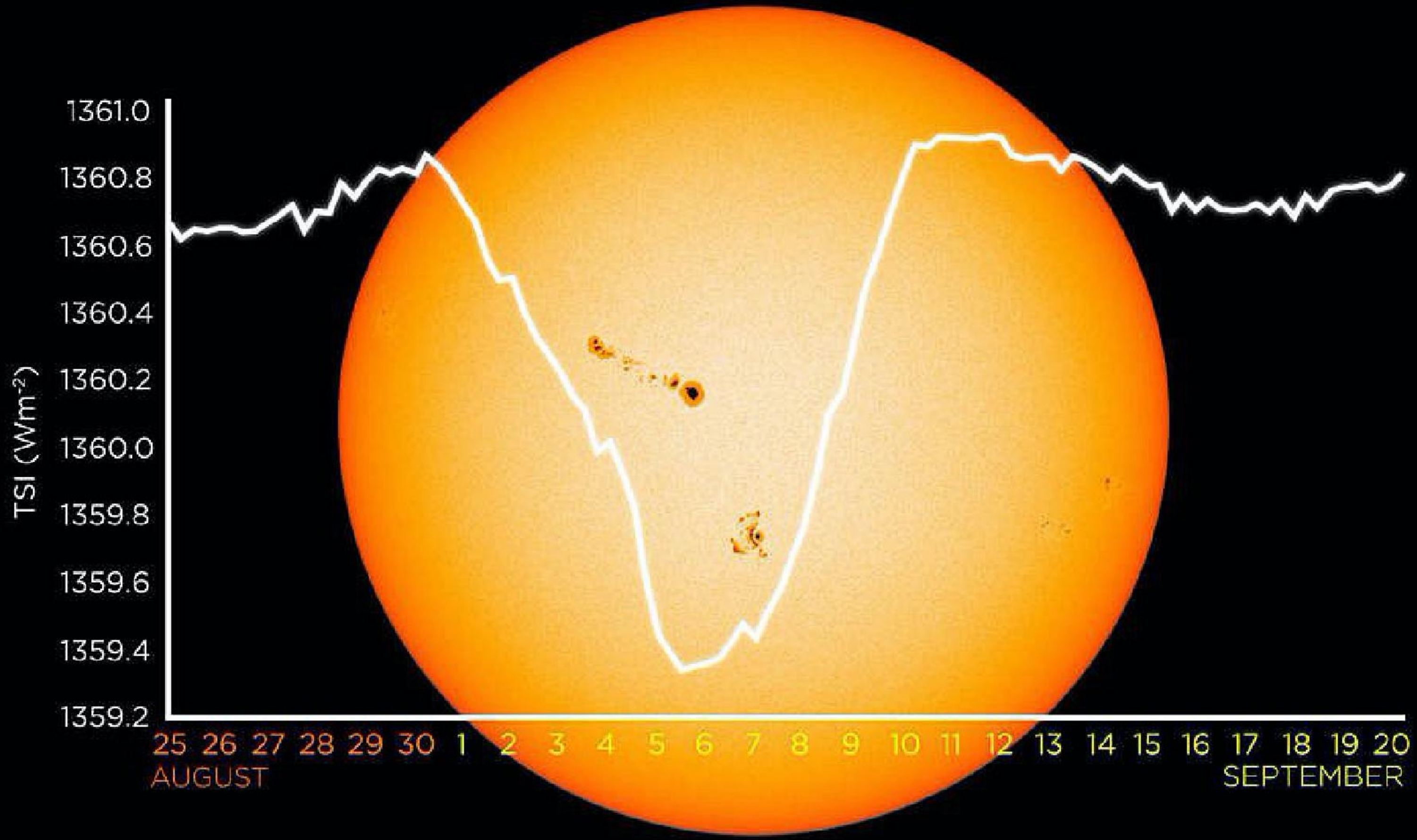
How does it work: passive remote sensing, radiometry, and polarimetry

Brian Cairns

Acknowledgements:
OceanicOpticsBook, Michael Mishchenko

PACE class @ UMBC
1-5 Aug 2022



passive remote sensing, radiometry, and polarimetry

Radiometry and Polarimetry:

1. Basic definitions and useful formulas
2. Calibration summary
3. Remote sensing – the toolbox you need

1. Basic definitions and useful formulas (radiance)

$$L(\boldsymbol{x}, \boldsymbol{t}, \boldsymbol{\xi}, \boldsymbol{\lambda}) = \frac{\boldsymbol{\Delta Q}}{\boldsymbol{\Delta t \Delta A \Delta \Omega \Delta \lambda}} \text{ W/m}^2\text{/nm/sterad}$$

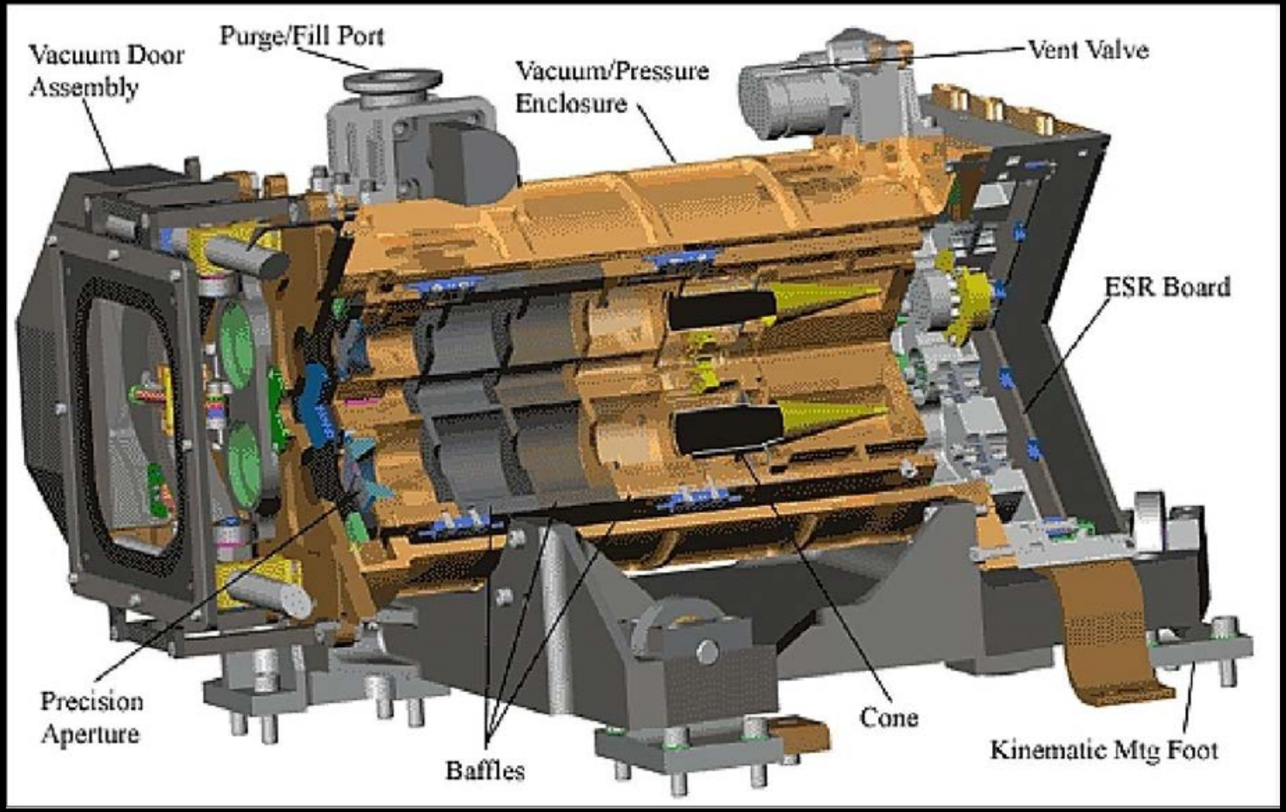
- The radiance *L* is the radiant energy in an observed radiation field per interval of time Δt, area ΔA, solid angle $\Delta\Omega$ and wavelength interval $\Delta\lambda$. In the Oceanic Optics book it's noted that "Spectral radiance is the fundamental radiometric quantity of interest in hydrologic optics because all other radiometric quantities can be derived from the radiance."

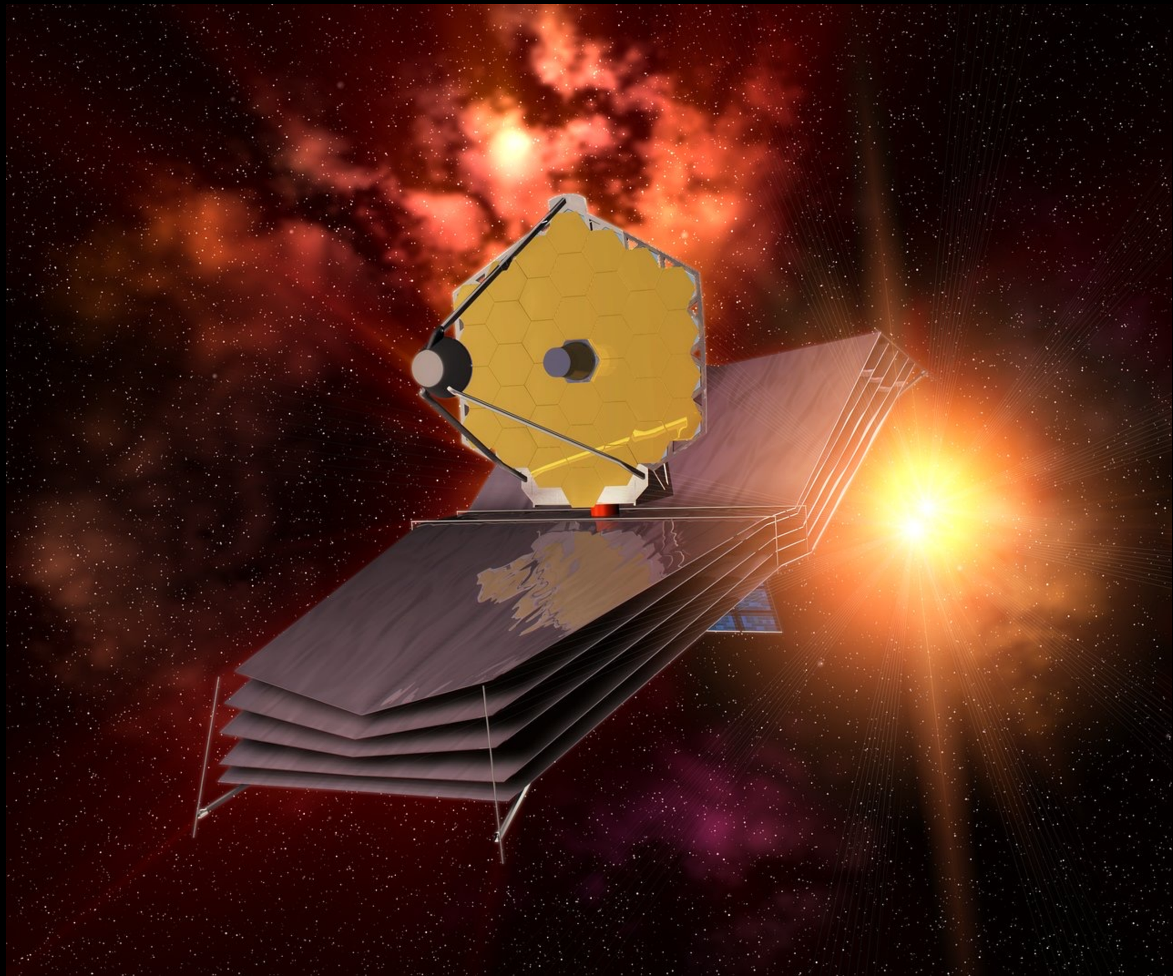
- The spectral radiance is also the fundamental radiometric quantity of interest for passive satellite remote sensing since what is measured by almost all satellite sensors is a marginal integral of the radiance over the integration time, entrance pupil size, field of view and spectral sampling chosen for the remote sensing problem at hand. The spectral radiance is also the radiometric quantity that has a propagator. Irradiances do not!

Two examples with very different marginal integrals over the radiances they observe.

Total Irradiance Monitor entrance area 5×10^{-5} m^2, integration times of ~ 100 sec, ~ 2×10^{-4} Sterad (0.5 deg), detects entire spectrum

James Webb Space Telescope, mirror 33 m^2, deep field integrations of ~ 1 day, various different spectrometers and 7.4×10^{-10} Sterad (0.1 arcsec resolution).



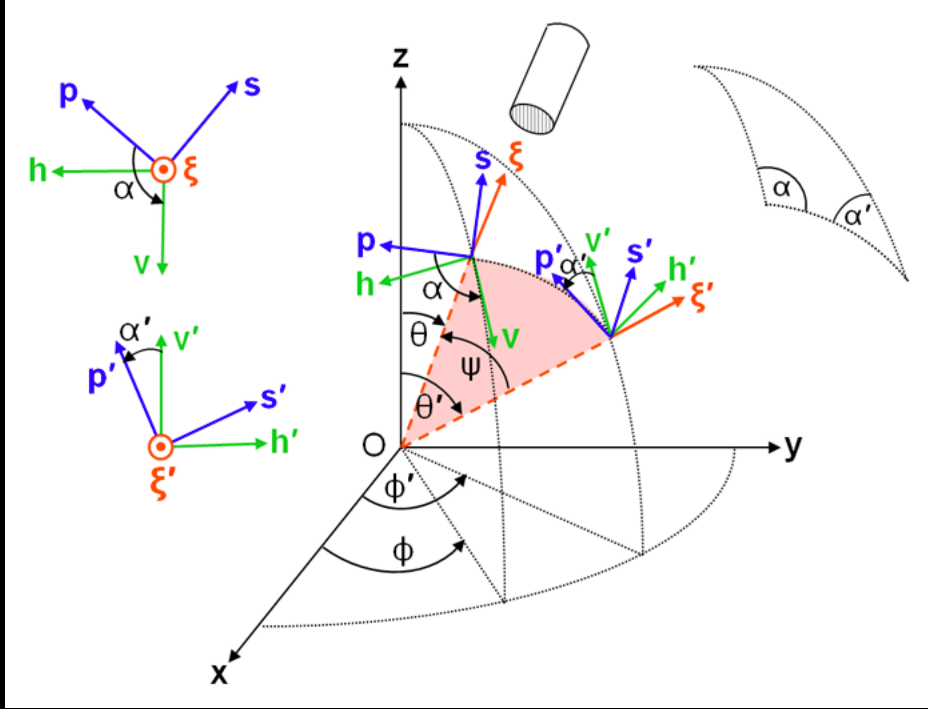


1. Basic definitions and useful formulas (planar irradiance)

$$E_d(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) = \int_{\varphi=0}^{2\pi}\int_{\theta=0}^{\pi/2} L(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\xi},\boldsymbol{\lambda})|\boldsymbol{cos\theta}|\boldsymbol{sin\theta d\theta d\varphi}$$

$$E_u(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) = \int_{\varphi=0}^{2\pi}\int_{\theta=\pi/2}^{\pi} L(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\xi},\boldsymbol{\lambda})|\boldsymbol{cos\theta}|\boldsymbol{sin\theta d\theta d\varphi}$$

Use geometry and conventions from Oceanic Optics Book. It's free, it's good and is generally in agreement with most of the other major texts.



$$E_{d,Net}(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) = E_d(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) - E_u(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda})$$

$$\frac{\partial T}{\partial x} = \frac{1}{\rho C_P}\frac{\partial E_{D,Net}}{\partial z}$$

1. Basic definitions and useful formulas (scalar irradiance)

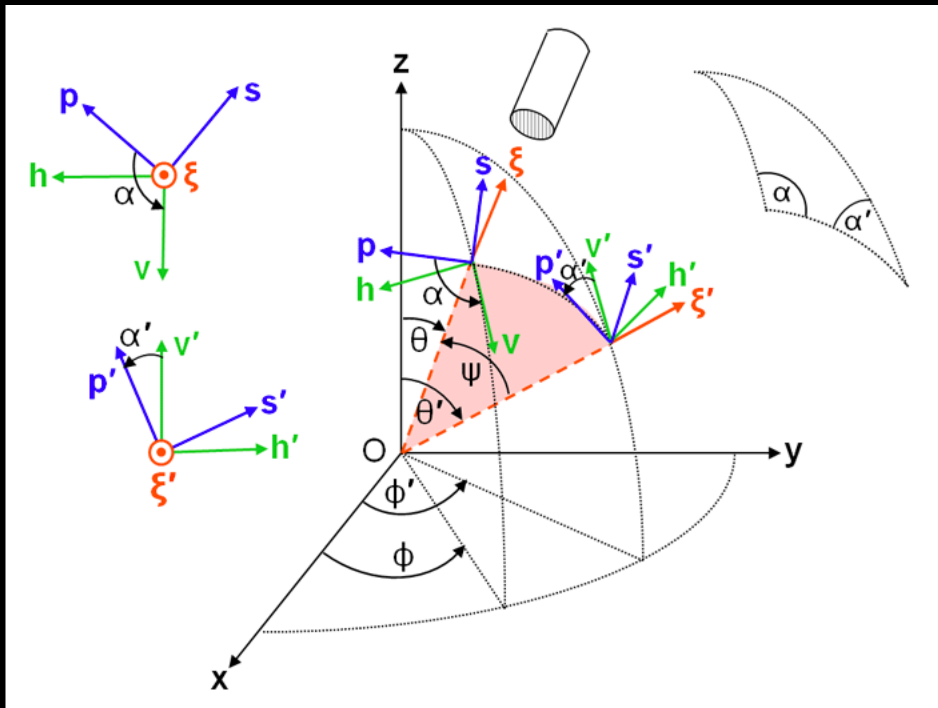
$$E_{od}(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) = \int_{\varphi=0}^{2\pi}\int_{\theta=0}^{\pi/2} L(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\xi},\boldsymbol{\lambda})\boldsymbol{sin\theta d\theta d\varphi}$$

$$E_{ou}(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) = \int_{\varphi=0}^{2\pi}\int_{\theta=\pi/2}^{\pi} L(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\xi},\boldsymbol{\lambda})\boldsymbol{sin\theta d\theta d\varphi}$$

Use geometry and conventions from Oceanic Optics Book. It's free, it's good and is generally in agreement with most of the other major texts.

$$E_o(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) = E_{od}(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda}) + E_{ou}(\boldsymbol{x},\boldsymbol{t},\boldsymbol{\lambda})$$

$$PAR(\boldsymbol{x},\boldsymbol{t}) = \int_{\boldsymbol{400\ nm}}^{\boldsymbol{700\ nm}} E_{\boldsymbol{o}}(x,t,\lambda)\frac{\lambda}{hc}d\lambda$$



1. Basic definitions and useful formulas (scalar irradiance)

$$\xi . \nabla L(\boldsymbol{x}, \boldsymbol{t}, \xi, \boldsymbol{\lambda}) = \mathbf{0}$$

$$L(\boldsymbol{x}, \boldsymbol{t}, \boldsymbol{\xi}, \boldsymbol{\lambda}) = L(\boldsymbol{x} + s\boldsymbol{\xi}, \boldsymbol{t}, \boldsymbol{\xi}, \boldsymbol{\lambda})$$

Radiance in free space is constant along a ray (except when it’s not e.g. when there are coherent fields and interference is an issue) and at least for incoherent sources, like lamps and the sun, the irradiance has a $1/r^2$ decrease with distance from the source.

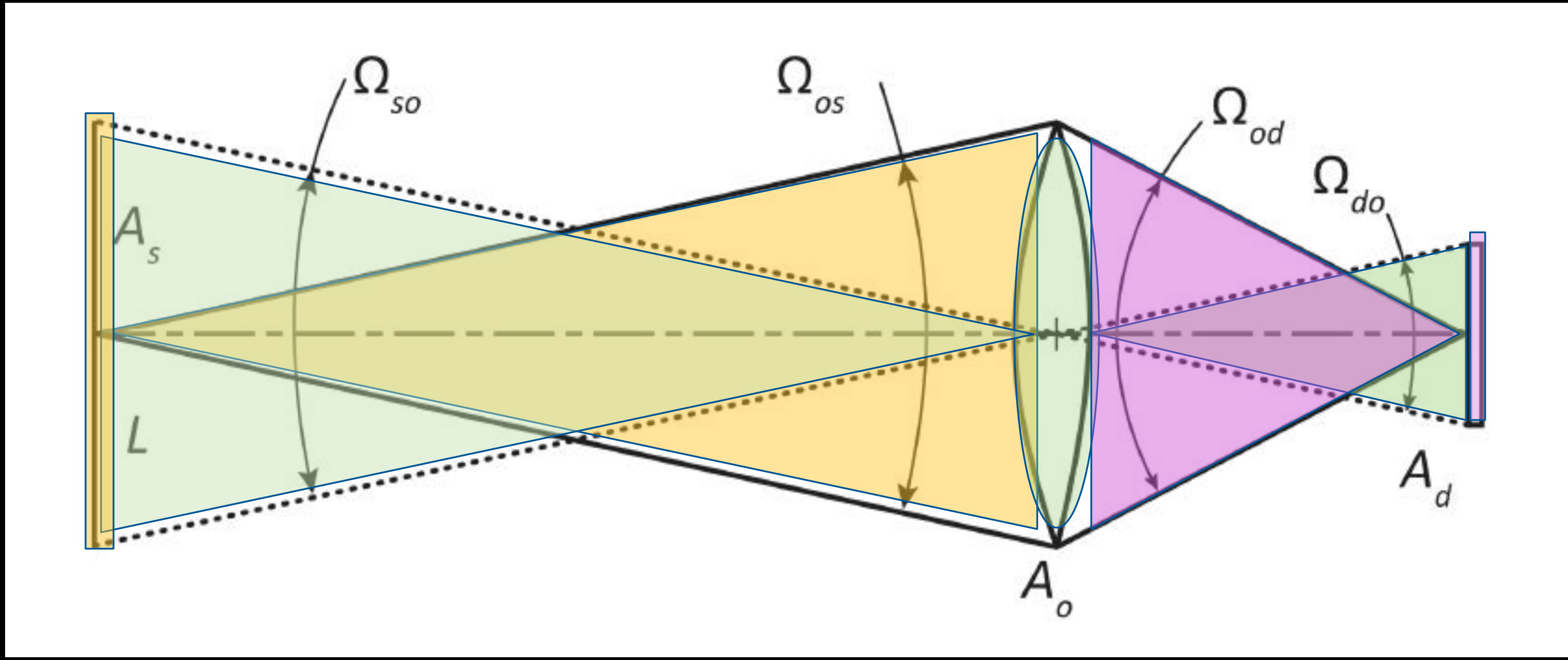
$$E(\boldsymbol{x}, \boldsymbol{t}, \boldsymbol{\lambda}) = L(\boldsymbol{x}, \boldsymbol{t}, \xi, \boldsymbol{\lambda}) d\Omega$$

$$d\Omega = \frac{dA}{r^2}$$

$$E(\boldsymbol{x}, \boldsymbol{t}, \boldsymbol{\lambda}) = \frac{L(\boldsymbol{x}, \boldsymbol{t}, \boldsymbol{\xi}, \boldsymbol{\lambda}) dA}{r^2} = \frac{I(\boldsymbol{t}, \boldsymbol{\xi}, \boldsymbol{\lambda}) dA}{r^2}$$

1. Basic definitions and useful formulas (invariance of etendue)

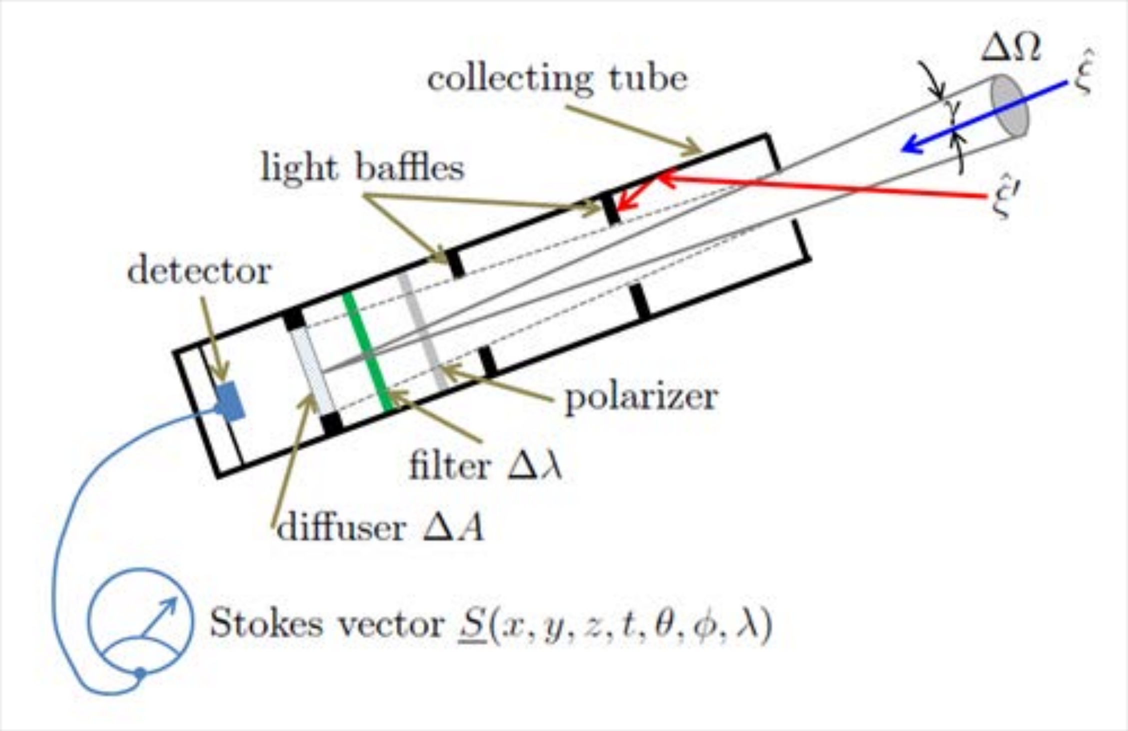
Important to remember that for a lossless system the etendue (the product of area and sold angle) is invariant, as you propagate light through the system. This means that if the front end is poorly designed there is no way to recover, but if it is well designed you should be able to maintain throughput throughout the system☺

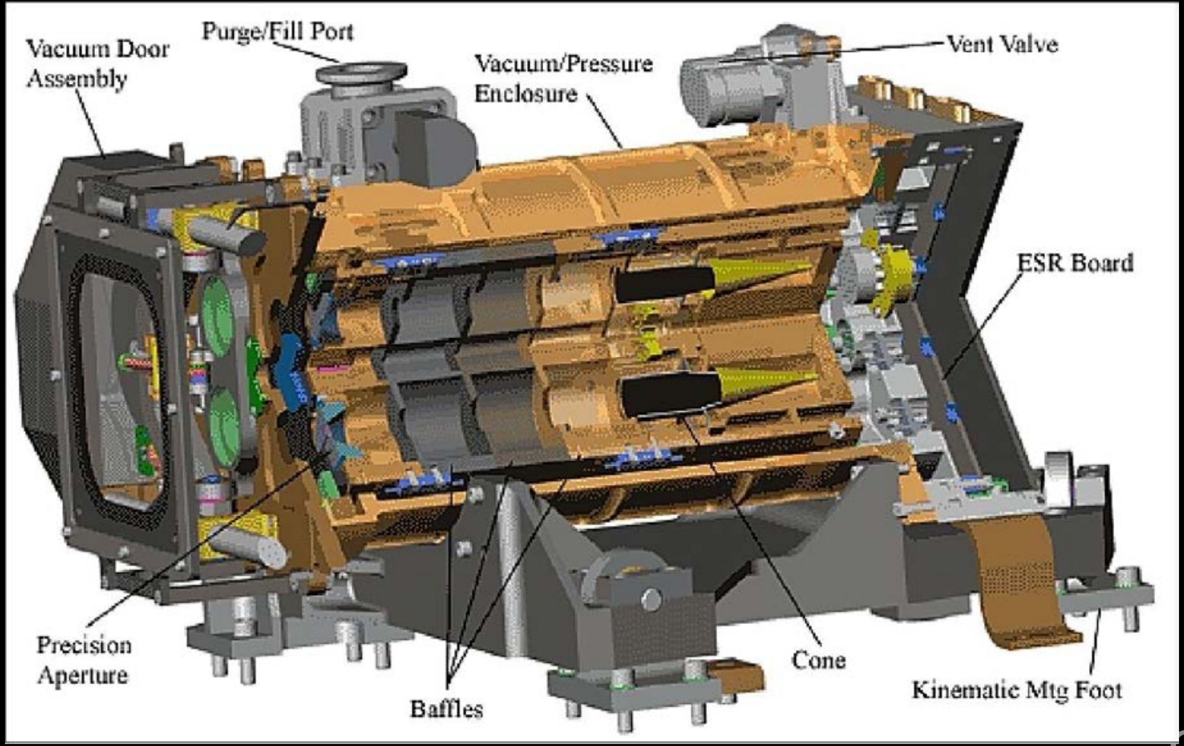


$$A_s\Omega_{os} = A_o\Omega_{so} = A_o\Omega_{do} = A_d\Omega_{od}$$

Passive remote sensing instruments span a huge range of designs and capabilities:

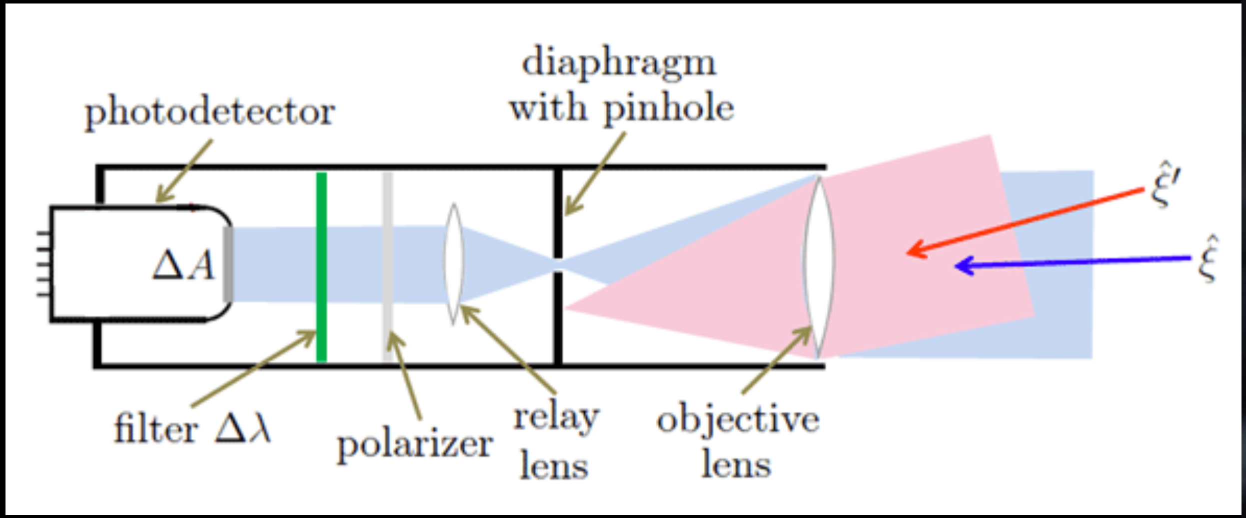
Simplest is probably a Gershun tube, no real optics, just an aperture, baffles and a detector; but such an instrument can make incredibly accurate measurements of the solar irradiance (100 ppm) with the right design and detectors

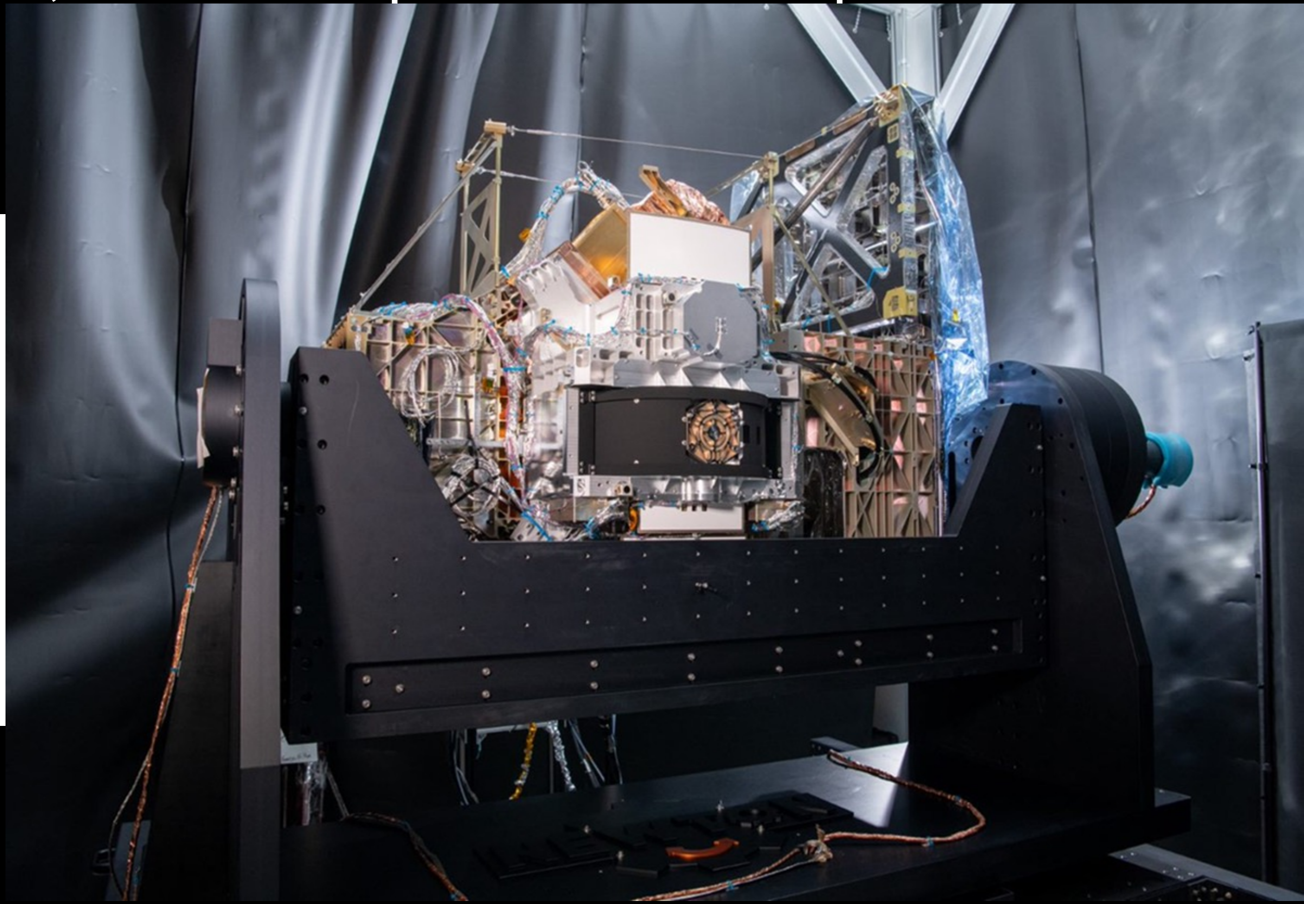




Passive remote sensing instruments span a huge range of designs and capabilities:

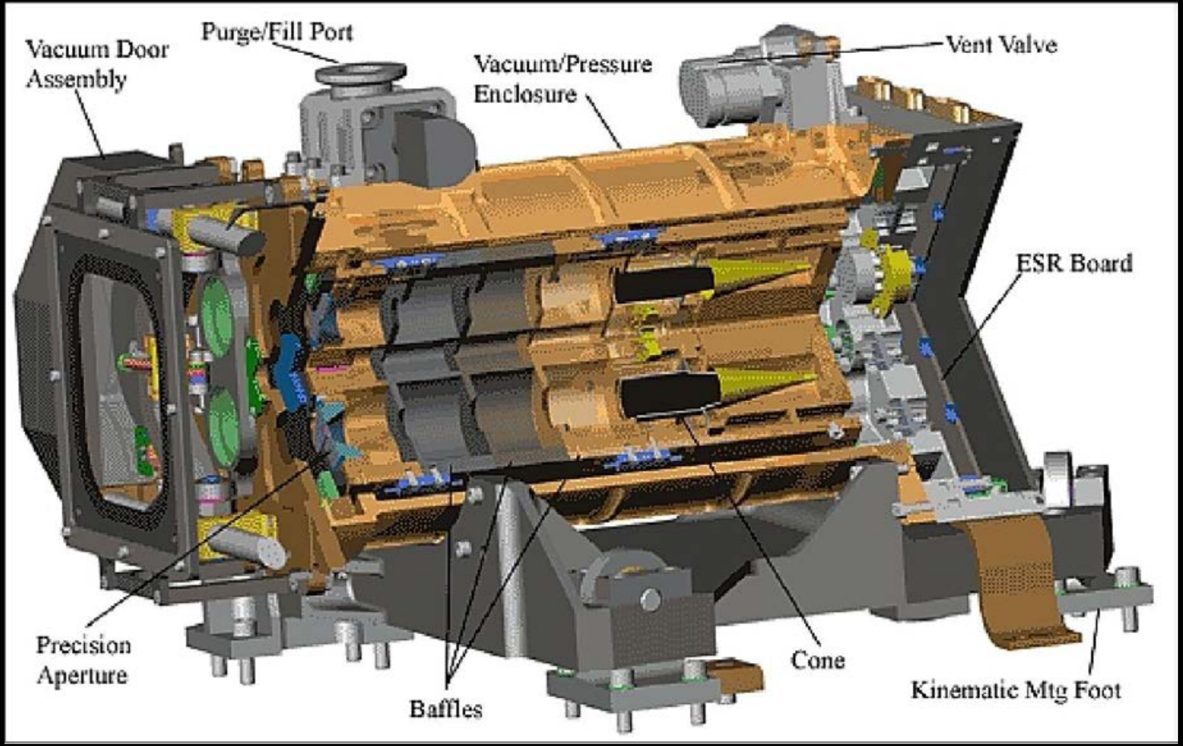
Another simple but common optical front end is a relay telescope. Provides a well collimated radiometer (WCR), which is the basis for almost every satellite sensor. OCI is a (rotating) relay telescope, with a complex back end spectral analysis and detection system.

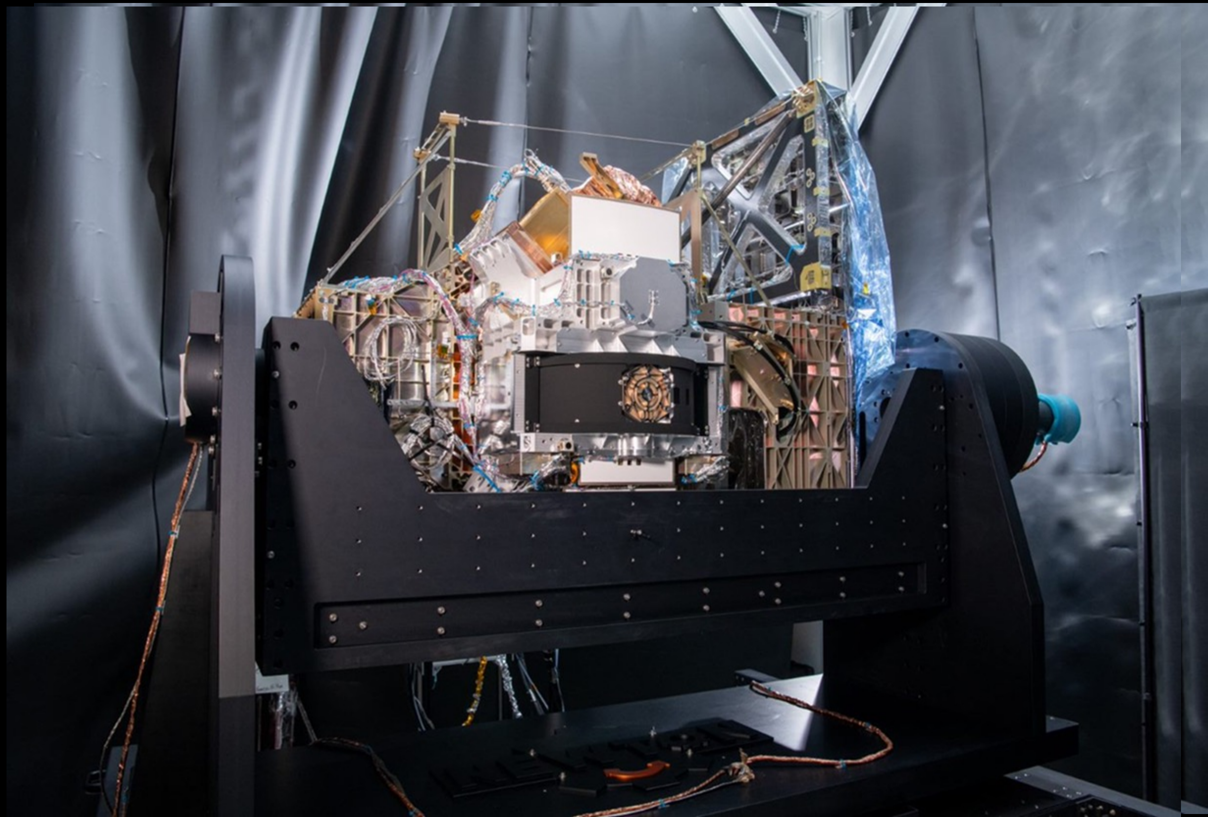




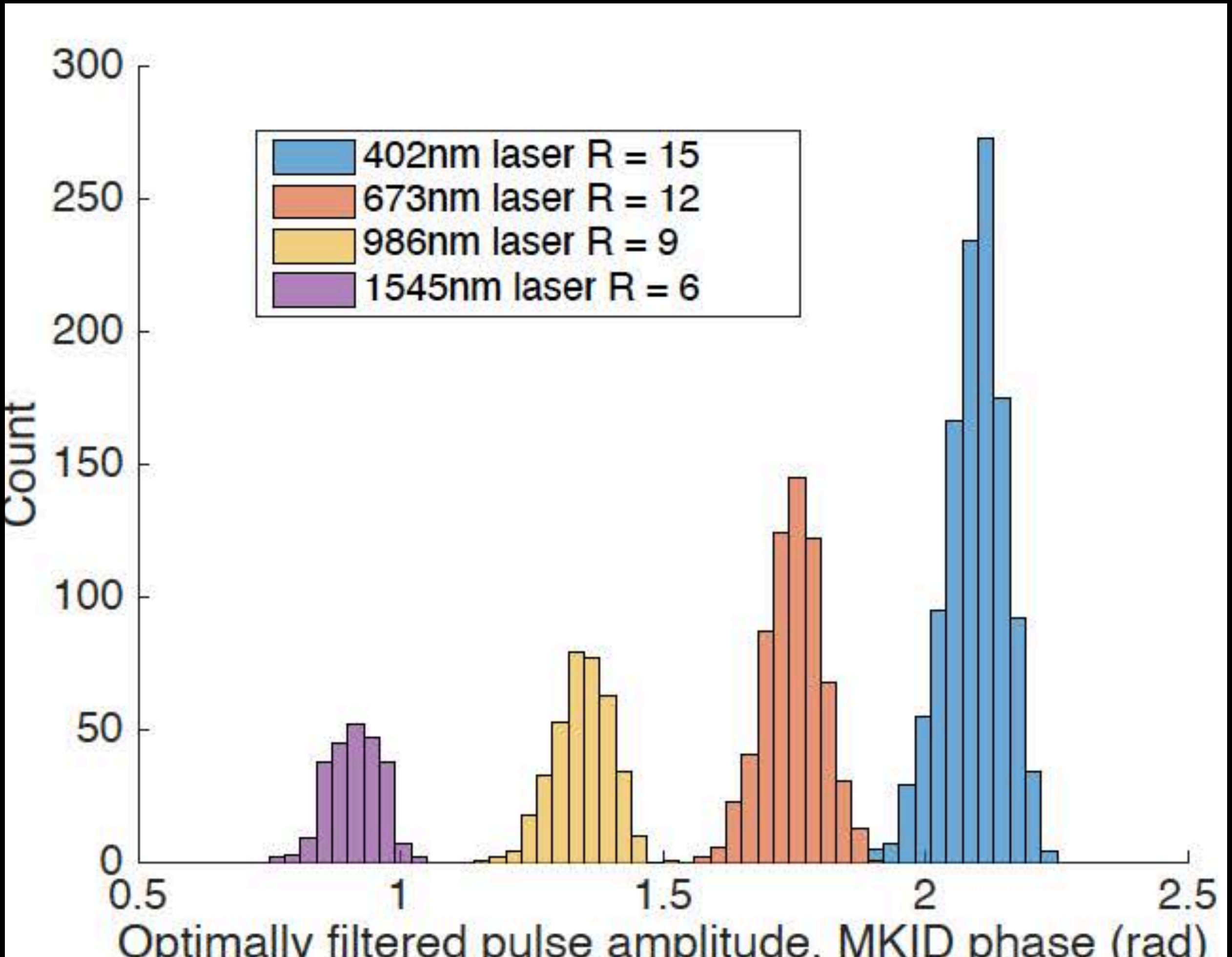
Passive remote sensing instruments span a huge range of designs and capabilities:

TIM and OCI provide good examples of the two main classes of light detectors, *thermal* and *quantum* detectors. Thermal detectors measure the *energy* of the incident field, quantum detectors respond to the *number* of incident photons, usually through the generation of photoelectrons.





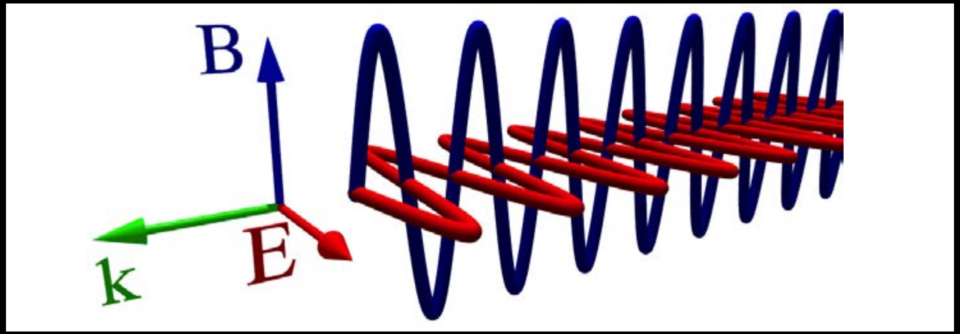
And there are now detectors such as Kinetic Inductance Detectors (KIDs) that can resolve single photons and their energy (color) – so you can make on chip spectrometers!

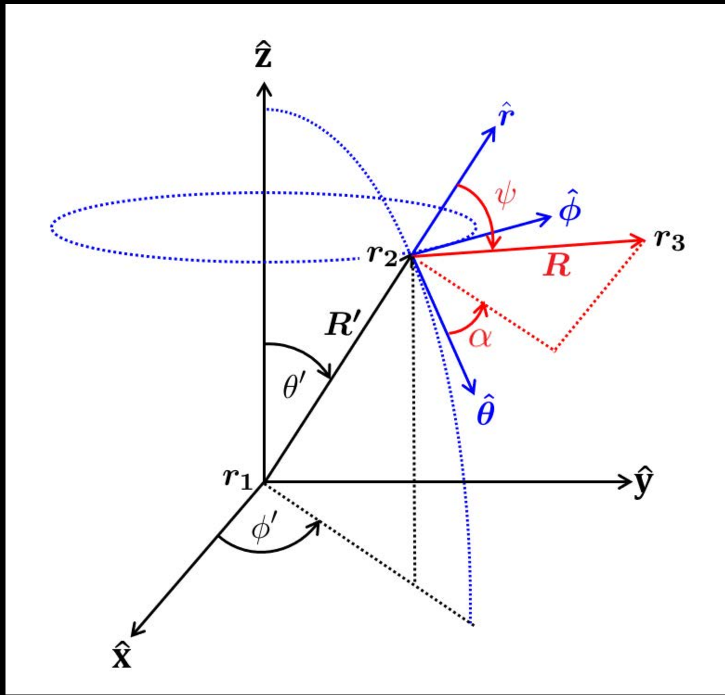


1. Basic definitions and useful formulas (Stokes vector)

$$\boldsymbol{S} = \begin{bmatrix} I \\ Q \\ U \\ V \end{bmatrix} = \frac{1}{2}\sqrt{\frac{\epsilon}{\mu}} \begin{bmatrix} E_{o\theta}E_{o\theta}^* + E_{o\phi}E_{o\phi}^* \\ E_{o\theta}E_{o\theta}^* - E_{o\phi}E_{o\phi}^* \\ -E_{o\theta}E_{o\phi}^* - E_{o\phi}E_{o\theta}^* \\ i\left(E_{o\phi}E_{o\theta}^* - E_{o\theta}E_{o\phi}^*\right) \end{bmatrix}$$

$$\boldsymbol{S} = \begin{bmatrix} I \\ Q \\ U \\ V \end{bmatrix} = \begin{bmatrix} intensity \\ linear\ 0° - linear\ 90° \\ linear\ 135° - linear\ 45° \\ circular\ left - circular\ right \end{bmatrix}$$





1. Basic definitions and useful formulas (Stokes vector)

The single scattered Stokes vector ***S*** in the meridional plane is given by the equation

$$\boldsymbol{S \sim R(\alpha) F(\psi)\, R(\alpha') S'}$$

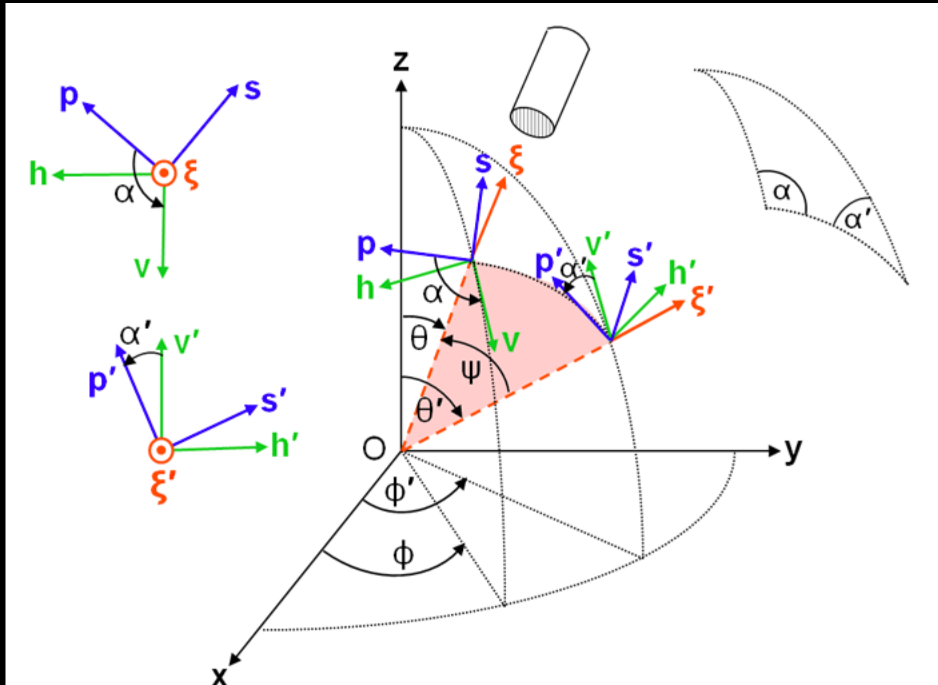
where **S'** is the incident Stokes vector, ***R*** is a rotation matrix that rotates the Stokes vector from meridional plane to scattering plane (and back) and ***F*** is the scattering matrix.

$$\boldsymbol{R(\alpha)} = \begin{bmatrix} 1 & 0 & 0 & 0 \\ 0 & \boldsymbol{cos2\alpha} & \boldsymbol{-sin2\alpha} & 0 \\ 0 & \boldsymbol{sin2\alpha} & \boldsymbol{cos2\alpha} & 0 \\ 0 & 0 & 0 & 1 \end{bmatrix}$$

with

$$\cos\alpha = (cos\theta' - cos\theta cos\psi)/(sin\theta sin\psi)$$

and ψ is the scattering angle.



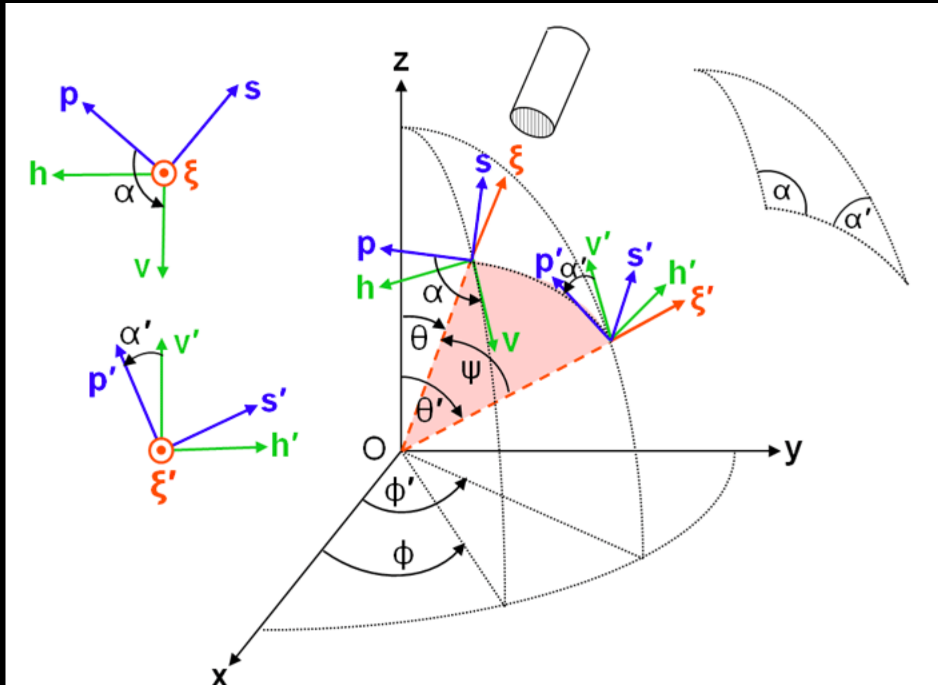
1. Basic definitions and useful formulas (Stokes vector)

Often (for statistically isotropic and mirror symmetric random particles) the structure of the scattering matrix is

$$\boldsymbol{F}(\psi) = \begin{bmatrix} F_{11}(\psi) & F_{12}(\psi) & 0 & 0 \\ F_{12}(\psi) & F_{22}(\psi) & 0 & 0 \\ 0 & 0 & F_{33}(\psi) & F_{34}(\psi) \\ 0 & 0 & -F_{34}(\psi) & F_{44}(\psi) \end{bmatrix}$$

so for unpolarized incident (sun) light it is useful to look at the Stokes vector in the scattering plane

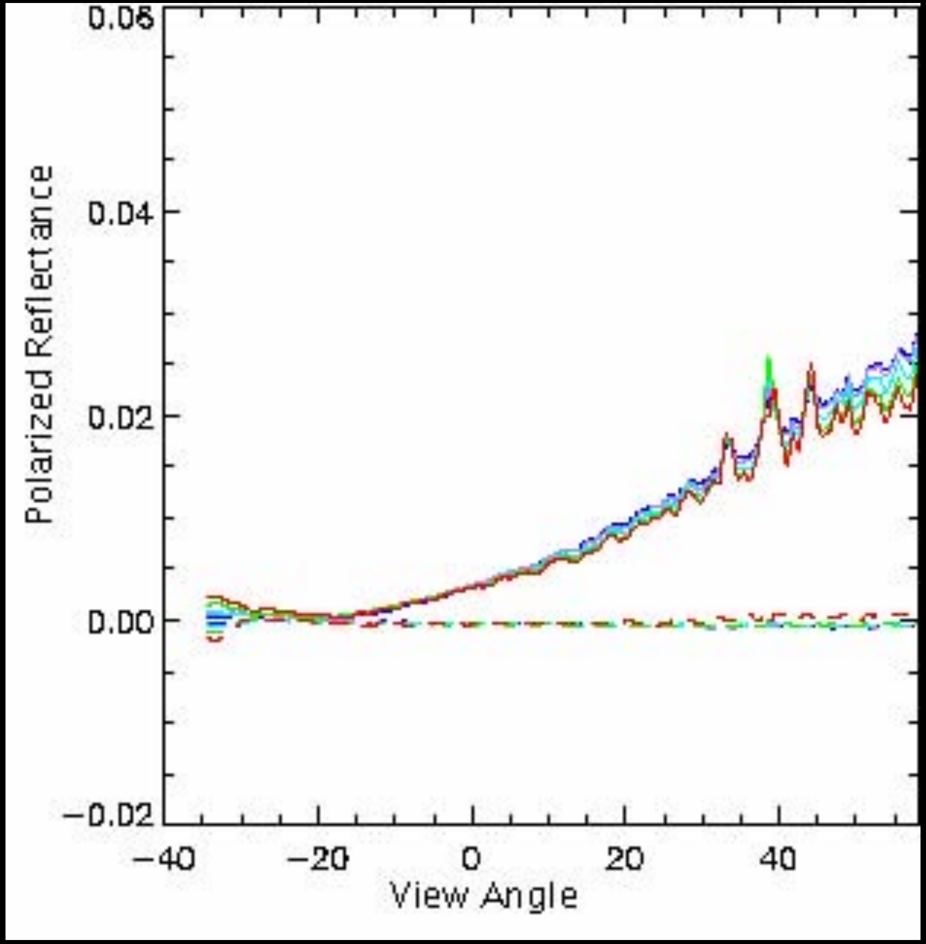
$$S_{scatt} = R(-\alpha) S \sim \begin{bmatrix} F_{11}(\psi \\ F_{12}(\psi) \\ 0 \\ 0 \end{bmatrix}$$



because the Stokes parameter U~0 and most of the information is in the Stokes parameter Q.

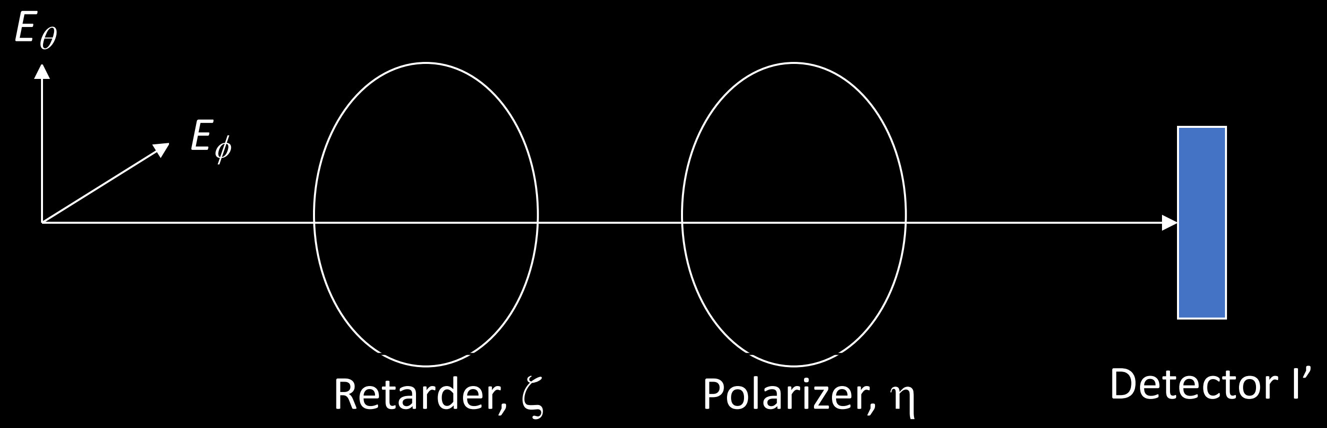
1. Basic definitions and useful formulas (Stokes vector)

An example of the benefits of rotating into the scattering plane from 200 m above Mexico City



1. Basic definitions and useful formulas (Stokes vector)

How could you measure the Stokes vector?



Put a retarder (changes phase of plane electromagnetic wave +ζ/2 along the θ–axis and –ζ/2 along the ϕ–axis) and a polarizer oriented at an angle η between a relay telescope and a detector.

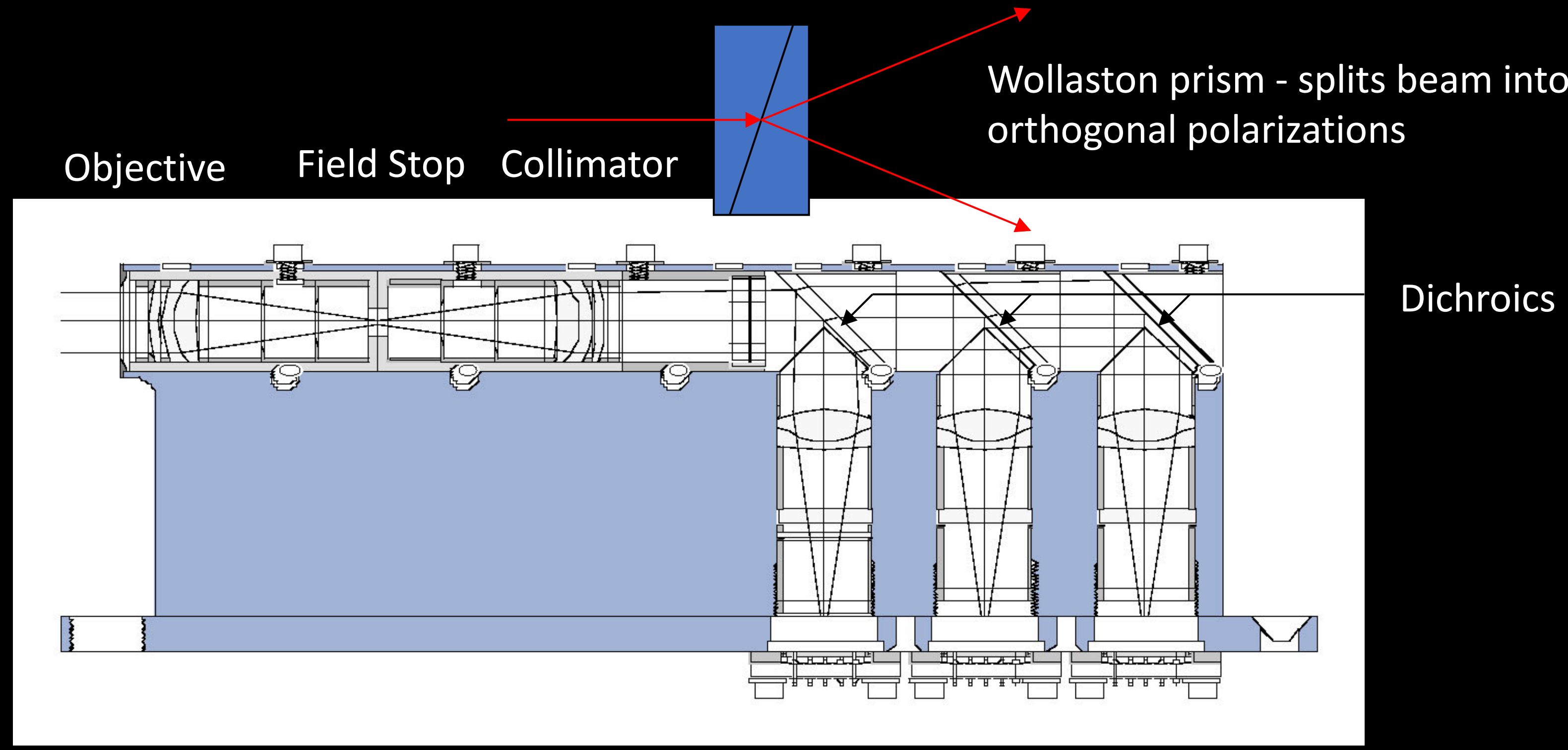
$$I'(\eta,\zeta) = \frac{1}{2}(I + Qcos2\eta - Usin2\eta cos\zeta - Vsin2\eta sin\zeta)$$

$$I = I'(0,0) + I'(90,0), \qquad Q = I'(0,0) - I'(90,0), \qquad U = I'(135,0) - I'(45,0), \qquad V = I - 2I'(45,90)$$

Generally, for Earth observations V is small and we only need to measure I, Q and U

1. Basic definitions and useful formulas (Stokes vector)

How could you measure the Stokes vector?



And in point of fact, with a pair of telescopes with Wollaston prisms oriented at 45° to one another you can simultaneously make the set of measurements for I, Q and U described on the previous slide in three spectral bands simultaneously. Throw in four more telescope and you can measure I, Q and U for nine spectral bands simultaneously. This is the Research Scanning Polarimeter.

1. Basic definitions and useful formulas (comment)

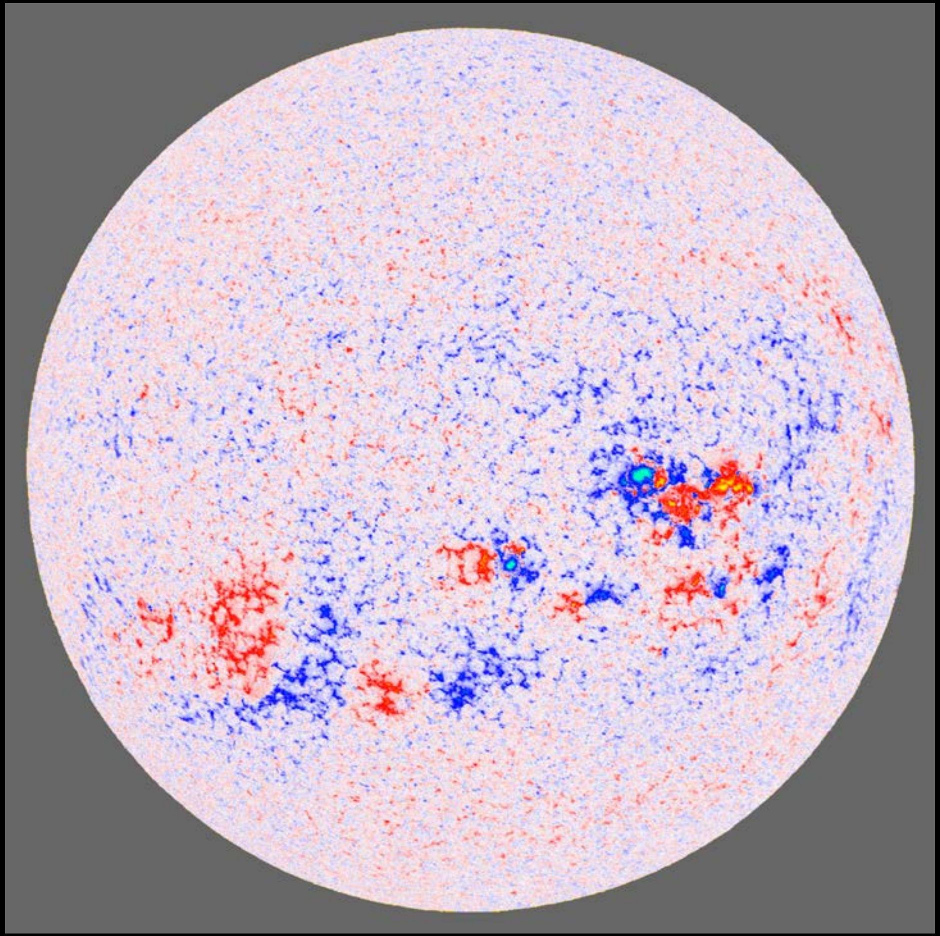
The definition of the Stokes vector given above has the dimension of an irradiance, but this can be generalized to a Stokes vector for the radiance.

Of more importance is that, radiometry has historically been a phenomenological theory that is not rigorously tied to Maxwell’s equations. This has been corrected to a large extent by Michael Mishchenko in his 2014 book, Electromagnetic Scattering by Particles and Particle Groups (Cambridge University Press) and subsequent papers in collaboration with Adrian Doicu.

In that work the radiance is directly linked to the two-point electric field dyadic correlation function and a Poynting Stokes Tensor, such that the radiance and the radiative transfer equation (RTE) used to propagate it are directly tied to the solution of Maxwell’s equations. As Mishchenko comments in his book “It is, in fact, remarkable that although the RTE has the formal mathematical structure of a kinetic equation descrbing particle transport, it follows directly from the electromagnetic ***wave*** theory.” - his emphasis.

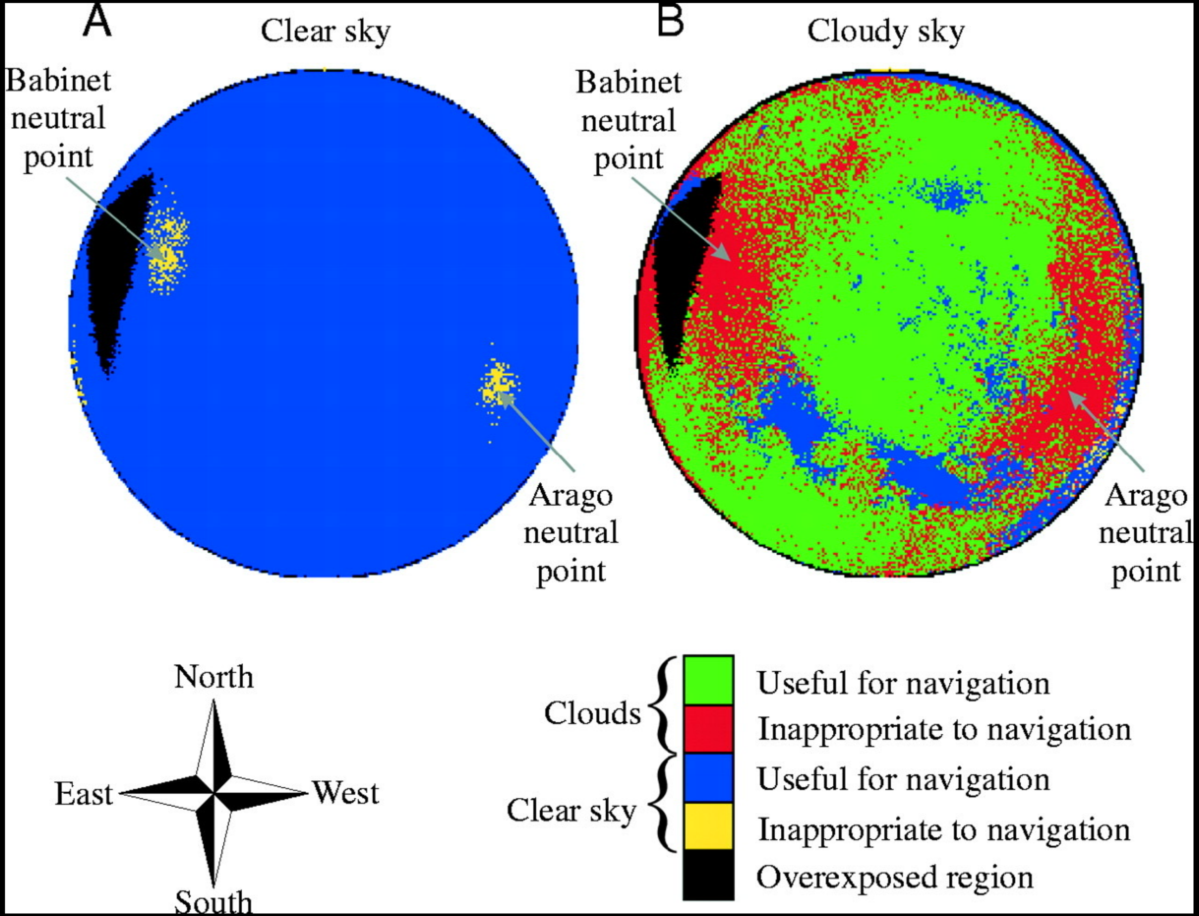
Aside:

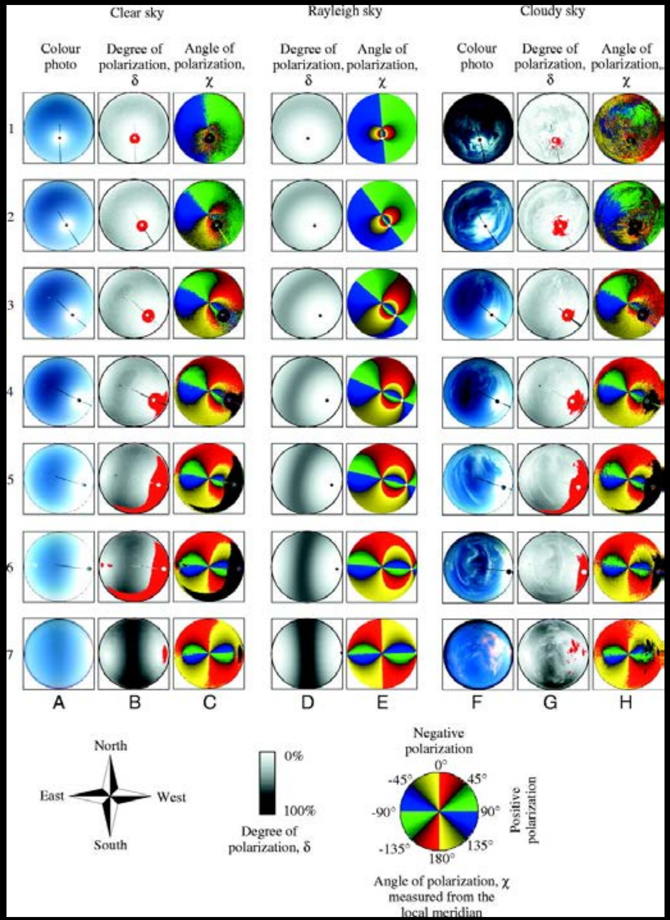
While the disk integrated solar Stokes vector is essentially unpolarized (and this is what we are interested in for remote sensing of the Earth) images of the solar disk itself show significant polarization features that can be used to map the Zeeman effect (longitudinal magnetic field) using observations of ***circular*** polarization.



Aside:

- Desert ants and honey bees use ultraviolet receptors in a specialized dorsal rim area of the eye as polarization analyzers that provide a polarization “compass” even under cloudy skies. Ultraviolet and polarization sensitivity is common in many animals, so PACE is looking at the world in a more natural way!





Aside:

- Polarization observations have also been used for remote sensing of other planets dating back to the early years of the 20th century. These observations were explained in 1974 by Hansen and Hovenier as sulfuric acid (refractive index of 1.45) clouds (with droplets of 1.1 μm radius) in the upper atmosphere of Venus. https://doi.org/10.1175/1520-0469(1974)031<1137:IOTPOV>2.0.CO;2

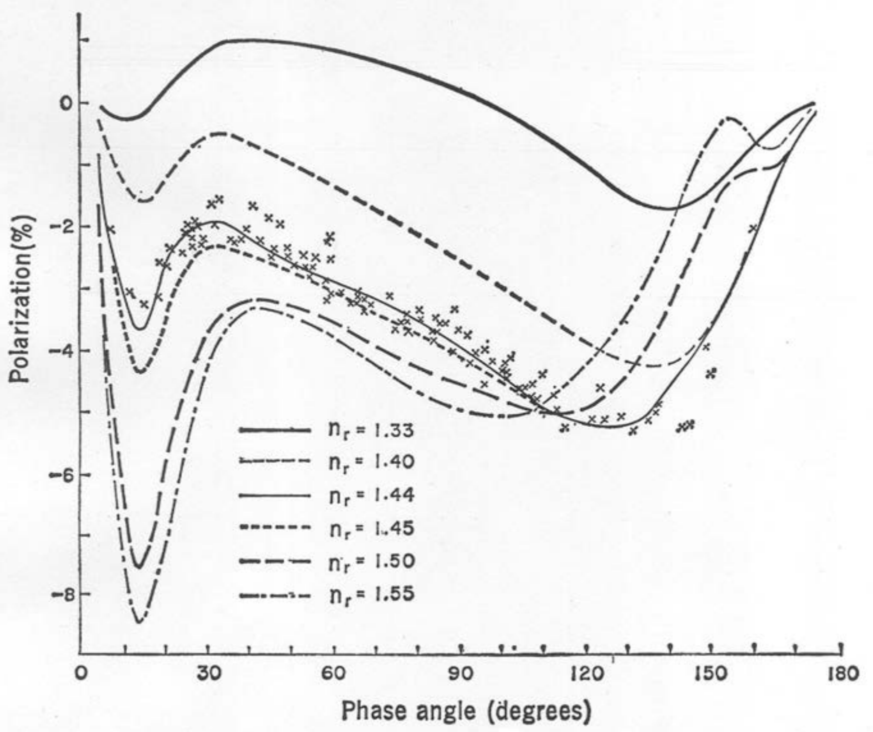


Fig. 2. The crosses show the observations of Coffeen and Gehrels of the polarization of Venus at a wavelength of 0.99 μ (2). For each refractive index n_r the theoretical calculations are for the particle size giving best agreement with the observations: $\bar{r}$ = 0.6, 0.8, 1.1, 1.1, 1.2, and 1.2 μ, respectively, beginning with $n_r = 1.33$. The curves for $n_r = 1.44$ and 1.45 are indistinguishable for phase angles greater than 110°. The albedo of Venus is assumed to be ~ 90 percent at $\lambda = 0.99\ \mu$.

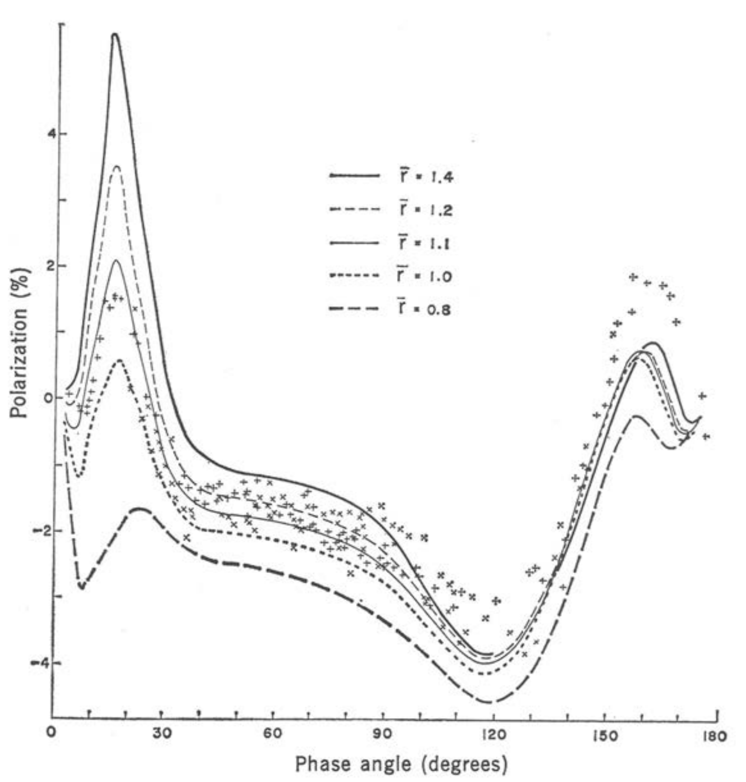
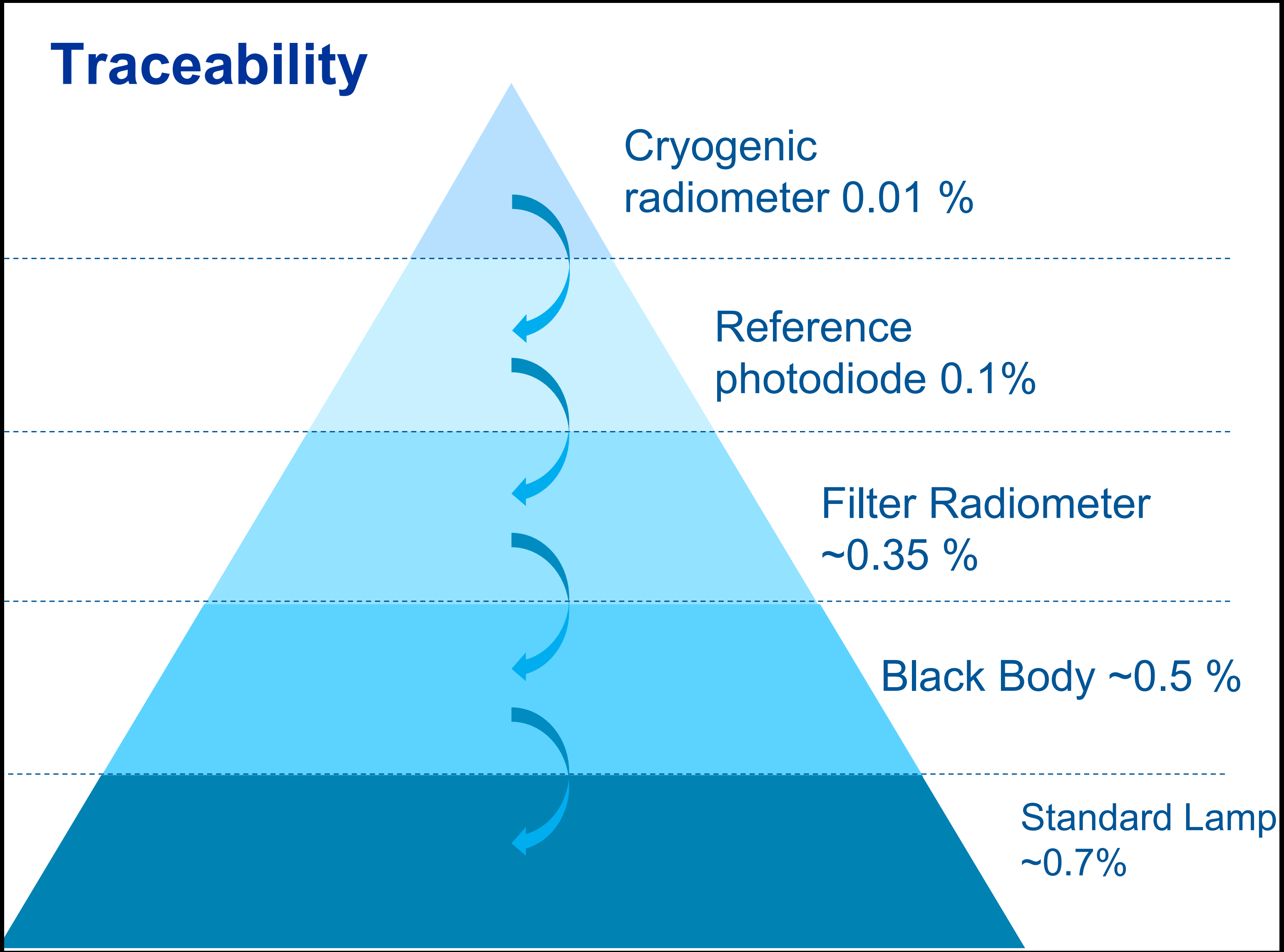


Fig. 4. The pluses represent the visual observations of Lyot (1), and the crosses are the intermediate-bandwidth observations of Coffeen and Gehrels at $\lambda = 0.55\ \mu$ (2). The theoretical curves are for $n_r = 1.45$ with several values for the mean radius $\bar{r}$. The albedo of Venus is assumed to be ~ 87 percent, and the Rayleigh scattering determined in the ultraviolet region ($f_R = 0.045$) is included after reduction by a λ^{-4} law.

2. Calibration summary

Primary and preferred secondary standards for radiometric calibration are now detector based, but source-based standards (lamps/integrating spheres) are still more readily available.

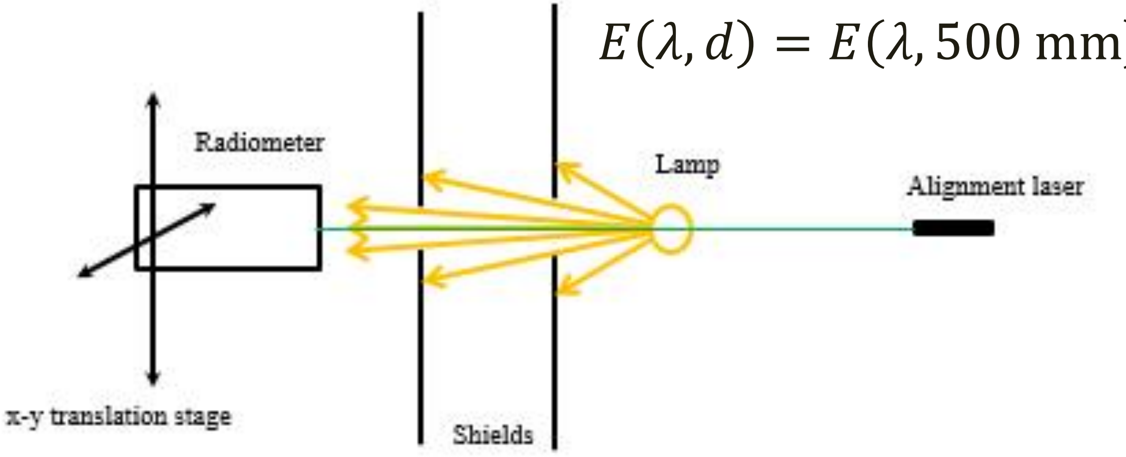


2. Calibration summary

Lamps are most widely available calibration source. 1 KW FEL lamp is bright and hot and does not have a spectrum anything like the sun. Only get calibration at one level….

Irradiance

$$E(\lambda, d) = E(\lambda, 500\ \mathrm{mm}) \left(\frac{500\ \mathrm{mm}}{d}\right)^2$$



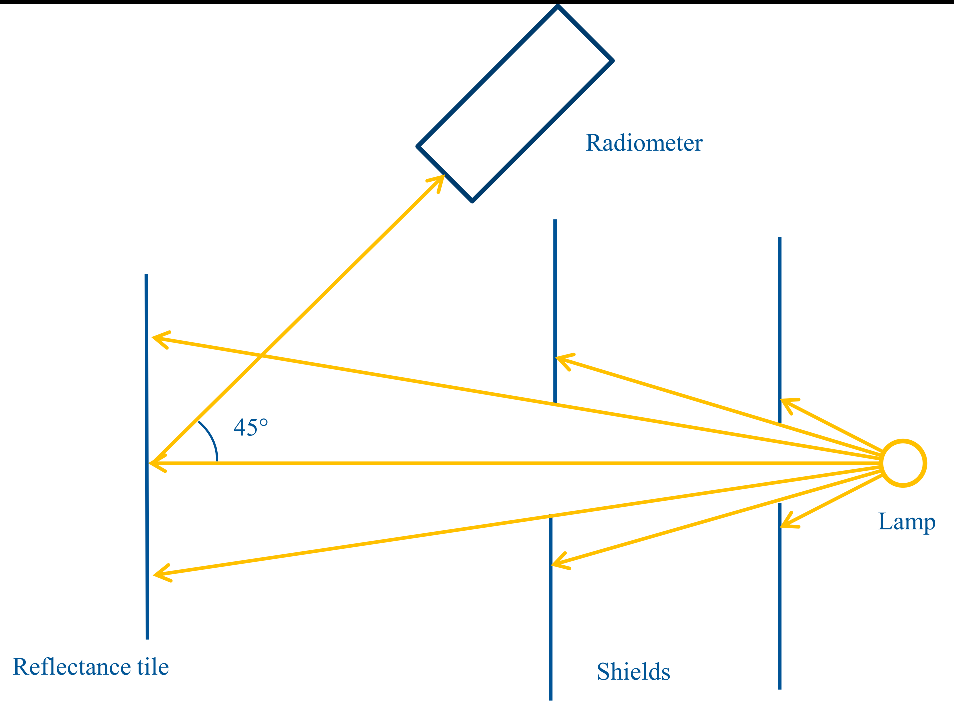
Bialek, IOCCG 2016 SLS, Villefranche Sur Mer

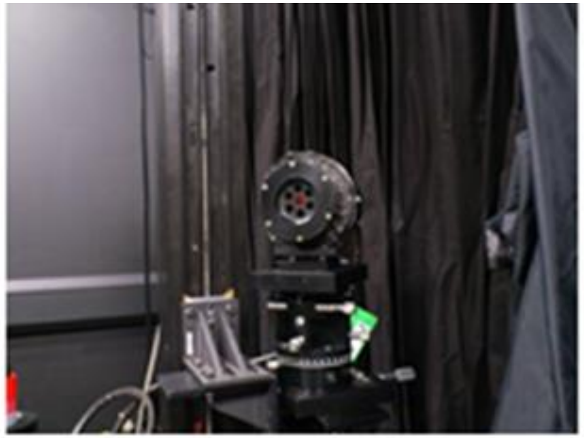
2. Calibration summary

Lamps are most widely available. An integrating sphere with multiple lamps is a good (but expensive) alternative

Radiance

Lamp - tile





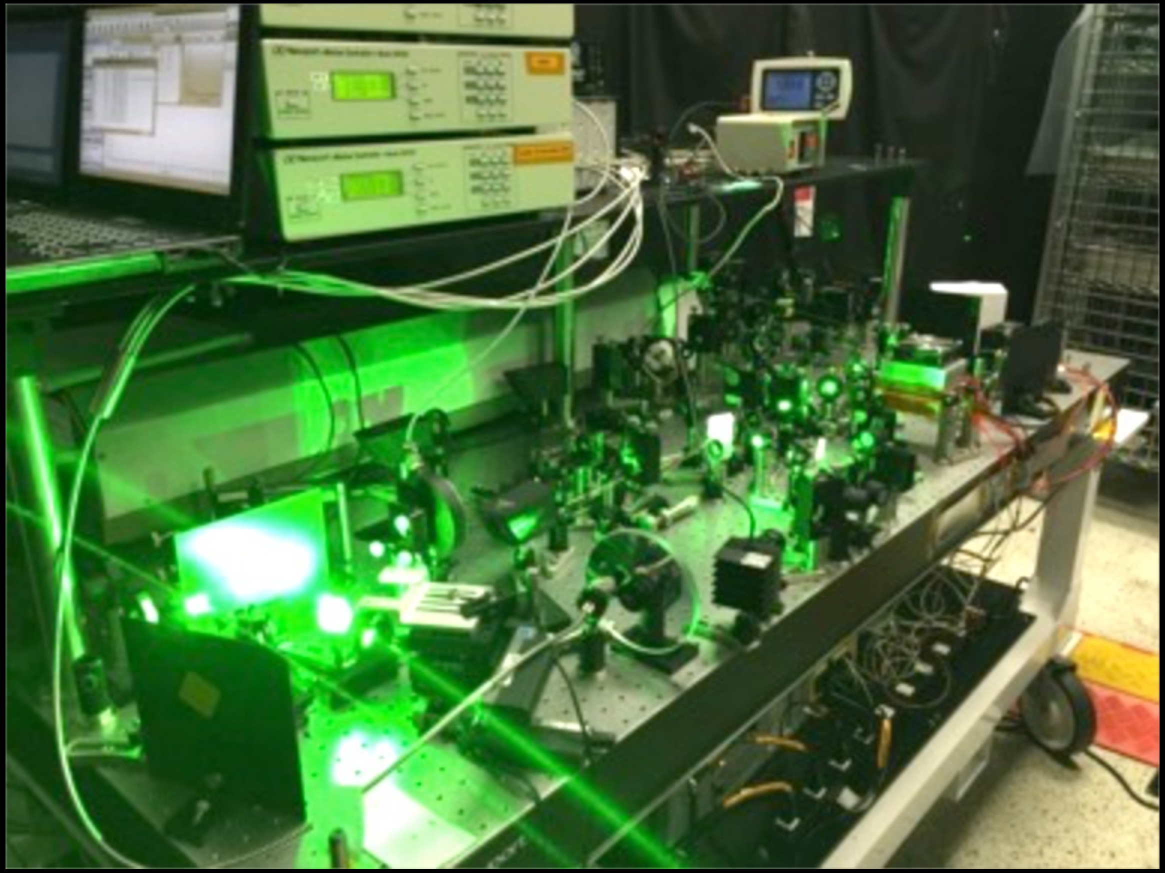
$$L_s = \frac{E_{FEL}\beta_{0:45}}{\pi}\frac{d_{cal}^2}{d_{use}^2}$$

Bialek, IOCCG 2016 SLS, Villefranche Sur Mer

2. Calibration summary

Lamps are most widely available, but a tunable laser system combined with a reference photodiode provides a much more accurate and capable calibration system. Only problem is expense and availability, so primarily used for satellite instruments (currently).

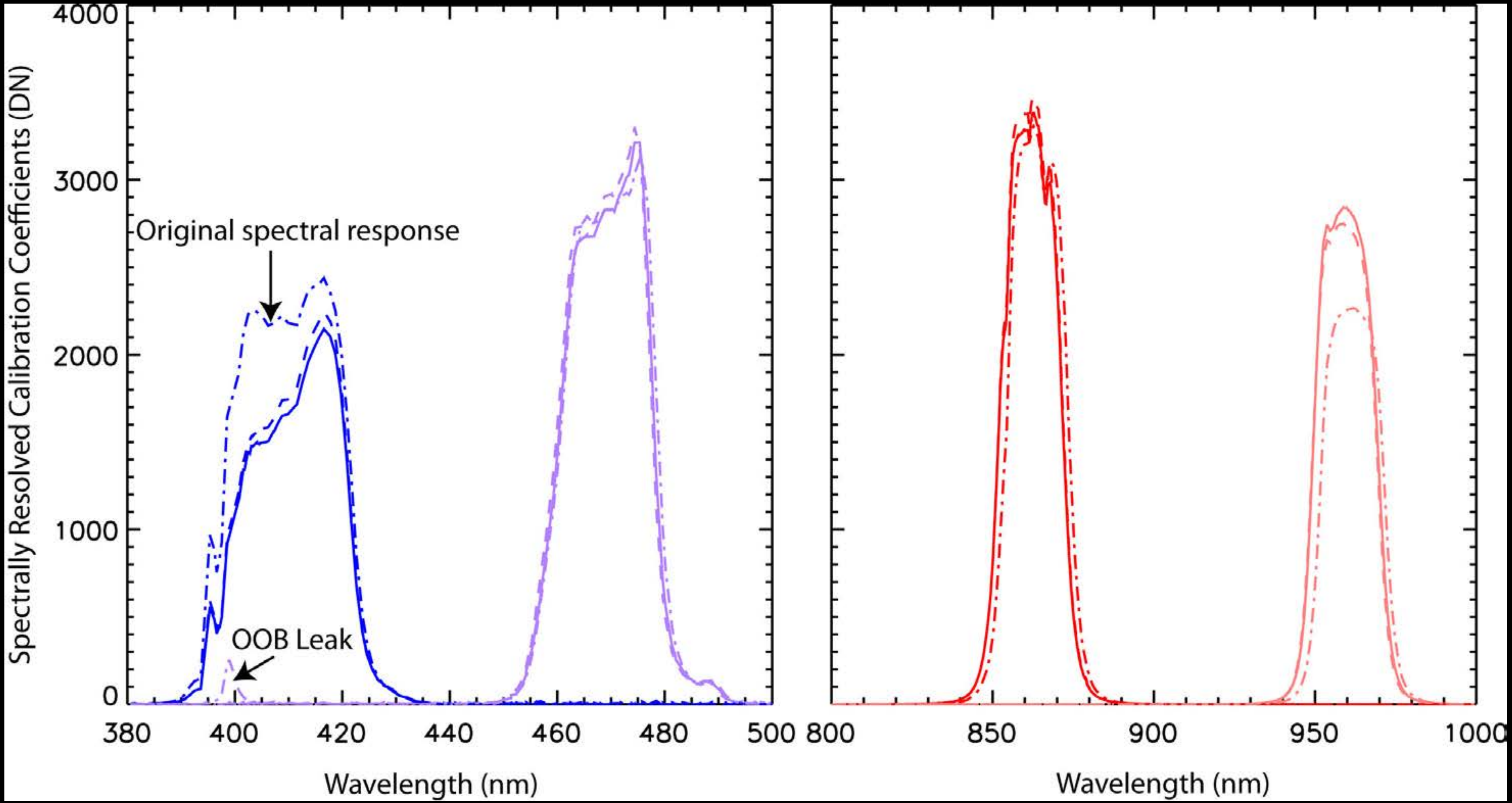
Tunable laser light is injected into an integrating sphere that is observed by the instrument being calibrated. The sphere is also monitored by reference photodiodes so the radiance calibration is directly traceable to the primary reference standard. The primary reference standard is a cryogenically cooled electrical substitution radiometer that links the radiometric standard to that for electrical power.



The Goddard Laser for Absolute Measurement of Radiance (GLAMR) system has been used to calibrate OCI.

2. Calibration summary

Lamps are most widely available, but a tunable laser system combined with a reference photodiode provides a much more accurate and capable calibration system. Get not only absolute radiometry at 0.2% accuracy but also a complete spectral characterization that can readily detect (or rule out) spectral leaks.



3. Remote Sensing

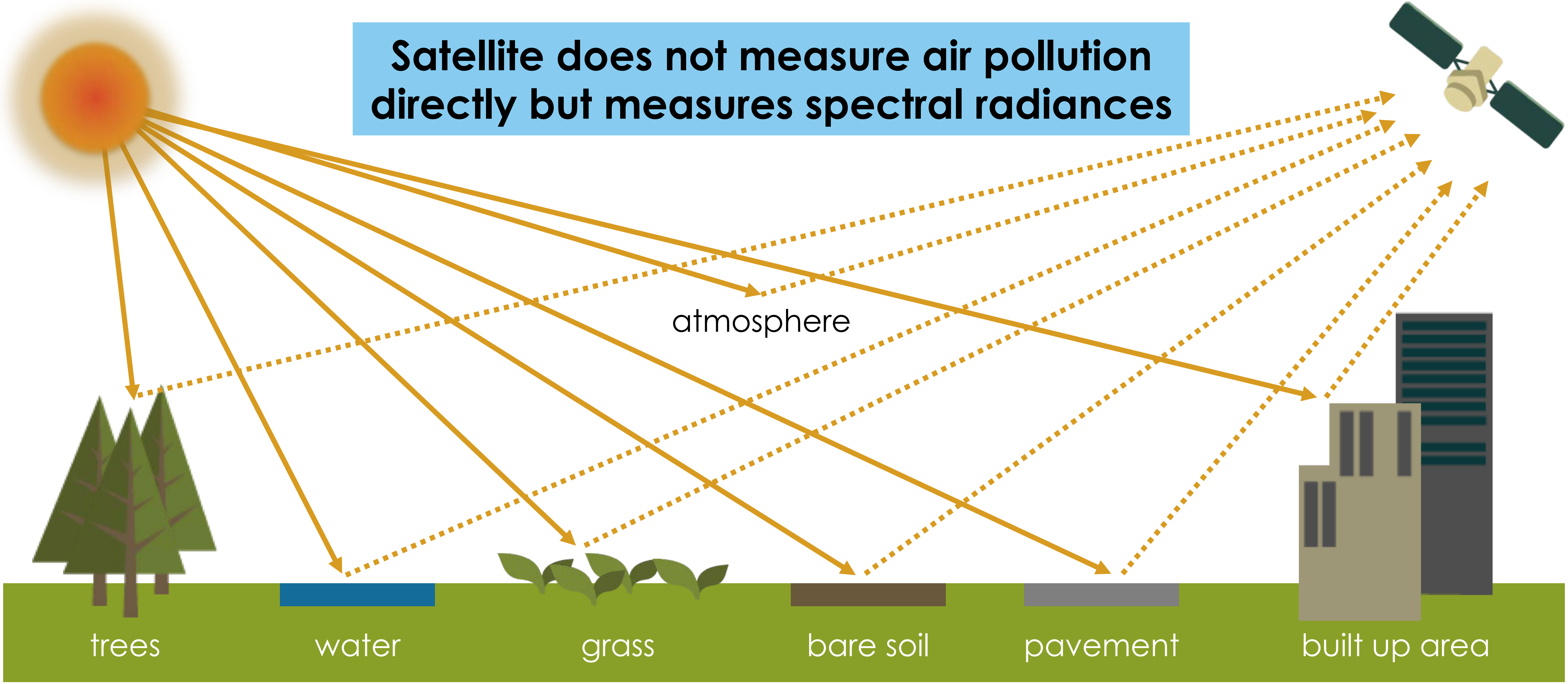
The Earth is the most visually spectacular planet in the solar system.

Blue skies, blue (and other color) oceans, wide range of land surfaces, white clouds and aerosols (small particles that are not made of liquid or ice water) with all sorts of colors.

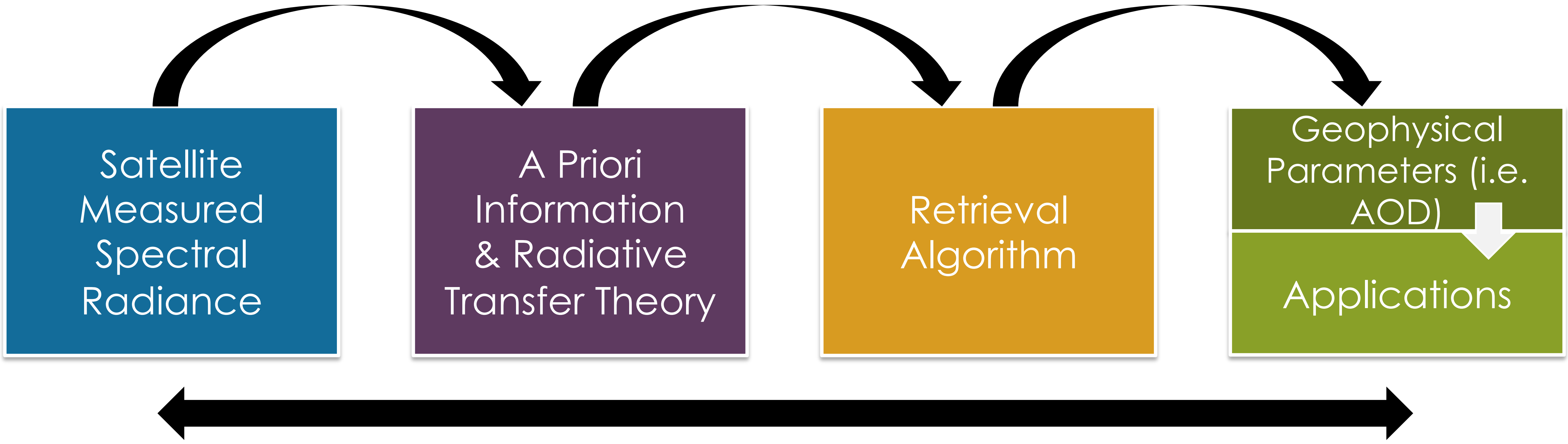
Remote sensing requires that you have good models of how all these elements of the Earth system scatter and absorb light so that you can make inferences about what you are looking at.

What do satellites measure ?

Satellite does not measure air pollution directly but measures spectral radiances

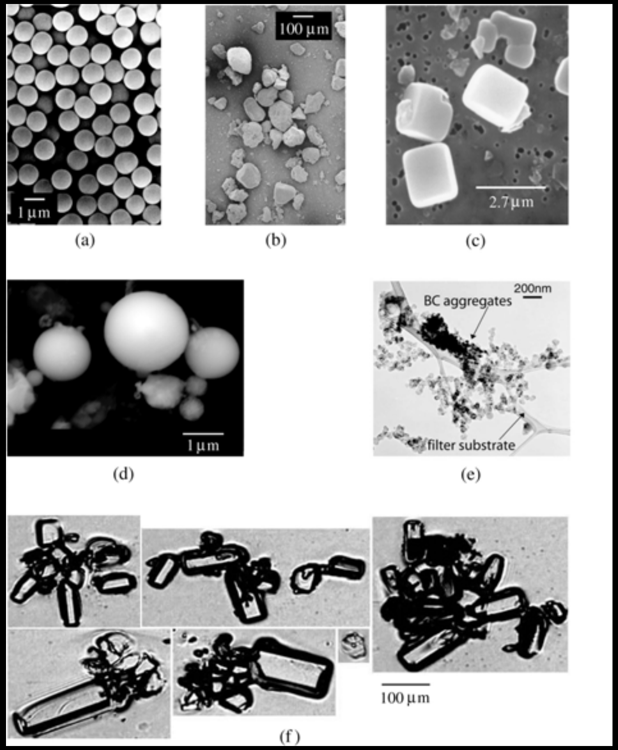


The Remote Sensing Process



3. Remote Sensing

Remote sensing takes the remotely sensed (polarized) radiance (field) and infers the distribution of scatterers and absorbers and their sizes and compositions that generated it. With plenty of assumptions, or prior information.



Particle size, shape and composition determine optical properties

$$L(\boldsymbol{x}, \boldsymbol{t}, \boldsymbol{\xi}, \boldsymbol{\lambda}) = F(\boldsymbol{a}) + \boldsymbol{\varepsilon}$$

Radiance = Solution of RTE + noise (of all kinds)

Optical properties together with concentration profiles of particles and gases (***a***) determine observed radiation field

3. Remote Sensing

Remote sensing takes the remotely sensed (polarized) radiance (field) and infers the distribution of scatterers and absorbers and their sizes and compositions that generated it. With plenty of assumptions, or prior information.

Retrieval is a (generally iterative) minimization where you find the state a_{ret}

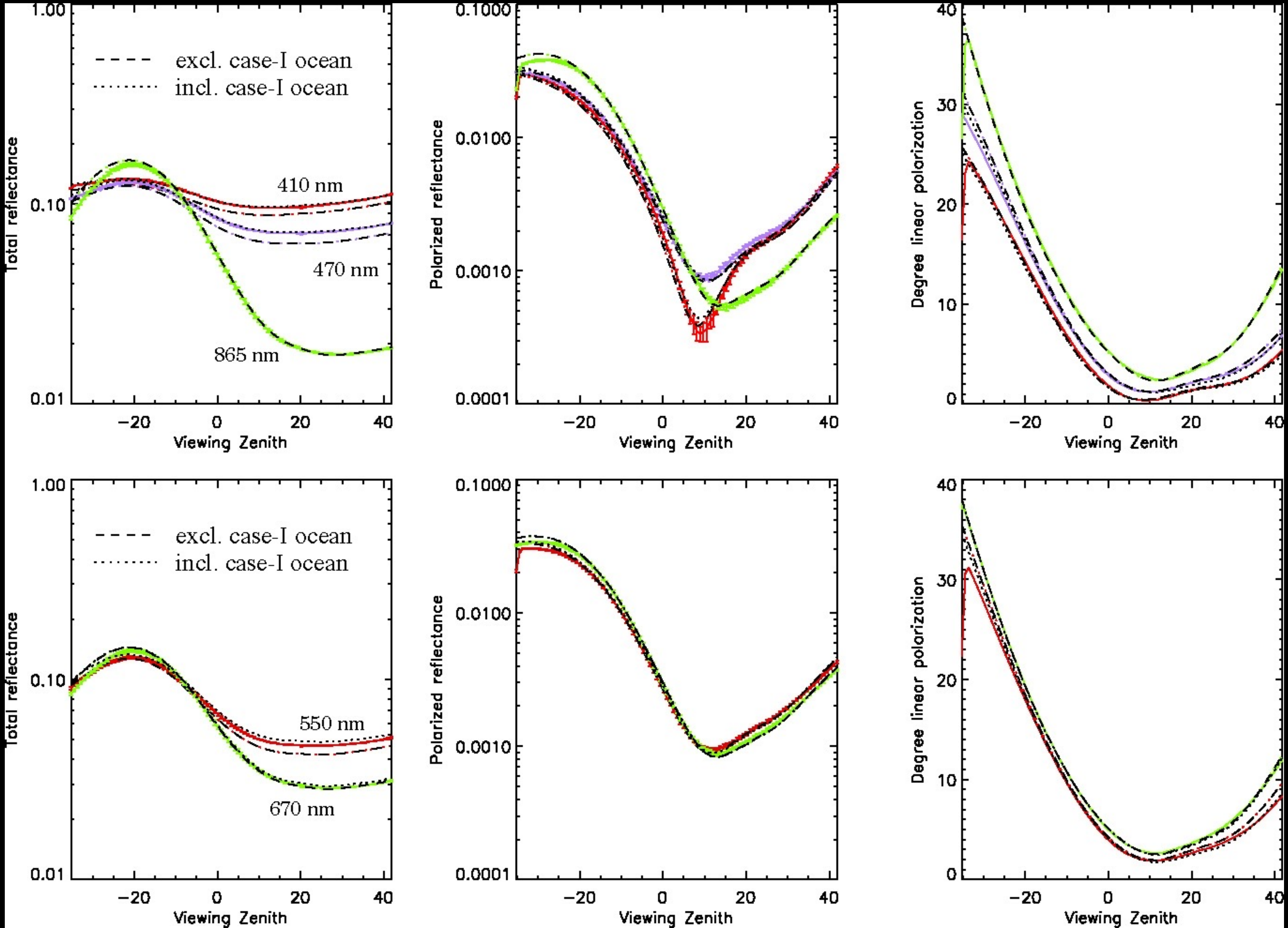
$$a_{\boldsymbol{ret}} = \boldsymbol{min}_a \left\| \boldsymbol{S}_{\boldsymbol{\varepsilon}}^{-1/2} [\boldsymbol{L}(\boldsymbol{x}, \boldsymbol{t}, \boldsymbol{\xi}, \boldsymbol{\lambda}) - F(\boldsymbol{a})] \right\| + \boldsymbol{\Phi}(\boldsymbol{a})$$

that best matches the observations without disagreeing with the constraint function $\boldsymbol{\Phi}$. The constraint function can be a prior probability distribution, the norm of the retrieval vector (Philips-Tikhonov regularization), or a function of the steps in the iteration (Levenberg-Marquardt method) Sometimes retrieval schemes will transition between these methods e.g. using LM initially to prevent the iteration diverging.

There are entire books about this e.g. C. Rodgers, Inverse Methods for Atmospheric Sounding (World Scientific, 2000).

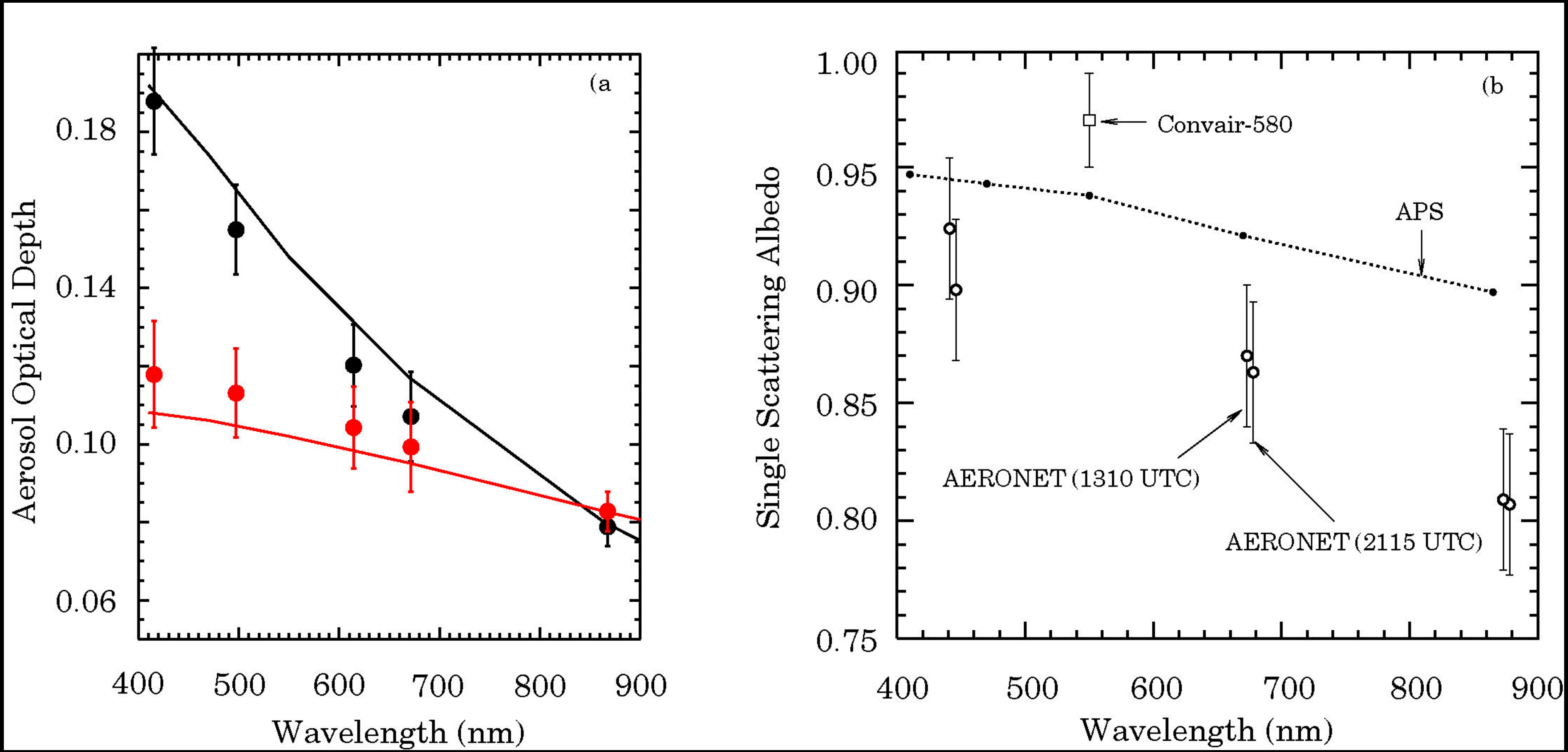
3. Remote Sensing

The result of the iteration is that you fit your observations……



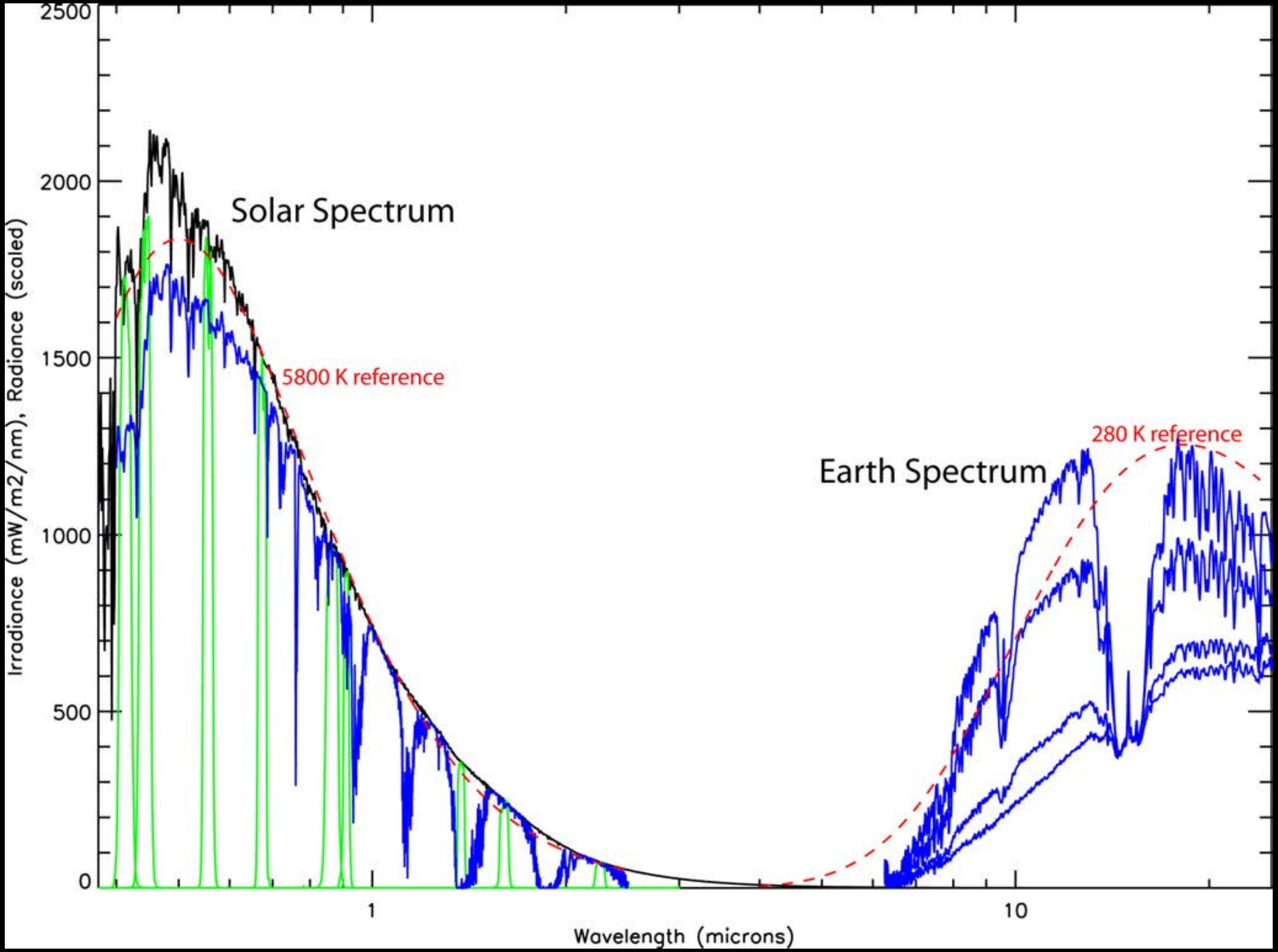
3. Remote Sensing

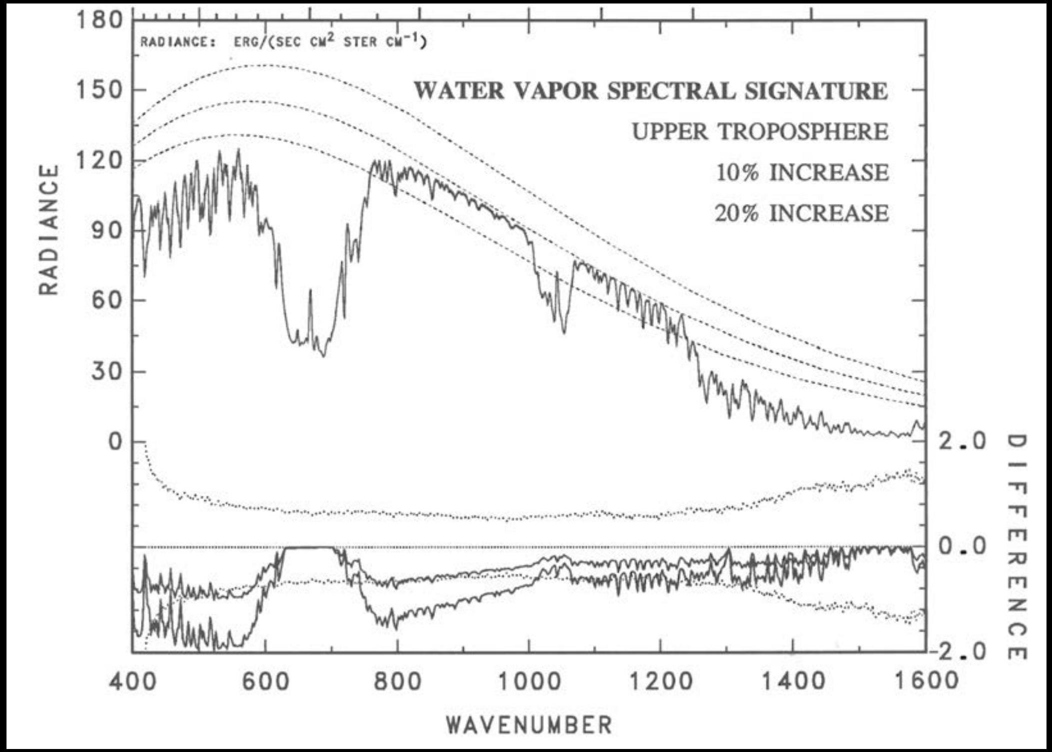
and get a set of retrieved parameters that are consistent with the observations and any prior constraints……



3. Remote Sensing

While the Earth's energy (im)balance is between incoming solar radiation and outgoing longwave radiation emitted by the Earth, here we will focus just on the reflected solar radiation.



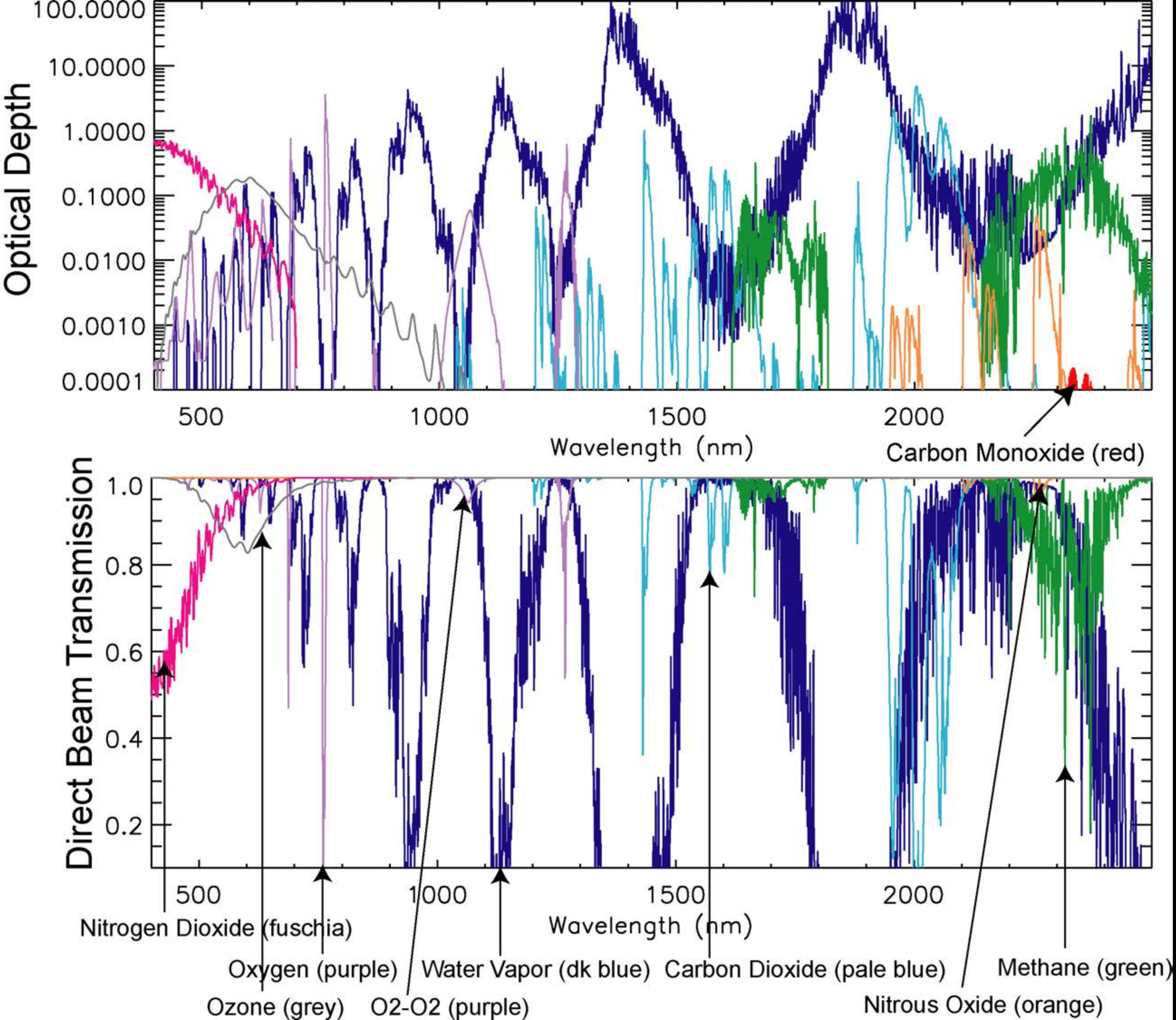


IR spectra on left are from Nimbus IV IRIS in 1970. Warming has been closing the "dirty window" λ>20 μm, but measurements of this spectral region will not happen again until next year with PreFIRE.

3. Remote Sensing

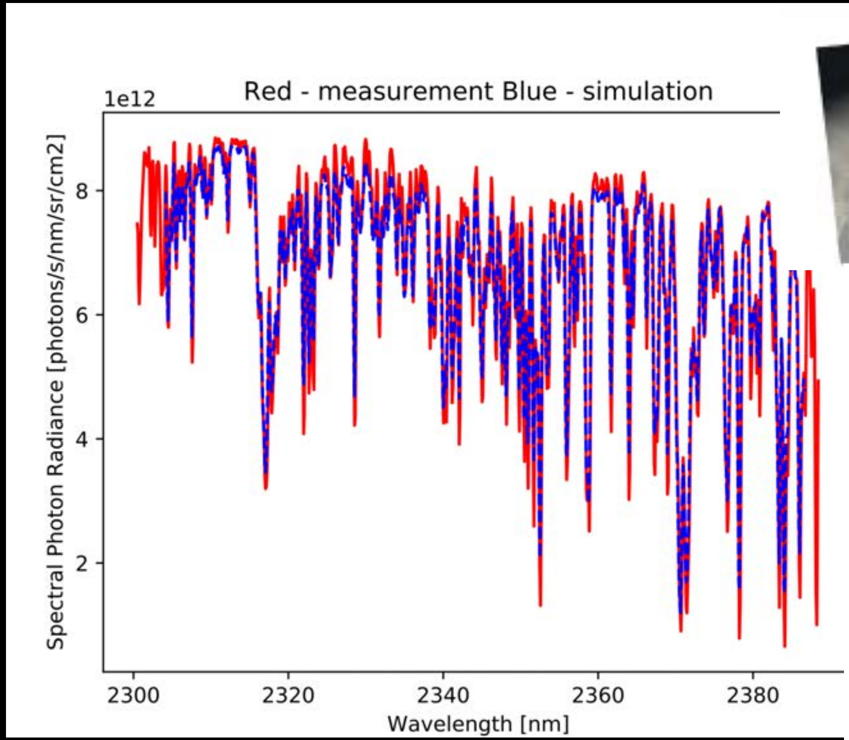
Absorbing gases may be the target of interest (e.g. ozone, nitrogen dioxide, sulfur dioxide, formaldehyde, carbon monoxide, carbon dioxide, methane etc.), or an irritating thing that gets in the way of seeing what you are really interested in.

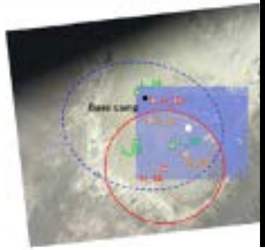
Origin of all gaseous absorption calculations is HITRAN (https://hitran.org/), but there are packages that provide simpler ways than doing everything yourself (e.g. http://www.libradtran.org/doku.php)



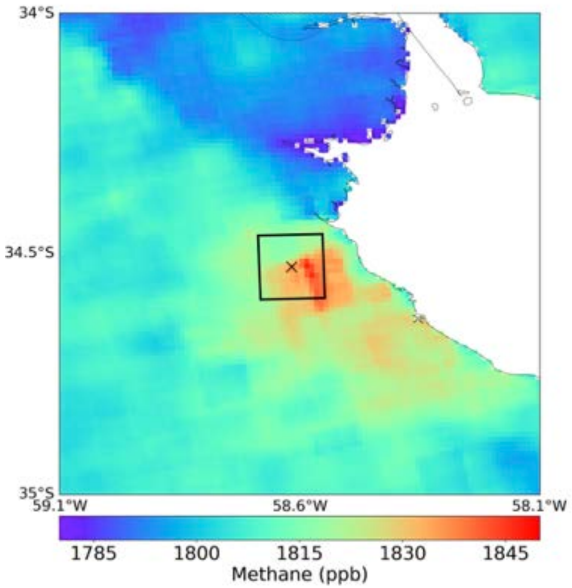
3. Remote Sensing

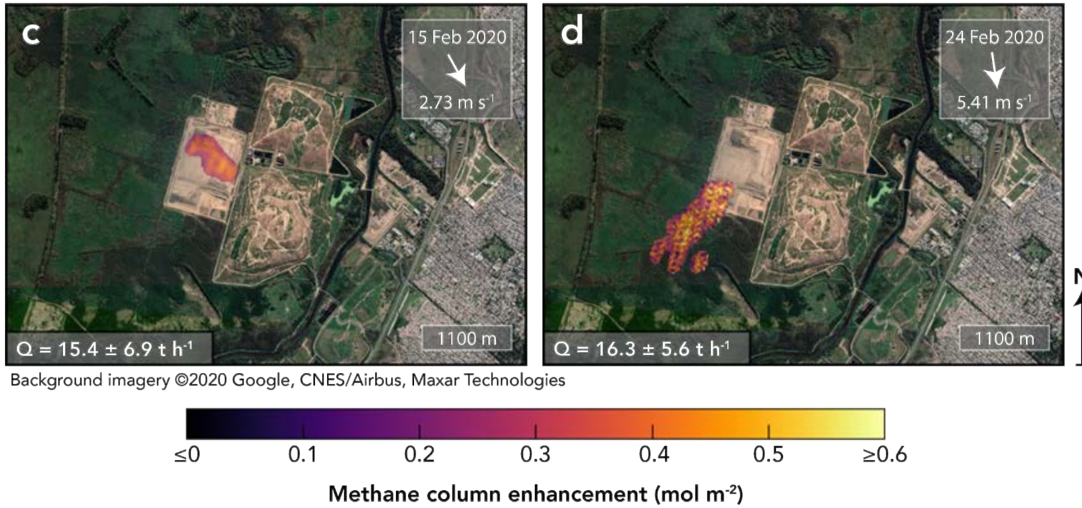
<u>Absorbing gases:</u> instruments focused on trace gas retrievals generally have higher spectral resolution and coarser spatial resolution than instruments focused on clouds and aerosols.





Combining daily global coverage at ~ 5 km resolution from TropOMI with targeted measurements from GHGSat (25 m resolution) allows sources of methane to be identified and source strengths determined.





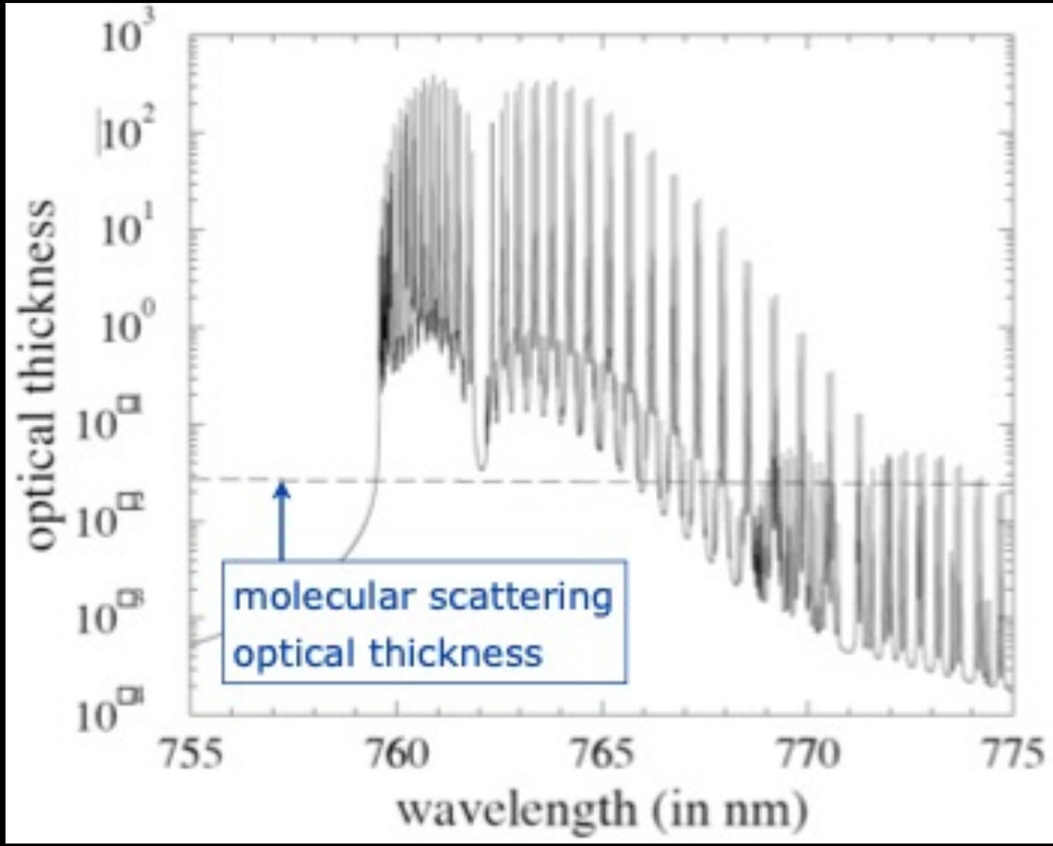
Maasakkers et al, in prep.

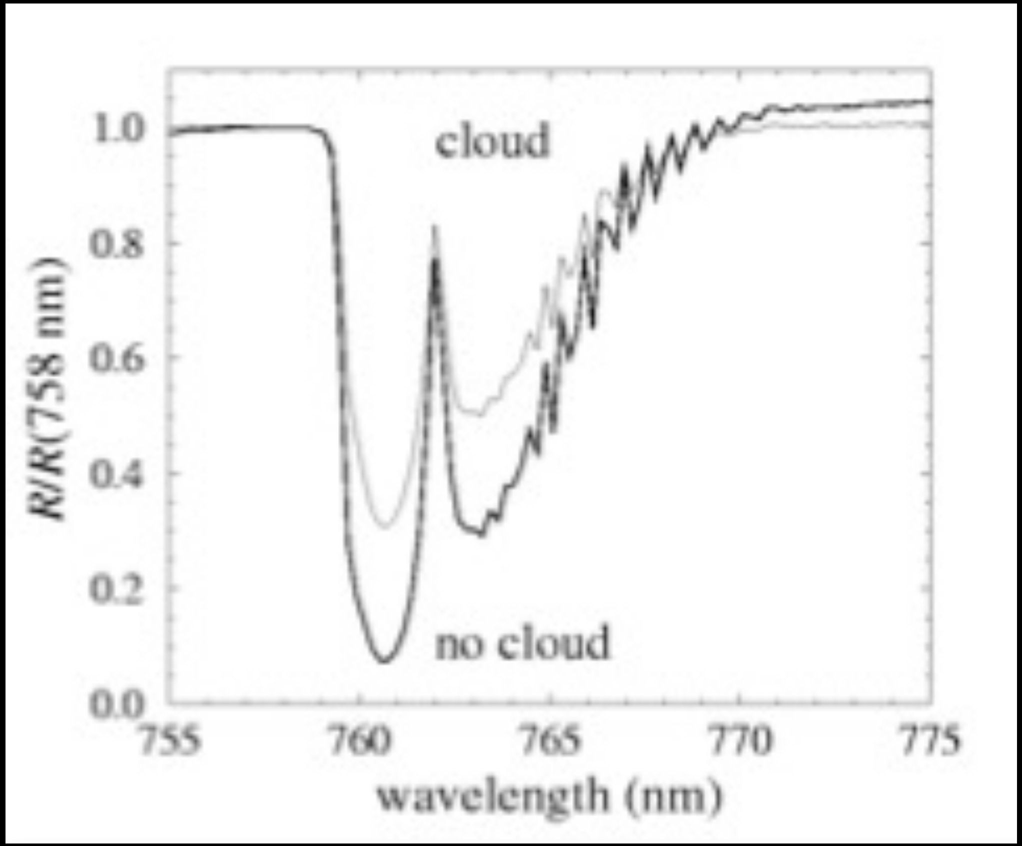
https://eartharxiv.org/repository/view/2985/ Methane source shown in figure is Norte III landfill in Buenas Aires, Argentina.

3. Remote Sensing

<u>Absorbing gases</u> can in some cases (particularly for well mixed gases) be a useful tool to characterize the target you are really interested in.

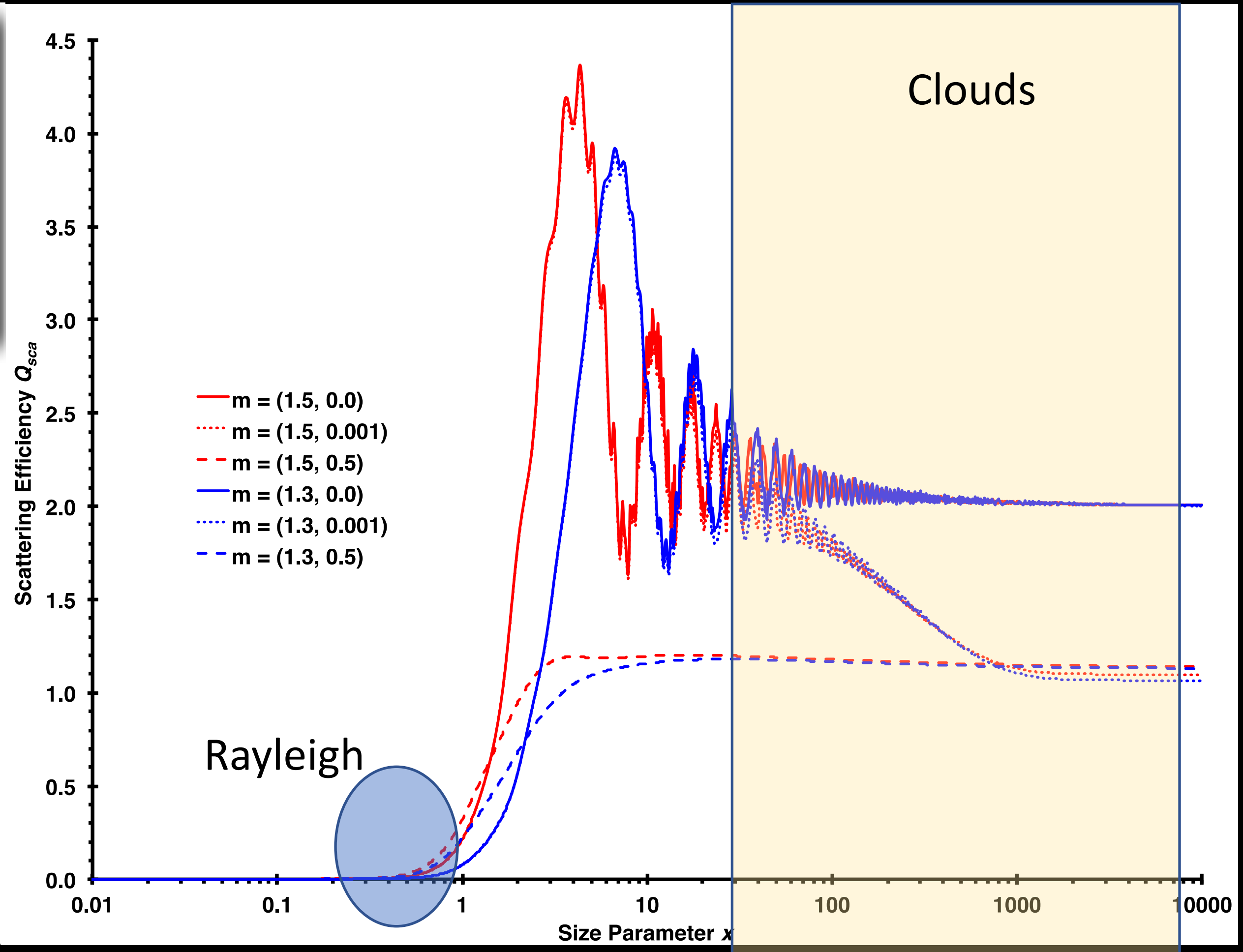
More on cloud top pressure on Wednesday in L10.





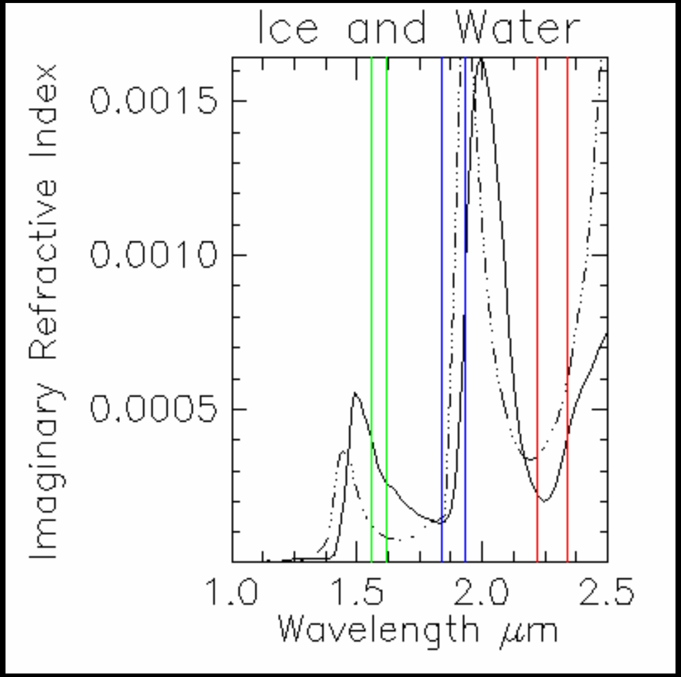
3. Remote Sensing

Particle scattering: Aside from absorbing gases we need to think about how particles of different sizes scatter. Very small particles scatter with a wavelength dependence of λ^{-4} (blue sky), whereas very large particles scatter with no wavelength dependence (clouds are white). The size parameter x in the right figure is $x=2\pi r/\lambda$ (from Moosmüller and Ogren, Atmosphere 2017, **8**, 133).

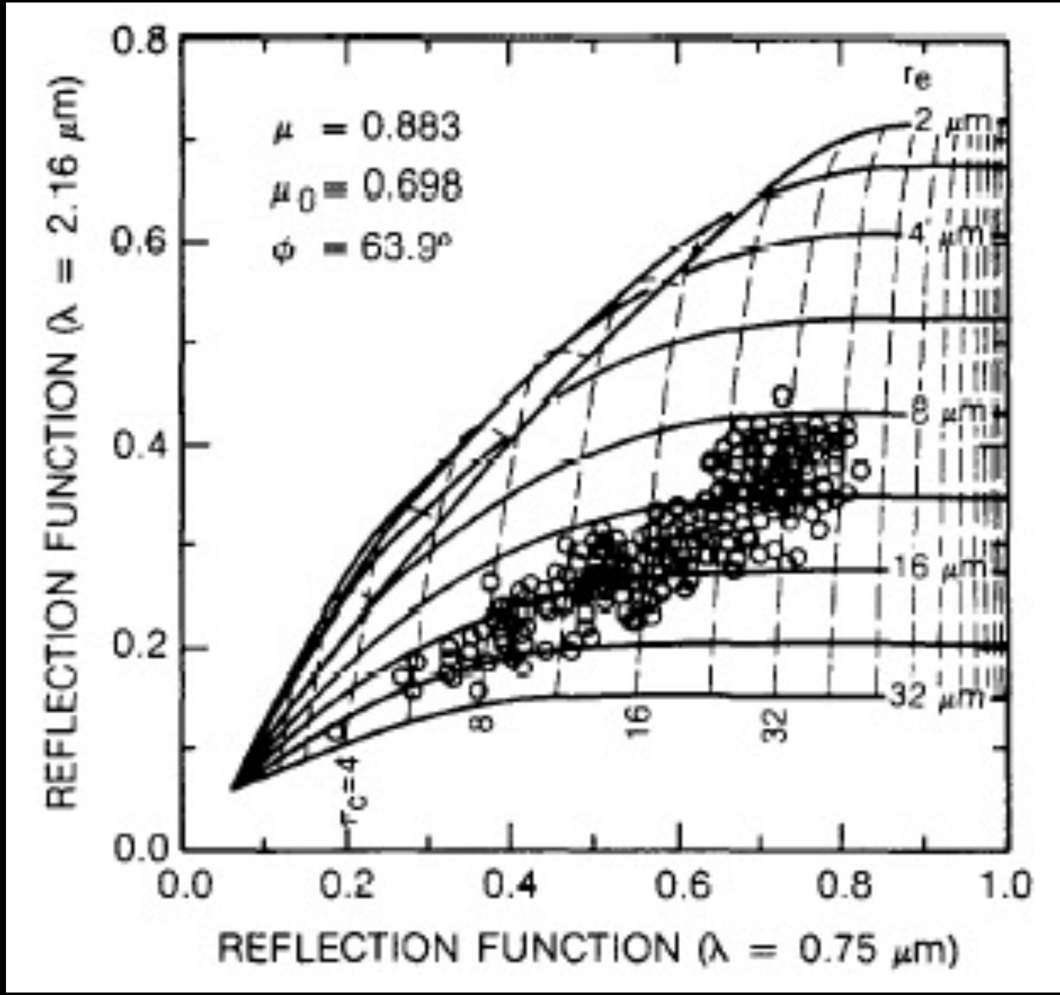


3. Remote Sensing

Particle scattering: In order to calculate the scattering properties of particles you need to know what their refractive indices are and how big they are, or make some good assumptions. Clouds are the easiest because they are made of solid or liquid water and liquid drops are spherical☺

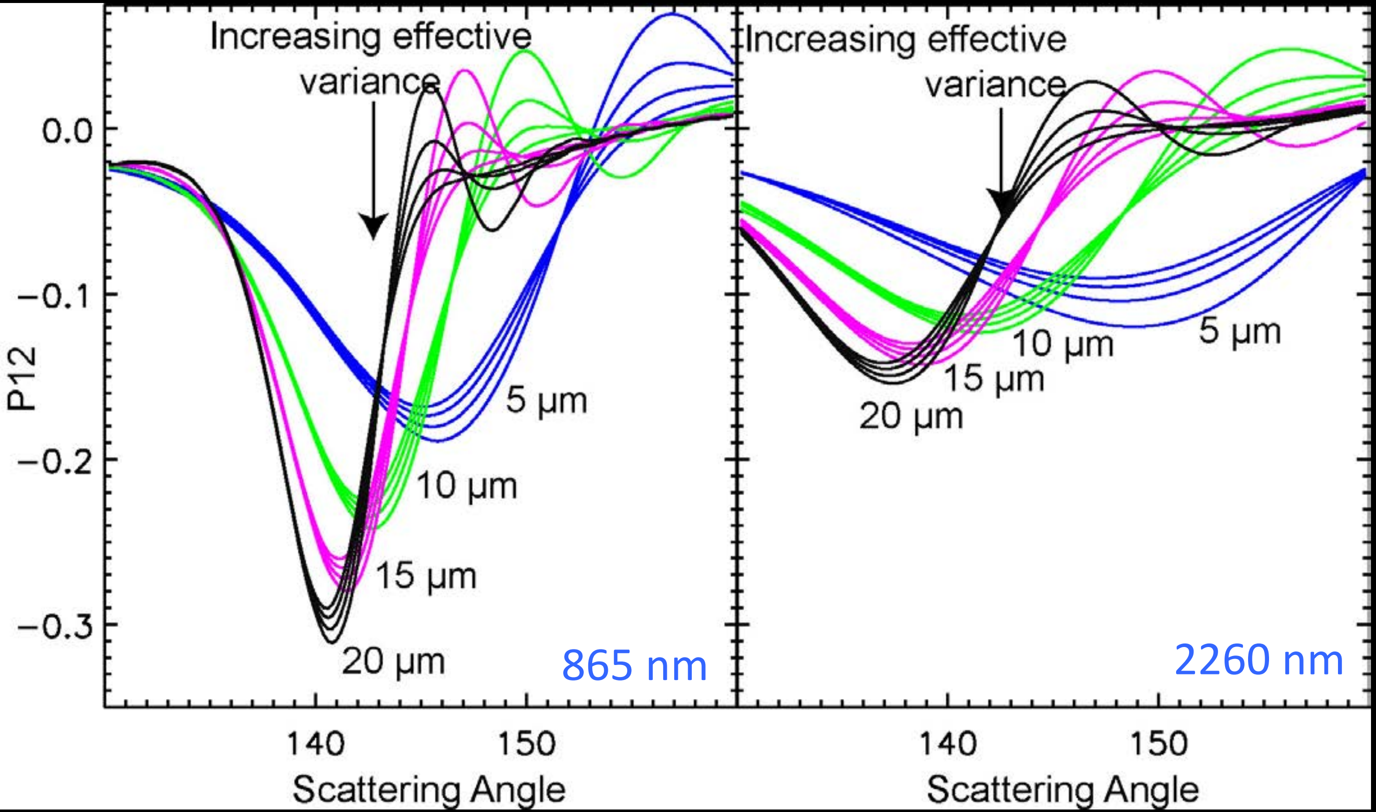


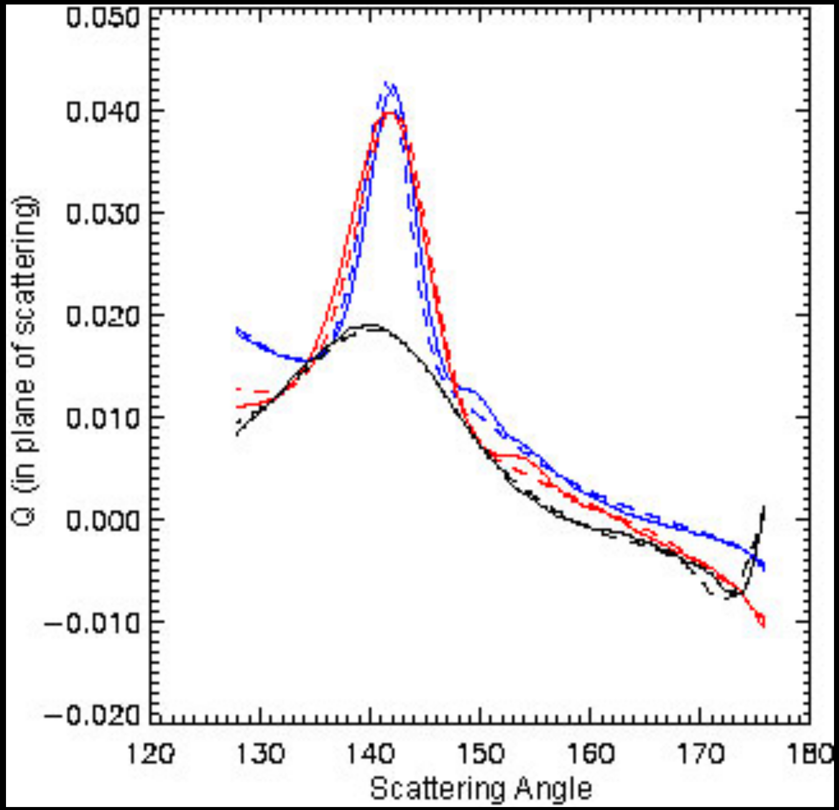
Imaginary indices of ice (solid line) and liquid water (dot-dashed line). For large weakly absorbing particles $C_{abs} \sim V n_i/\lambda$ and $C_{ext} \sim 2A$ so fraction absorbed $\sim V/A^*(n_i/\lambda)$



3. Remote Sensing

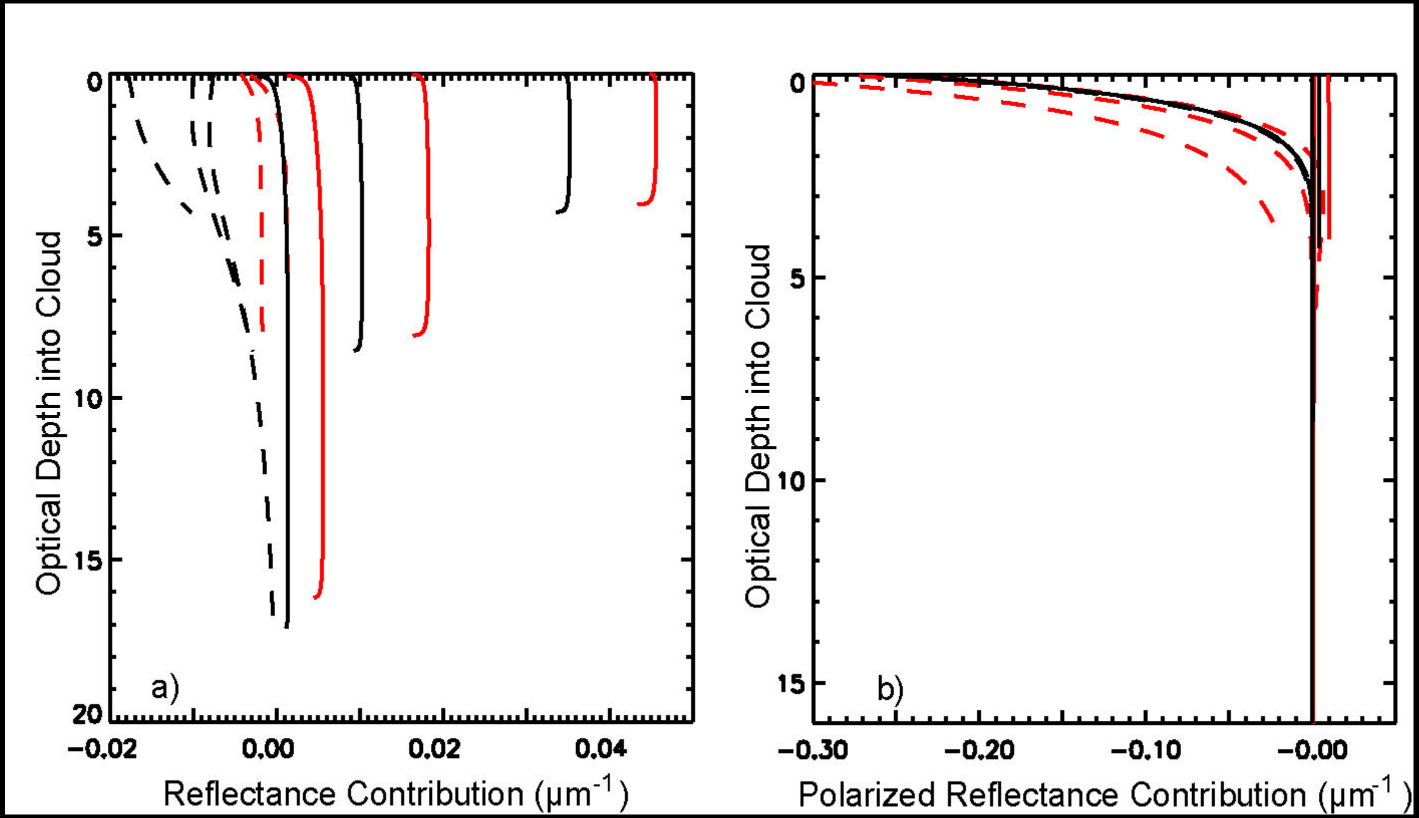
- **Particle scattering:** light scattered by spherical water drops also exhibits cloud bows that can be used to determine the droplet size distribution. Stokes parameter Q in the plane of scattering is dominated by the Mie phase matrix element P12 (single scattering) so retrievals insensitive to heterogeneity, or 3D effects. No restrictions on cloud optical thickness. If there is a cloud bow, the droplet size can be determined.

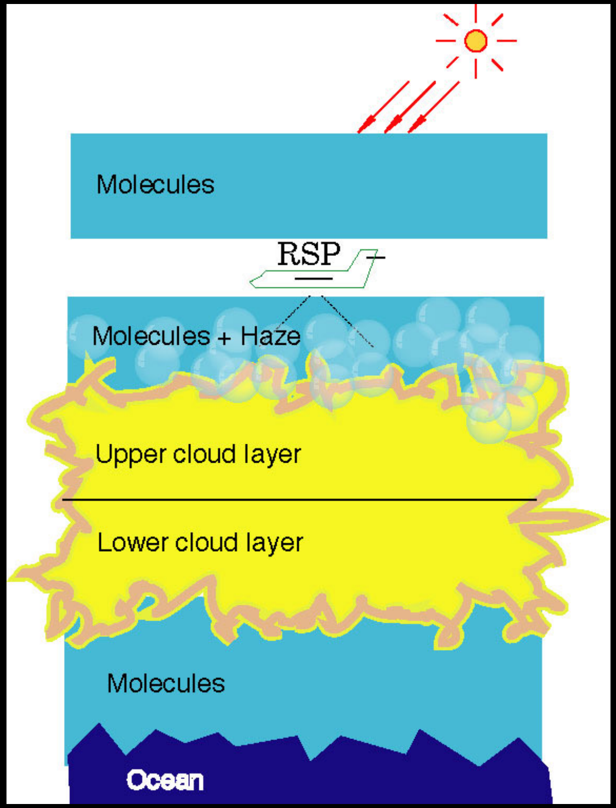




3. Remote Sensing

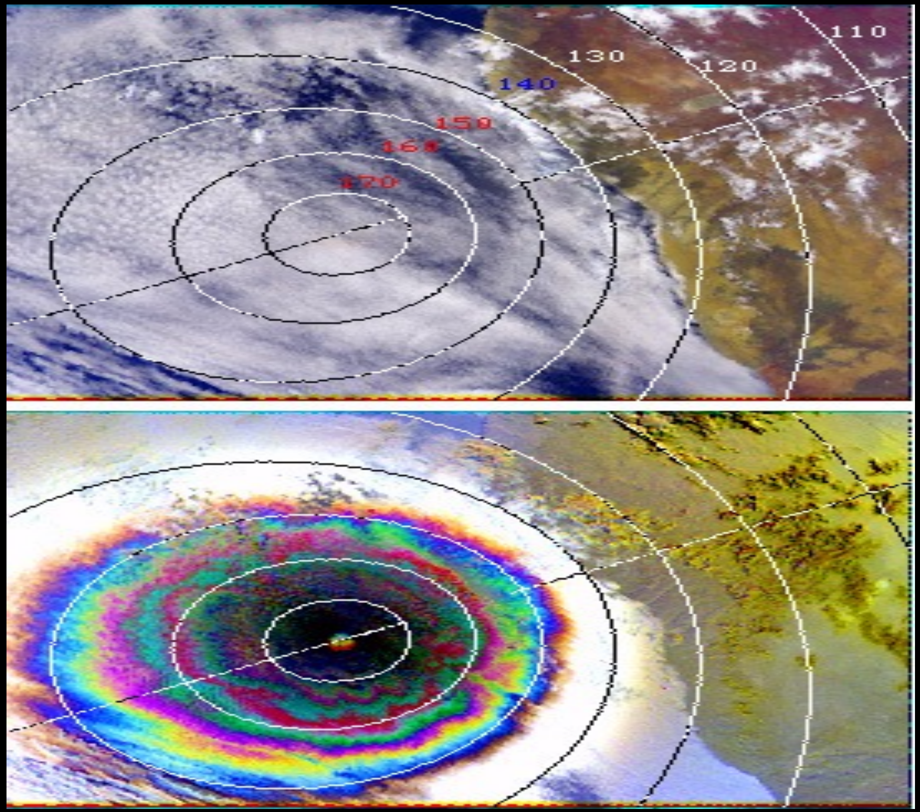
- **Particle scattering:** important to remember that the parts of a cloud that contribute to the signals in the total and polarized reflectance are generally different, so a more sophisticated model of a could may be needed to interpret the observations.



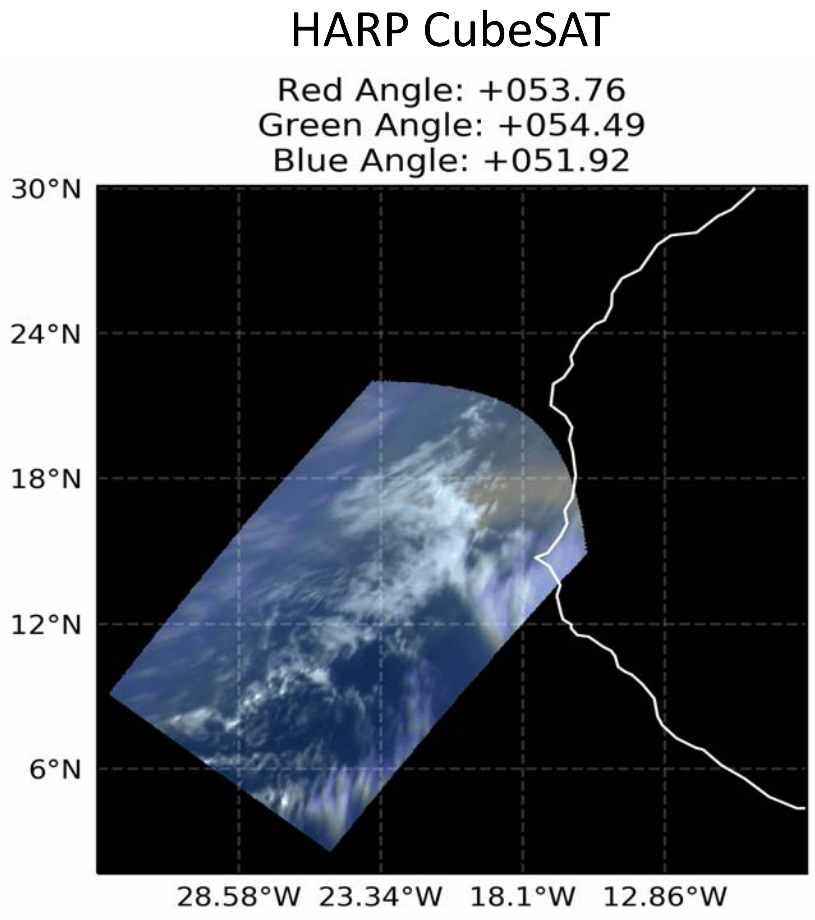


3. Remote Sensing

- **Particle scattering:** to observe the cloud bow you do of course need multi-angle observations. This capability is provided by HARP2 on PACE – more in L9 with Lorraine!



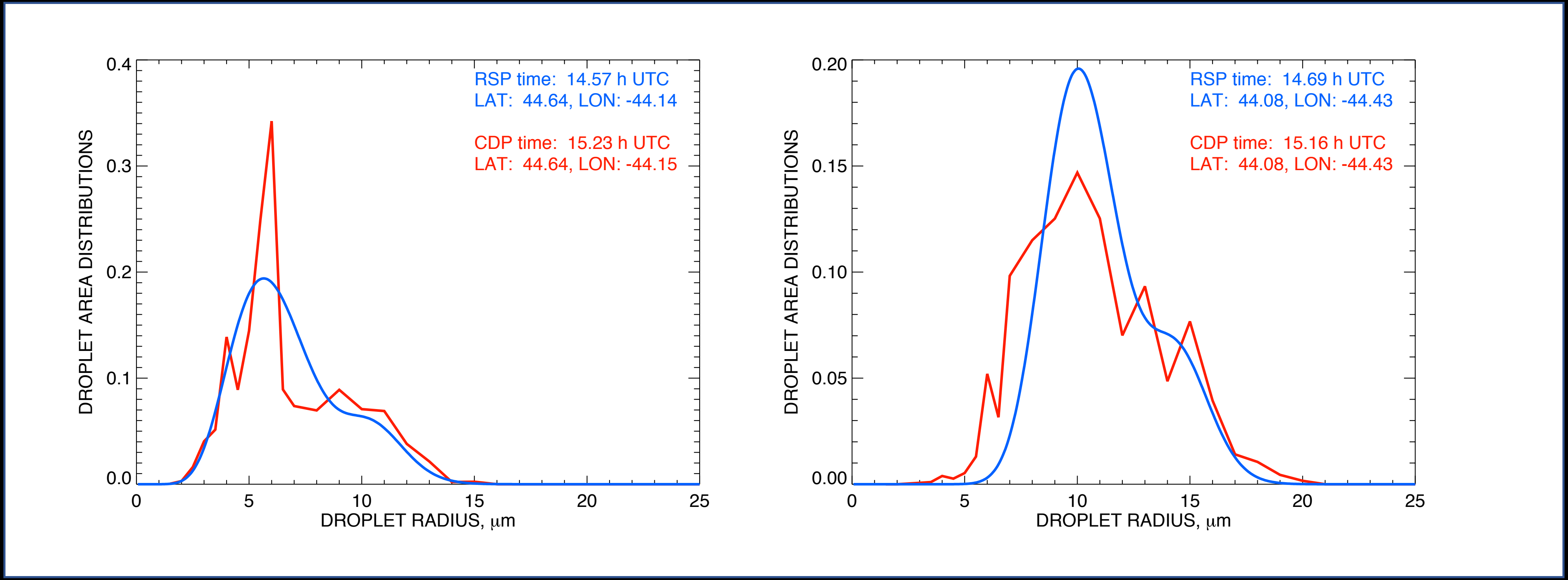
Could be done for cloud fields with POLDER, image courtesy of F. M. Bréon, LSCE, France



Courtesy of Vanderlei Martins, UMBC: https://esi.umbc.edu/files/2020/07/HARP-Dust-Story-3-png.ppsx

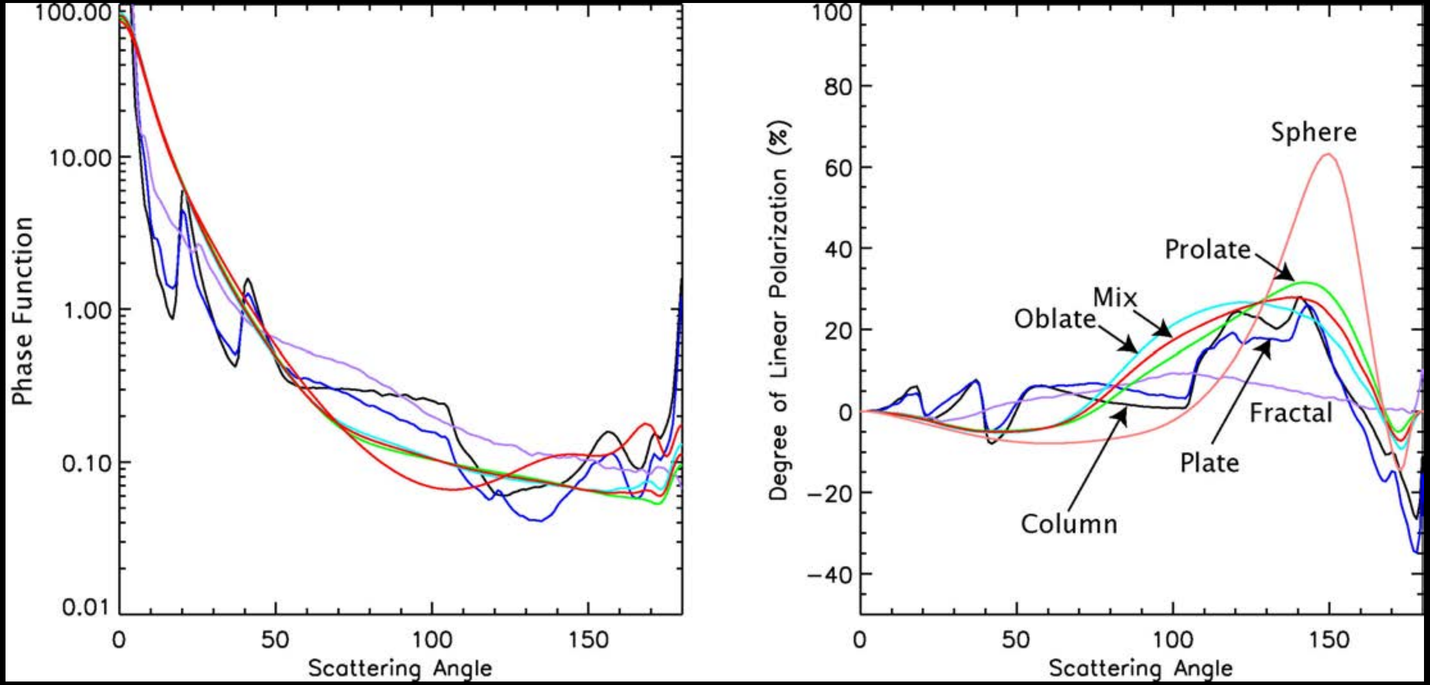
3. Remote Sensing

- **Particle scattering:** and when you make the multi-angle cloudbow measurements you can get the droplet size distribution. Allows you to see the development of drizzle near cloud top (autoconversion) that will cause the cloud to rain.



3. Remote Sensing

- **Particle scattering:** not all particles are spheres and shape affects the distribution of scattered radiation. Certainly an issue for dust aerosols and ice clouds and to some extent for sea salt too, although only at low relative humidities.



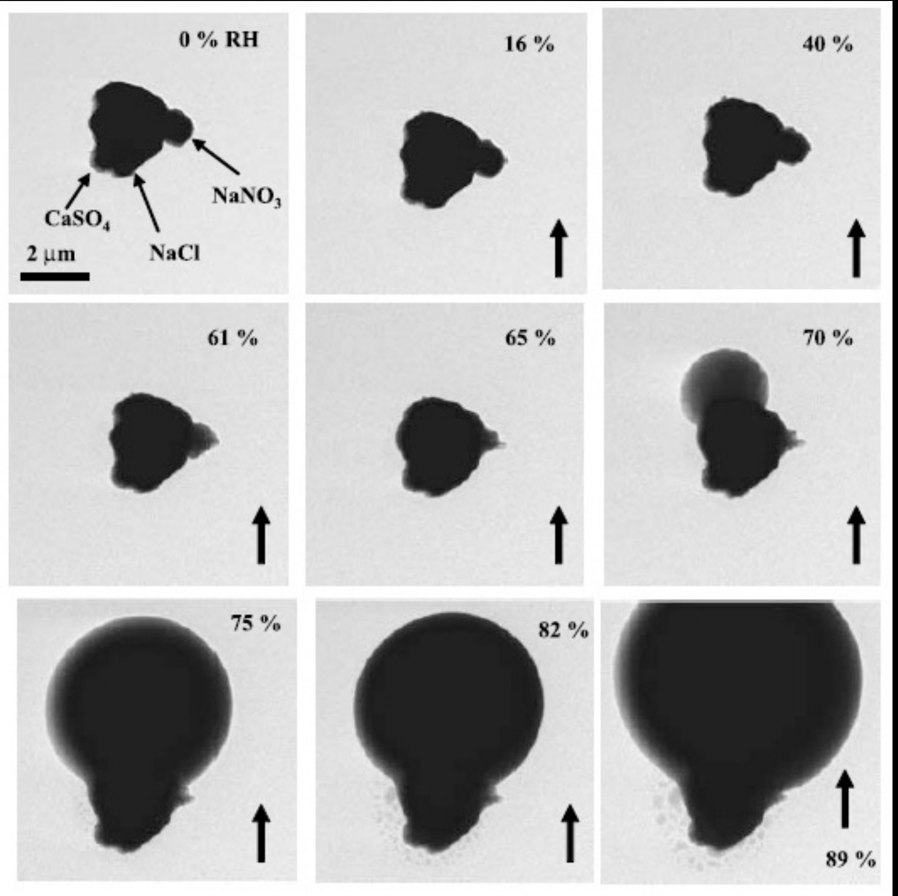
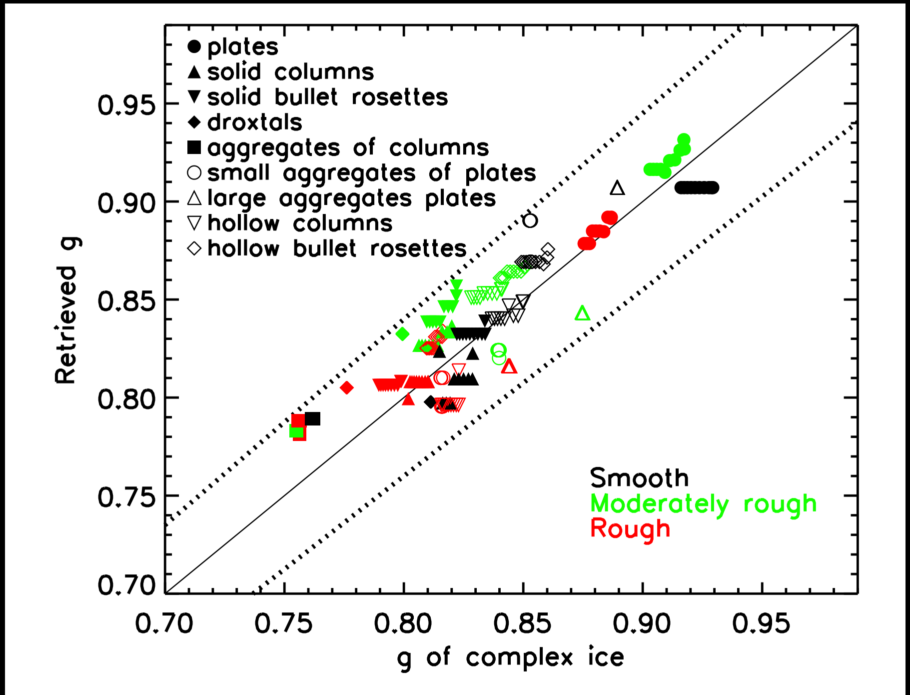


Figure 8. Images of an ~3.6 μm NaCl particle with an attached $NaNO_3$ crystal as RH is increased from 0 to 89%.

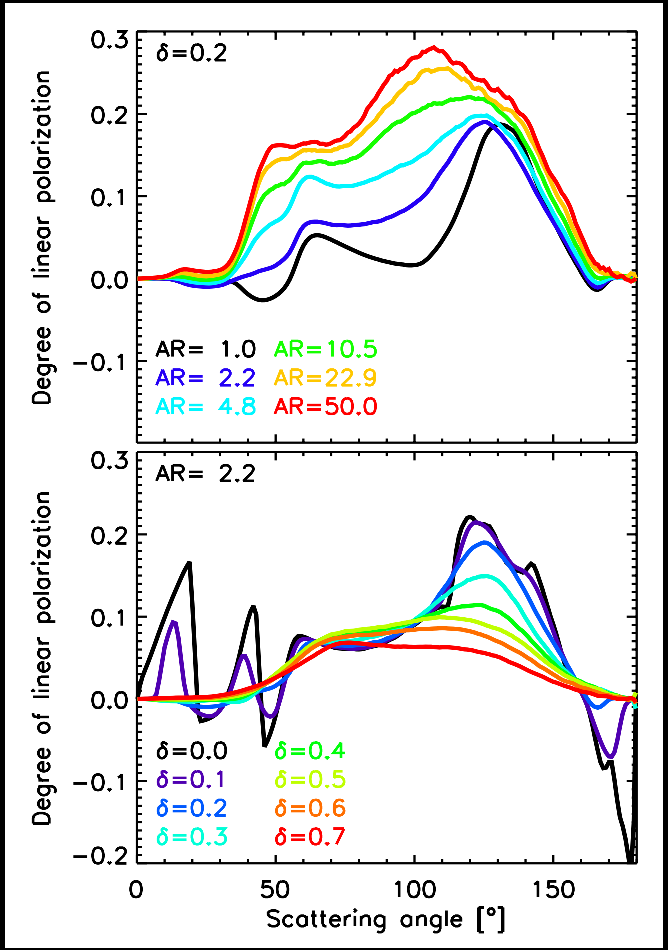
Spheroids are a commonly used model for scattering by dust that seems to work quite well for passive remote sensing: Dubovik et al. (2006), J. Geophys. Res., 111, D11208

3. Remote Sensing

- **Particle scattering:** and ice has its own bestiary of dendrites, bullet rosettes, aggregates, columns, plates. But for remote sensing it's preferable to have properties that we can retrieve that have a continuous dependence on a given variable(s) rather than a categorical identification – especially when it's difficult to enumerate all the categories.

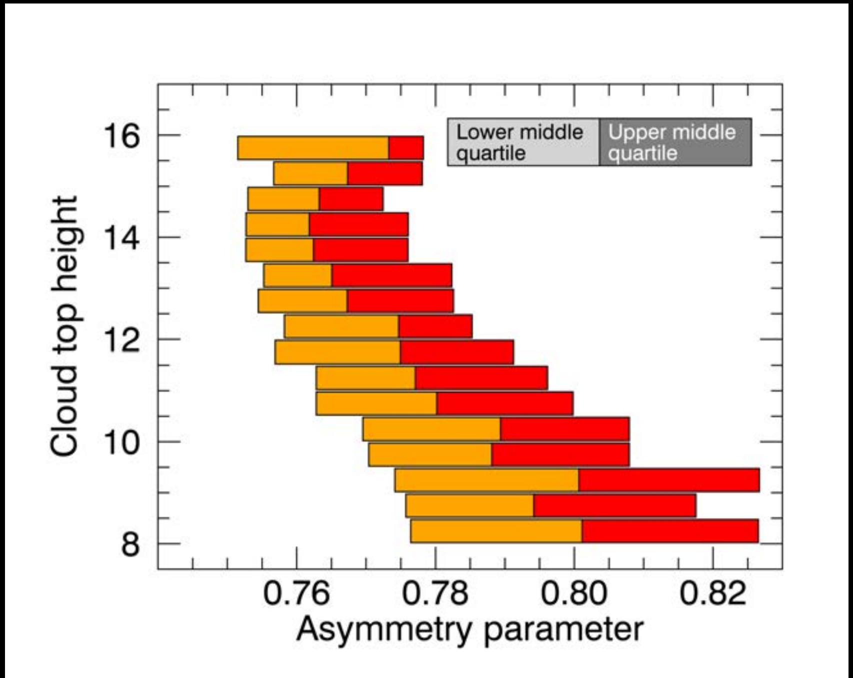


Roughened hexagonal plates/columns with a range of aspect ratios and roughnesses can serve as a good proxy for complex ice.



3. Remote Sensing

- **Particle scattering:** use HARP2 to determine aspect ratio and roughness and then correct OCI ice cloud optical depth. This is easily done using the asymmetry parameter for the retrieved ice crystal habit and the assumed asymmetry parameter used in the OCI retrieval since the reflectance actually depends on the reduced optical depth ($\tau(1-g)$). $\tau_{corr}=\tau_{OCI}*(1-g_{OCI})/(1-g_{corr})$



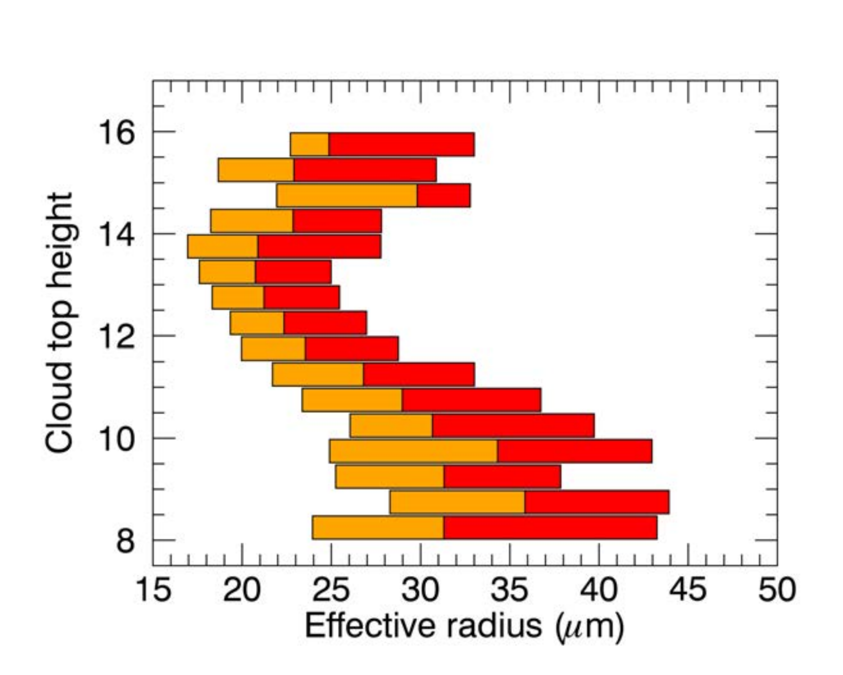
-75°C

-58°C

-52°C

-37°C

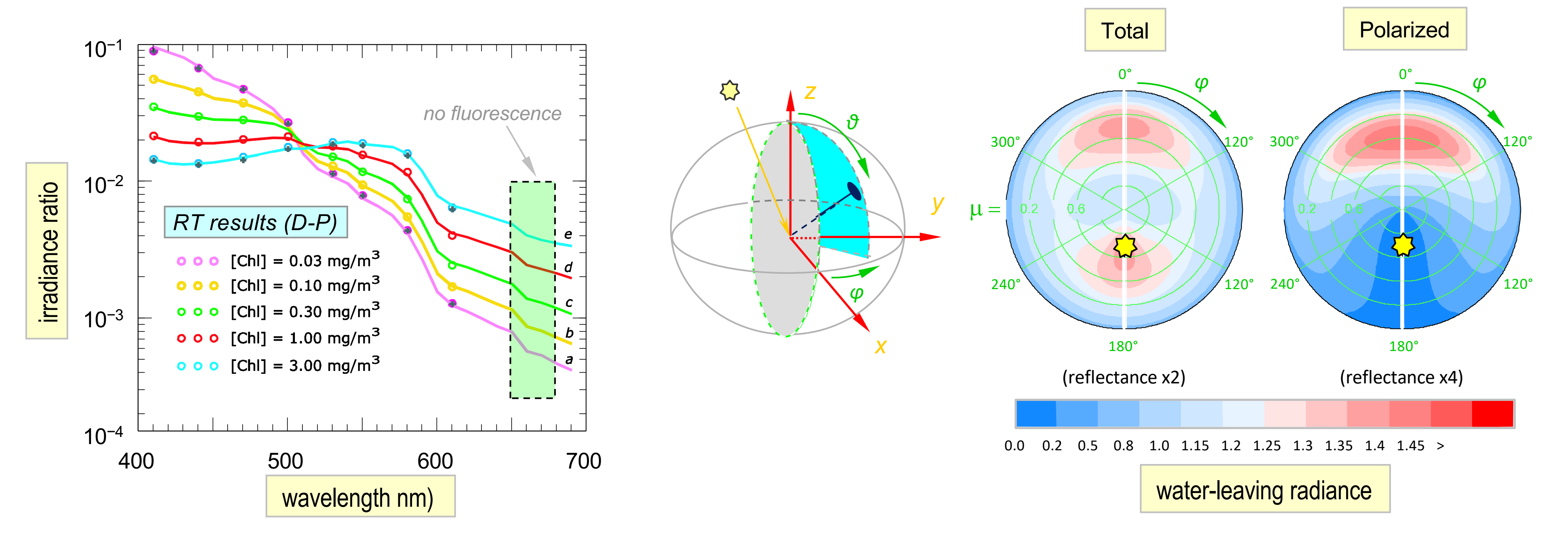
-20°C



Using a single ice crystal habit would create a vertically varying bias in the optical depth and ice water path estimate.

3. Remote Sensing

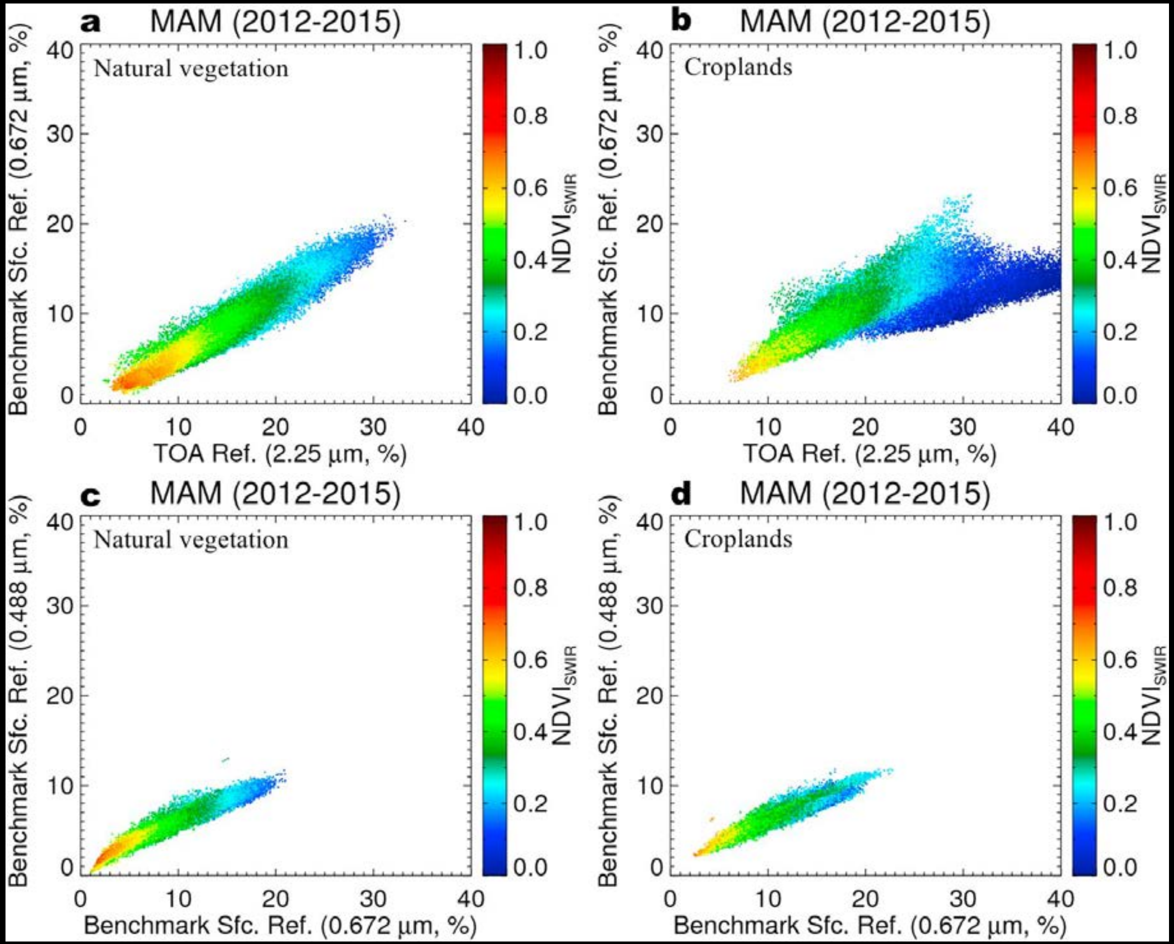
- **Surface properties (ocean):** if you are going to use total and polarized reflectances together for aerosol remote sensing then you do need a model for the surface. There are now a range of models for ocean body scattering that are available (e.g. Pengwang Zhai from this class, and Jacek.Chowdhary@nasa.gov). Models are generally parameterized on [Chl] (with more recent extensions to include independently varying [CDOM], spectral dependence of CDOM absorption and backscatter) and are constrained to match available polarization and irradiance ratio observations.



Chowdhary *et al.* (*AO*, 2006; *RSE* 2011)

3. Remote Sensing

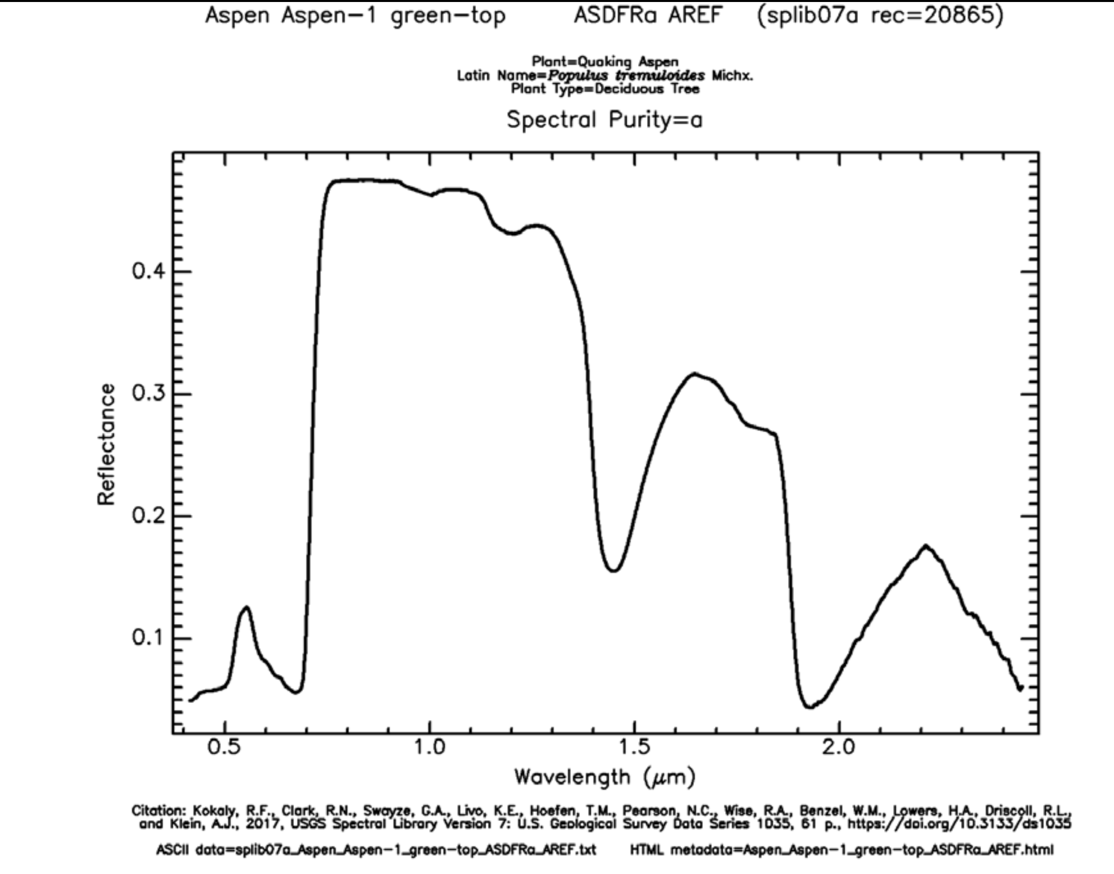
- **Surface properties (land):** if you are going to use total and polarized reflectances together for aerosol remote sensing then you do need a model for the surface. For vegetated surfaces regression relations between longer wavelength (2.13/2.25 μm) bands and shorter wavelength bands are used.

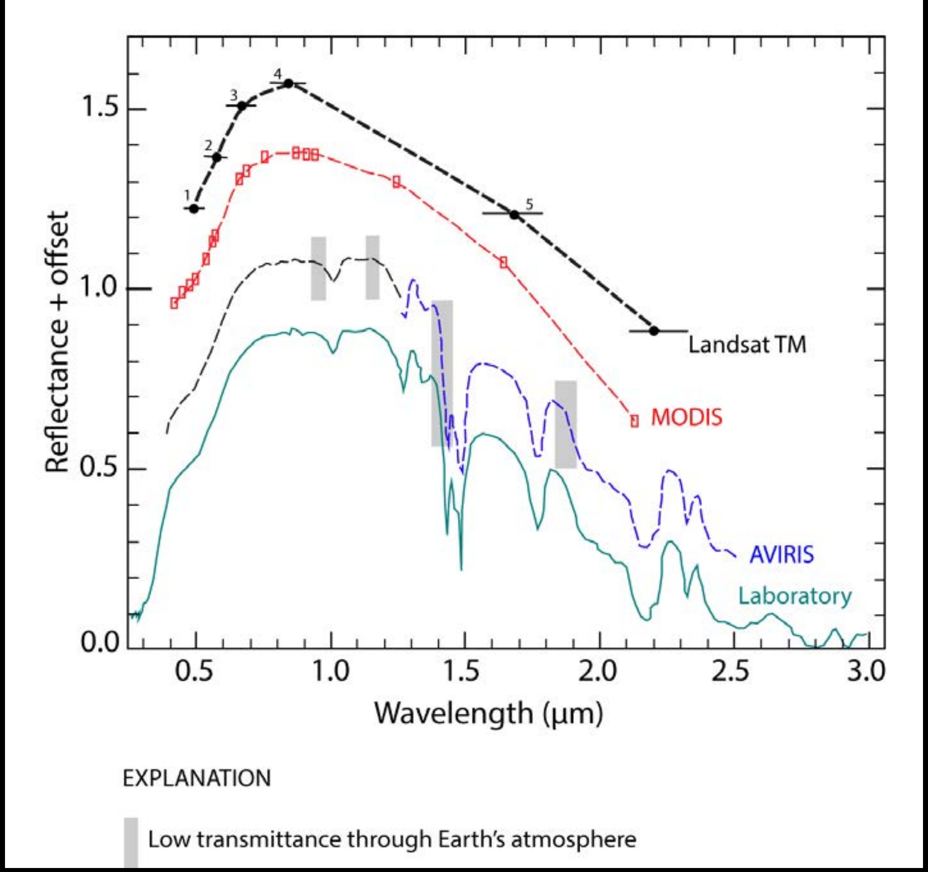


Kaufman *et al.* https://doi.org/ 10.1109/36.628795, Hsu et al. https://doi.org/10.1029/2018JD029688

3. Remote Sensing

- **Surface properties (land):** and if someone atmospherically corrects the observations to give a surface reflectance (e.g. Alexei Lyapusting with the MAIAC algorithm https://doi.org/10.5194/amt-11-5741-2018) there are spectral libraries that are available to help identify the materials in the scene you are observing (e.g. https://speclib.jpl.nasa.gov/ and https://pubs.er.usgs.gov/publication/ds1035)

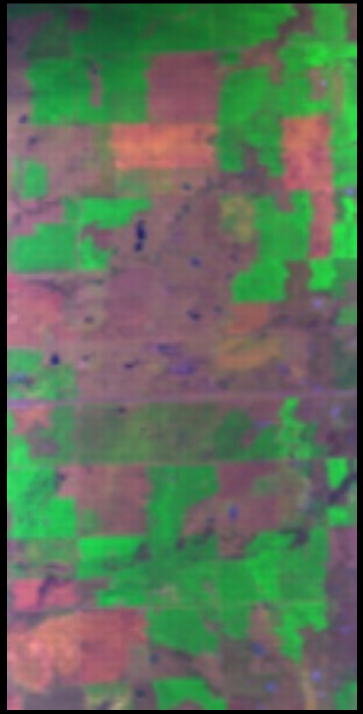


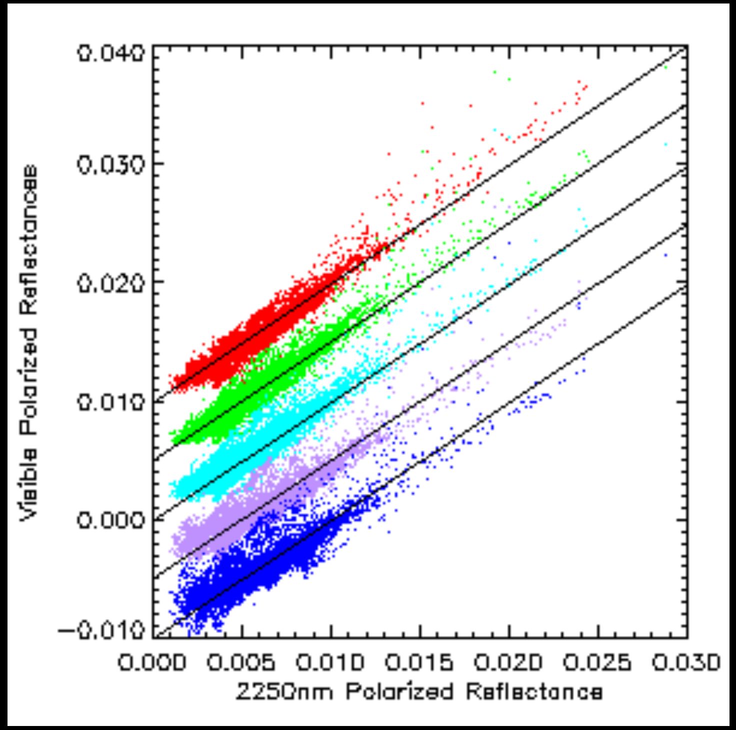


3. Remote Sensing

- **Surface properties (land):** if you are going to use total and polarized reflectances together for aerosol remote sensing then you do need a model for the surface. For most natural surface researchers have found that the surface polarized reflectance is grey. This is ascribed to the principal source of polarization by surfaces being front facet (Fresnel) reflection. his has led to surface polarized reflectance being modeled using equations of the form $R_p=A(\mathbf{s}_{\mathrm{view}},\mathbf{s}_{\mathrm{illum}})F_p(\Omega)$ where F_p is the polarized reflectance generated by an isotropically distributed collection of Fresnel reflectors, the function A accounts for fractional coverage, shadowing etc. and $\mathbf{s}_{\mathrm{view}}$, $\mathbf{s}_{\mathrm{illum}}$ are unit vectors defining the viewing and illumination geometries respectively.

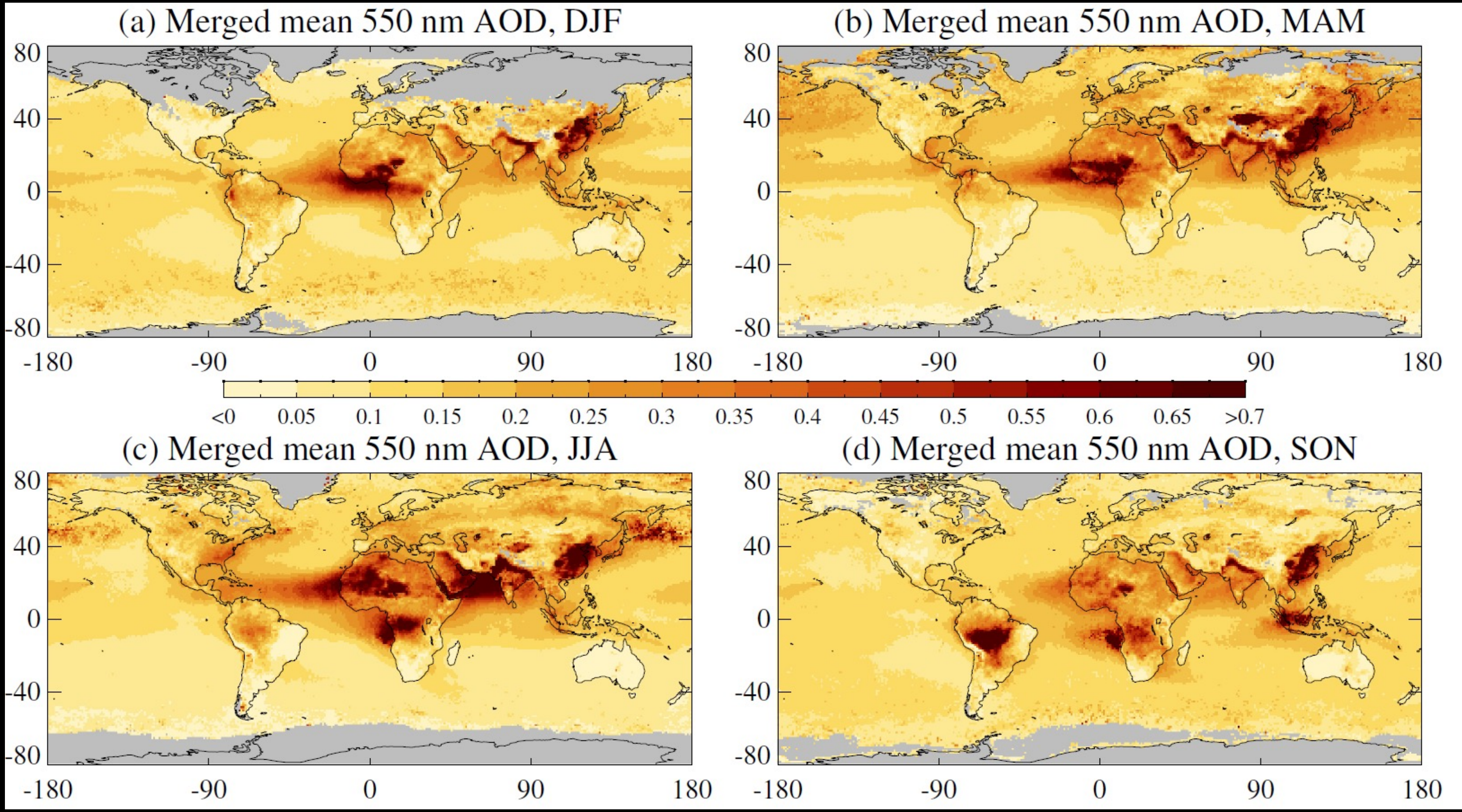
Waquet *et al.* doi:10.1364/AO.48.001228,
Litvinov et al. doi:10.1016/j.jqsrt.2012.06.027,
Maignan et al. doi: 10.1016/j.rse.2009.07.022





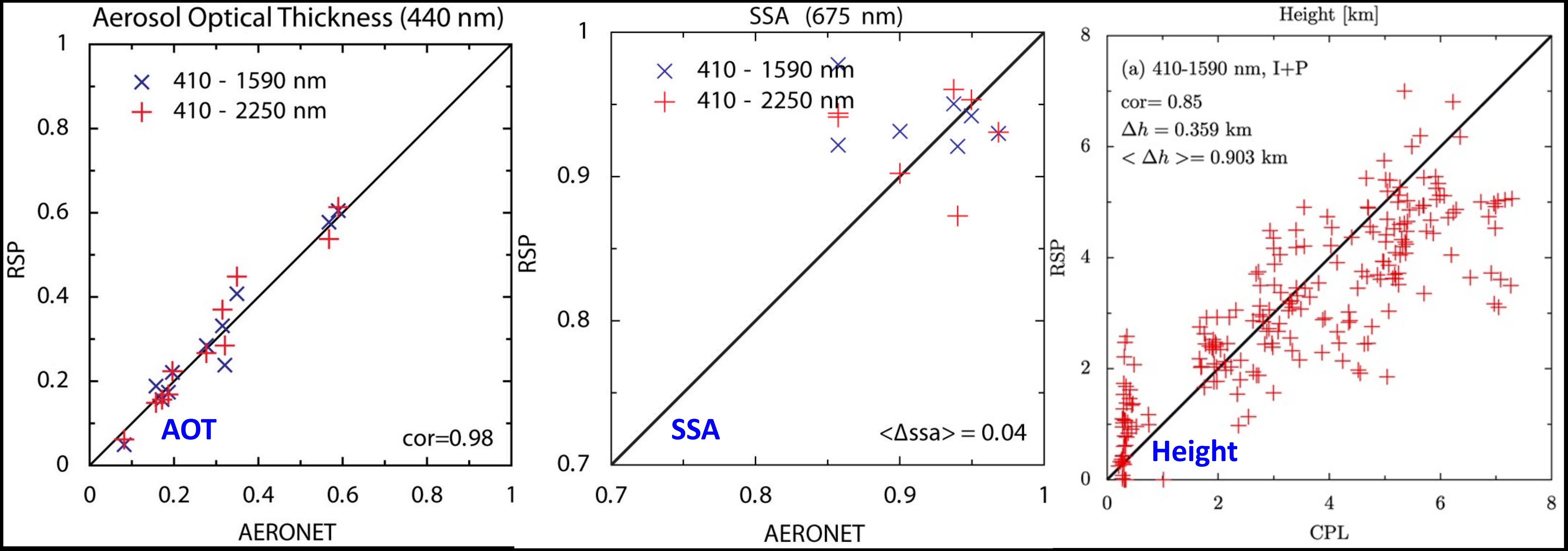
3. Remote Sensing

- **Aerosols:** the surface models and scattering models of small particles when incorporated into radiative transfer models can be used to construct aerosol retrieval algorithms. Total column integrated extinction (Aerosol Optical Depth) and fine mode fraction are the primary products from current broad swath imagers (more in L10 with the aerosol experts Lorraine Remer and Andrew Sayer).



3. Remote Sensing

- **Aerosols:** multi-angle polarimeter retrieval schemes extend the set of retrieved aerosol parameters to single scattering albedo, layer height, size distribution, real refractive index and the parameters of a surface model e.g. [Chl] in the case of open oceans.



Wu, L. et al. 2016: doi:10.1002/2016GL069848.
Wu, L. et al., 2015:doi:10.5194/amt-8-2625-2015.

Take Home Messages

- **Remote sensing uses observed radiances to make inferences about the scatterers and absorbers that are present in the sytem**
- **In order to make such inferences you need to have good models of how particles (and gases) scatter and absorb**
- **You also need good measurements of the spectral variation of radiance (e.g. OCI), and, ideally for aerosols, the angular variation of radiance and its polarization state (e.g. HARP2 and SPEXone)**
- **If you have such models and measurements you can infer particle shape (e.g. aspect ratio of ice, detection of nonspherical dust), size and composition and the amount of stuff that is there (optical depths, concentrations). You can also say something about the physical depth of what is there (e.g. cloud top, aerosol layer depth)**