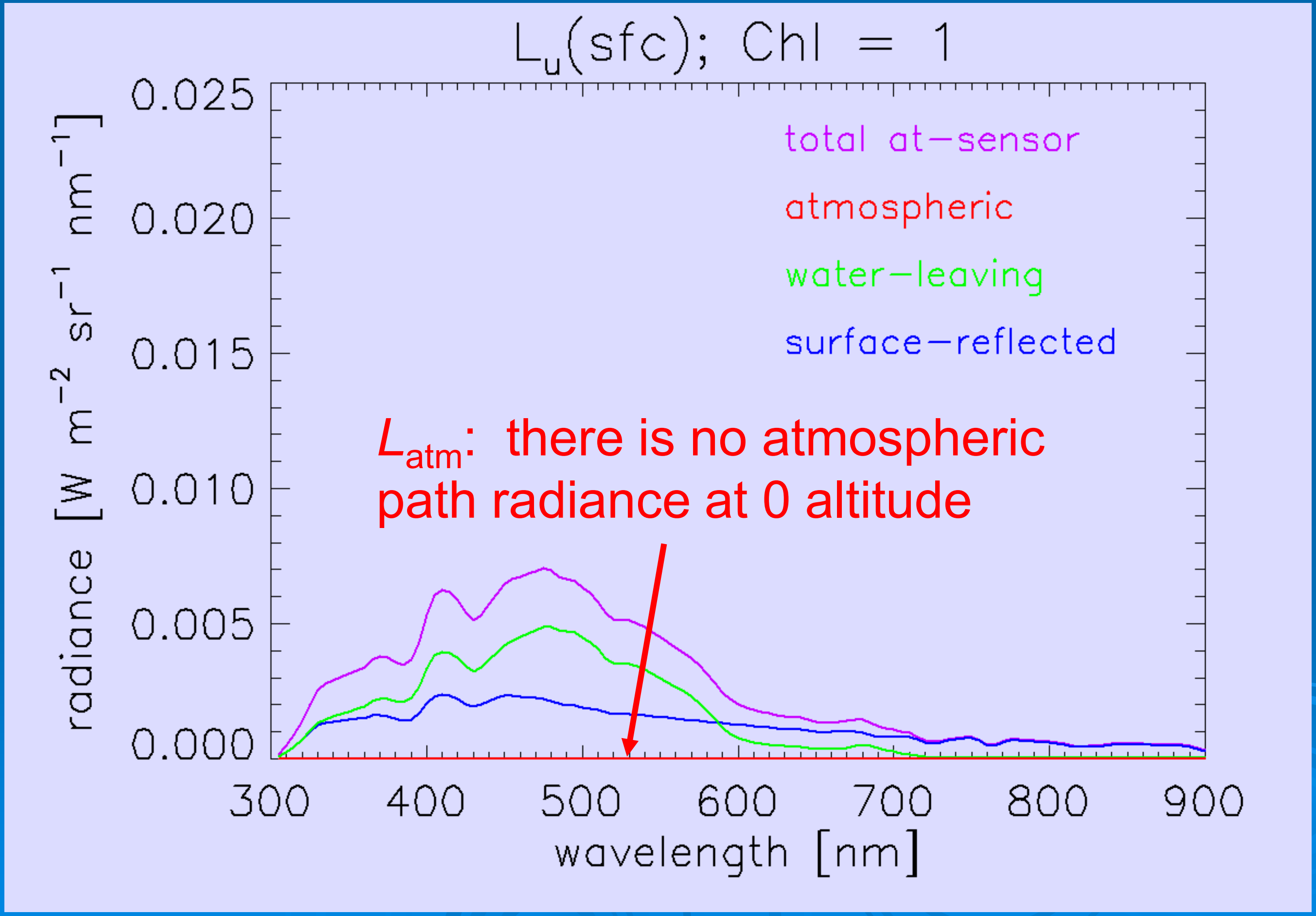
Example for Case 1 Water

Upwelling (nadir-viewing) radiances just above the sea surface



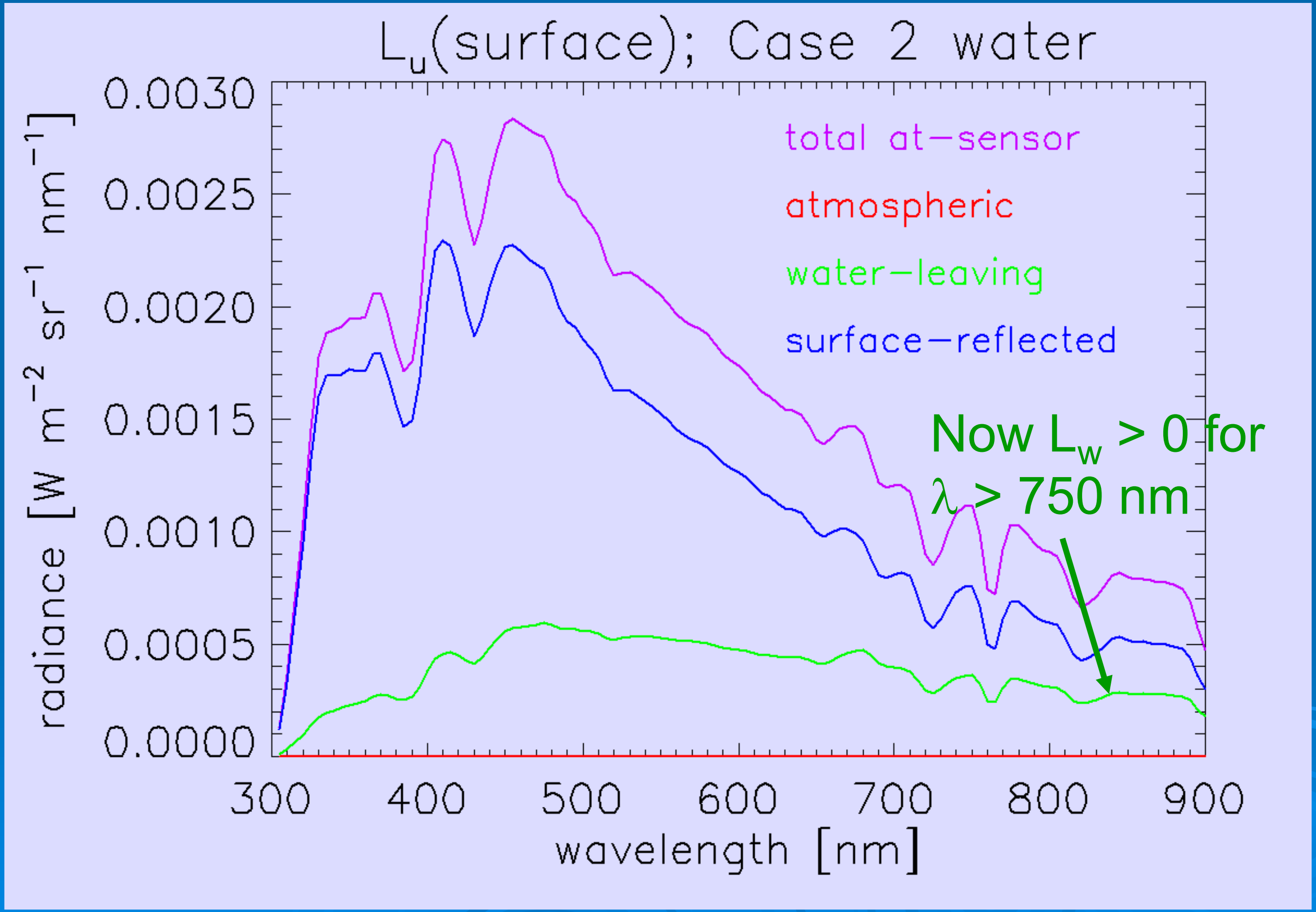
Example for Case 1 Water

Upwelling (nadir-viewing) radiances just above the sea surface



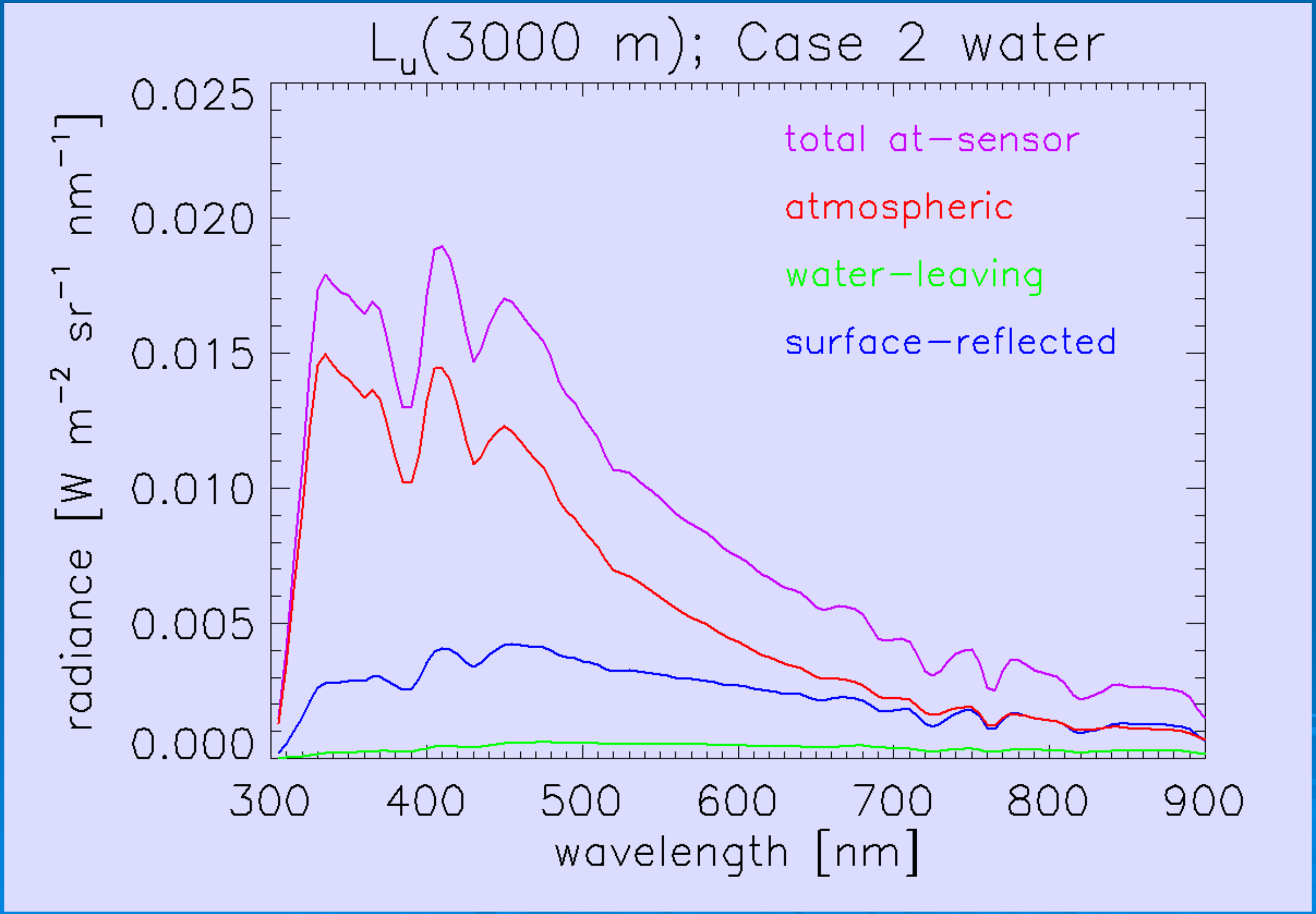
Example for Case 2 Water

Upwelling (nadir-viewing) radiances just above the sea surface



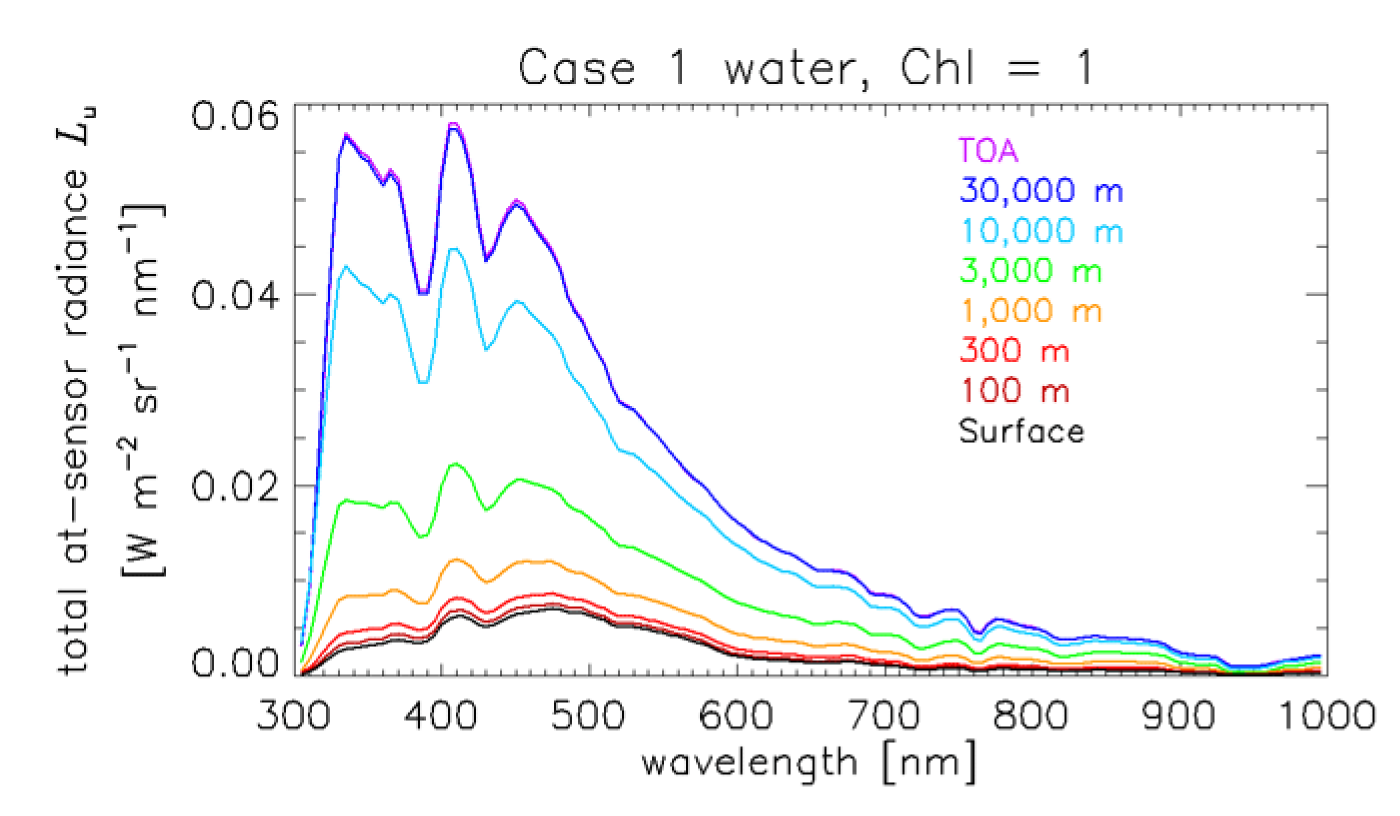
Example for Case 2 Water

Upwelling (nadir-viewing) radiances at 3,000 m altitude



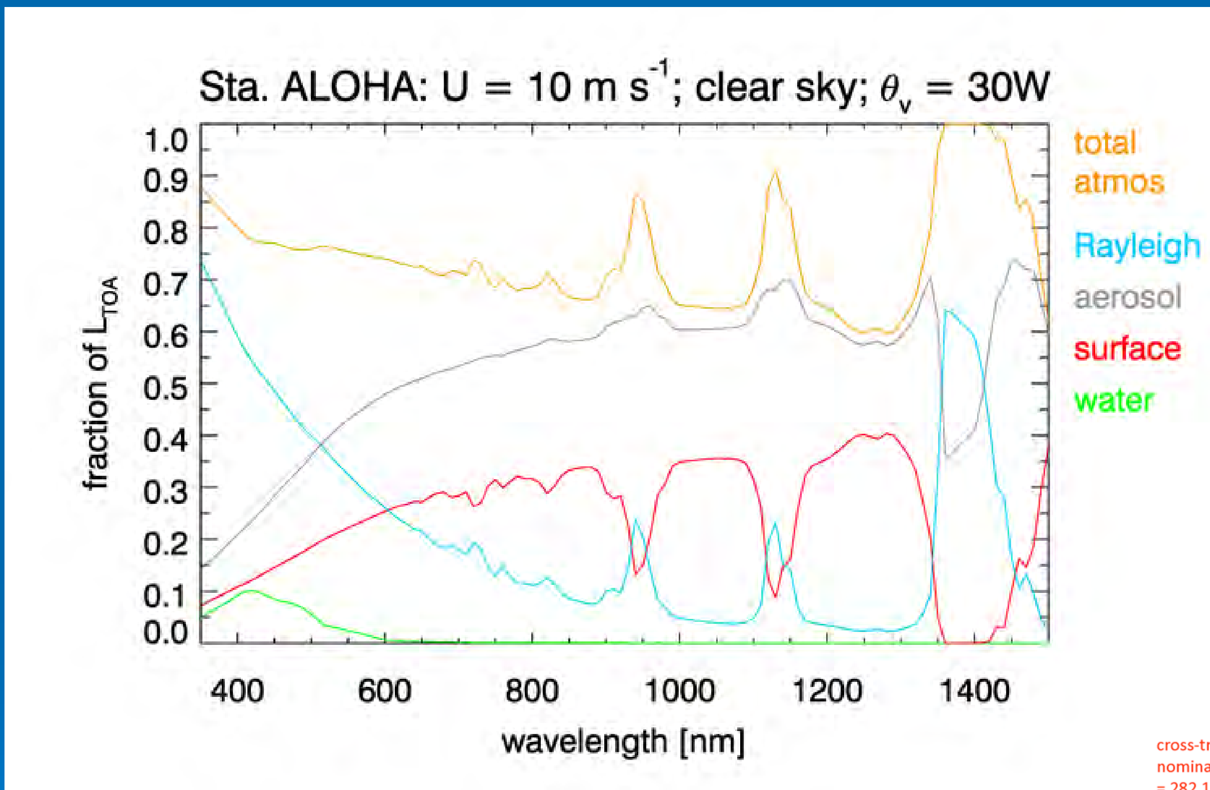
Total At-sensor Radiances at Various Altitudes

Most airborne remote sensing is done from altitudes of 1,000 to 10,000 m. Atmospheric path radiance is very important

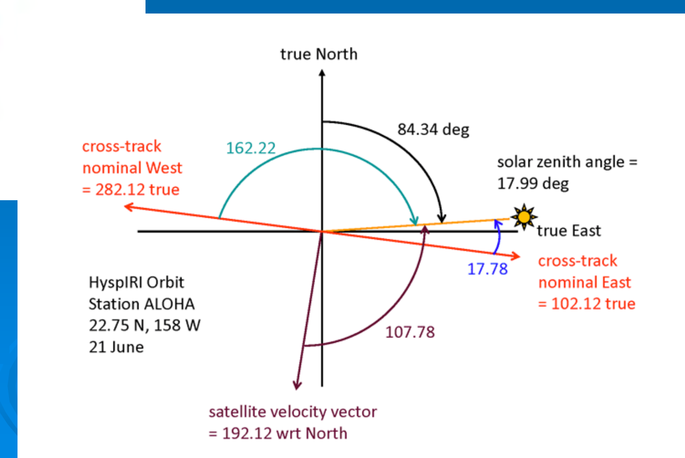


Fractional Contributions to TOA Radiance

L_w is at most 10% of L_{TOA} at visible wavelengths.

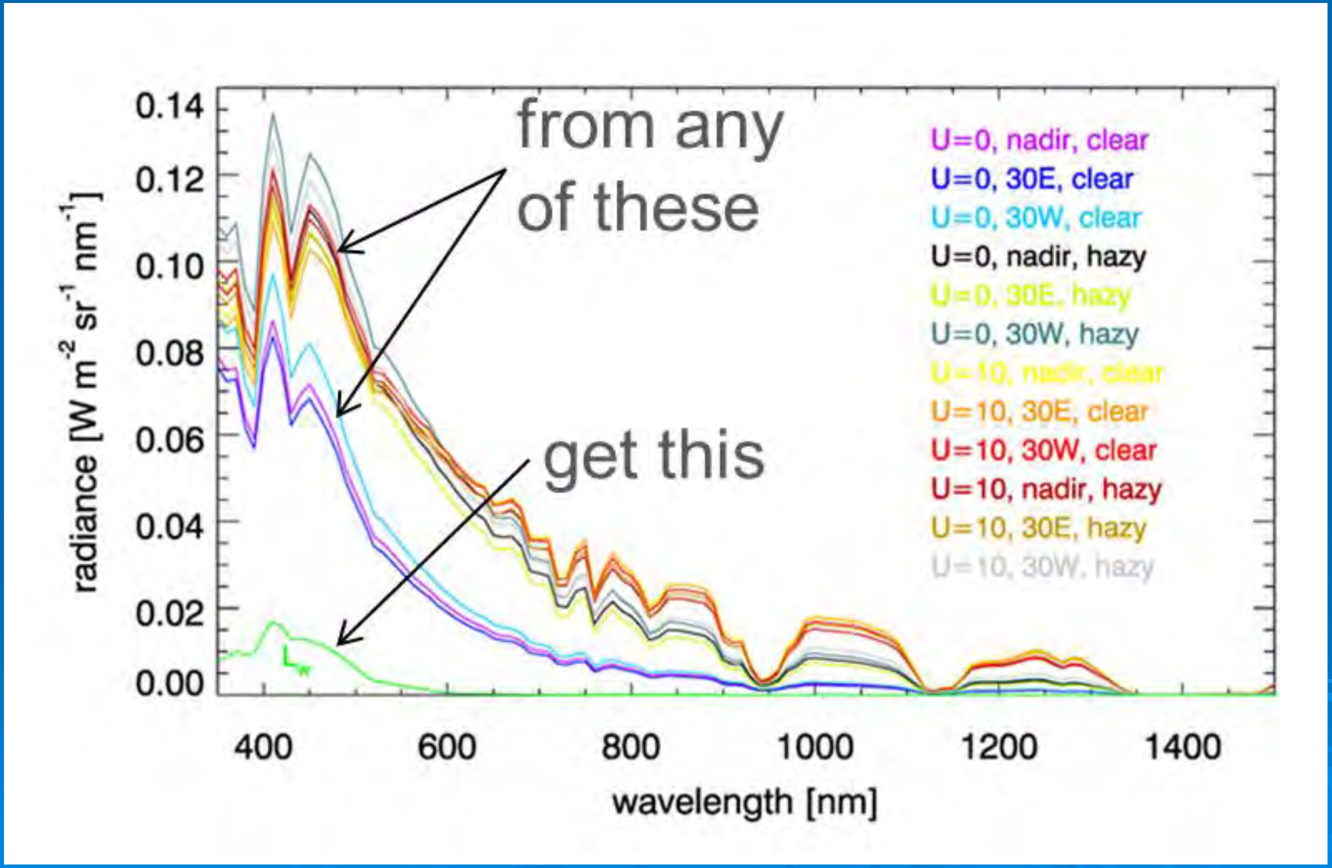


Runs for conditions at Station Aloha (Hawaii) and HyspIRI satellite orbit.



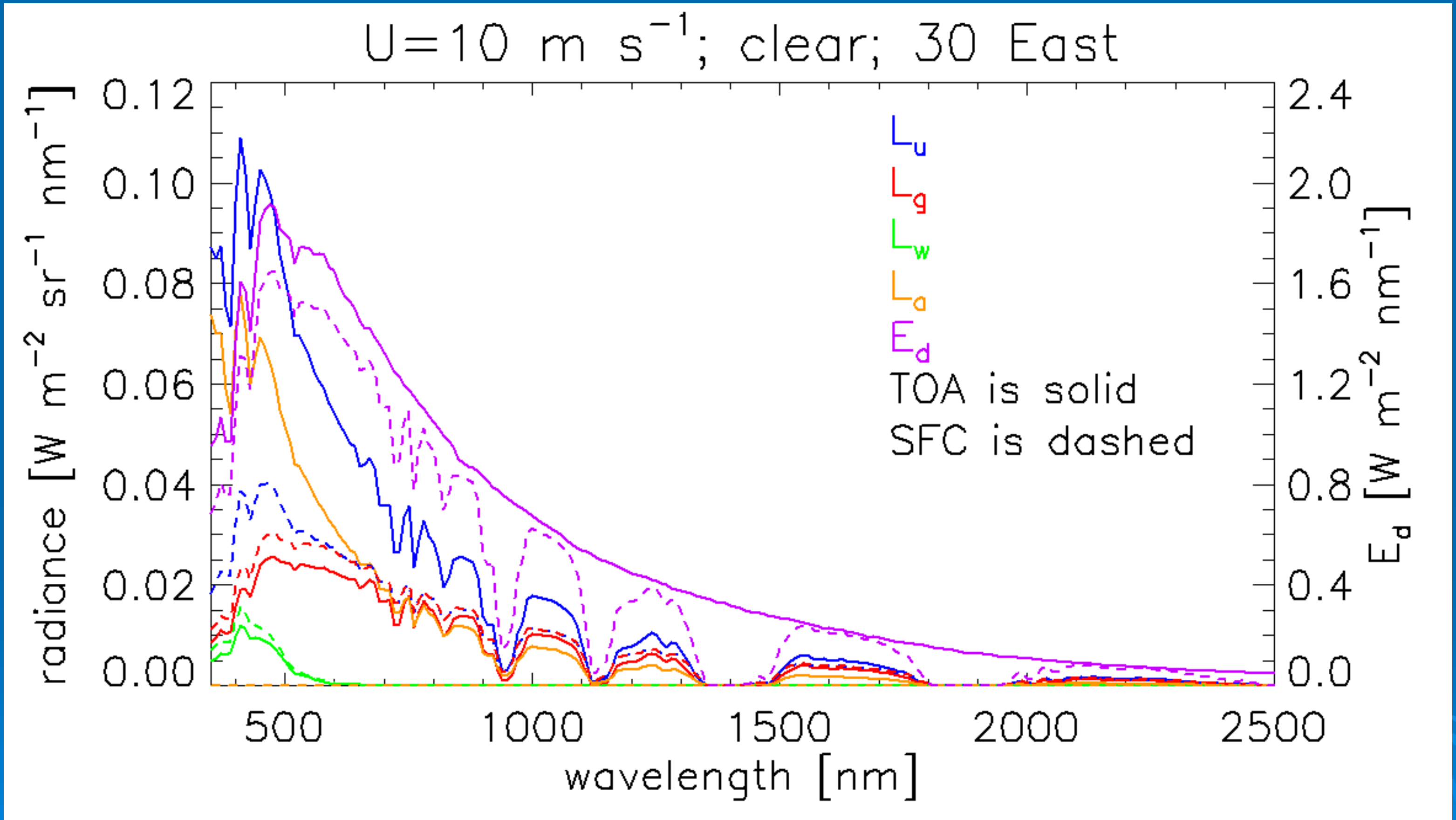
One L_w and Many TOA Radiances

The Problem: Each of these TOA radiances corresponds to the same water-leaving radiance



Contributions to the TOA radiance

The Level 1b TOA data are in radiance units

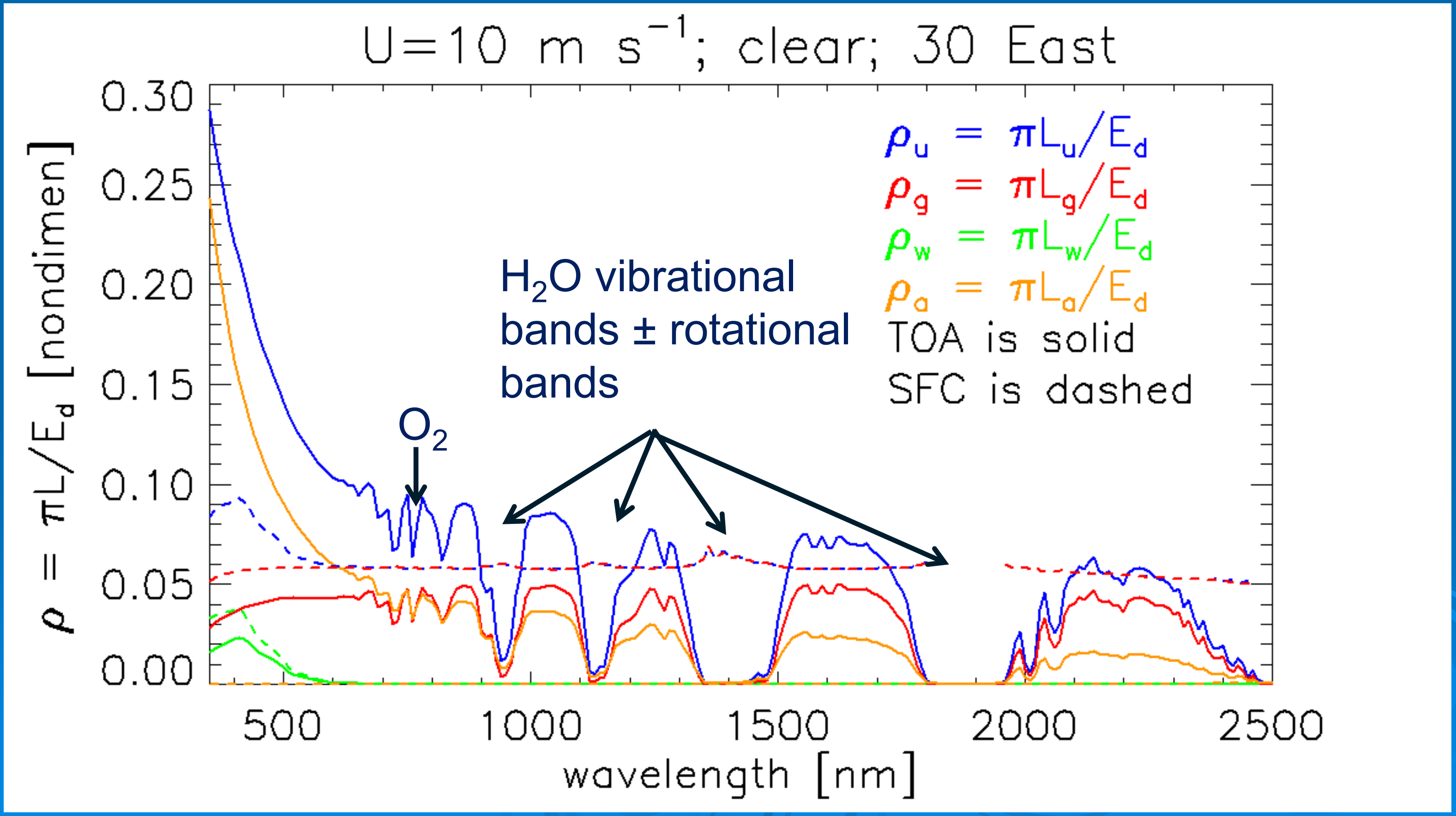


Contributions to the TOA Reflectance

Atmospheric correction is now usually done with nondimensional reflectance $\rho = \pi L_u / E_d$

Retrieval algorithms always use reflectances.

The data of the previous figure, plotted as reflectance:



Terrestrial vs Ocean Remote Sensing

Ocean remote sensing is much more difficult than terrestrial remote sensing.

Land is much brighter than water, so the total TOA radiance is much larger over land, and the atmospheric contribution to the total is relatively less, so that atmospheric correction is easier. Sensor signal-to-noise ratio is greater over land.

Some terrestrial problems do not even require atmospheric correction

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drop the bomb here

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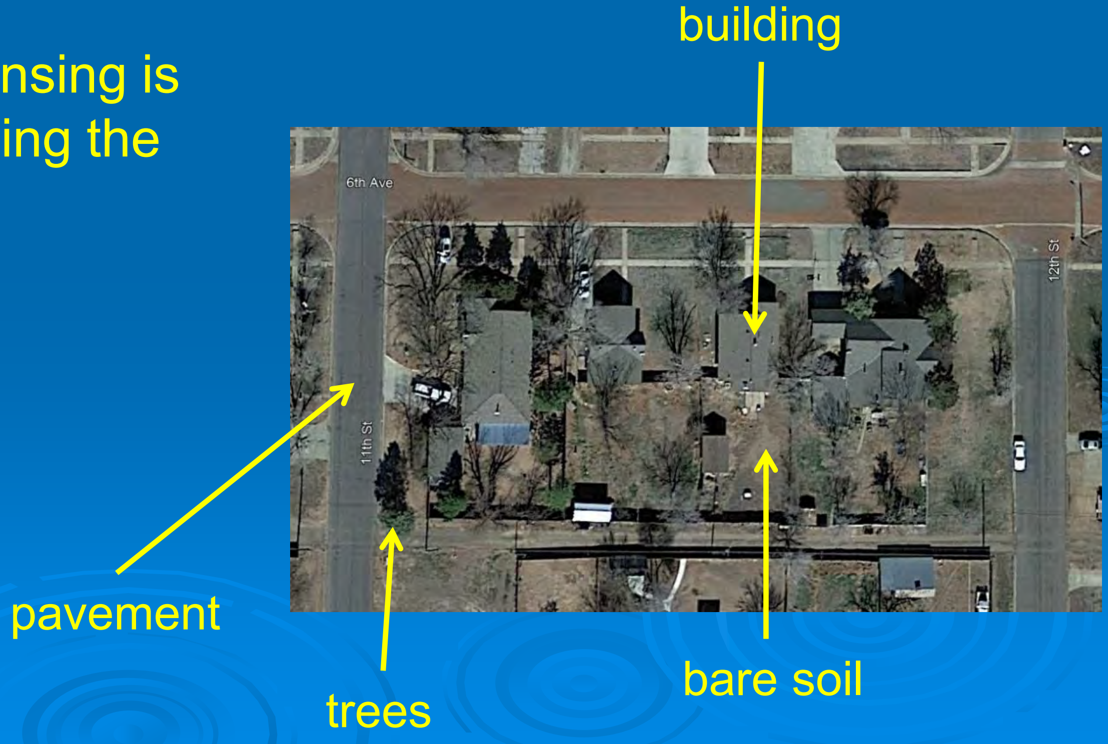
Much terrestrial remote sensing is concerned only with mapping the type of surface (thematic mapping), after minimal atmospheric correction.



Terrestrial vs Ocean Remote Sensing

Ocean remote sensing wants to retrieve in-water properties (chl, CDOM, or mineral concentration) or bottom properties (depth, bottom type), which is complicated by surface effects (glint), and the water itself when mapping bathymetry or bottom type.

Supervised classification techniques developed for thematic mapping of land types do NOT work for mapping of bottom types. See https://www.oceanopticsbook.info/view/remote-sensing/level-2/thematic-mapping or OOB Sec. 9.4

What Others Say

"Curt, you'll never be able to do that."

-- A famous atmospheric radiative transfer guru, after my lecture at a MODTRAN conference about the atmospheric correction accuracy needed for ocean remote sensing.

What I Say

“Curt, you’ll never be able to do that.”

-- A famous atmospheric radiative transfer guru, after my lecture at a MODTRAN conference about the atmospheric correction accuracy needed for ocean remote sensing.

“The man who says something cannot be done is often interrupted by someone doing it.”

-- An old saying

“The man who says something cannot be done is often interrupted by a woman doing it.”

-- A new saying

You Need Something To Do on Rainy Weekends

Rocking Chair University; http://www.haltaylor.com/

~130 hours later...

For more on making this chair, see
https://ann-and-curt.smugmug.com/Rocking-Chair