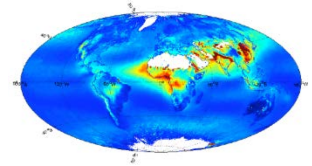
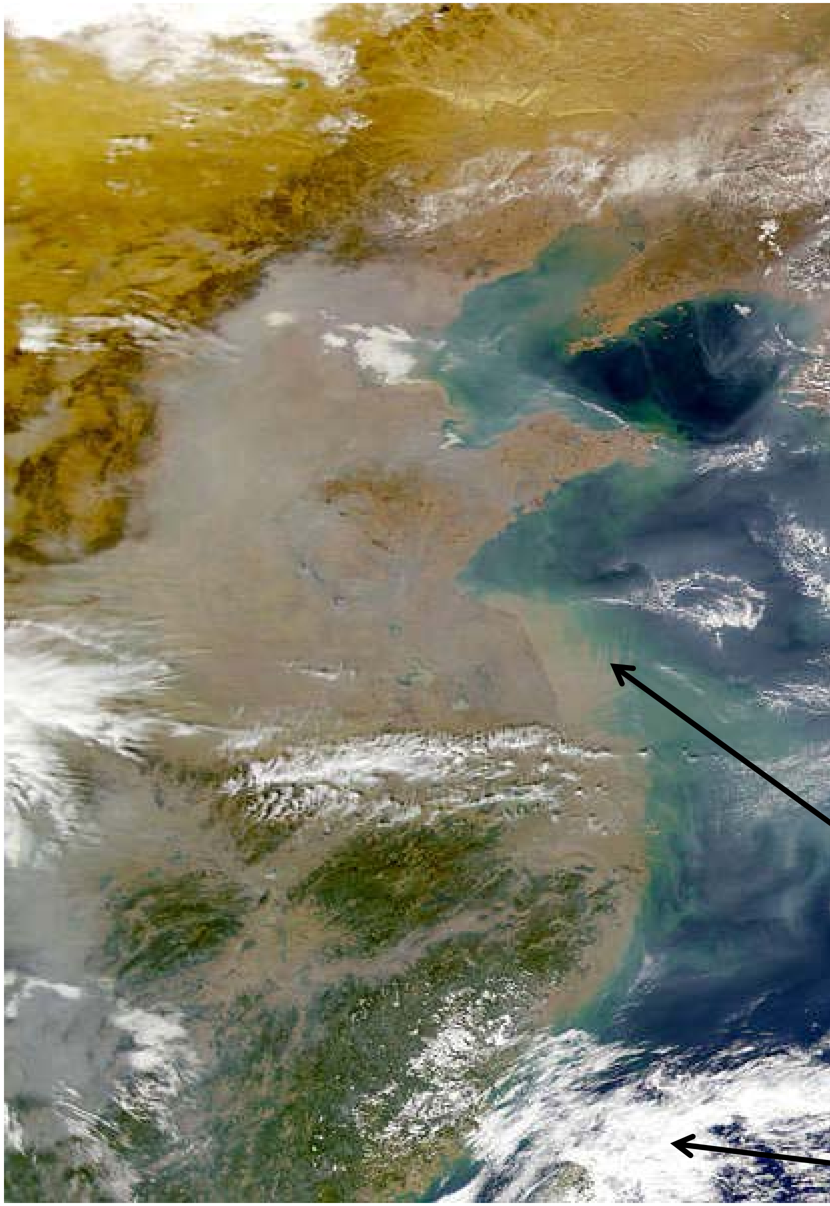


# SECOND YEAR: 2604
# REMOTE SENSING

## *PASSIVE SOUNDING OF AEROSOLS AND CLOUDS*

**Prof. HARTMUT BOESCH**

**Earth Observation Science Group, Dept. of Physics and Astronomy, University of Leicester**



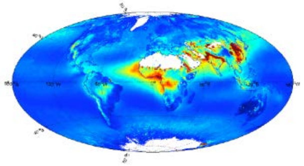
SeaWiFS image of eastern China

Beijing has completely disappeared under the (pollution) haze.

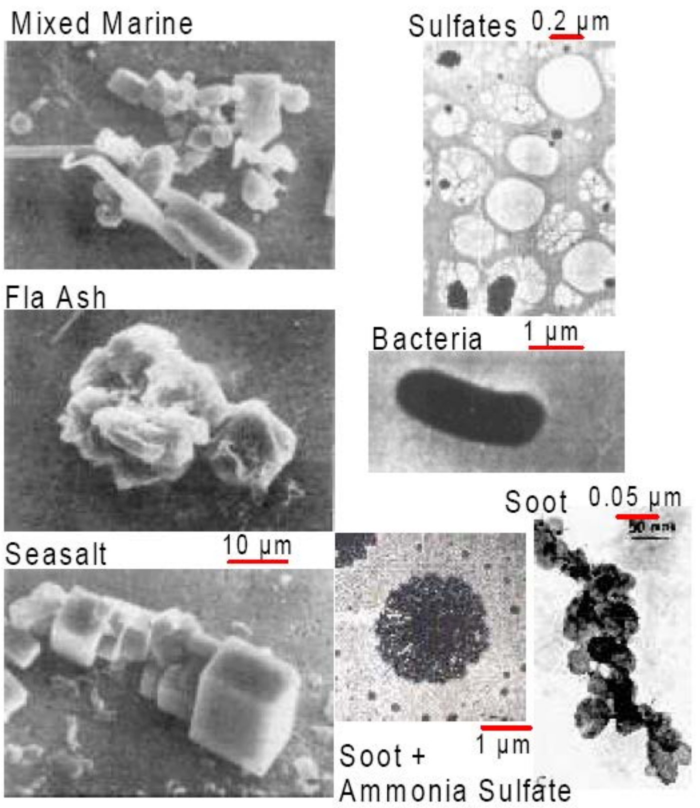
**Aerosol**

**Clouds**

# WHAT ARE AEROSOLS ?



- Aerosols are small particles in atmosphere carried by the air
- Aerosols can be
  - Solid
  - Liquid
  - Mixed phase state
  - Arbitrary form (spherical ... non-spherical)
  - Internally and/or externally mixed
- Main Sources:
  - combustion
  - gas to particle conversion ($H_2SO_4$)
  - wind blown dust / sand
  - pollens
  - sea salt
- Atmospheric Concentrations:
  - 10 .. $10^7$ $cm^{-3}$, decreasing with altitude
- Size:
  - $10^{-4}$ $\mu m$ .. $10^2$ $\mu m$




**Microphysical properties are largely variable !**

# AEROSOL SIZE DISTRIBUTION

- **Aerosols can vary dramatically depending on formation process, source and location/time**
  - **Size range varies by 5 orders of magnitude**
  - **Concentration of particles can vary 7 orders of magnitude**

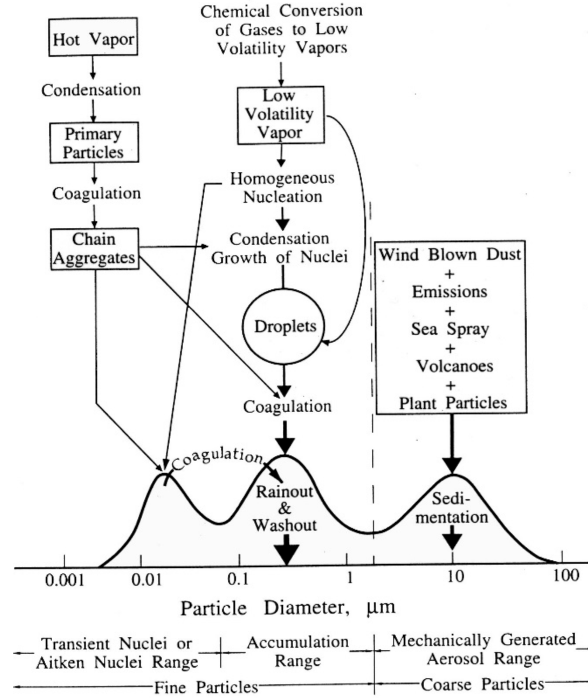
- **Size distributions is used to describe aerosol**

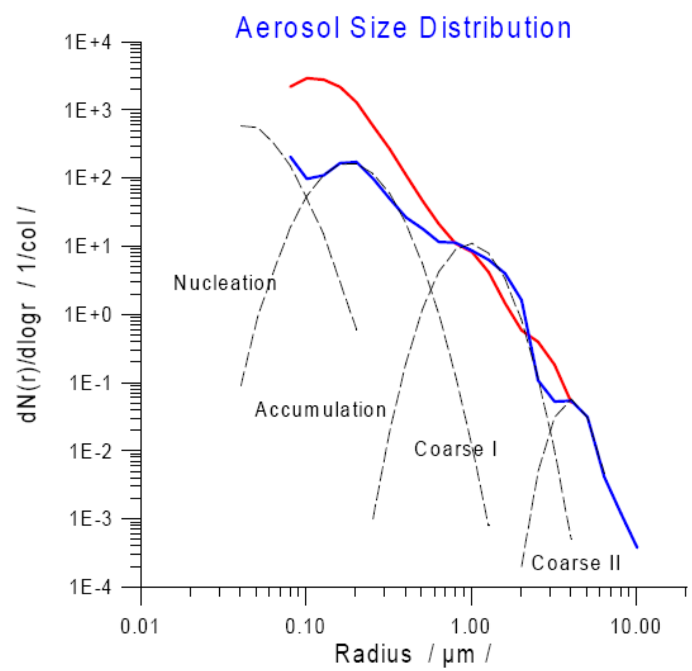
$$\frac{dN}{dr} = 1/r \frac{dN}{d(\log r)}$$

**Number Size Distribution (# particles per unit volume N per unit size, units: μm$^{-1}$ cm$^{-3}$ )**

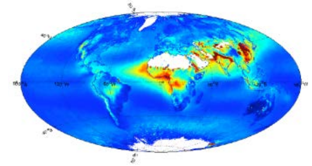
- **Volume size distribution**

$$\frac{dV}{dr} = \frac{dV}{dN}\frac{dN}{dr} = \frac{4}{3}\pi r^3 \frac{dN}{dr}$$



# AEROSOL SIZE DISTRIBUTION



- **Simple power law (Junge distribution)**

$$n_N(\log r) = \frac{dN}{d(\log r)} = C\, r^{-\beta}$$

  - **with *C* the concentration and size parameter *β* (*β* ~ 2 – 4)**
  - **describe particles in size range 0.1 -2 µm**

- **More general description given by Log-normal size distribution**

$$n_N(\log D_p) = \sum_{i=1}^{n} \frac{N_i}{(2\pi)^{1/2} \log \sigma_i} \exp\left( - \frac{\left(\log D_p - \log \overline{D}_{pi}\right)^2}{2 \log^2 \sigma_i} \right)$$

$N_i$: number concentration in mode i

$\sigma_i$: (geometric) standard deviation (unitless)

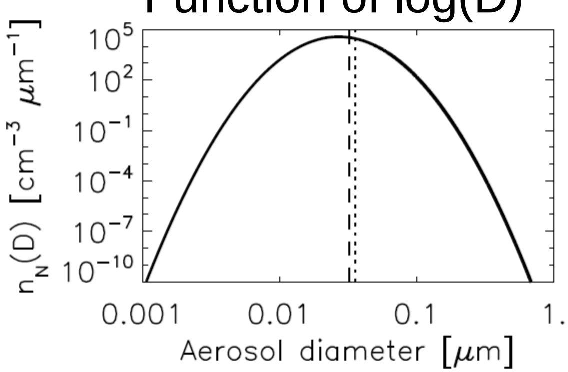
$D_{pi}$: mode (median) diameter

Log-Normal size Distribution

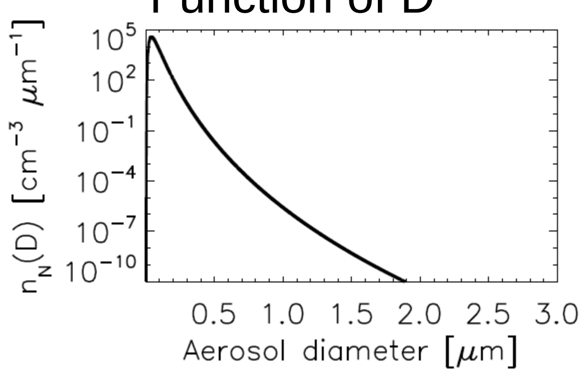
$$\overline{D_{pi}} = 0.05 \mu m$$

$$\sigma = 1.6$$

Function of log(D)




Function of D



# Example: Aerosol Distributions

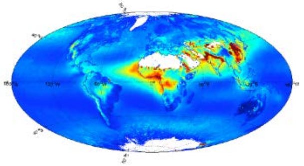


**TABLE 7.3 Parameters for Model Aerosol Distributions Expressed as the Sum of Three Log-Normal Modes**

| | Mode I | | | Mode II | | | Mode III | | |
|---|---|---|---|---|---|---|---|---|---|
| Type | $N$ ($cm^{-3}$) | $D_p$ ($\mu m$) | $\log\sigma$ | $N$ ($cm^{-3}$) | $D_p$ ($\mu m$) | $\log\sigma$ | $N$ ($cm^{-3}$) | $D_p$ ($\mu m$) | $\log\sigma$ |
| Urban | $9.93 \times 10^4$ | 0.013 | 0.245 | $1.11 \times 10^3$ | 0.014 | 0.666 | $3.64 \times 10^4$ | 0.05 | 0.337 |
| Marine | 133 | 0.008 | 0.657 | 66.6 | 0.266 | 0.210 | 3.1 | 0.58 | 0.396 |
| Rural | 6650 | 0.015 | 0.225 | 147 | 0.054 | 0.557 | 1990 | 0.084 | 0.266 |
| Remote continental | 3200 | 0.02 | 0.161 | 2900 | 0.116 | 0.217 | 0.3 | 1.8 | 0.380 |
| Free troposphere | 129 | 0.007 | 0.645 | 59.7 | 0.250 | 0.253 | 63.5 | 0.52 | 0.425 |
| Polar | 21.7 | 0.138 | 0.245 | 0.186 | 0.75 | 0.300 | $3 \times 10^{-4}$ | 8.6 | 0.291 |
| Desert | 726 | 0.002 | 0.247 | 114 | 0.038 | 0.770 | 0.178 | 21.6 | 0.438 |

*Source:* Jaenicke (1993).

**Urban Aerosols**

**Mode I**

**$D_N$ = 0.013 micron**

**$D_V = D_N \exp(3\log^2\sigma)$=0.013 micron*1.13=0.016 micron**

**Mode II**

**$D_N$=0.014micron**

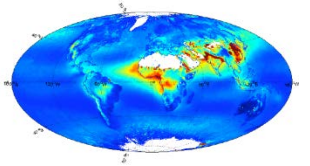
**$D_V$=0.05 micron**

**Mode III**

**$D_N$ =0.05 micron**

**$D_V$ ==0.07 micron**

# Recap: Interaction of Aerosols with EM Radiation



**The 'observable' interaction of light with aerosols is described by 3 parameters (macrophysical optical properties):**

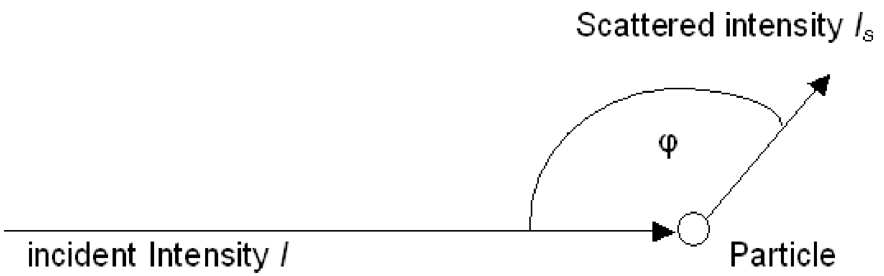
- **extinction coefficient $k_{ext}$: intensity loss by absorption and scattering through aerosols**

$$dI(\lambda) = -I_0(\lambda)\, k_{ext}(\lambda)\, ds \qquad [k_{ext} = 1/\mathrm{m}]$$

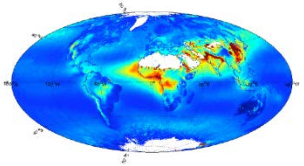
- **single scattering albedo ω: ratio of scattering coefficient and extinction coefficient (ω < 1 for absorbing aerosols):**

$$\omega = \frac{k_{sca}}{k_{ext}}$$

- **phase function p: directional dependence of scattered intensity**



# Summary: Optical Properties of Aerosols



- **For atmospheric aerosol mixture, the optical properties are integrated over all particles:**

**extinction coefficient**

$$k_{ext}(\lambda) = \int_0^\infty \pi r^2 Q_{ext}(\lambda, r, m) n(r)\, dr \qquad k_{sca}(\lambda) = \int_0^\infty \pi r^2 Q_{sca}(\lambda, r, m) n(r)\, dr$$

**single scattering albedo**

$$\omega(\lambda) = \frac{k_{sca}(\lambda)}{k_{ext}(\lambda)}$$

**phase function:**

$$P(\theta, \lambda) = \int_0^\infty \pi r^2 F(\theta, \lambda, r, m) n(r)\, dr$$

- **All quantities depend on wavelength and physical-chemical properties:**
  - **refractive index m (real and imaginary part)**
  - **Particle size distribution dN/dlog(r)**
  - **Particle shape**
- **For spherical particles: efficiency factors per particle $Q_{ext}$, $Q_{sca}$ and F can be calculated from electrodynamic theory using Maxwell equations (Mie theory)**

# AOD AND ANGSTROM

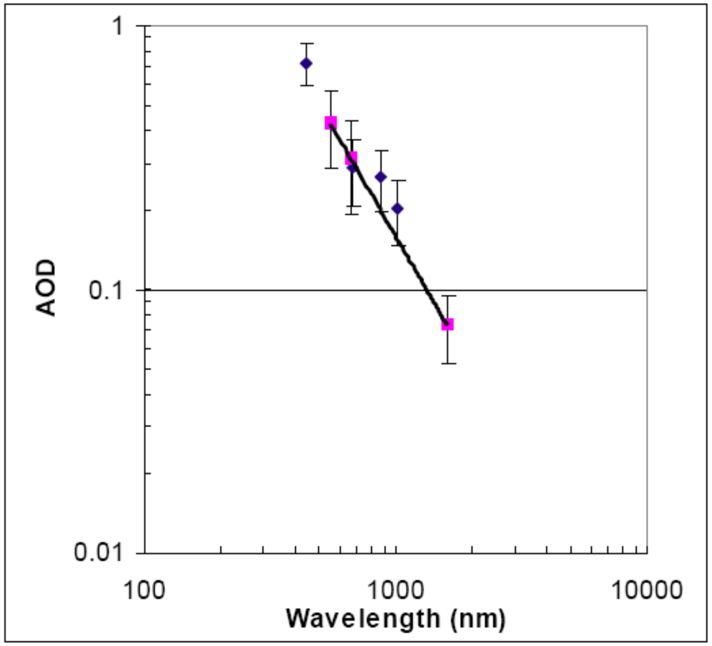
- The integral along vertical height is called **aerosol optical depth** AOD

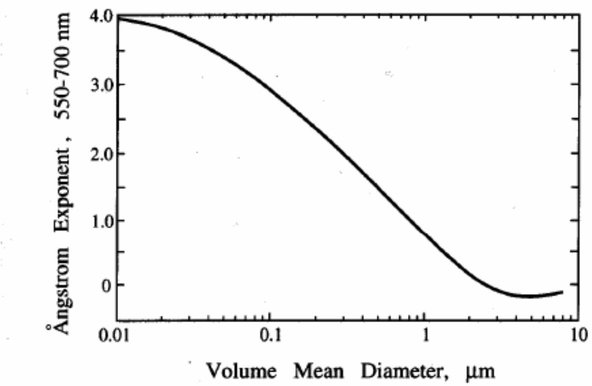
$$AOD = \int_0^{TOA} k_{ext}(z)dz$$

- Wavelength dependence of AOD is defined by aerosol size distribution - **Angstrom relation:**

$$AOD(\lambda) = const \times \lambda^{-\alpha}$$

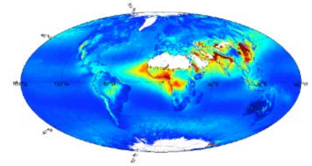
**Angstrom coefficient α**




RE 22.9 Ångstrom exponent for a log-normally distributed water aerosol ($\sigma_g = 2.0$) with tive index $m = 1.33 - 0i$ in the wavelength range $\lambda = 550$ to 700 nm (courtesy of J. A. ).

# Aerosol Extinction



$$k_{ext}(\lambda) = \int_0^\infty \pi r^2 Q_{ext}(\lambda, r, m) n(r)\, dr$$ **with** $$n(r) = const \times r^{-(\gamma+1)}$$ **(Power law)**

**Thus** $$k_{ext}(\lambda) = const \int_0^\infty Q_{ext}(\lambda, r, m) r^{1-\gamma}\, dr$$

**Change of integration variable to** $x = 2\pi r / \lambda$

$$k_{ext}(\lambda) = const \left(\frac{\lambda}{2\pi}\right)^{2-\gamma} \int_0^\infty Q_{ext}(\lambda, r, m) x^{1-\gamma}\, dx$$ **or** $$k_{ext}(\lambda) \propto \left(\frac{\lambda}{2\pi}\right)^{2-\gamma}$$

**Assume $k_{ext}$ depends only on distribution of aerosol number density**

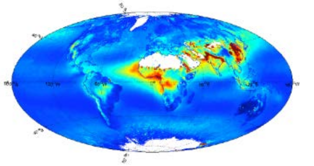
$$AOD(\lambda) = \int k_{ext}(z) dz \propto H \left(\frac{\lambda}{2\pi}\right)^{2-\gamma}$$ **with** $$H = \int h(z) dz = \int \frac{n(z)}{\int n(z) dz} dz$$

**Compare Angstrom relation** $AOD(\lambda) \propto \lambda^{-\alpha}$ **thus** $\alpha = \gamma - 2$

**Wavelength dependence of AOD is determined by size distribution !**

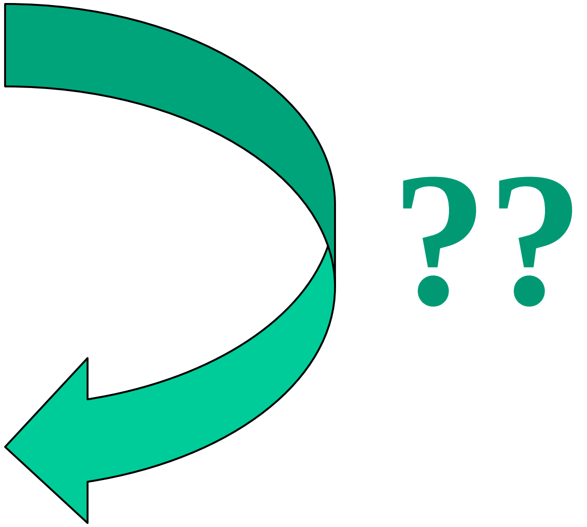
- **$\alpha \rightarrow 4$ small particle ((k$_{ext}$ -> $\lambda^{-4}$, Rayleigh)**
- **$\alpha \rightarrow$ ~1.5 typical value for aerosols**
- **$\alpha \rightarrow 0$ large particles (k$_{ext}$ -> const, extinction paradox)**

# REMOTE SENSING OF AEROSOLS



**Observable quantities**

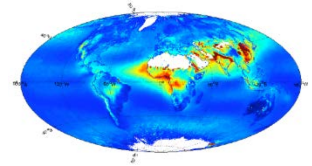
- **Spectral extinction along the light path**
- **Radiation as function of viewing angle**
- **(Degree of polarisation)**

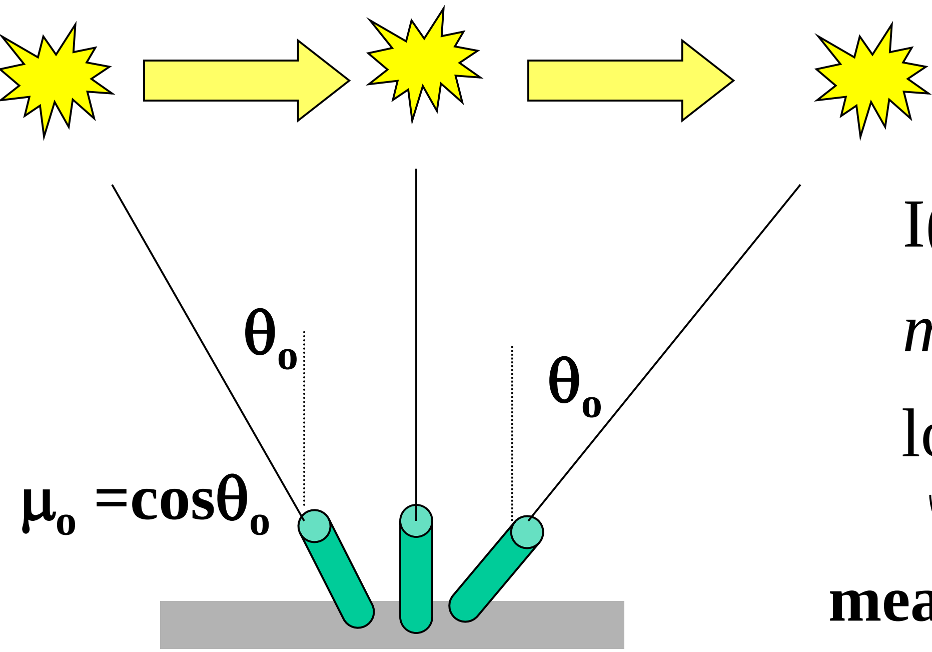



**What do we want?**

- **AOD**
- **Angstrom coefficient**
- **Size (distribution)**
- **Single scattering albedo**
- **Aerosol type (or composition)**
- **Vertical distribution (very limited information in passive observations)**

# Determination of the aerosol optical depth from the ground





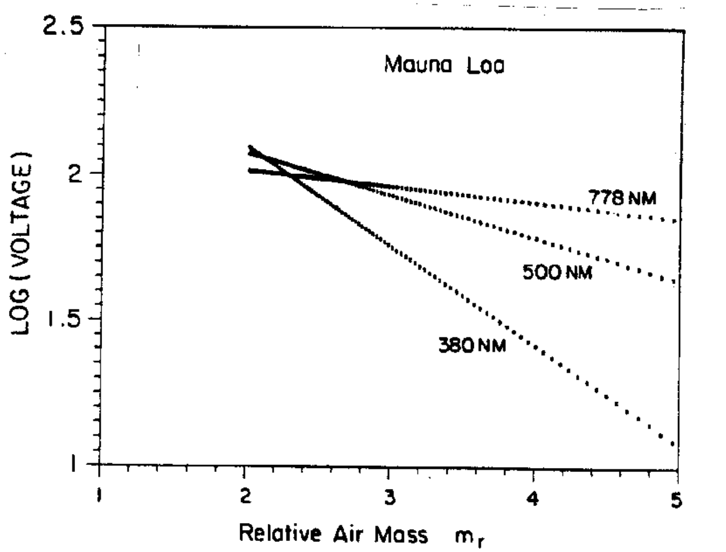

**Lambert Beer Law**

$$\mathrm{I}(\lambda) = I_0(\lambda)\exp(-\mathrm{A}OD(\lambda)\, m_r)$$

$$m_r \approx 1./\cos\theta_\oplus$$

$$\underbrace{\log(\mathrm{I}(\lambda))}_{\textbf{measurement}} = \underbrace{\log(I_0(\lambda))}_{\textbf{intercept}} - \mathrm{A}OD(\lambda)\, m_r$$

**measurement** **intercept** **slope variable**

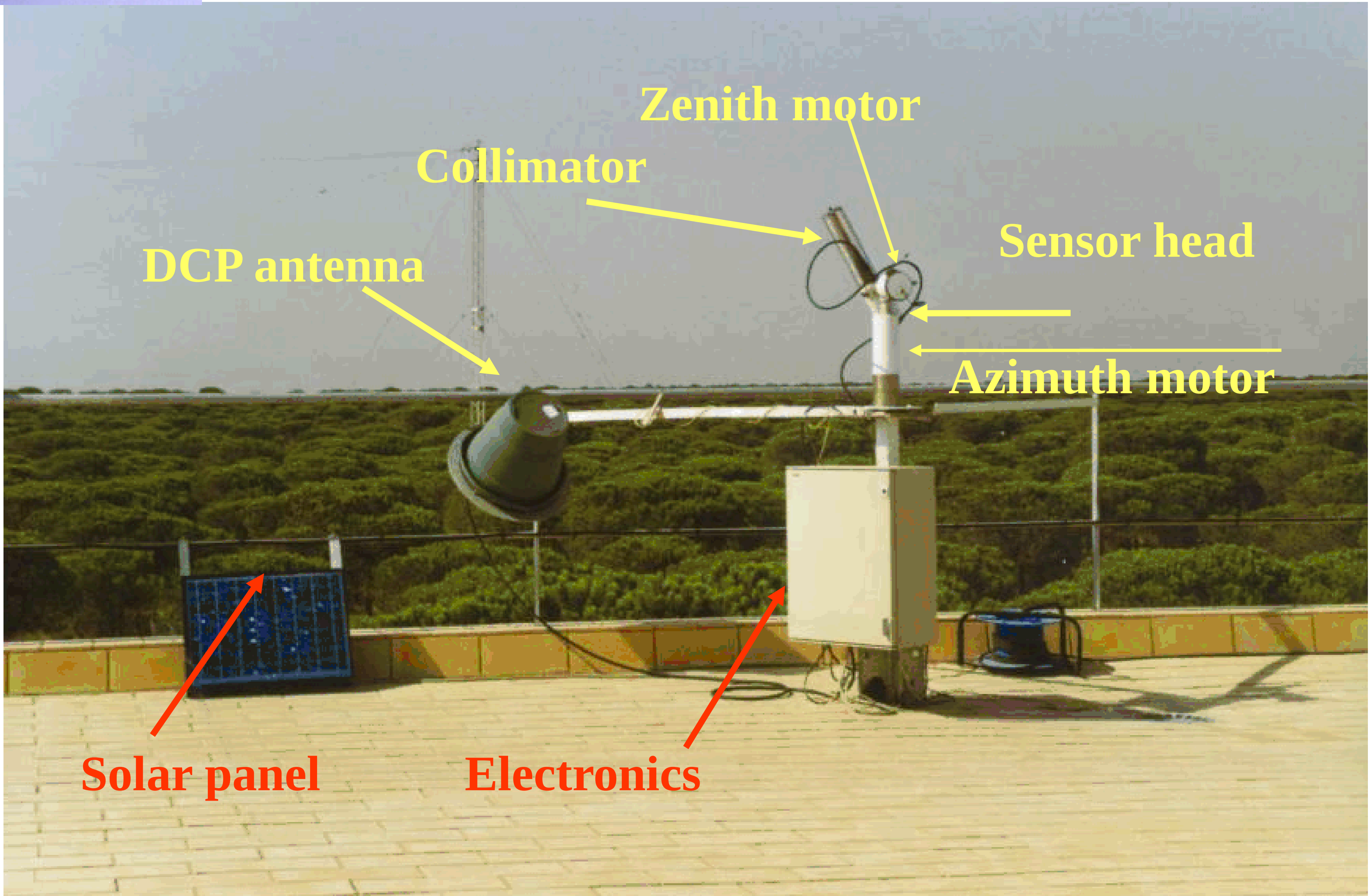



**Measurement at multiple λ allows derivation of Angstrom coefficient**

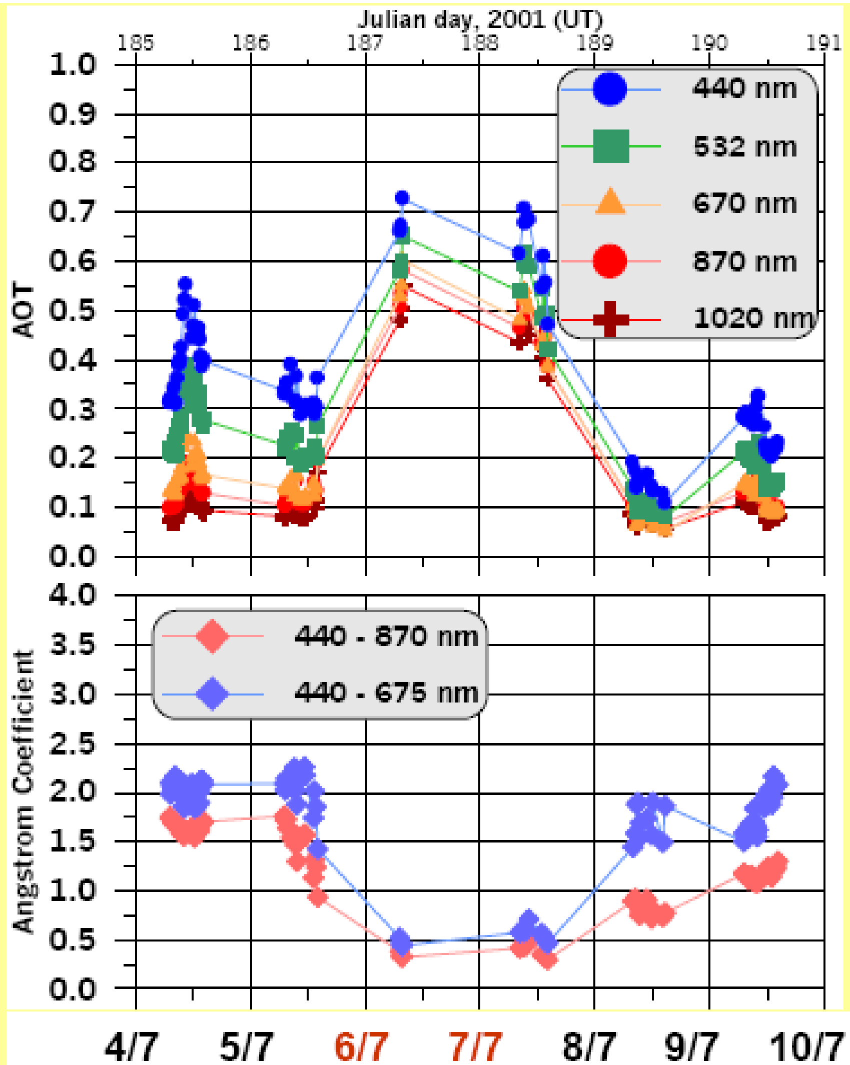
**In reality optical depth from Rayleigh and potential gas absorption need to be taken into account:**

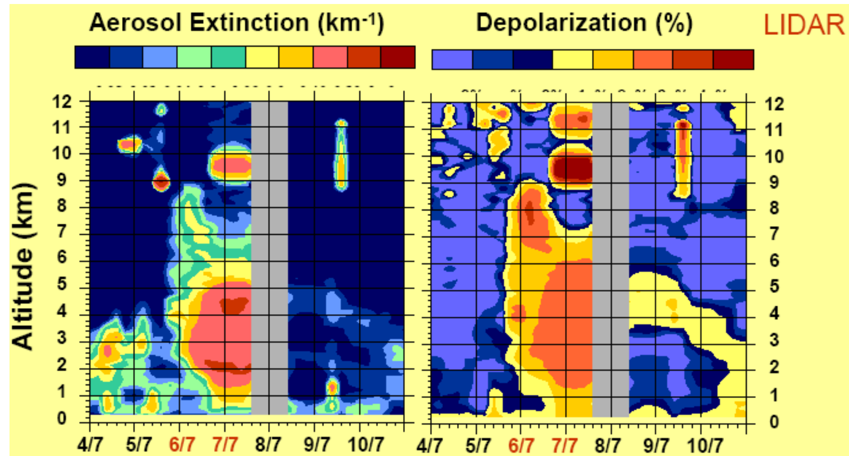
**AOD = τ(meas.) – τ(Ray) –τ(gas)**

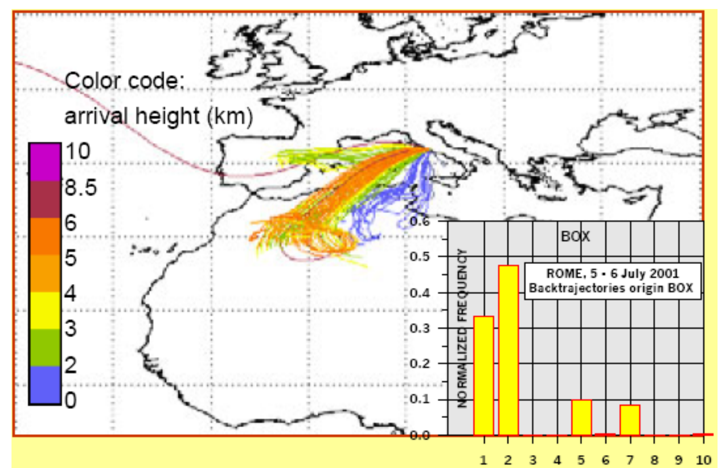
# Sun-photometer Cimel 318A (#243)
# PHOTONS-AERONET



# Saharan Dust Event

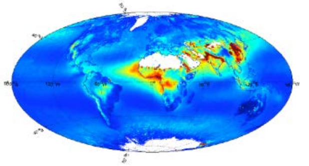


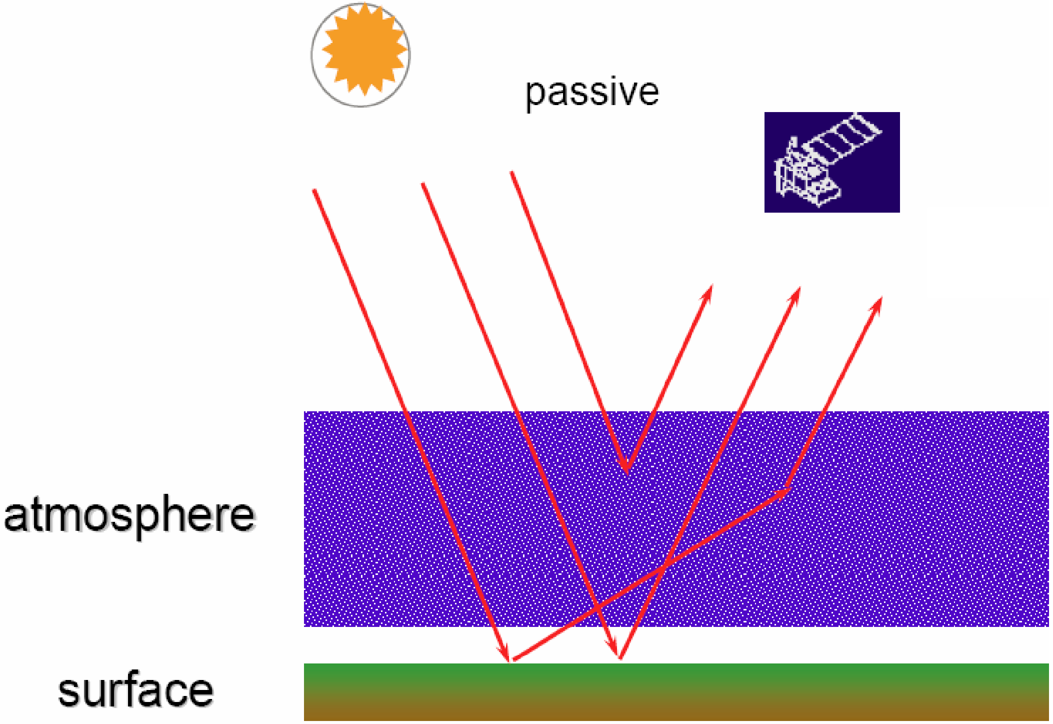




**Backtrajectories analyses**

# AEROSOL MEASUREMENT FROM SPACE






As usual we measure a **signal** composed of contributions from:

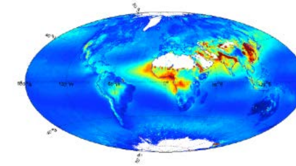
- ❑ Atmospheric constituents
  - Clouds
  - Aerosols
  - Gases
- ❑ Surface reflections
- ❑ Interaction between surface and atmosphere

Depending on what we want to know, e.g.:

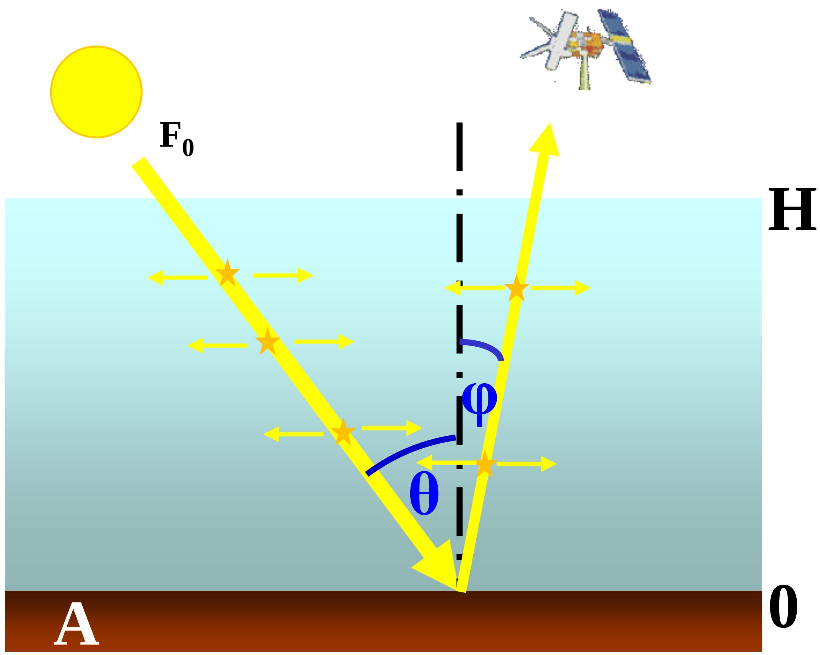
- ❑ Surface properties require atmospheric correction
- ❑ Atmospheric properties require elimination of surface effects

Note: Absorption effects can be minimized by choosing wavelength without gaseous absorption

# A View from Space

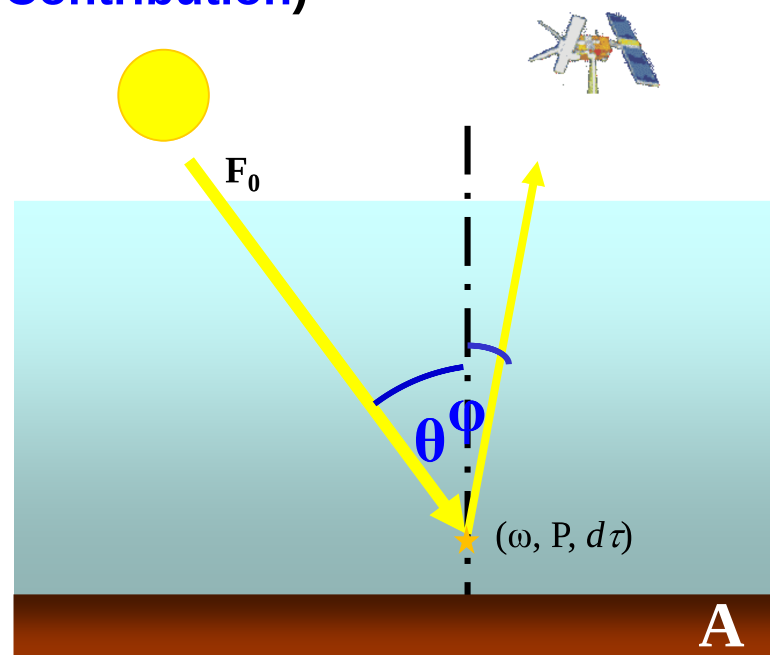


**Sunlight reflected from surface (Surface Contribution)**




$$I = \frac{AF_0 \cos\theta}{\pi} e^{-\tau/\cos\theta} e^{-\tau/\cos\varphi}$$

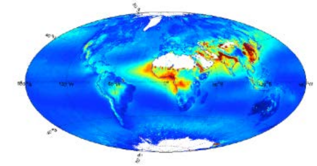
**Sunlight scattered by aerosols (Atmospheric Contribution)**



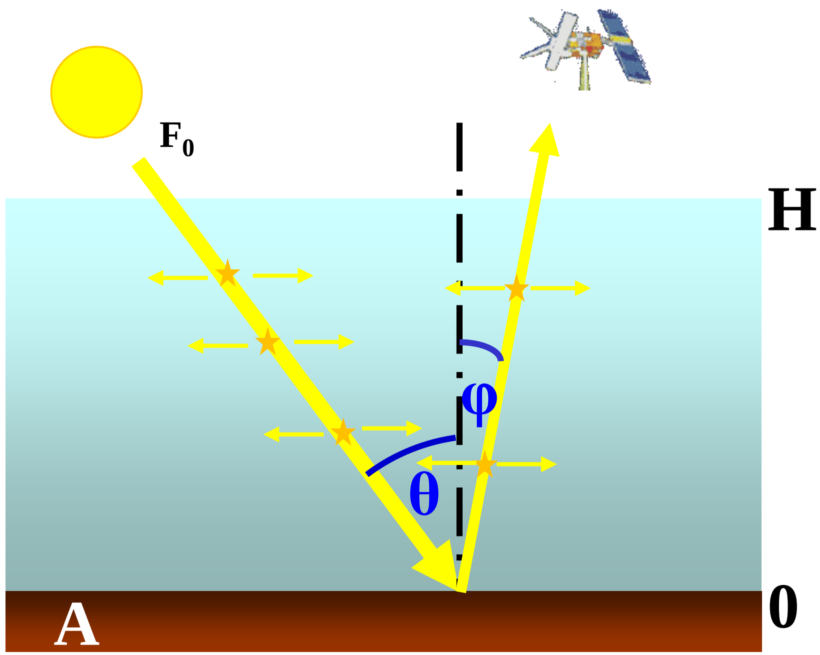

$$dI = \frac{\varpi F_0 P(\theta, \phi_0; \varphi, \phi)}{4\pi} d\tau$$

**Only layer of aerosols with dτ (single scattering approximation)**

# A View from Space



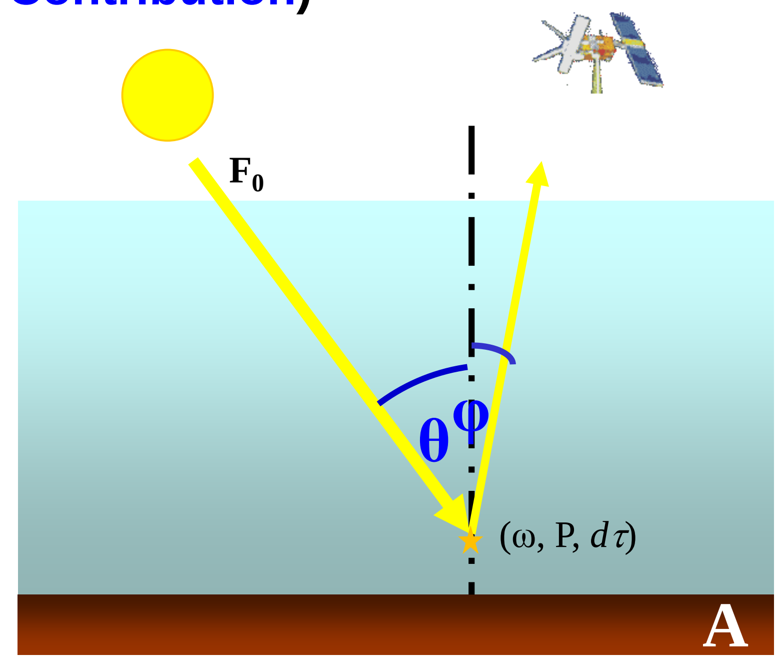
**Sunlight reflected from surface (surface Contribution)**




$$I = \frac{AF_0 \cos\theta}{\pi} e^{-\tau/\cos\theta} e^{-\tau/\cos\varphi}$$

**Aerosols will reduce *observed intensity***

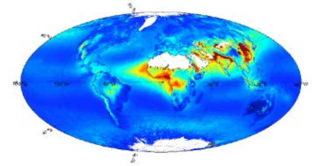
**Sunlight scattered by aerosols (Atmospheric Contribution)**




$$dI = \frac{\varpi F_0 P(\theta, \phi_0; \varphi, \phi)}{4\pi} d\tau$$

**Aerosols will increase observed intensity**

# BASIS FOR AEROSOL REMOTE SENSING



## Relative effects of AOD and single scattering albedo ω on satellite reflection as function of surface reflectance

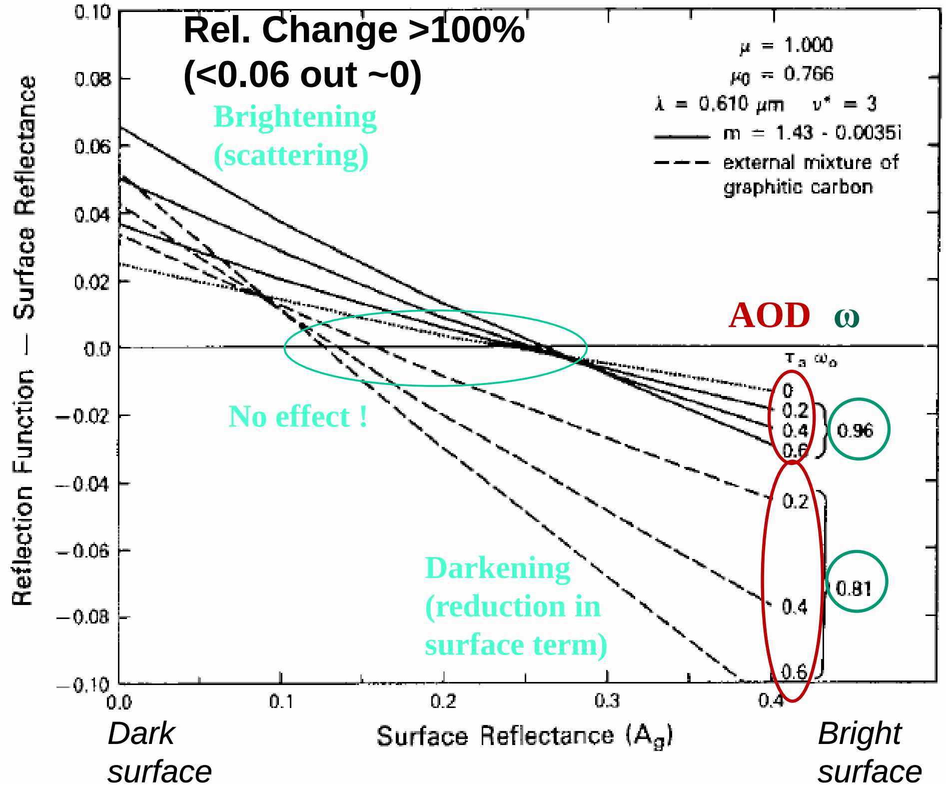
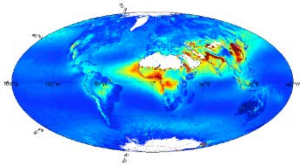



FIG. 1. Difference between the reflection function and surface reflectance at 0.61 $\mu$m as a function of surface reflectance for various values of the aerosol optical thickness $\tau_a$ and single scattering albedo $\omega_0$. The solid and dashed lines correspond to an aerosol-ladened atmosphere having a Junge size distribution of the form $n(r) \propto r^{-4}$. Results apply to nadir observations ($\mu = 1$) when $\mu_0 = 0.766$ ($\theta_0 = 40°$). For surface reflectance below a given critical value ($A_g^c$), the aerosol effect is positive ($A_g^c = 0.25$ for $\omega_0 = 0.96$) and above this value the effect is negative (adapted from Fraser and Kaufman 1985).

**Rel. Change ~10-20% (>0.1 out of 0.4)**

- **Maximum sensitivity** to AOD occurs **over dark surfaces**
- **For surfaces brighter than $A_g = 0.1$, sensitivity is much reduced and depends on aerosol absorption.**
- **Aerosol effect can be positive or negative (brighten or darken the scene)**

# STRATEGIES FOR SURFACE



**Reflection measured by satellite:**

$$R(AOD,\omega) = \underbrace{R_{atm}(AOD,\omega)}_{\text{Pure atmosphere term}} + \underbrace{\frac{A_g \mathcal{T}_{down}(AOD,\omega)\mathcal{T}_{up}(AOD,\omega)}{1 - A_g r_{atm}(AOD,\omega)}}_{\text{Multiple scattering + surface}}$$

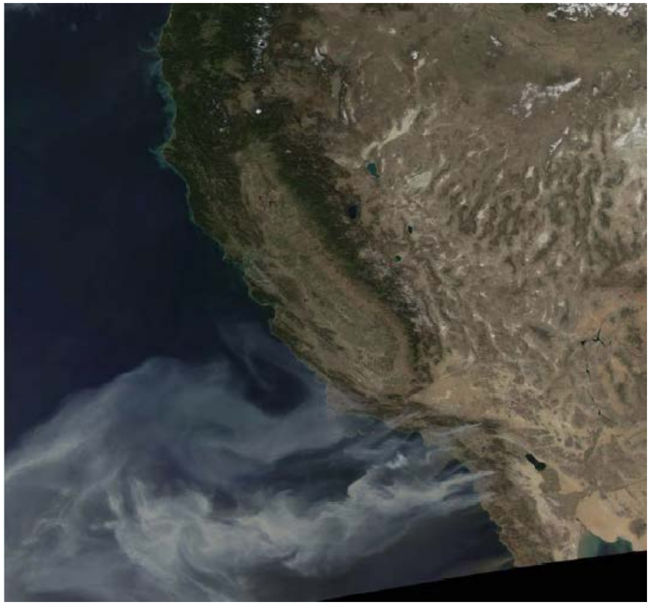
Pure atmosphere term

Multiple scattering + surface

$\mathcal{T}$, R: atmospheric Transmission and Reflectance
A: surface reflectance
r: spherical albedo of atmosphere

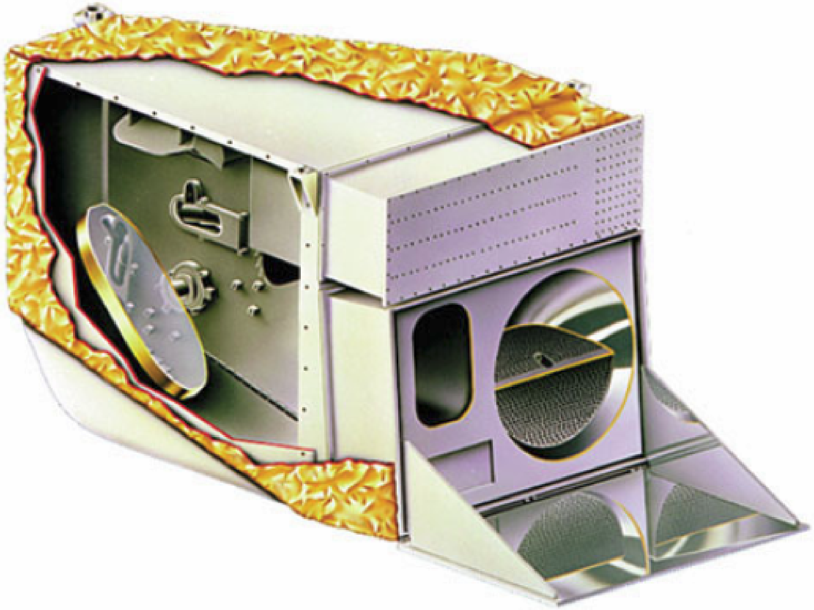
**Single (complex) equation with many unknowns**

1. **Measure over surfaces $A_g \rightarrow 0$ (ocean, dark vegetation):**
   - **Second term is zero**
   - **Large sensitivity to AOD**
2. **Estimate $A_g$ from some other measurements:**
   - **combination of dark and bright surfaces can be used to measure single scattering albedo**
3. **We can increase the number of equations:**
   - **Multiple wavelengths**
   - **Multiple angles**
   - **(Polarization)**



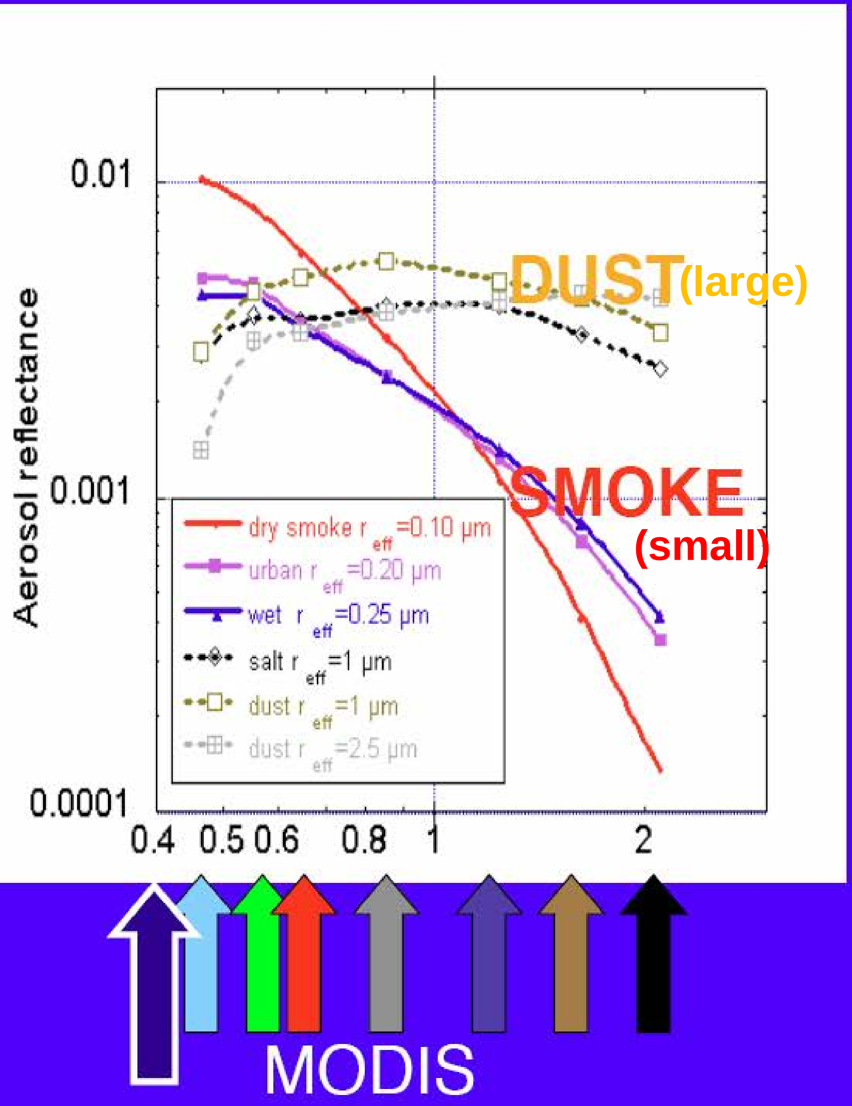
# MODIS INSTRUMENT (Again)

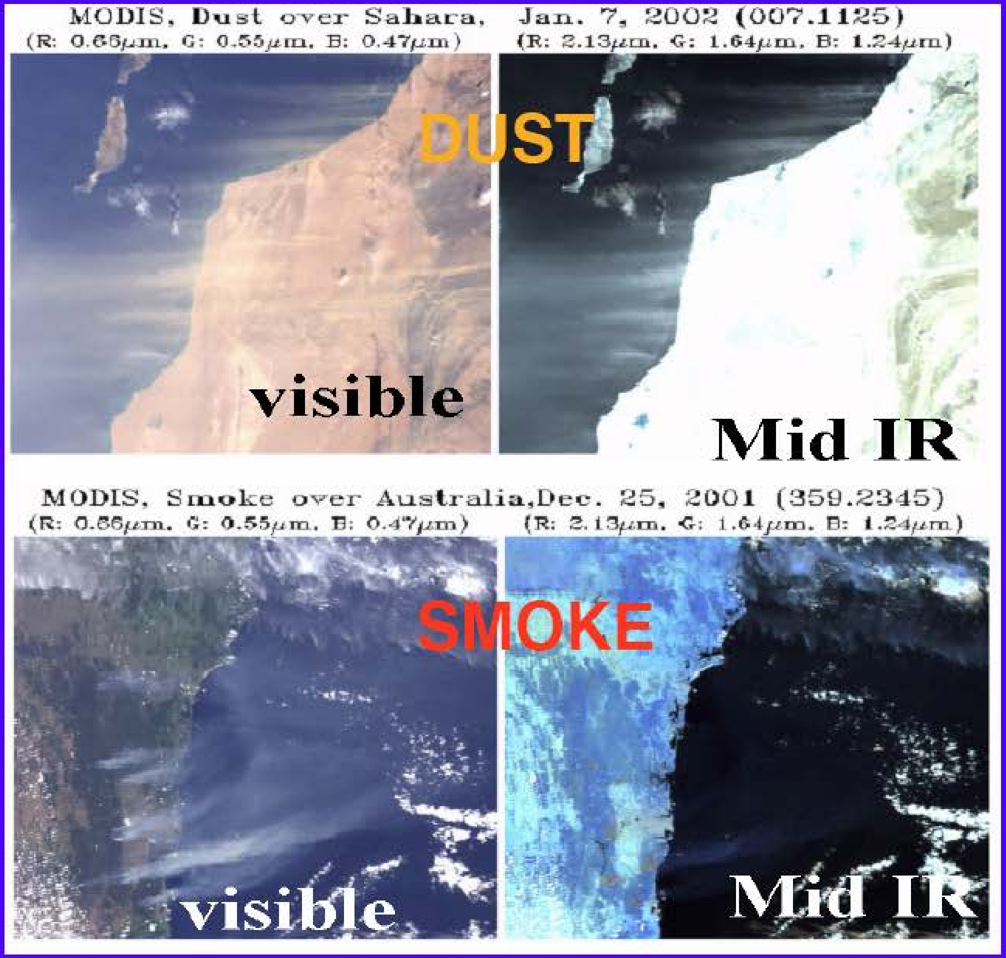
- NASA, Terra & Aqua
  - launched 1999, 2002
  - 705 km polar orbits, descending (10:30 a.m.) & ascending (1:30 p.m.)
- Sensor Characteristics
  - 36 spectral bands (490 detectors) ranging from 0.41 to 14.39 μm
  - Two-sided paddle wheel scan mirror with 2330 km swath width
  - Spatial resolutions:
    - 250 m (bands 1 - 2)
    - 500 m (bands 3 - 7)
    - 1000 m (bands 8 - 36)
  - 2% reflectance calibration accuracy
  - onboard solar diffuser & solar diffuser stability monitor
  - 12 bit dynamic range (0-4095)



**Multi-wavelength but single view instrument !**

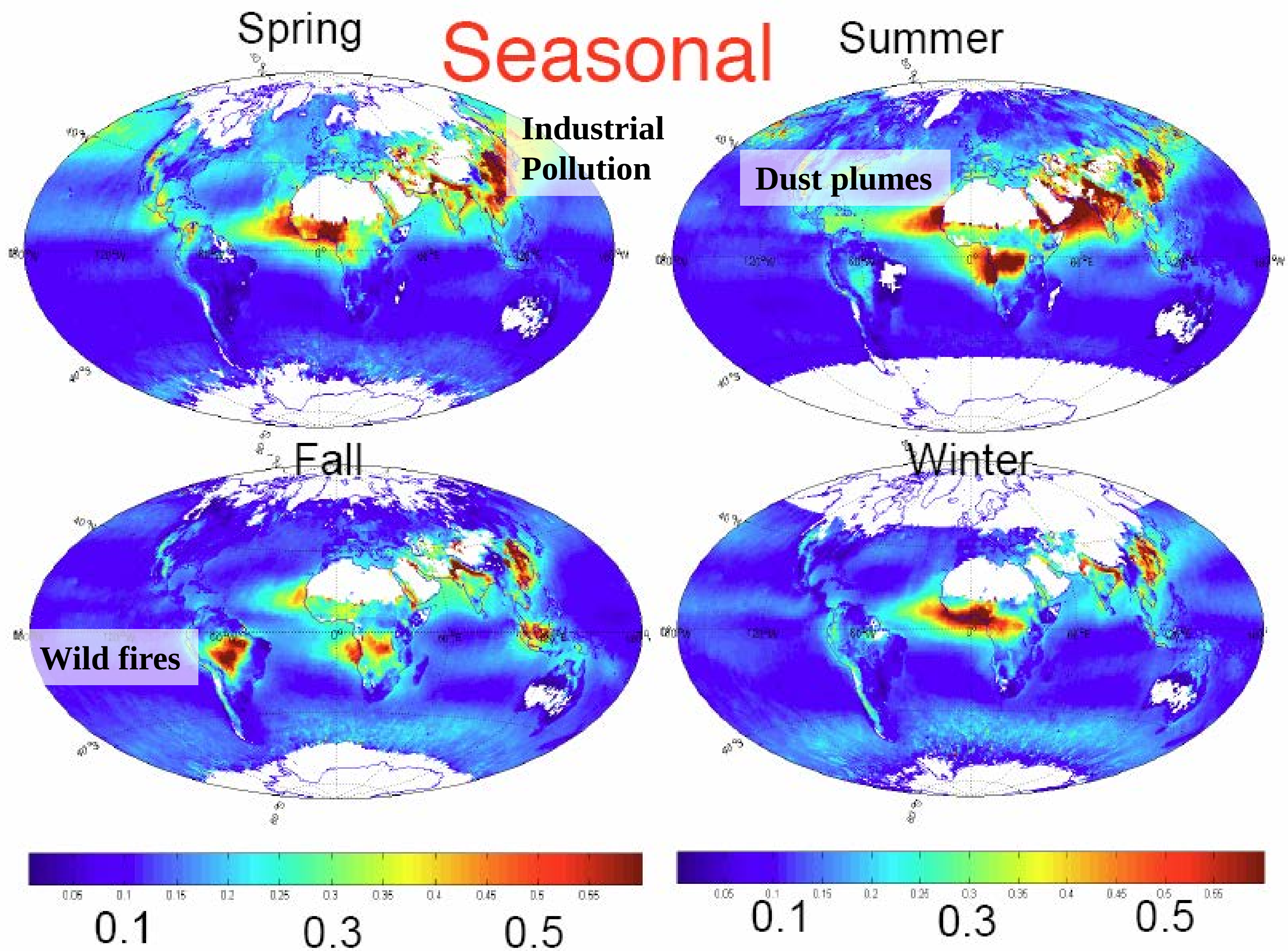
# Spectral optical properties of aerosol




Y. Kaufman

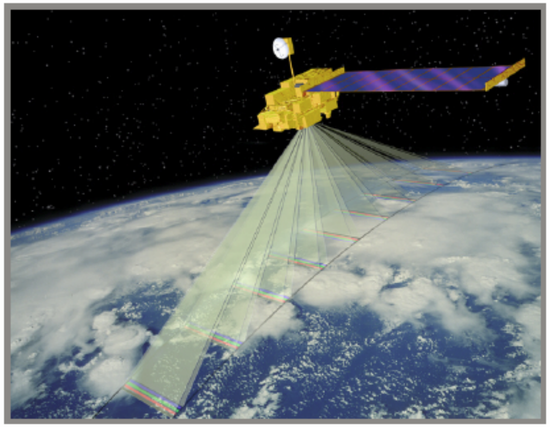
# MODIS AEROSOL OPTICAL DEPTH

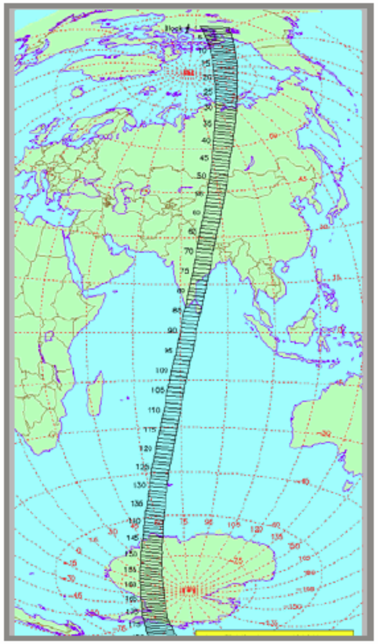


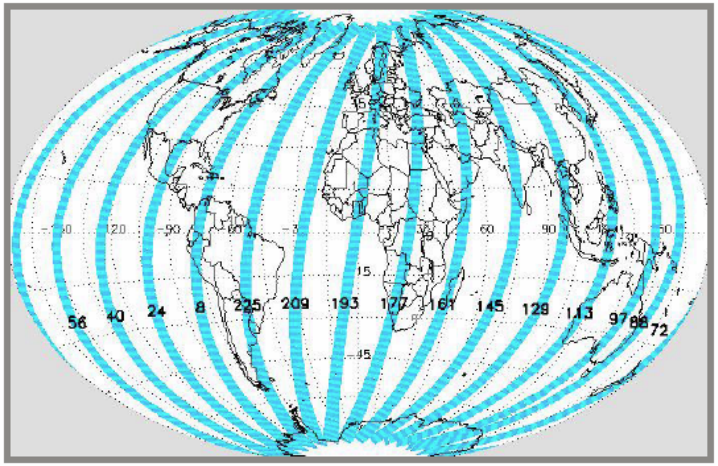

**No AOD over desert and snow/ice surfaces !**

# Multiangle Imaging SpectroRadiometer (MISR)

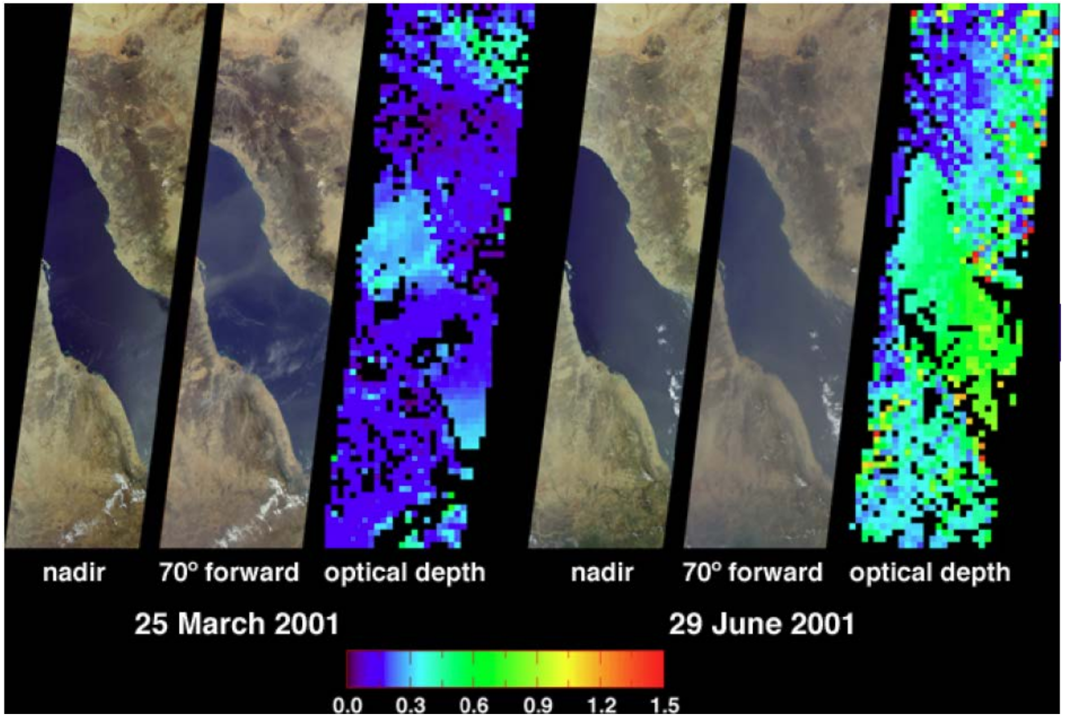
- On board satellite TERRA
- 9 view angles at earth surface:
  - ±70.5º. ±60.0º, ±45.6º, ±26.1º, nadir
- Four spectral bands at each angle:
  - 446 nm (Blue)
  - 558 nm (Green)
  - 672 nm (Red)
  - 866 nm (NIR)
- Global Mode:
  - 275 m sampling resolution for nadir camera and red band of other cameras
  - 1.1 km for the other channels
  - 400-km swath
  - Global coverage: 9 days at equator, 2 days at poles
- Continuous data retrieval since Feb 2000.







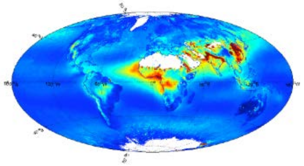
# AEROSOL RETRIEVALS WITH MULITPLE VIEW

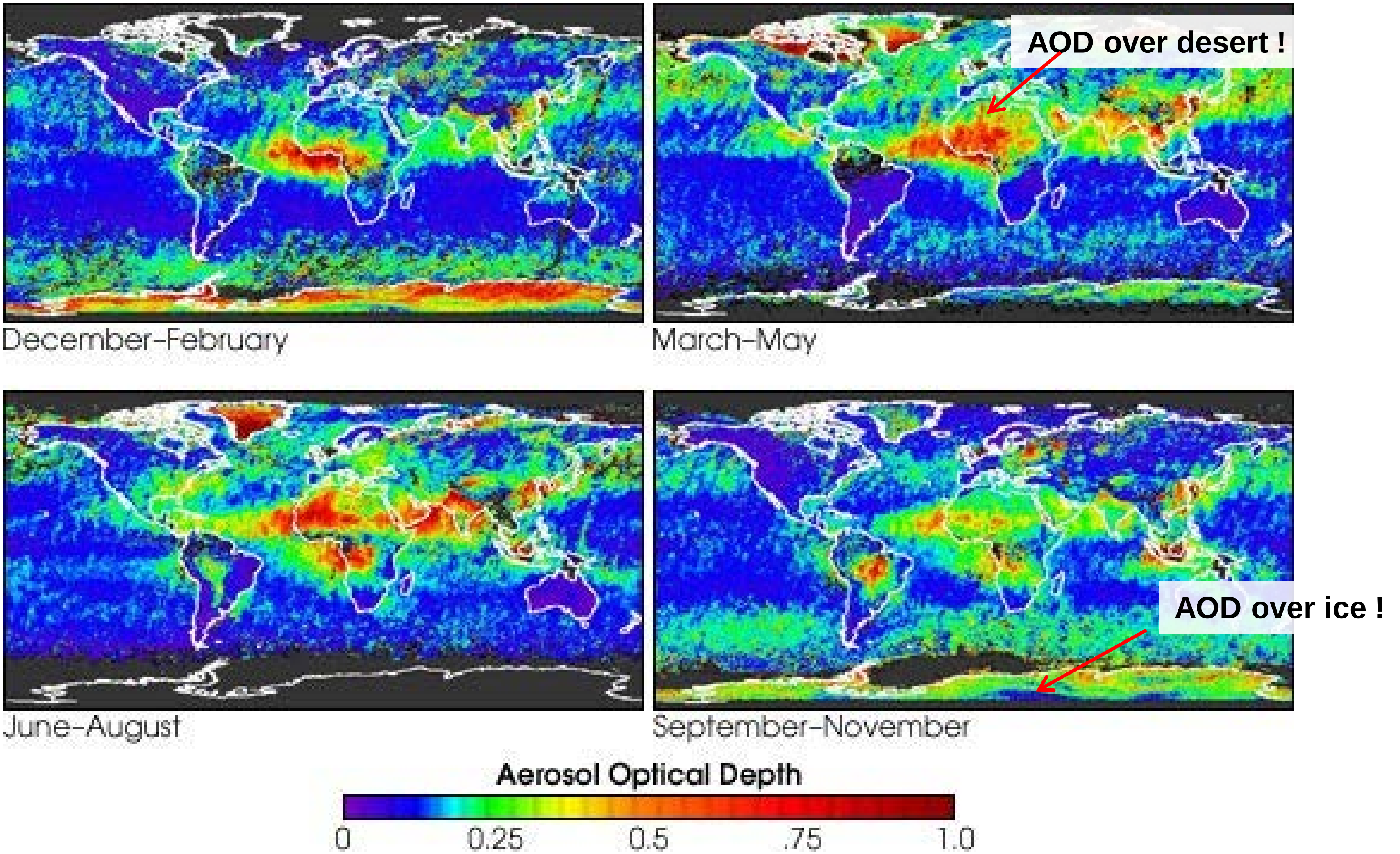



MISR image over red sea

- **Nadir view: short path through atmosphere and large sensitivity to surface**
- **Forward-viewing (70.5-degree camera): increased line-of-sight through atmosphere and aerosol (thin haze) is more apparent**
- **Surface effects can be separated from aerosol effects and aerosol retrieval over bright surface possible**
- **Need to assume Lambertian surface or BRDF independent of wavelength**

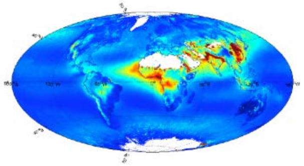
# MISR AEROSOL OPTICAL DEPTH



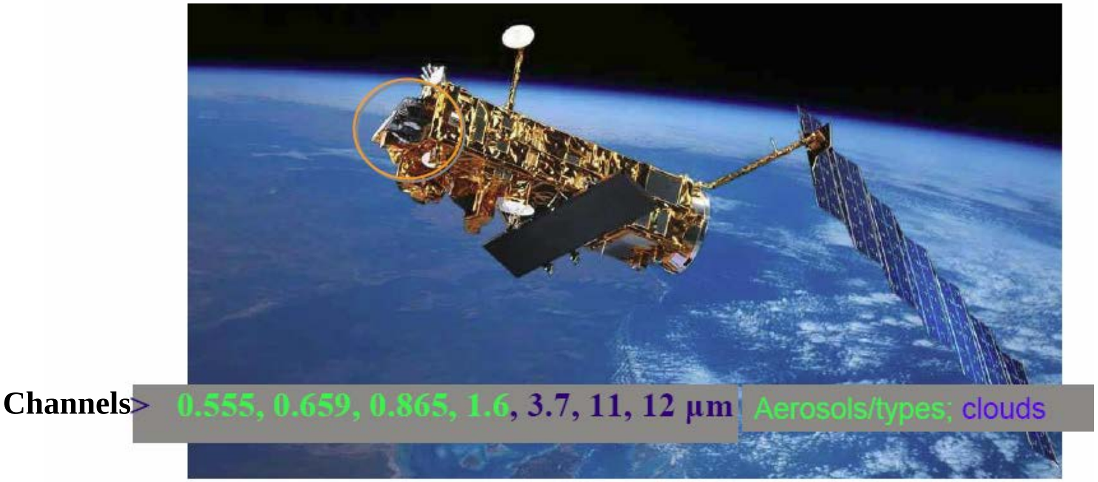


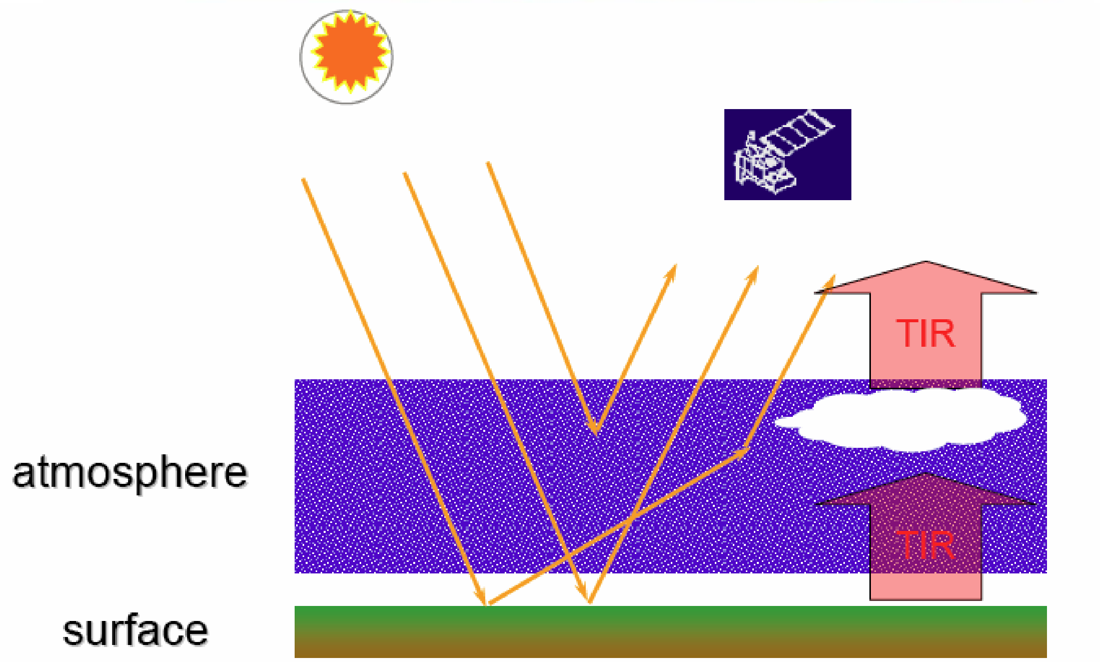

**MISR can measure AOD over bright scenes !**

# CLOUD EFFECTS



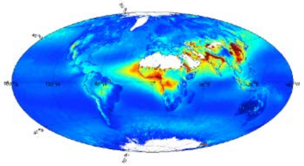
**AATSR on ENVISAT**




- **Clouds will scatter and absorb light in visible**
  - **Potentially large source of errors for aerosol and trace gas retrievals**
  - **Clouds are often optically thick (ie .surface effects can be ignored)**
- **Infrared channels:**
  - **Much less sensitive to aerosol**
  - **Allow detection of clouds**
  - **Help with very large aerosols (desert dust)**

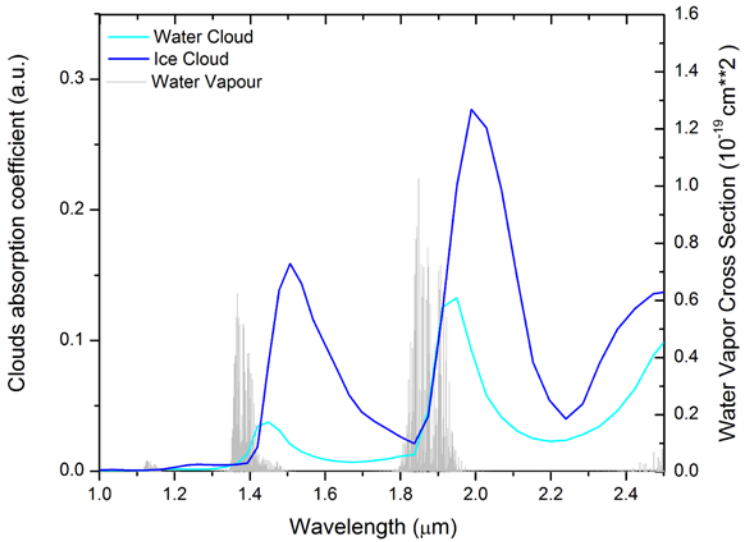
# CHARACTERICS OF CLOUDS



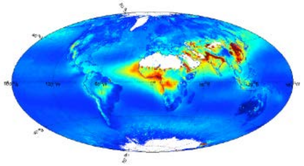
- Clouds consist of **liquid or frozen water particles**
- Water **cloud droplets are spherical** (Mie theory applies) with effective radius of a few µm
- **Ice particles can have many different shapes** with effective diameter of several tenth of µm
- Cloud particles are just like aerosols but
  - Particles are large
  - High number density thus optically thick so that surface effects are small or zero
  - Clouds have distinct absorption features (water absorption)
- Main parameters of interest: Cloud optical depth, effective radius, phase, liquid (ice) absorption and cloud height

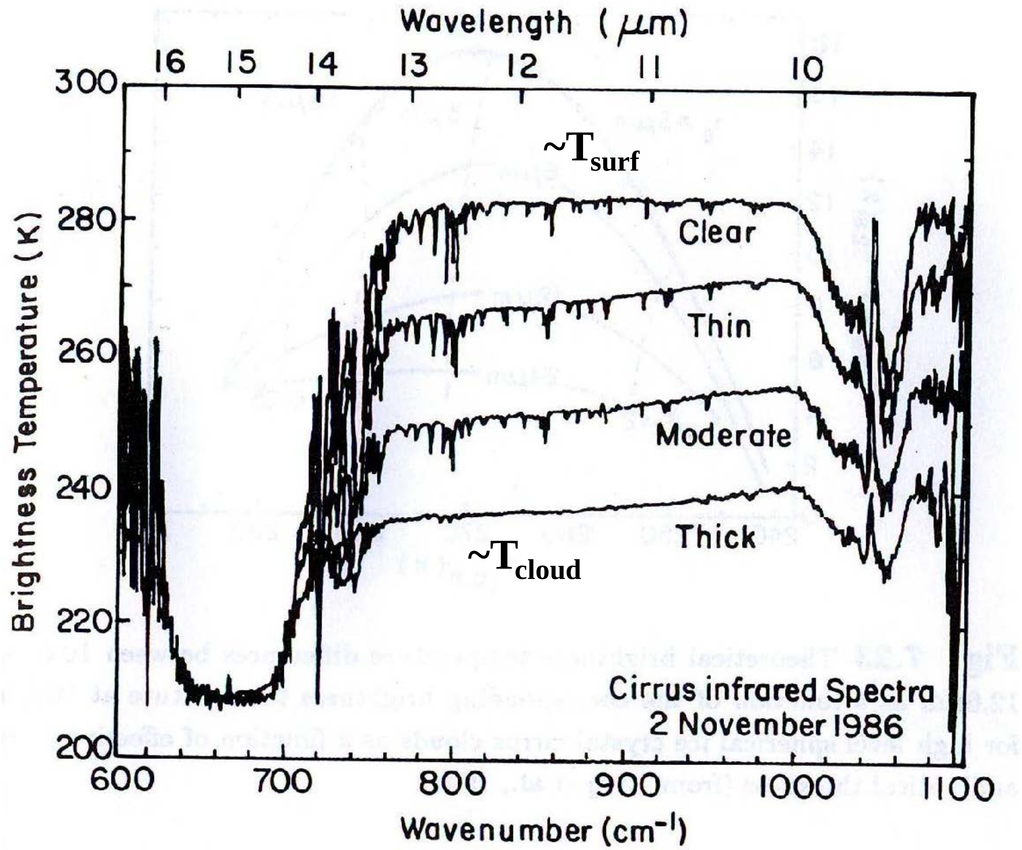
**Ice Particles**

**Absorption by water in different phases**



# CLOUD EFFECT IN INFRARED






**Clouds absorb in IR (continuum absorption) and they reduce surface contribution in window range**

**Schwarzschild Eqn (Single layer model):**

**(for λ without strong gaseous absorbers)**

$$I(\lambda) = \varepsilon_s(\lambda) \times B(\lambda, T_s) \times \mathcal{T}_{cloud}(\lambda) + (1 - \mathcal{T}_{cloud}(\lambda)) \times B(\lambda, T_{cloud})$$

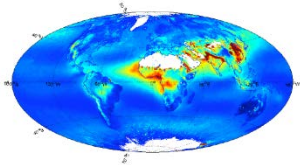
**Surface contribution**

**For thick cloud: transmission $\mathcal{T}_{cloud}(\lambda) \rightarrow 0$:**

$$I(\lambda) = B(\lambda, T_{cloud}) \quad \text{-> } T_{cloud} \sim 240K$$

**-> Measurement gives estimate of cloud top temperature (or height)**

# MODIS Cloud Channels



- 11 MODIS bands are utilized to derive cloud properties
- Visible and near-infrared bands:
  - daytime retrievals of cloud optical thickness and effective radius
- Thermal infrared bands:
  - determination of cloud top properties, including cloud top altitude, cloud top temperature, and thermodynamic phase

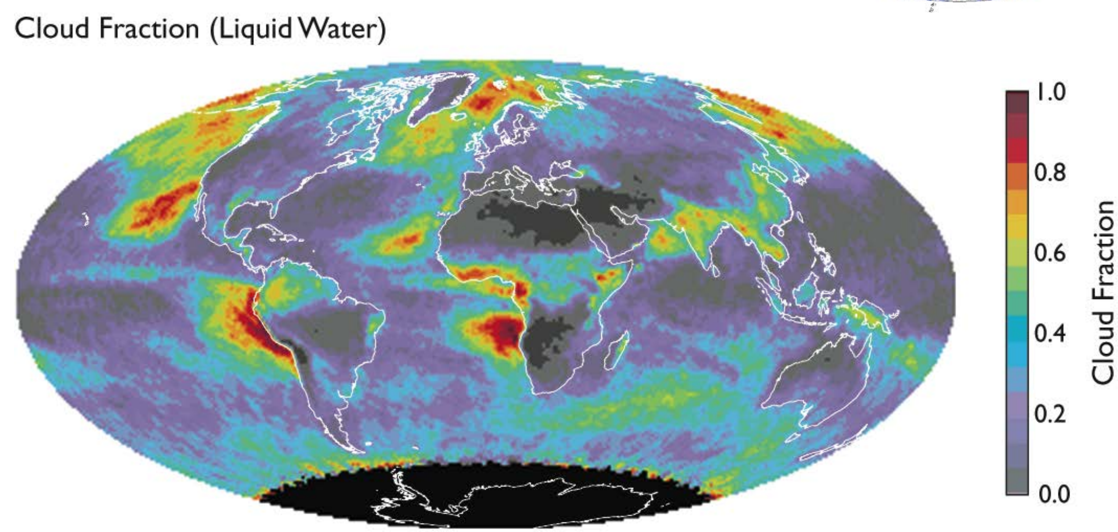
| Channel | λ (μm) | Δλ (μm) | Resolution (m) | Atmospheric Purpose |
|---|---|---|---|---|
| 1 | 0.645 | 0.050 | 250 | cloud optical thickness over land |
| 2 | 0.858 | 0.035 | 250 | cloud optical thickness over ocean |
| 5 | 1.240 | 0.020 | 500 | cloud optical thickness over snow & ice |
| 6 | 1.640 | 0.025 | 500 | snow/cloud discrimination, thermodynamic phase |
| 7 | 2.130 | 0.050 | 500 | cloud effective radius |
| 20 | 3.750 | 0.180 | 1000 | cloud effective radius; surface temperature |
| 26 | 1.375 | 0.030 | 1000 | thin cirrus detection |
| 29 | 8.550 | 0.300 | 1000 | thermodynamic phase |
| 31 | 11.030 | 0.500 | 1000 | thermal emission correction; cloud height |
| 32 | 12.020 | 0.500 | 1000 | thermodynamic phase |
| 33 | 13.335 | 0.300 | 1000 | cloud height |

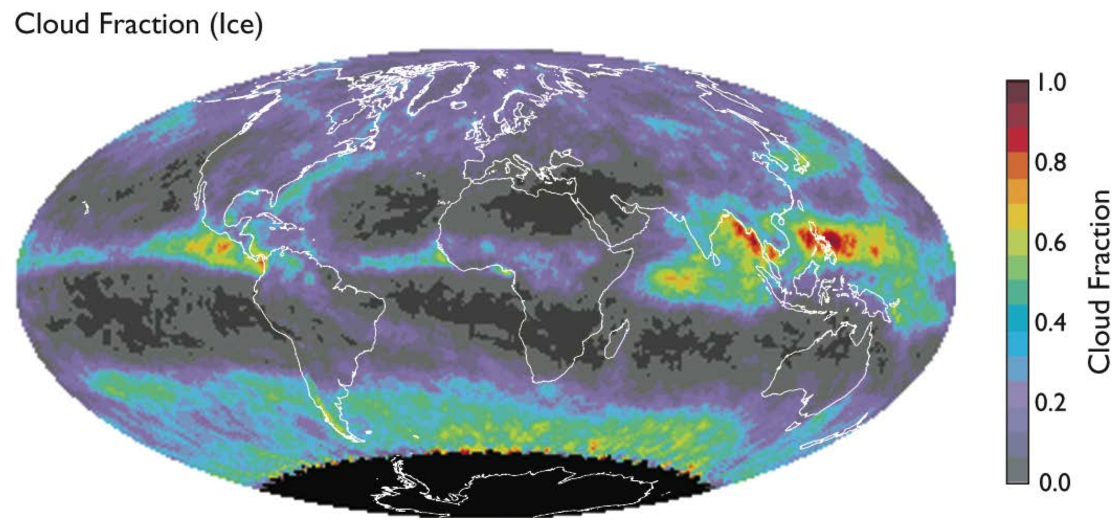
# Monthly Mean Cloud Fraction by Phase

**July 2006 – Terra**

- **Liquid water clouds**
  - **Marine stratocumulus regions**
    - ✓ **Angola/Namibia**
    - ✓ **Peru/Ecuador**
    - ✓ **California/Mexico**

- **Ice clouds**
  - **Tropics**
    - ✓ **Indonesia & western tropical Pacific**
    - ✓ **ITCZ**
  - **Roaring 40S**



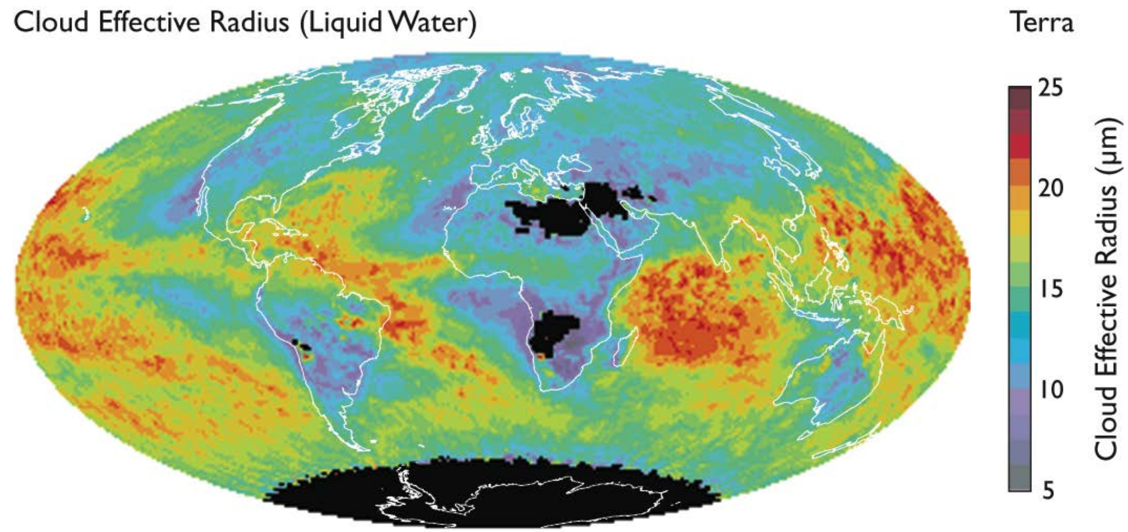



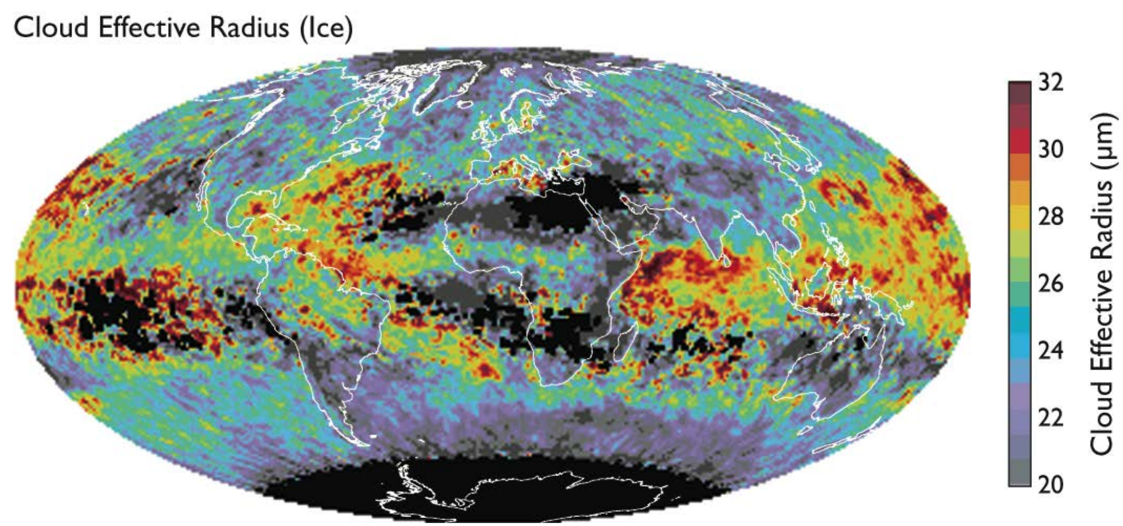
# Monthly Mean Cloud Effective Radius

**July 2006 – Terra**

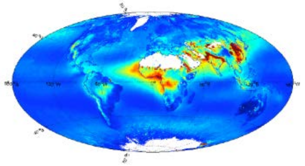
- **Liquid water clouds**
  - **Larger drops in SH than NH**
  - **Larger drops over ocean than land**
    - ✓ **Due to cloud condensation nuclei (aerosols)**

- **Ice clouds**
  - **Larger in tropics than high latitudes**
    - ✓ **Anvils**
  - **Small ice crystals at top of deep convection**



# What do you need to know?



- Characteristics of atmospheric aerosols and cloud particles
- Aerosol size distribution
- Aerosol optical properties and their dependence on microphysical parameters (size distr., refractive index)
- Angstrom relation
- Determination of AOD
- Basics of aerosol remote sensing
- Surface effects and how to deal with them
- Multi wavelength and multi-view instruments for aerosols and clouds
- Clouds in visible/near-IR and thermal infrared

# INCOMPLETE LIST OF AEROSOL SATELLITE SENSORS

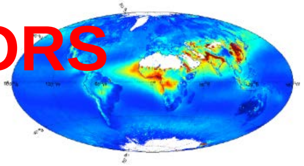


TABLE 3. Techniques for remote sensing of global aerosol properties from space. Here, $\tau_a$ denotes aerosol optical thickness, $\omega_0$ the single scattering albedo, $r_e$ the aerosol particle effective radius, and $n_c(r)$ the columnar aerosol size distribution.

| Technique | Aerosol property | Relevant satellite sensors | Principal assumptions | References |
|---|---|---|---|---|
| Ocean | | | | |
| reflectance–one channel | $\tau_a$ | AVHRR | Size distribution, ocean reflectance, whitecaps, calibration, $\omega_0$ | Griggs (1975); Mekler et al. (1977); Quenzel and Koepke (1984); Fraser et al. (1984); Stowe et al. (1997) |
| reflectance–multiple channels | $\tau_a$, $n_c(r)$ | GLI, MERIS, MISR, MODIS, OCTS, POLDER, SeaWiFS, AVHRR, TOMS, OMI | Type of size distribution, ocean reflectance, whitecaps, calibration, $\omega_0$ | Durkee et al. (1986, 1991); Tanré et al. (1997); J. R. Herman et al. (1997); Fukushima and Toratani (1997); Higurashi and Nakajima (1999) |
| Land | | | | |
| reflectance | $\tau_a$ | AVHRR, TOMS, OMI | Size distribution, no change in surface reflectance, $\omega_0$ (to derive $\tau_a$) | Fraser et al. (1984); Kaufman et al. (1990a,b); Fraser (1993) |
| reduction in contrast | $\tau_a$ | AVHRR, GLI, MODIS, POLDER | No change in surface reflectance, bidirectional reflectance, $\omega_0$, asymmetry factor | Tanré et al. (1988b); Tanré and Legrand (1991); Holben et al. (1992) |
| dark targets over dense dark vegetation | $\tau_a$ | AVHRR, GLI, MERIS, MISR, MODIS, POLDER, SeaWiFS | Reflectance of dark targets, size distribution, $\omega_0$ | Kaufman and Sendra (1988); King et al. (1992); Kaufman et al. (1997a,b) |
| thermal contrast | $\tau_a$ | AVHRR, GLI, MODIS | | Legrand et al. (1989); Tanré and Legrand (1991); Ackerman (1997); |
| Land–ocean contrast | $\tau_a$, $\omega_0$, $r_e$ | AVHRR, GLI, MODIS, POLDER | Size distribution and surface bidirectional reflectance | Kaufman and Joseph (1982); Fraser and Kaufman (1985); Kaufman et al. (1990a); Nakajima and Higurashi (1997) |
| Angular distribution of reflectance | $\tau_a$, $n_c(r)$ | ATSR-2, AATSR, MISR, POLDER | Surface bidirectional reflectance | Martonchik and Diner (1992); Wang and Gordon (1994); Veefkind and de Leeuw (1997) |
| Polarization | $\tau_a$, $n_c(r)$ | POLDER | Reflectance and polarization of underlying surface, spherical particles | Leroy et al. (1997); Mishchenko and Travis (1997); M. Herman et al. (1997) |