Atmospheric Radiation and its Interaction with Other Physical Processes: Representation in Numerical Models

Yu Gu

Department of Atmospheric and Oceanic Sciences
Joint Institute for Regional Earth System Science and Engineering
University of California, Los Angeles, Los Angeles, California

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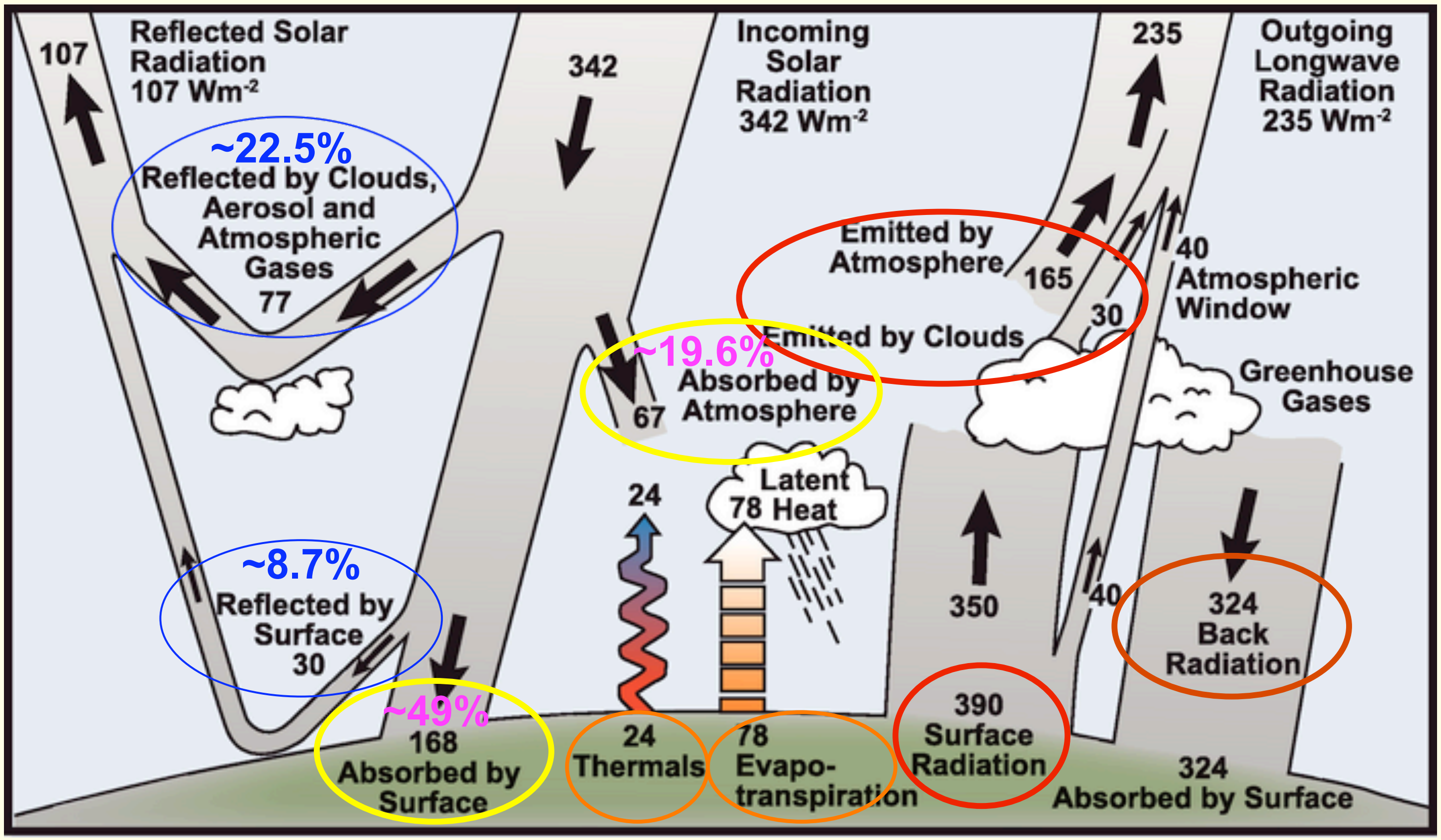
Outline

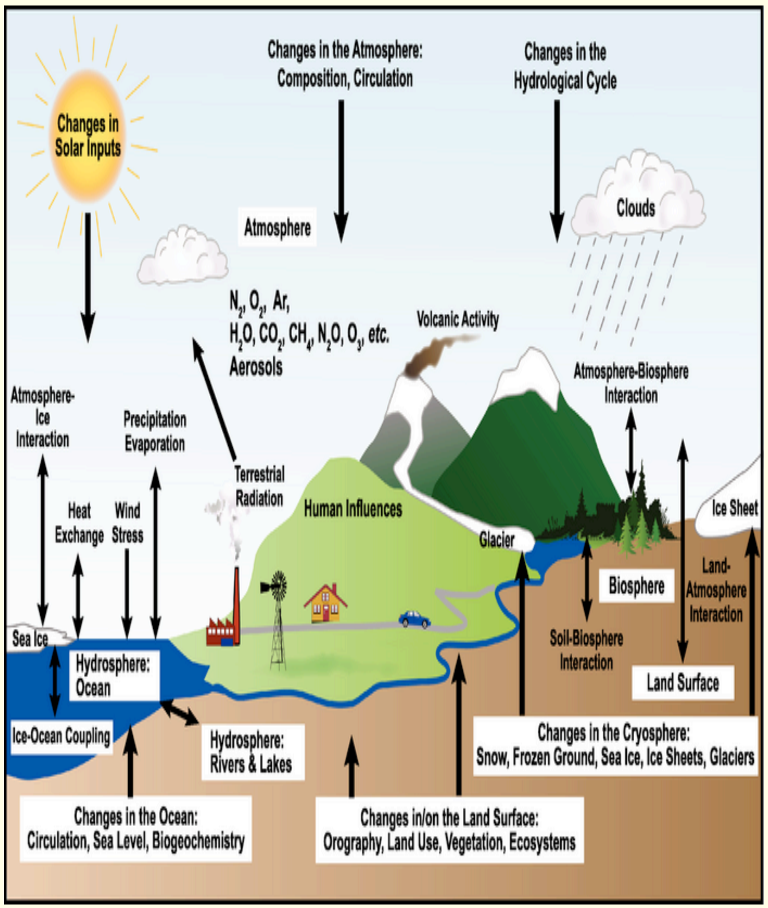
- Fundamentals of Radiative Transfer
- Interaction with Other Physical Processes
- Radiation Schemes in WRF
- Future Work

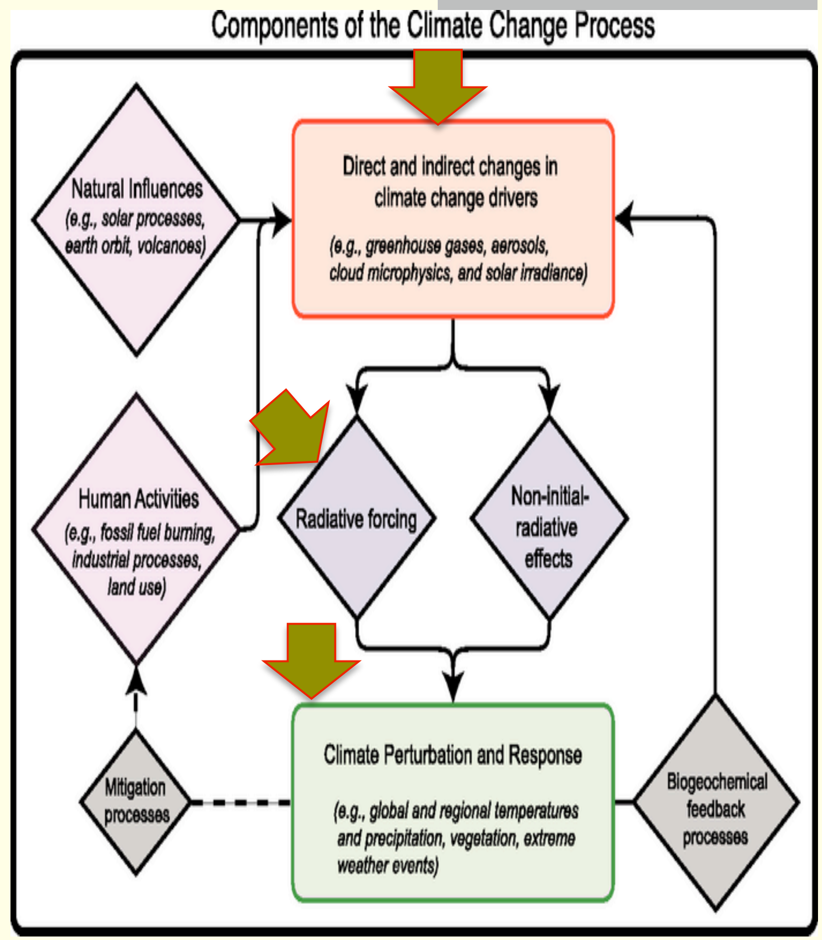
Why Radiation is Important



Estimate of the Earth's annual and global mean energy balance.
Source: Kiehl and Trenberth (1997).

Climate System & Radiative Forcing



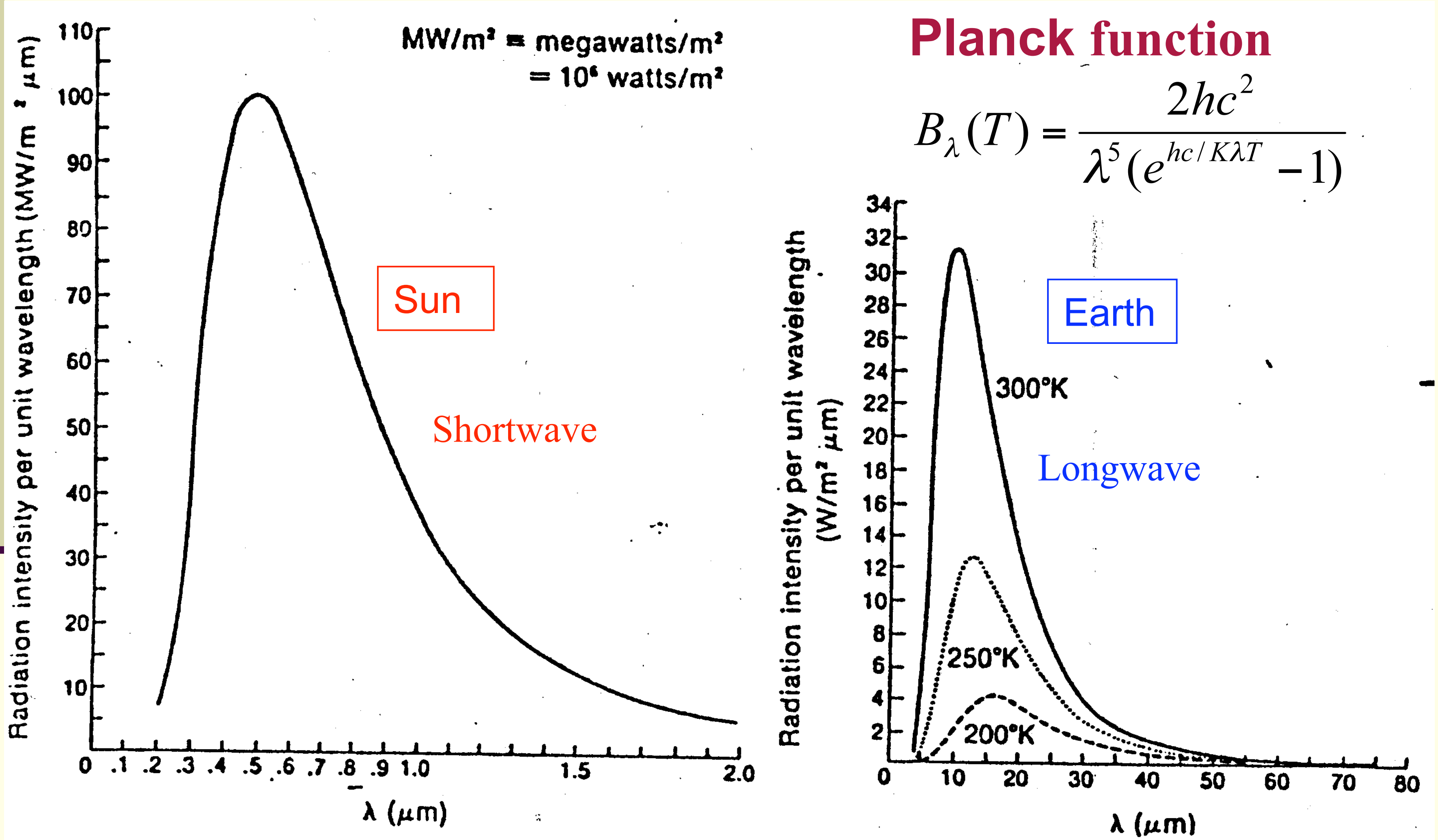


Schematic view of the components of the climate system, their processes and interactions (IPCC, 2007).

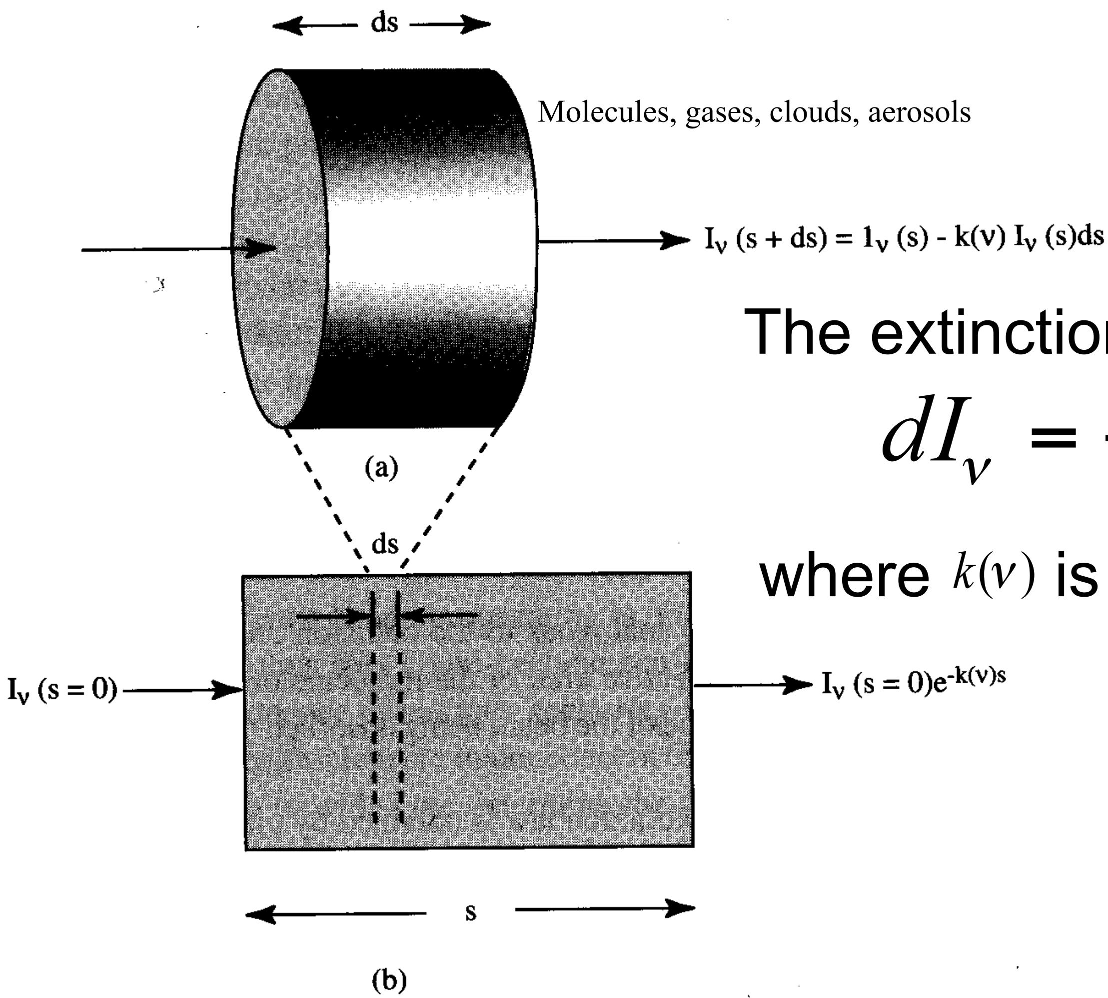
Source of Energy: Comparison of solar and earth's blackbody intensity

Planck function

$$B_\lambda(T) = \frac{2hc^2}{\lambda^5\left(e^{hc/K\lambda T} - 1\right)}$$



The Extinction law



The extinction law can be written as

$$dI_\nu = -k(\nu)\,\rho_a I_\nu ds$$

where $k(\nu)$ is the extinction cross section

Basic Equation of Radiative Transfer

- Source function coefficient, j_ν: describe the ***emission and scattering*** *process*

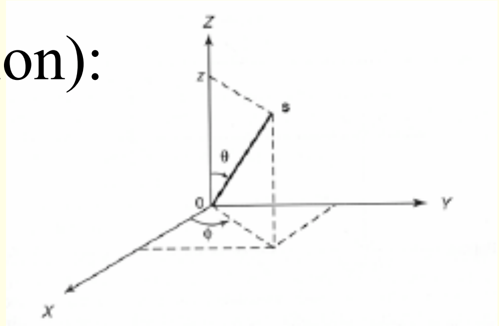
- Combining the extinction law with the definition of the source function coefficient – **Basic radiative transfer Eq**:

$$dI_\nu = \underbrace{-k(\nu)\rho_a I_\nu ds}_{\text{(Extinction)}} + \underbrace{j_\nu \rho_a ds}_{\text{(Source)}}$$

- The ratio $j_\nu/k(\nu)$ is known as the source function J_ν

$$\boxed{\frac{dI_\nu}{k(\nu)\rho_a ds} = -I_\nu + J_\nu}$$

1-D Equation of Radiative Transfer



- **Plane-parallel assumption** (vertical direction):

  $ds = dz/\cos\theta = dz/\mu$

  where θ is the solar zenith angle, $\mu = \cos\theta$

- The total extinction over a finite distance (path length) - ***optical depth***

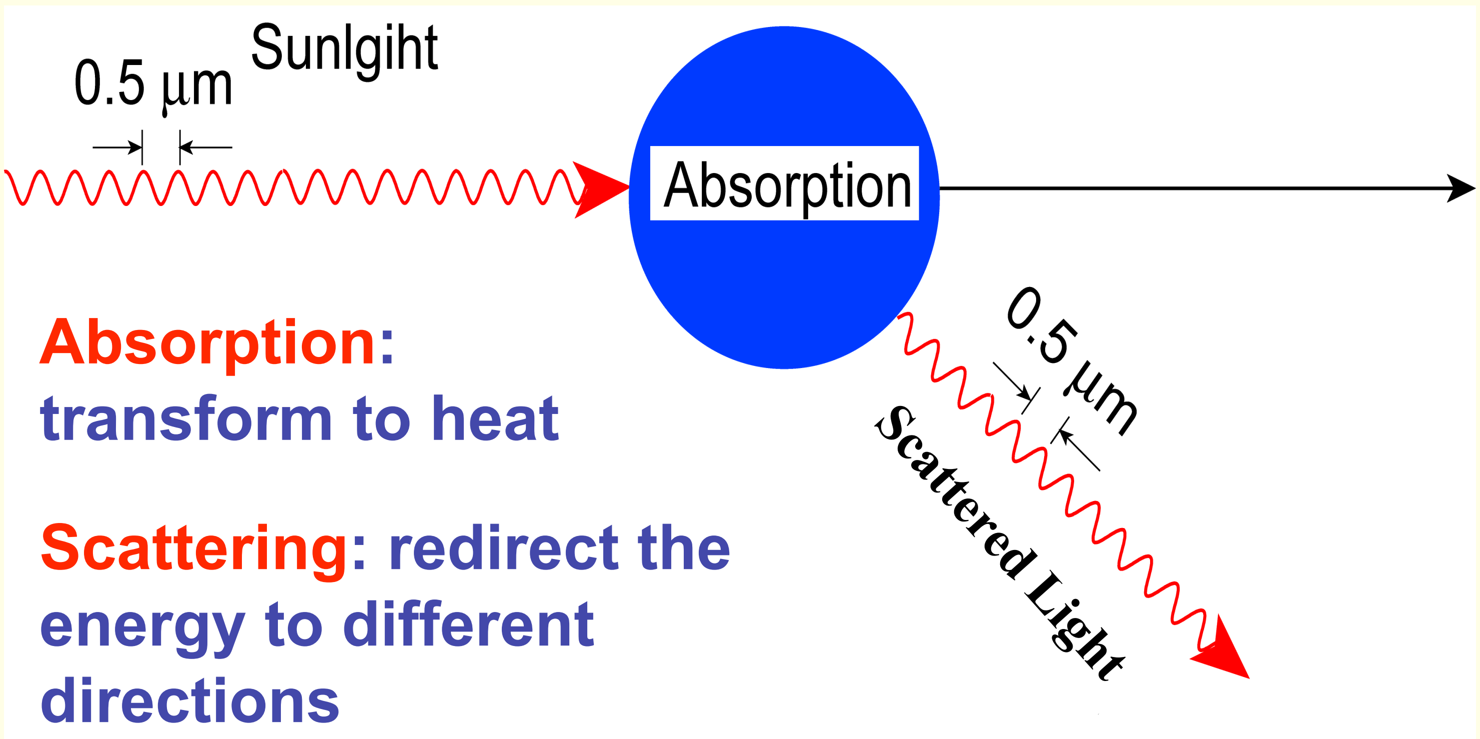
$$\tau(\nu) = \int_{z}^{z_\infty} k_\nu(z')\rho_a(z')dz' \quad \text{or} \quad d\tau = -k_\nu(z)\rho_a(z)dz$$

The basic equation of radiative transfer:

$$\boxed{\mu \frac{dI(\tau,\mu,\phi)}{d\tau} = I(\tau,\mu,\phi) - J(\tau,\mu,\phi)}$$

Extinction: Light Scattering and Absorption

- Extinction consists of two distinct processes, scattering and absorption, hence $$\tau(\nu) = \tau_s(\nu) + \tau_a(\nu)$$



Scattering Process

- Two types of scattering are considered –

1) Molecular scattering (Rayleigh) by particles with sizes much smaller than the wavelength – The scattered intensity is inversely dependent on the wavelength to the fourth power $I_\lambda \sim 1/\lambda^4$

2) Scattering from particles whose sizes are comparable to or larger than the wavelength (Mie or Lorenz-Mie) – Solution to Maxwll's equation which takes the form of analytical infinite series

Source Function

- Three factors contribute to the source function**:**

❑**Emission** which can be expressed by Planck function

❑**Single scattering** of direct solar irradiance at TOA, which is associated with the exponential attenuation to the level τ

❑**Multiple scattering** of diffuse intensity

- Two parameters are introduced to represent single scattering and multiple scattering:

❑**Phase function** $P(\cos\theta)$ or **asymmetry factor** g: represents the angular distribution of scattered energy as a function of direction

❑**Single-scattering albedo** ϖ**:** defined as the ratio of the scattering cross section to the extinction (scattering plus absorption) cross section

Fundamental Parameters in Radiative Transfer – Single-Scattering Properties

- Extinction coefficient
- Single-scattering albedo ϖ:
- Scattering phase function $P(\boldsymbol{cos}\theta)$ or Asymmetry factor g

These parameters are functions of the incident wavelength, particle size and shape, and refractive index with respect to wavelength (real part m_r ~ scattering property; imaginary part m_i ~ absorption property)

Solve Radiative Transfer

- Exact Method: Discrete-ordinates method:
  - The solution of the radiative transfer equation with multiple scattering involves the following integral over angle

$$\int_{-1}^{1} I(\tau,\mu)d\mu = \int_{0}^{1} I^{+}(\tau,\mu)d\mu + \int_{0}^{1} I^{-}(\tau,\mu)d\mu$$

(Upward) (Downward)

  - Discretize the integration:

$$\int_{-1}^{1} I(\tau,\mu)d\mu = \sum_{j=1}^{m} w_j' I(\tau,\mu_j)$$

where w'_j is a quadrature weight, and μ_j is the discrete ordinate

- Approximations of discrete-ordinates method:
  - Two-stream approximation
  - Four-steam approximation

Two-Stream Approximation

- Two-stream approximation: replace the angular dependent quantities I(μ) by their **averages** over the upward and downward directions, respectively.

$$\int_{-1}^{1} I(\tau,\mu)d\mu = I^{+}(\tau) + I^{-}(\tau)$$

- A pair of coupled differential equations:

$$\overline{\mu}^{+}\frac{dI^{+}(\tau)}{d\tau} = I^{+}(\tau) - \frac{\varpi}{2}I^{+}(\tau) - \frac{\varpi}{2}I^{-}(\tau) - (1-\varpi)B$$

$$-\overline{\mu}^{-}\frac{dI^{+}(\tau)}{d\tau} = I^{-}(\tau) - \frac{\varpi}{2}I^{+}(\tau) - \frac{\varpi}{2}I^{-}(\tau) - (1-\varpi)B$$

Analytic solutions can be obtained

- Accuracy: errors generally greater than 5%-10% (Liou et al. 1988)

Four-Steam Approximation

- Four-Stream Approximation: two radiative streams in the upward direction and two streams in the downward direction based on the double Gaussian quadratures

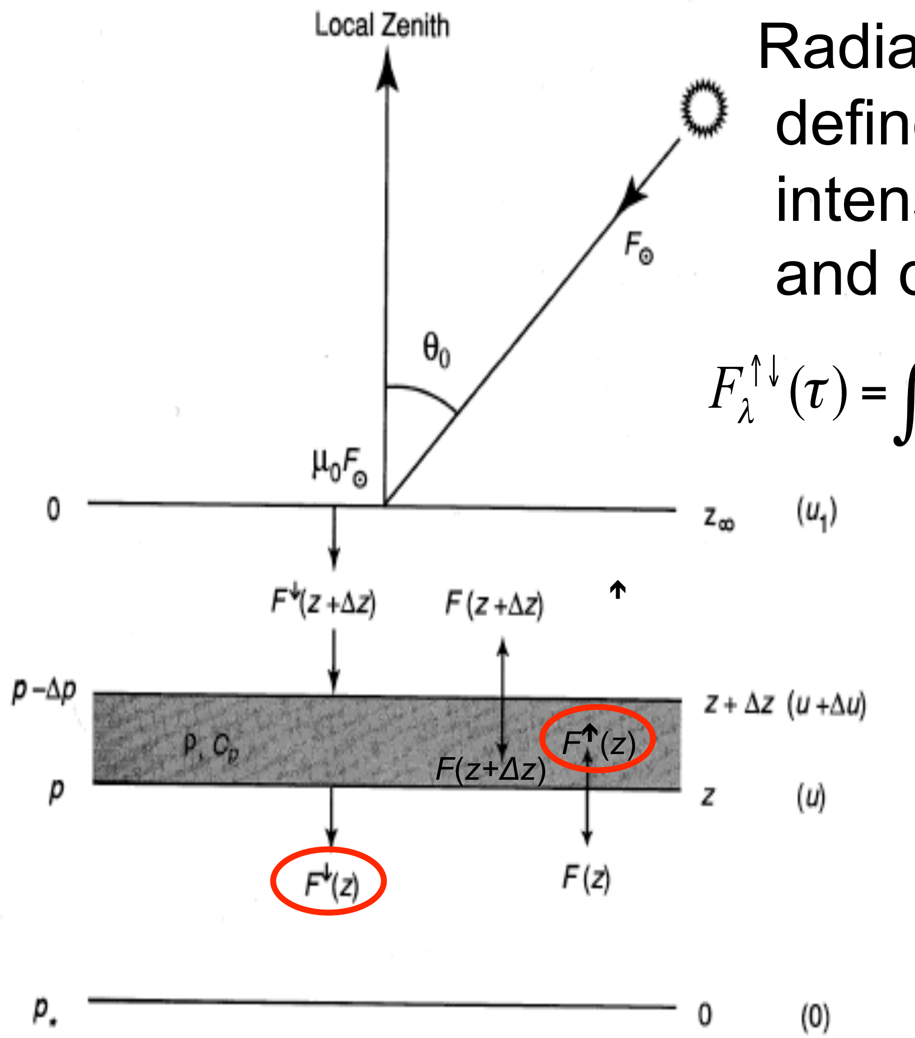
$$\int_{-1}^{1} I(\tau,\mu)\,d\mu = I^{+}_{1,2}(\tau) + I^{-}_{1,2}(\tau)$$

- Four equations; Analytic Solution

- Delta Function Adjustment for Forward Peak in the Phase Function: Similarity Principle; Simple, Convenient, and Accurate Method for Radiative Transfer

- Accuracy: errors within about 5% (Liou et al. 1988)

Radiative Fluxes



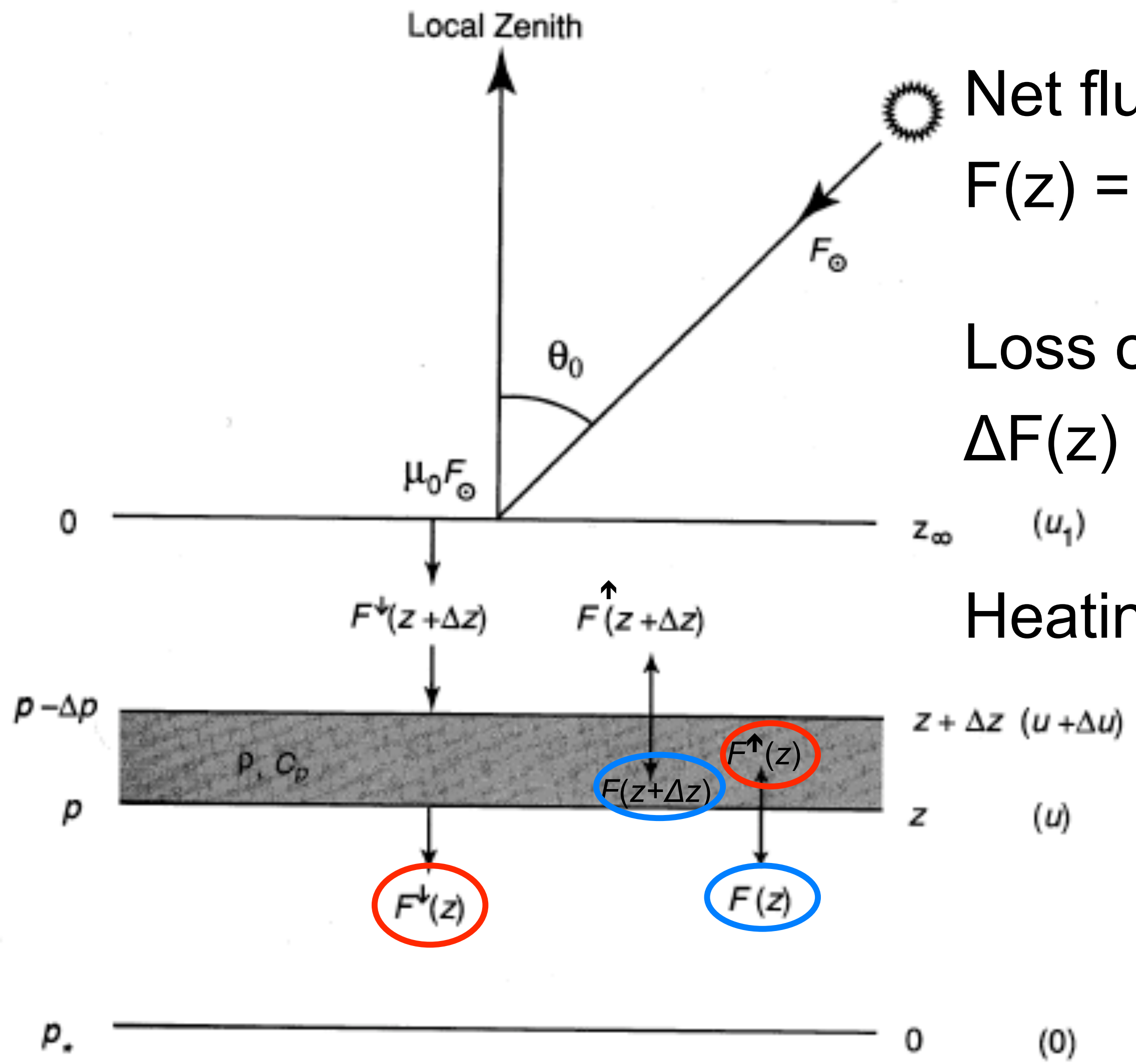
Radiative flux for a given wavelength is defined by the integration of radiation intensity over the hemisphere (upward and downward):

$$F_\lambda^{\uparrow\downarrow}(\tau) = \int_0^{2\pi}\int_0^{\pm 1} I_\lambda(\tau,\mu,\phi)\mu d\mu d\phi = 2\pi\int_0^{\pm 1} I_\lambda(\tau,\mu,\phi)\mu d\mu$$

Total flux for all wavelengths at a given height z:

$$F^{\uparrow\downarrow}(z) = \int_0^{\infty} F_\lambda^{\uparrow\downarrow}(\tau)d\lambda$$

Radiative Heating Rates



Net flux at a given height z:

$F(z) = F^{\downarrow}(z) - F^{\uparrow}(z)$

Loss of net flux by absorption:

$\Delta F(z) = F(z) - F(z+\Delta z)$

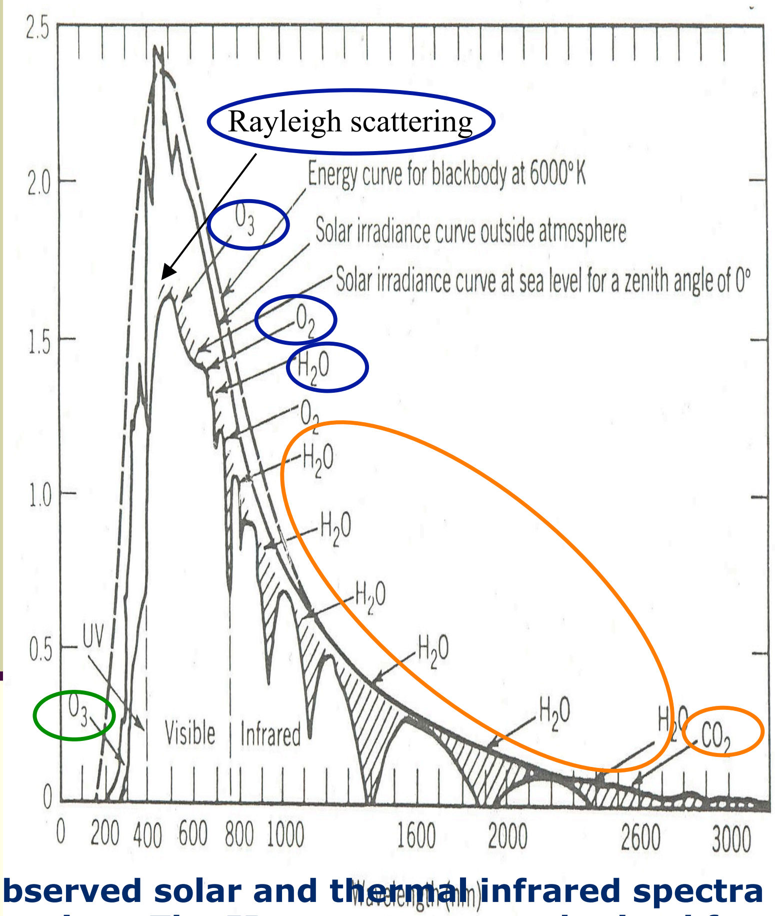
Heating rate for a layer Δz:

$$\boxed{\frac{\partial T}{\partial t} = -\frac{1}{\rho C_P}\frac{\Delta F(Z)}{\Delta Z}}$$

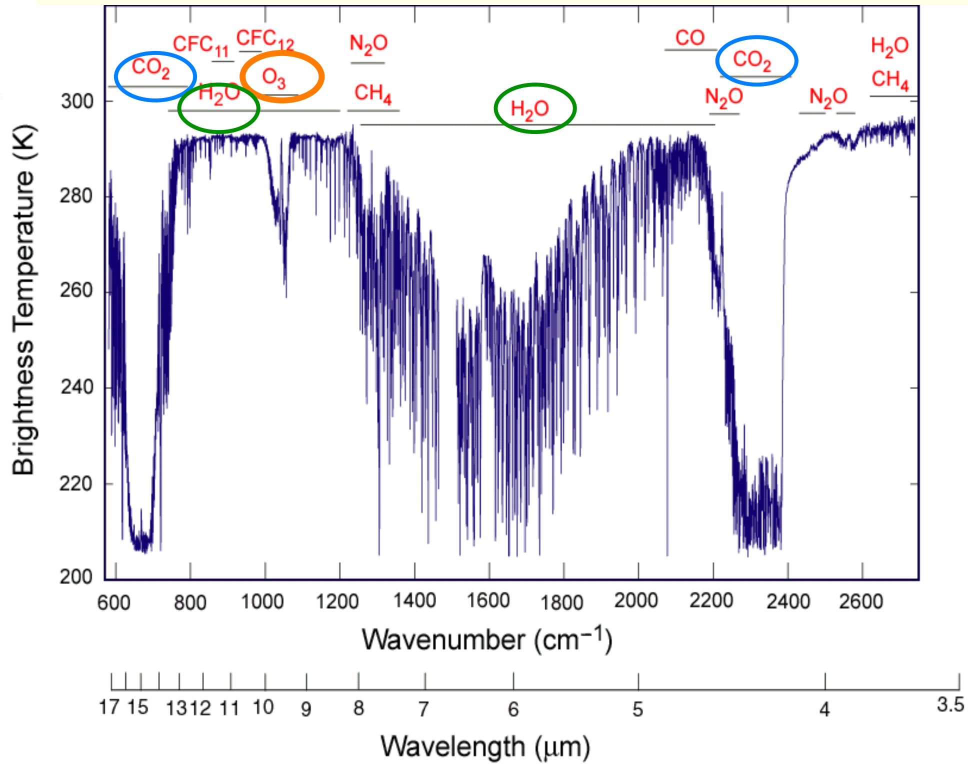
Radiation & Other Physical Processes

- **Gaseous Absorption**
- **Cloud-Radiation Interaction**
- **Aerosol-Cloud-Radiation Interactions**

Gaseous Absorption



Thermal IR Radiation Spectrum



Observed solar and thermal infrared spectra displaying principal absorbing gases and their spectral location. The IR spectrum was obtained from the Scanning High-resolution Interferometer Sounder (S-HIS), which measured the emitted thermal radiation between 3.3 and 18 μm, onboard the NASA ER-2 aircraft over the Gulf of Mexico southeast of Louisiana, on April 1, 2001 (courtesy of Allen Huang and Dave Tobin of the University of Wisconsin).

Absorption Coefficient

- Function of wavenumber, pressure, and temperature

$$k(\nu, p, T) = \sum_i S_i(T) f_i(\nu, p, T)$$

S_i: Line intensity for the ith absorption line

f_i: Normalized line shape

- Look-up tables

- Parameterization (Chou and Kouvaris 1986)

$$\ln k(\nu, p, T) = \sum_{n=0}^{2} a_n(\nu, p)(T - 250)^n$$

Line-by-Line → Band Model → k-Distribution

- Line-by-Line: Divide the full spectrum into nearly monochromatic wavenumber intervals ($\delta\nu = 10^{-4} \sim 10^{-3}$ cm^{-1}) and calculate and integrate the atmospheric fluxes and heating rates for each of these intervals – Very time consuming.

- Band Model: Simplify the integration over wavenumber using an equivalent width, defined as the width of an infinitely strong line of rectangular shape, which would have been the same as the actual absorption of a single line

- k-Distribution: Replace the wavenumber integration by grouping the absorption coefficient

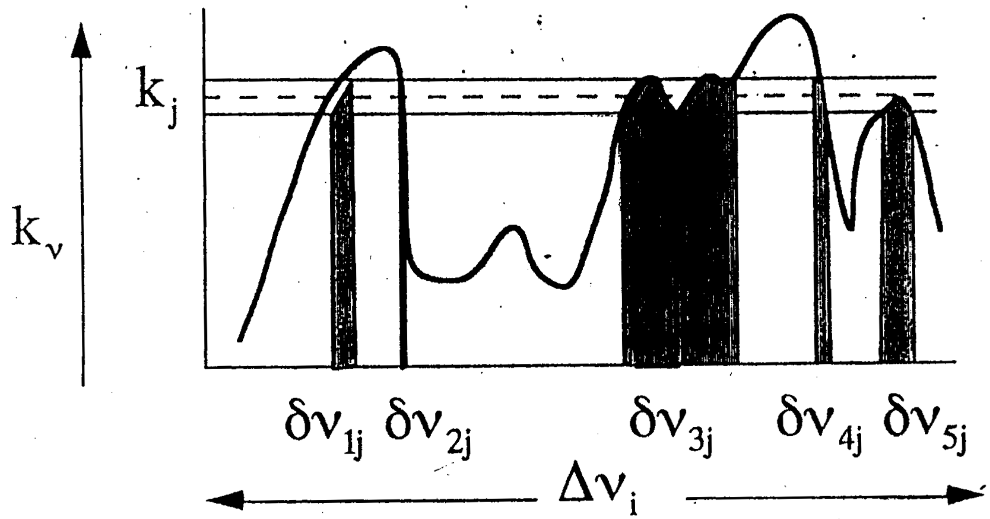
- Correlated k-distribution: extension of k-distribution to nonhomogeneous atmosphere by assuming a simple correlation of k-distributions at different temperatures and pressures

k-Distribution Technique

$$I_i \equiv \frac{1}{\Delta\nu_i} \int I_\nu \, d\nu$$

Since B_ν is $\approx$ constant in an interval,

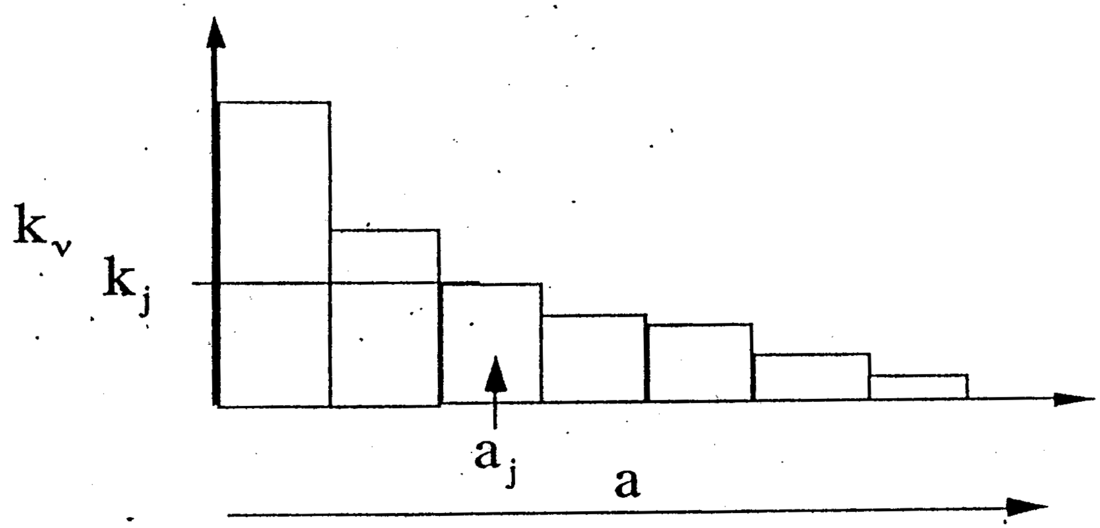
I_ν is the same when k_ν is the same



$$a_j = \sum_k \frac{\delta\nu_{kj}}{\Delta\nu} = \text{fraction of } \Delta\nu_i \text{ for which } k_\nu \cong k_j$$

k-Distribution Technique

Rearrange k_ν as a probability histogram



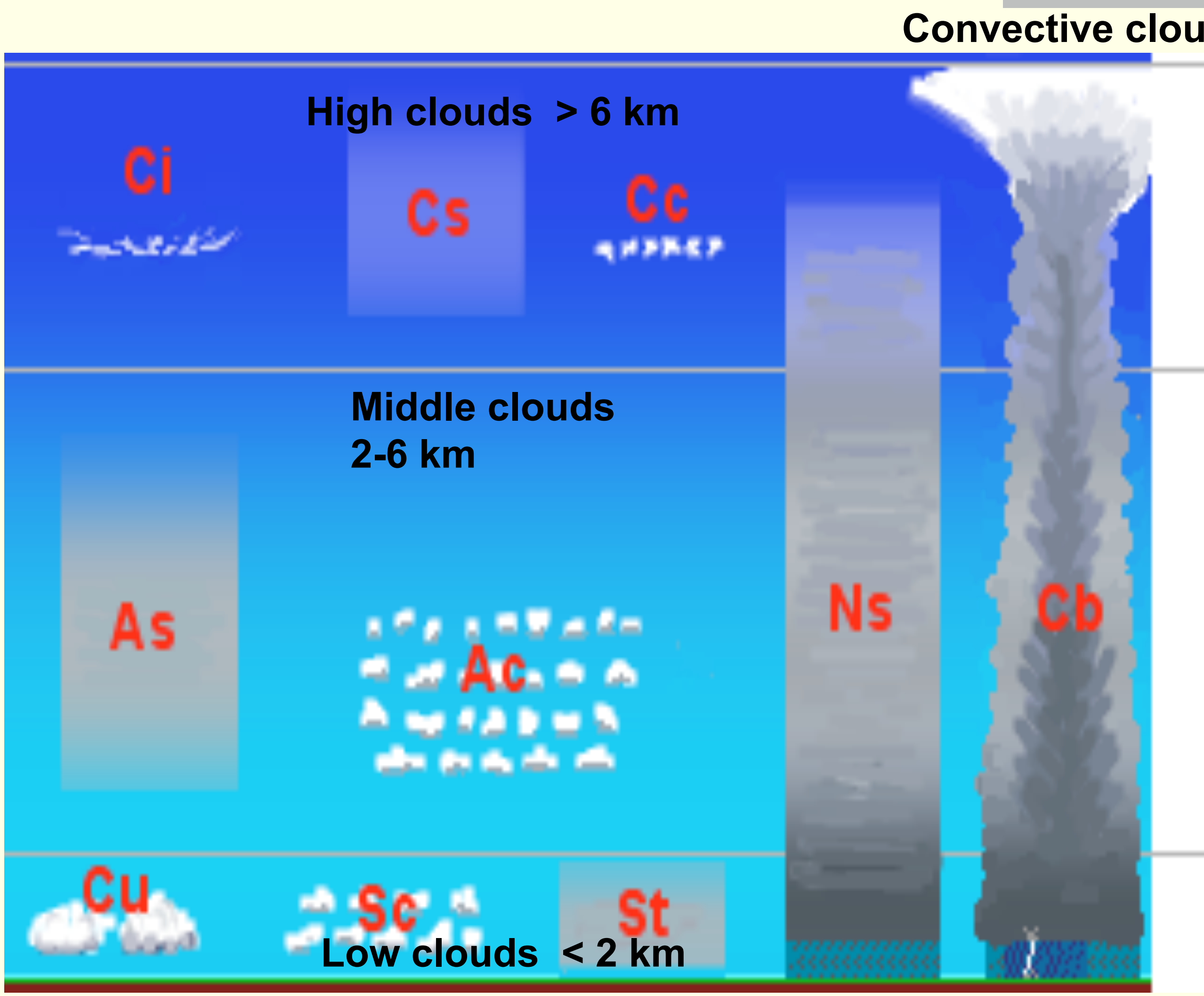
Now we need only solve M monochromatic problems

$$I_i = \sum_{j=1}^{M} a_j \, I(k_\nu = k_j)$$

A minimum number of quadrature points are required for the integration

One can obtain the histogram from:
(a) line-by-line calculations
(b) exponential sum fitting of band model transmission functions

Cloud-Radiation Interaction - Cloud Classification



Cloud/Radiation Processes in a Climate Model

Role of Cloud/Radiation in Climate

- Clouds provide one of the most important mechanisms for the vertical redistribution of momentum, moisture, and sensible and latent heat, and influence the coupling between the atmosphere and the surface as well as the radiative and dynamical-hydrological balance (Arakawa 1975)
- Clouds produce radiative heating and cooling that are comparable to the latent heat (e.g., Ramanathan 1987)
- Radiation, latent heating, and small-scale transports are of equal importance in cloud-climate problem, and these three cloud processes strongly interact with each other (e.g., Randall et al. 1989)

Difficulties in Cloud/Radiation Modeling

- Clouds interact with radiative, dynamical, and hydrological processes on various time scales. Relationship between clouds and other physical processes is still not clear
- Clouds are normally not resolved on the computational grid in GCMs or climate models

  → Cloud Fraction/Cloud Overlap problem
- Calculation of radiative heating/cooling in clouds is difficult due to the difficulties involved in the parameterizations for cloud single-scattering properties, especially ice clouds

Radiative Transfer in Clouds

- Reflect, absorb, and transmit solar and IR radiation; Emit IR radiation according to the temperature structure within the clouds

- Scattering and absorption of radiation by spherical water droplets can be exactly solved by the Mie theory.

- Ice clouds are largely composed of nonspherical particles. Unlike the scattering of light by spherical water droplets which is governed by Mie theory, the light scattering and absorption of the hexagonal ice crystals are extremely difficult to determine.

- Calculation of the single-scattering properties for clouds require a detailed light scattering program and information about particle size distributions and are very time consuming.

- Look-up tables or parameterizations

Mean Effective Radius/Size

• Characteristics of scattered light from clouds depend on the droplet/ particle size distribution *n(r)*

• The first parameter describing size distribution: some measure of the mean size

❑ Water clouds: spherical droplets scatter an amount of light in proportion to their cross-sectional area. A mean effective radius is defined using droplet cross section as a weight factor:

$$r_e = \int r\pi r^2 n(r)dr \Big/ \int \pi r^2 n(r)dr$$

❑ Analogous to the mean effective radius, a mean effective size representing ice crystal size distribution is defined based on the volume- and area-weighted ice crystal size distribution:

$$D_e = \int D \cdot LDn(L)dL \Big/ \int LDn(L)dL$$

Extinction Coefficient/Optical Depth

- Water clouds: From the definition of liquid water content (LWC), liquid water path (LWP) which is the vertically integrated LWC is given by:

$$LWP = \Delta z \cdot LWC = \Delta z \cdot \rho V = \Delta z \cdot \rho \cdot \frac{4}{3}\pi \int r^3 n(r) dr$$

- Optical depth for a given droplet size distribution is defined by

$$\tau = \Delta z \int \sigma_e n(r) dr = \Delta z \int Q_e \pi r^2 n(r) dr$$

Where the extinction cross section $\sigma_e = Q_e \pi r^2$, and Q_e is the efficiency factor which is a function of the droplet radius, wavelength and refractive index. For visible wavelength, $Q_e \cong 2$ for cloud droplets.

- Thus, the mean effective radius can be related to LWP and optical depth

$$\boxed{r_e \cong \frac{3}{2} LWP / \tau \qquad or \qquad \tau = \frac{3}{2} LWP / r_e}$$

Parameterization of Cloud Single-Scattering Properties

Following the similar procedure for ice clouds, in numerical models, cloud single scattering properties are normally parameterized in terms of LWP/IWP (or LWC/IWC) and r_e/D_e based on a number of droplet size distributions

Water

$$\tau(\lambda, x, y, z) = LWP(x, y, z)\left[a(\lambda) + b(\lambda)/r_e(x, y, z)\right]$$

$$1 - \varpi(\lambda, x, y, z) = c(\lambda) + d(\lambda) r_e(x, y, z)$$

$$g(\lambda, x, y, z) = e(\lambda) + f(\lambda) r_e(x, y, z)$$

Ice

$$\tau(\lambda, x, y, z) = IWP(x, y, z)\sum_{n=0}^{N} a_n(\lambda)/D_e^n(x, y, z)$$

$$1 - \varpi(\lambda, x, y, z) = \sum_{n=0}^{N} b_n(\lambda) D_e^n(x, y, z)$$

$$g(\lambda, x, y, z) = \sum_{n=0}^{N} c_n(\lambda) D_e^n(x, y, z)$$

- Coefficients are determined from numerical fittings based on detailed light scattering and absorption calculations for a range of droplet/particle size distributions, and also shapes for ice particles.
- LWP/IWP are predicted or diagnosed from numerical models
- r_e and D_e are prescribed or parameterized

More on Cloud-Radiation Interaction

- **Cloud Vertical Overlap**

  Not resolved on the computational grid in GCMs or climate models
  → Cloud Fraction/Cloud Overlap problem

- **Ice Crystal Size Effect**

  - Ice crystal size distribution is a function of temperature (Heymsfield and Platt 1984)
  - Significantly interact with radiation field (Gu and Liou 2000; Wu et al. 2000)

- **Cloud Inhomogeneity Effect**

  - Clouds are frequently finite in nature and display substantial variabilities (Minnis et al. 1993)
  - Cloud inhomogeneity plays a significant role in the heating rate profile (Gu and Liou 2001)
  - Cloud inhomogeneity leads to a smaller solar albedo (Cahalan et al. 1994; Kogan et al. 1995)
  - Adjust the optical depth to account for cloud horizontal inhomogeneity effect in numerical models

Cloud Vertical Overlap

Random Overlap:

- Divide the sky into sectors within which the cloud amount is either 0 or 1
- Radiative fluxes are calculated for each sector and then weighted by the respective cloud amount to obtain the all-sky gridbox fluxes
- 2^n (n= total number of layers of clouds) possible cloud configurations
- Neglect cloud geometry association and tend to overestimate total cloud cover

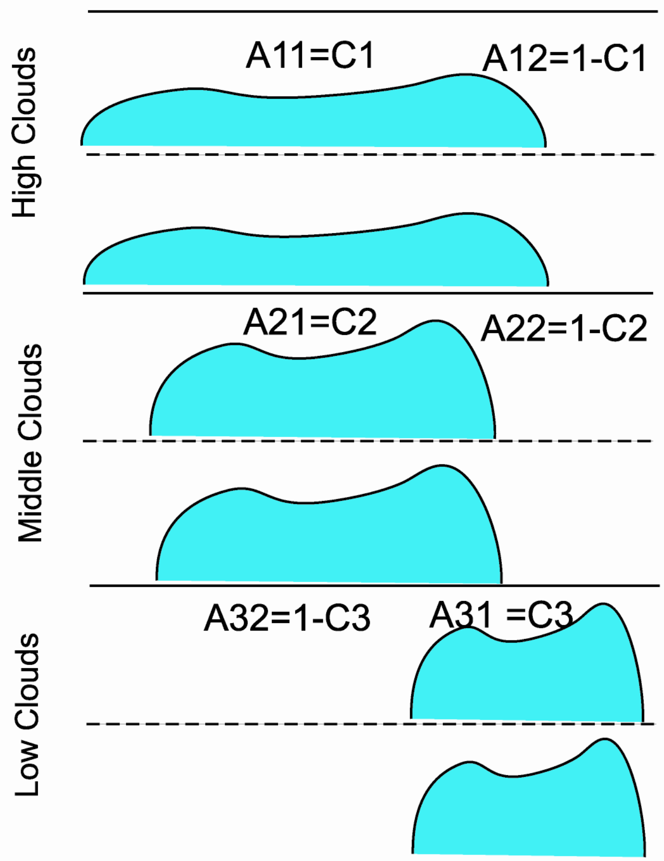
Maximum Overlap

- Clouds are closely associated and stack on each other
- 2 possible cloud configurations
- Underestimate total cloud cover

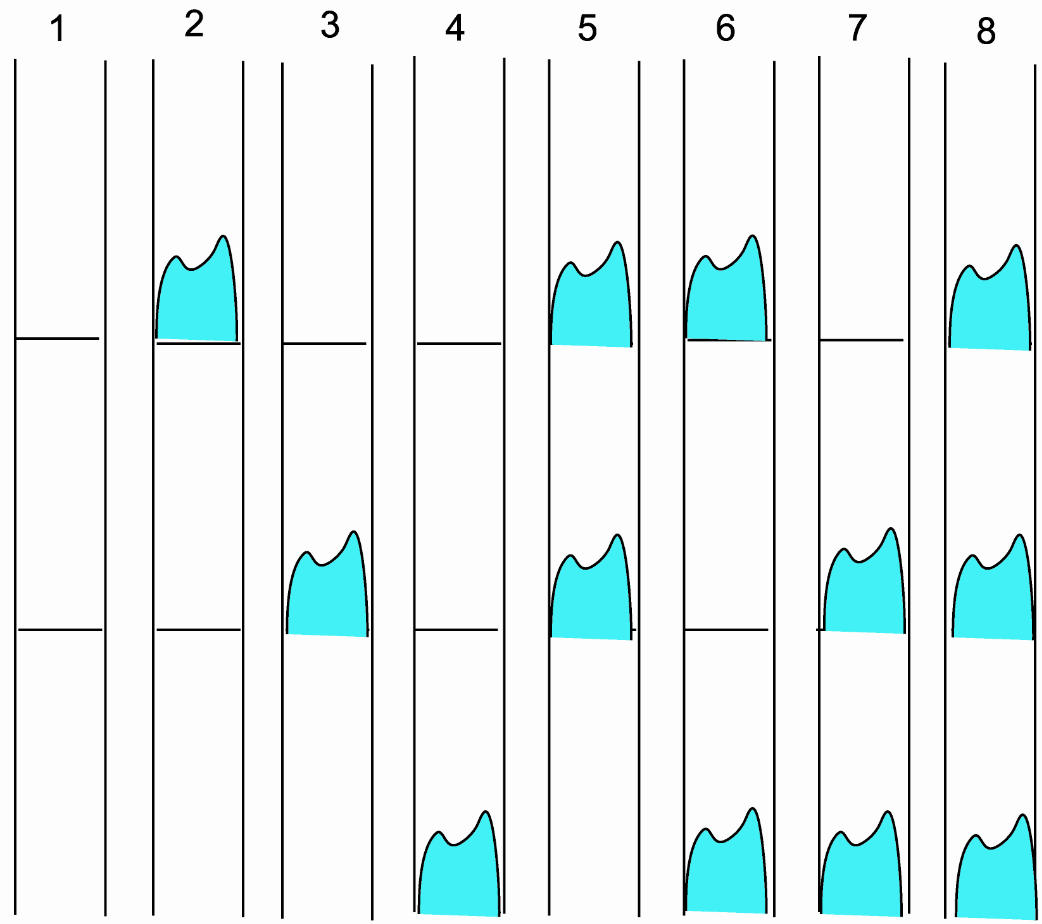
Maximum/Random Overlap:

- Clouds are grouped as regions (e.g., low, middle, and high; adjacent or discrete)
- Maximum overlap within each region
- Random overlap among clouds of different regions
- 2^N (N= total number of cloud regions) possible cloud configurations

Vertical Profile of Cloud Amounts



Possible Cloud Configurations



Fractional Area for Each Section

Area 1 = A12 * A22 *A32
Area 2 = A11 * A22 *A32
Area 3 = A12 * A21 *A32
Area 4 = A12 * A22 *A31

Area 5 = A11 * A21 *A32
Area 6 = A11 * A22 *A31
Area 7 = A12 * A21 *A31
Area 8 = A11 * A21 *A31

Observed Ice Crystal Size and Shape

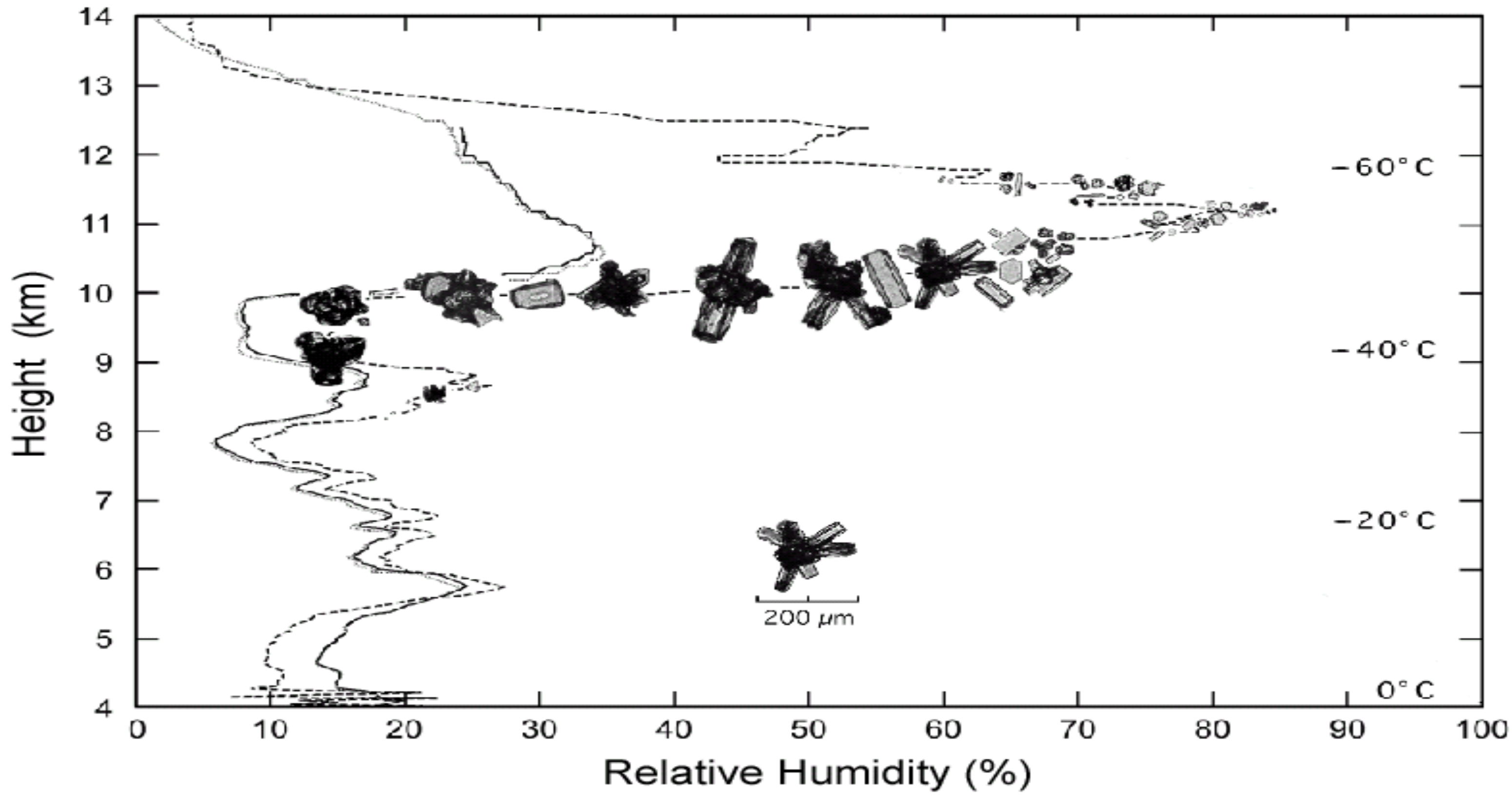
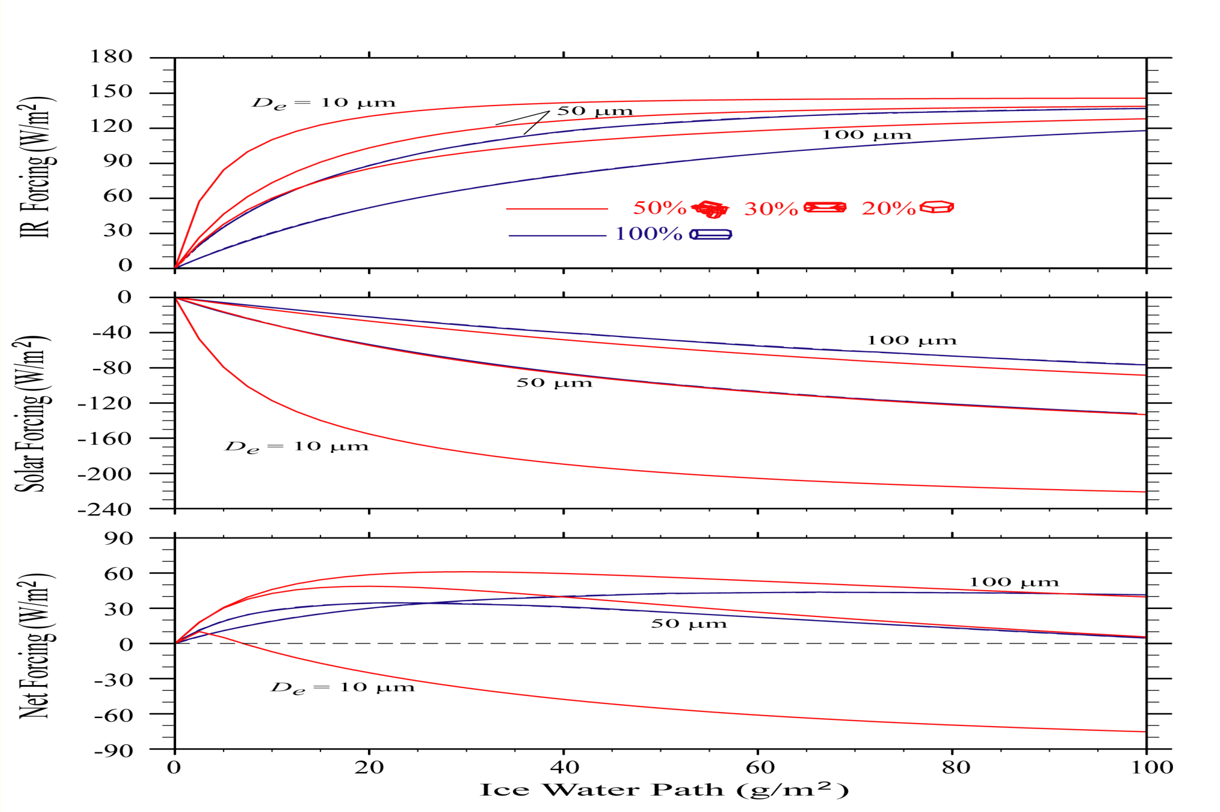


Fig. 5.3 Ice crystal size and shape as a function of height and relative humidity captured by a replicator balloon sounding system in Marshall, Colorado on November 10, 1994. The relative humidity was measured by cryogenic hygrometer (dashed line) and Vaisala RS80 instruments (solid line and dots). Also shown is temperature as a function of height (courtesy of Andrew Heymsfield of the National Center for Atmospheric Research).

Effect of Ice Crystal Size & Shape on Radiative Forcing

TOA: $CRF_{s,ir} = F_{s,ir(clear)} - F_{s,ir(cloud)}$, $CRF_{net} = CRF_{IR} + CRF_s$
(upward IR and solar fluxes)



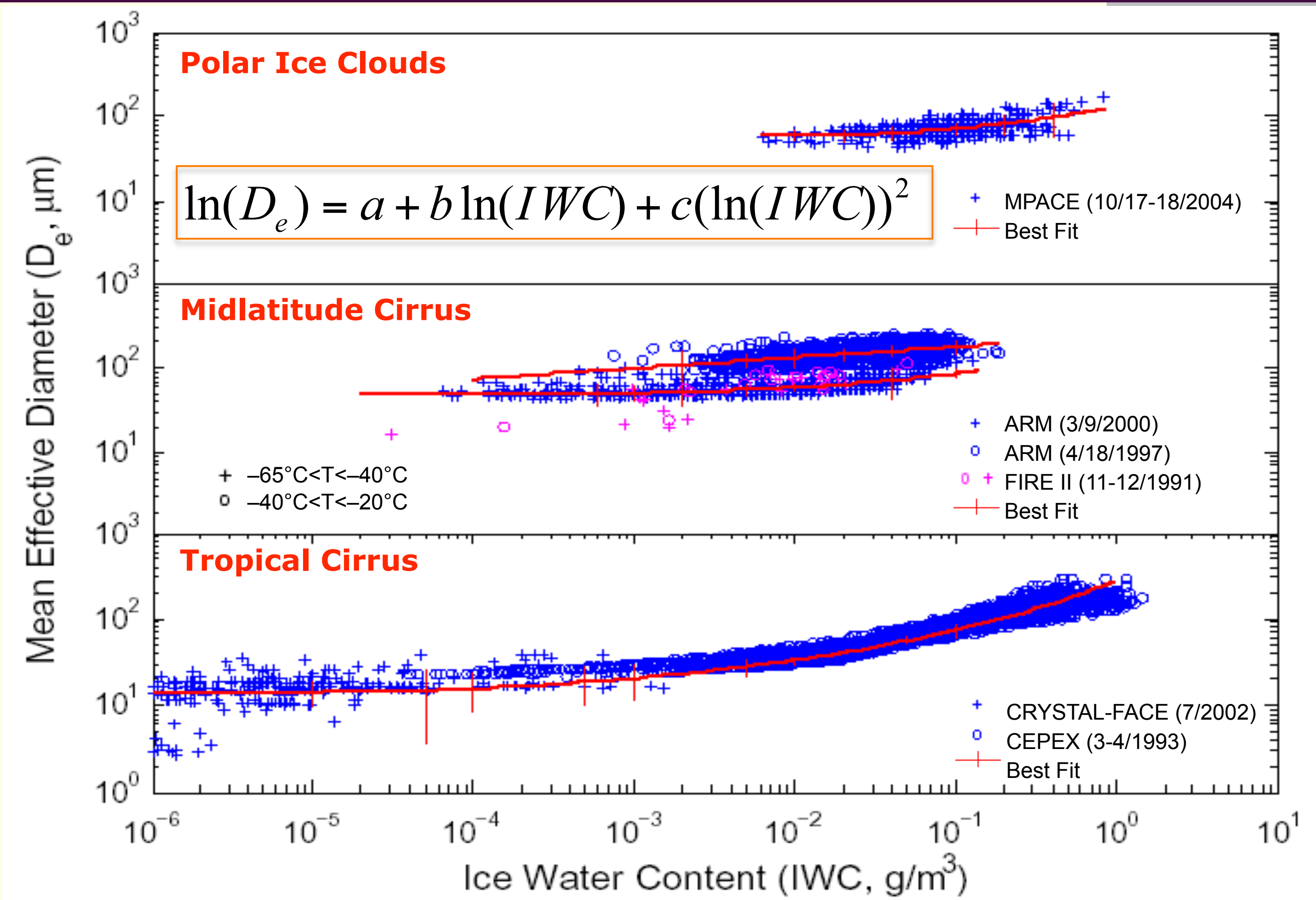
IR, Solar and net radiative forcings for cirrus clouds as functions of mean effective size D_e and ice water path for two ice crystal shape distributions in the standard atmospheric condition. The solar constant, solar zenith angle, surface albedo, cloud base height, and cloud thickness used in the calculations are 1366 W/m², 60°, 0.1, 9 km, and 2 km, respectively.

Parameterization of D_e

- A conventional approach has been to prescribe a mean effective ice crystal size in numerical models (e.g., Köhler 1999; Ho et al. 1998; Gu et al. 2003)

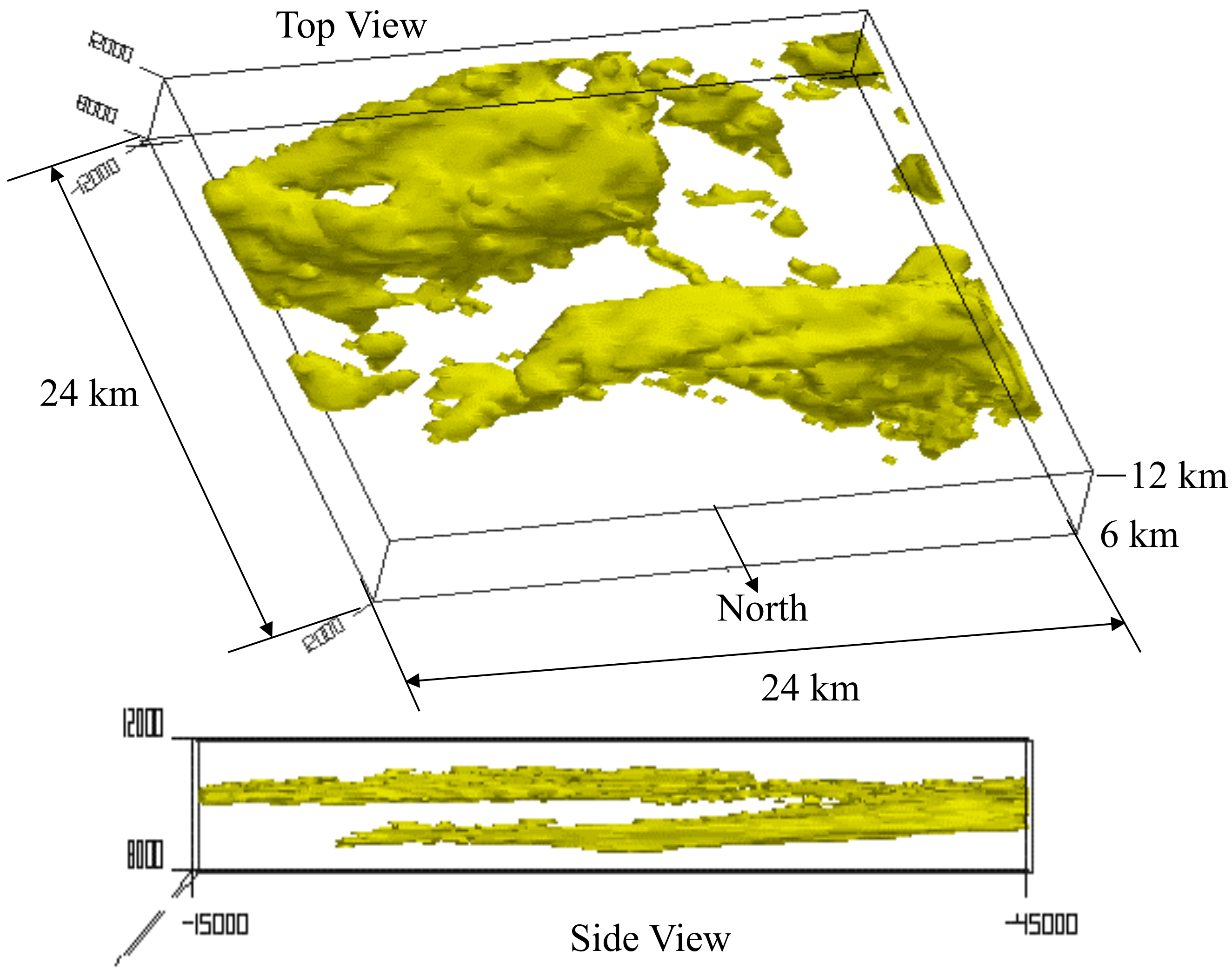
- Use IWC and/or temperature produced from models to determine a mean effective ice crystal size (Kristjánsson et al. 2005; Gu and Liou 2006; Ou and Liou 1995; Ou et al. 1995; McFarquhar et al. 2003; Liou et al. 2008).

IWC & Ice Crystal Effective Size (De) Correlation for Model Application



Mean correlation curves with standard deviations (vertical bars) for IWC and De for the tropics (bottom panel), midlatitude (middle panel), and Arctic region (top panel) (Liou, Gu et al. 2008)

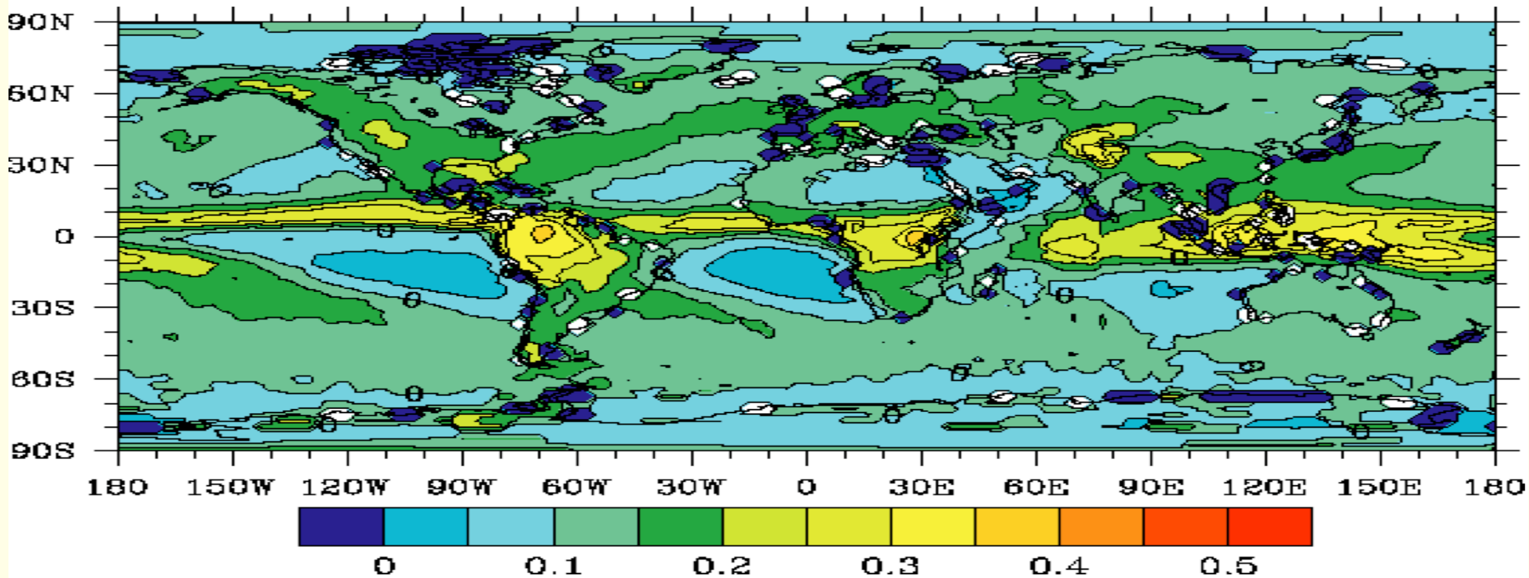
Cloud Inhomogeneity



3-D Cirrus Observations from Lidar
September 14, 1995 (E. Eloranta, Private communication)

Representing Cloud Inhomogeneity Effect in Climate Models

Inhomogeneity Parameter (χ) for High-level Clouds (Annual Mean, ISCCP)



- **The effects of cloud inhomogeneity on solar radiative transfer** can be accounted for approximately by re-scaling the area-mean optical parameters:

$$\tau_r = (1 - \chi)\,\tau_m$$

where τ_m is the linear average of the varying optical thickness over the area that may be predicted from GCM, and τ_r is the "radiative-average" that we want to use in the radiative transfer calculation to give the correct cloud albedo.

Commonly used value: $1 - \chi = 0.7$ or $\chi = 0.3$

Aerosol & Climate

<u>**Role of Aerosol in Climate**</u>

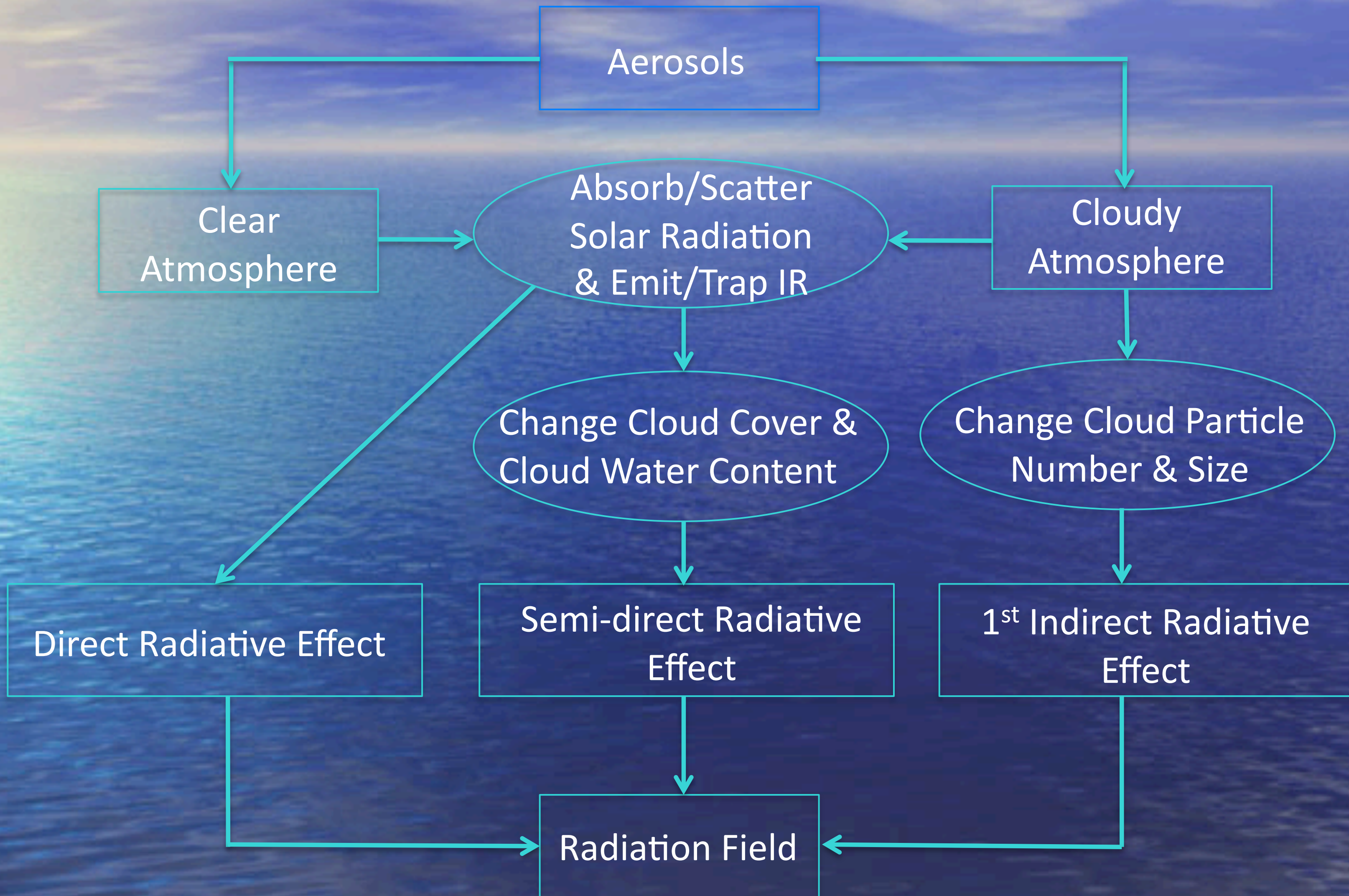
- Aerosol particles are an important atmospheric constituent that interacts with solar and IR radiation, and hence influence the energy balance of the earth-atmosphere system – Aerosol Direct Effect
- Water soluble aerosols can act as condensation or ice nuclei and initiate or modify the cloud formation process, which also attenuate atmospheric radiation – Aerosol Indirect Effect
- Aerosols are major contributors to thermodynamic exchange processes in the atmosphere

<u>**Difficulties in Aerosol Modeling**</u>

- The aerosol composition in the atmosphere is very complex and our knowledge of the physical and chemical properties of aerosols is still poor
- Atmospheric aerosols stem from localized and sparse sources but are transported by large general circulation – High spatial and temporal variability
- Calculation of radiative forcing due to aerosols is more complex than that due to gases due to their intricate interactions with other atmospheric physical processes.

(d'Almeida 1991)

Aerosol-Cloud-Radiation Interactions

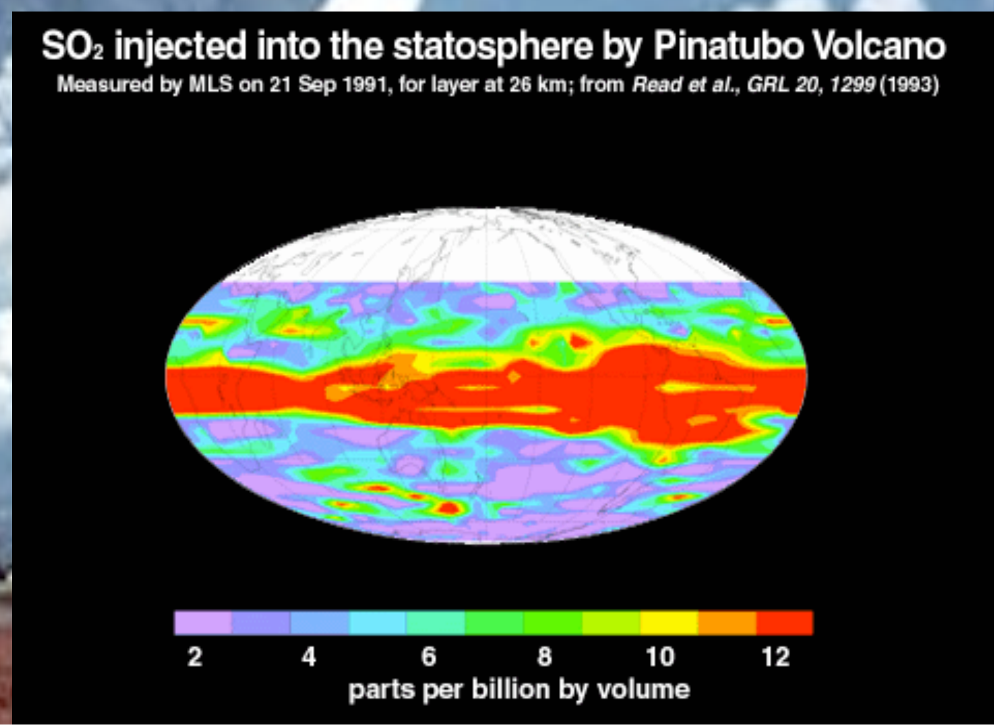


Aerosol Direct Climate Effect

- Reduce the amount of sunlight reaching the Earth's surface, thereby exerting a cooling influence over large regions
- Certain aerosols (most notably soot emitted from fossil fuel combustion, and wind-blown desert dust) absorb sunlight, resulting in a significant heating of the atmosphere in a manner somewhat analogous to the action of greenhouse gases

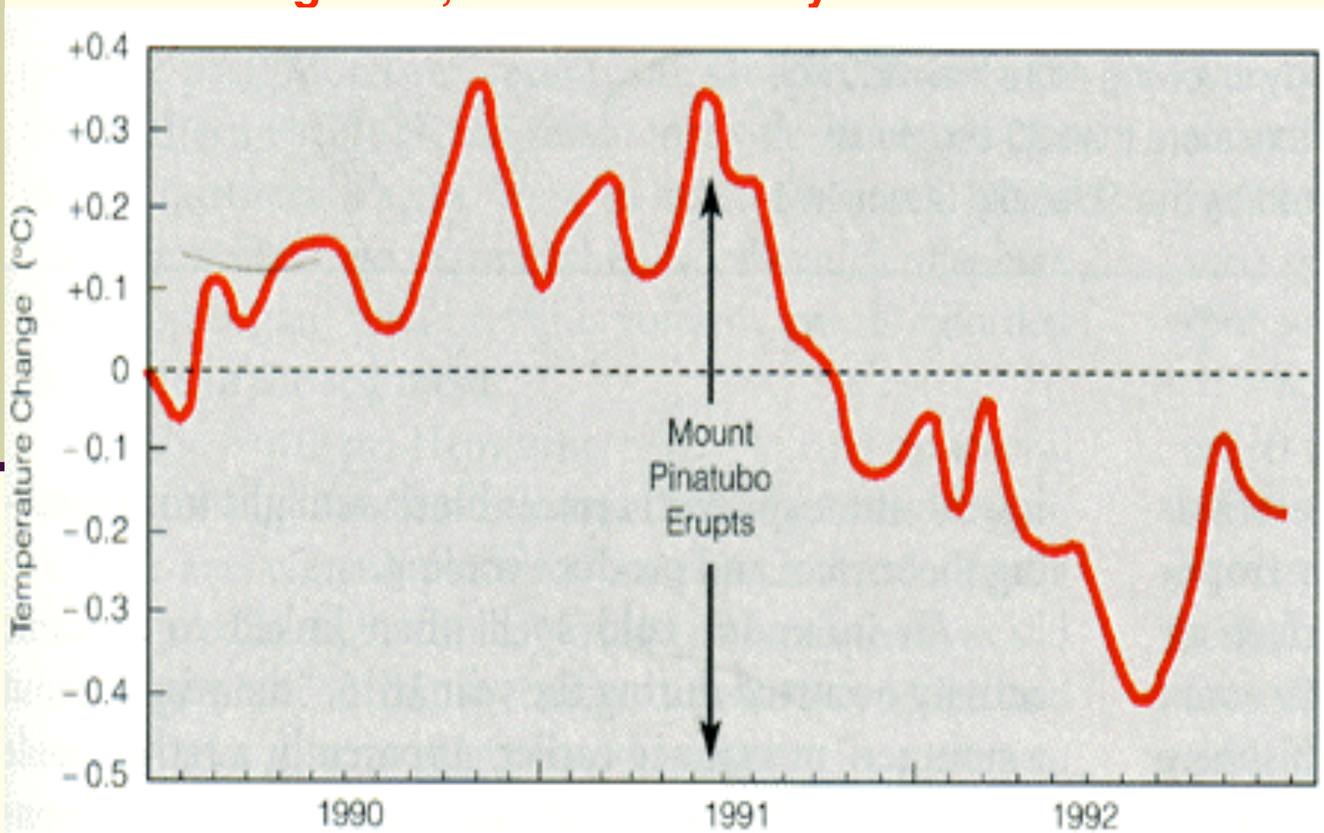
Volcano Eruption

Volcanoes spew enormous amounts of material into the atmosphere, including sulfur dioxide, which is transformed in the atmosphere into sulfate aerosols. These aerosols reflect sunshine.



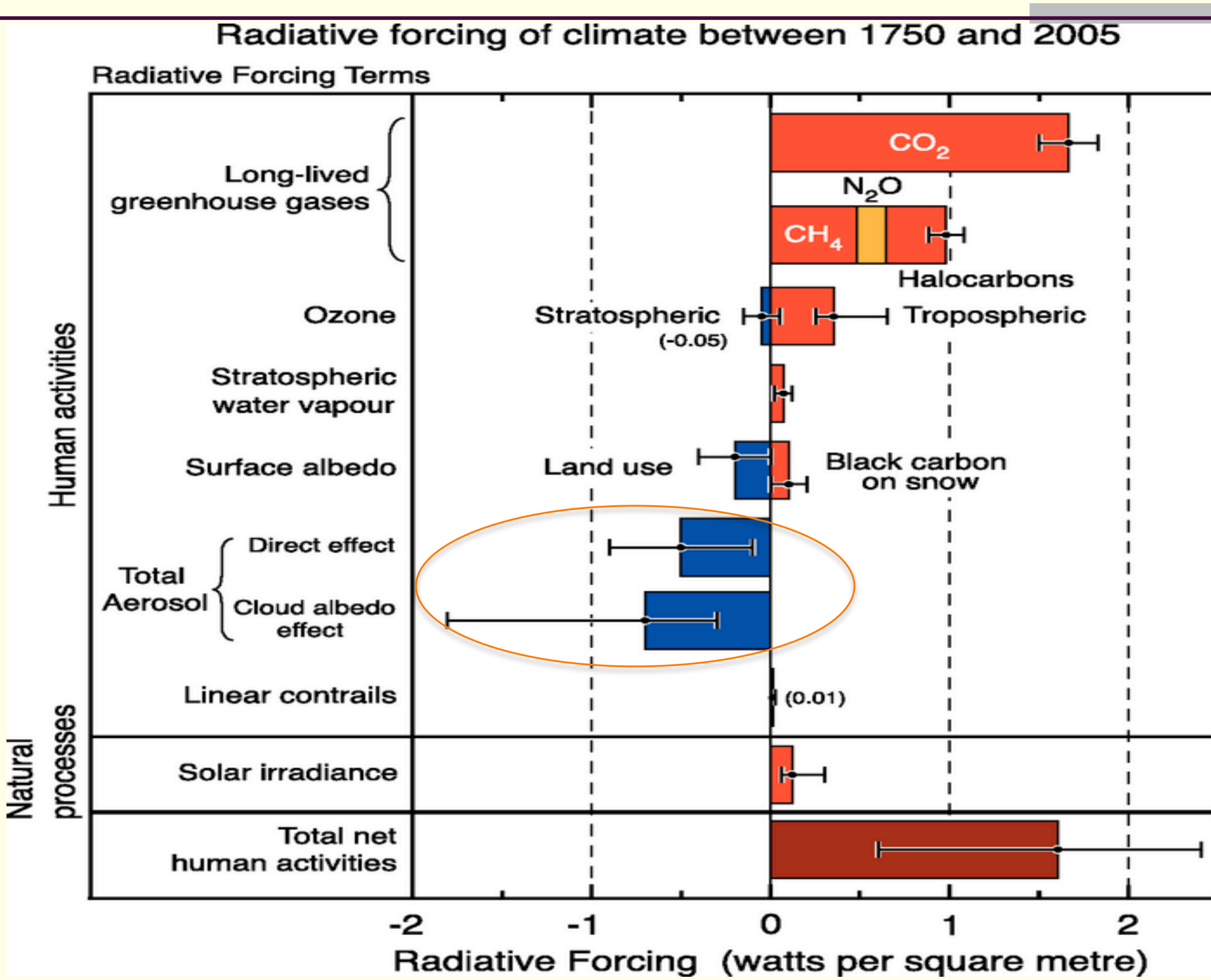
Radiative Forcing by Pinatubo Eruption (Sulfate)

In the case of the 1991 Mt. Pinatubo eruption, the sulfate aerosol cloud resulted in a radiative forcing of about -0.5 W/m^2, which lasted for a few months. This coincided with a global cooling on the order of a few tenths of degree C, which lasted a year or so.



Observed global air temperature anomaly associated with the Mt. Pinatubo eruption in 1991. (from Ahrens, Meteorology Today)

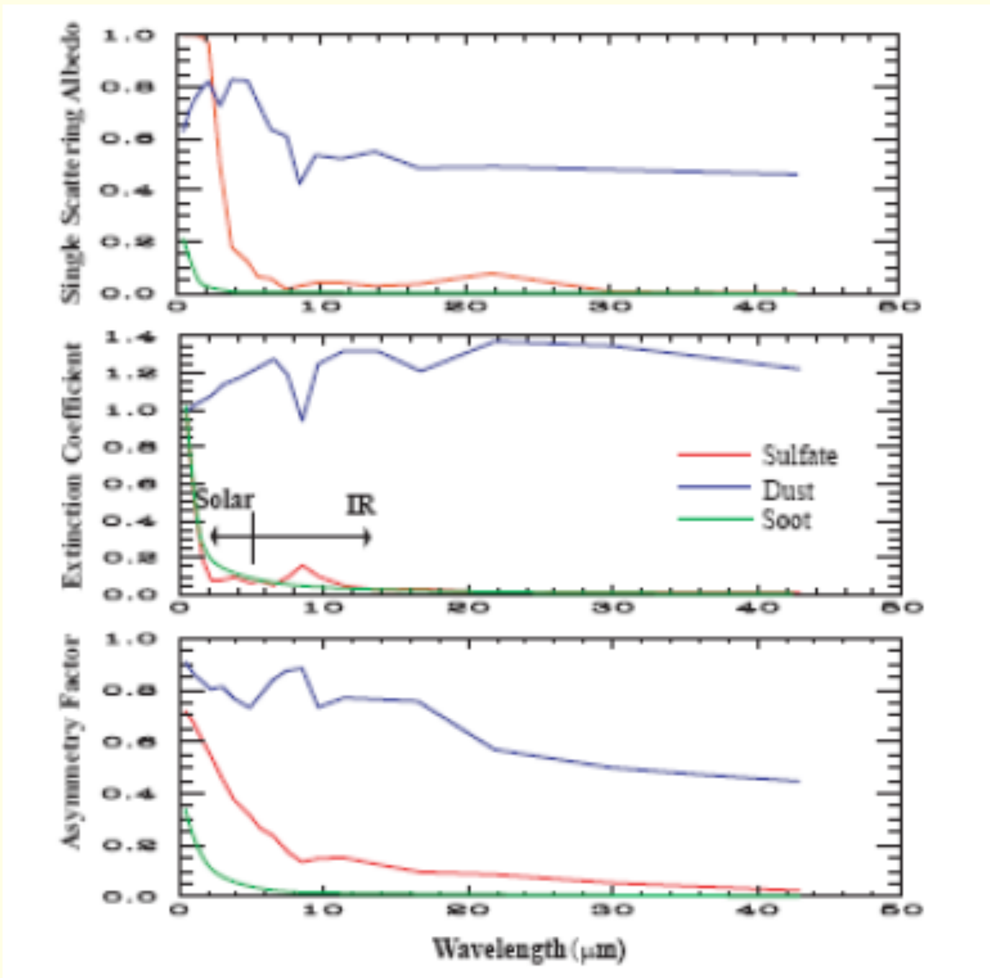
Radiative Forcing - IPCC Report 2007



Basic Factors that Determine the Aerosol Radiative Forcing

- **Aerosol Optical Depth**: affects the magnitude of the aerosol radiative forcing
- **Aerosol Single-Scattering Properties**: determine the sign of the forcing
- **Aerosol Compositions**: overall aerosol forcing depends on the composition of the aerosols and their mixing state.
- **Aerosol Vertical Distributions**: influence the layer AOD and compositions →affect the radiative forcings at the TOA and the surface → modify the vertical profile of radiative heating in the atmosphere → changing atmospheric stability and convection

Aerosol Single-Scattering Properties



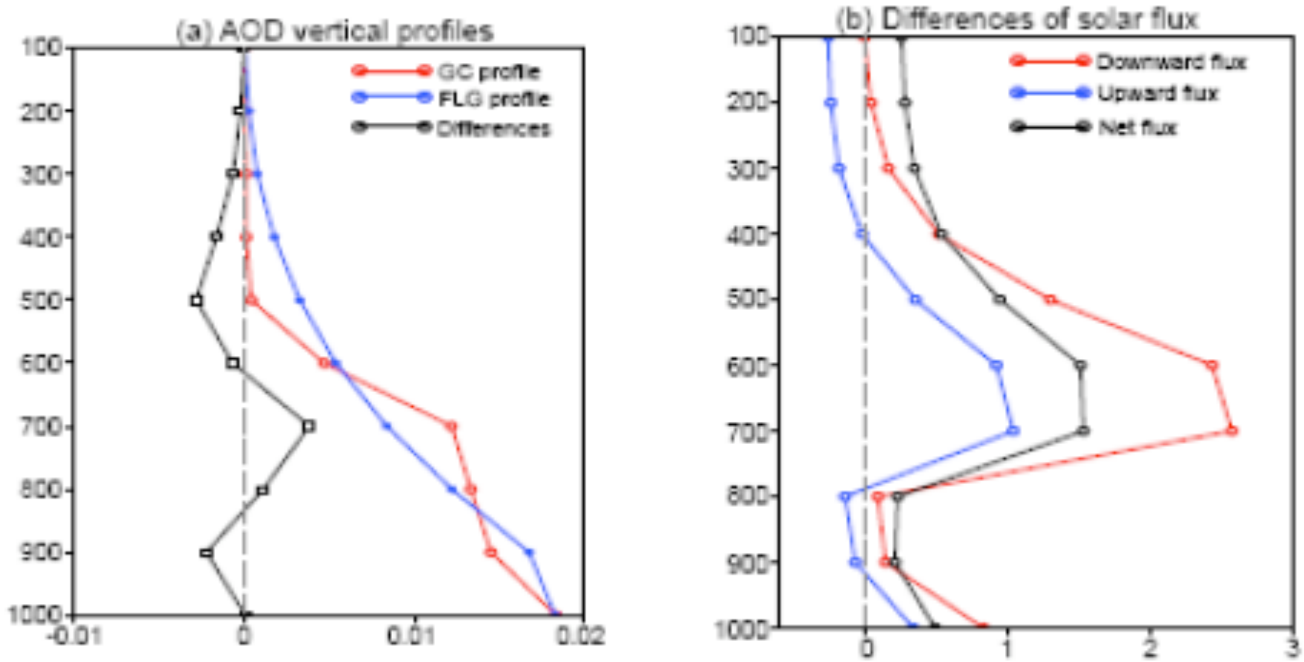
Single-scattering properties for spherical aerosols: Computed from the Lorenz-Mie theory in which the humidity effects are accounted for

Sulfate: Scatter

BC: Highly absorptive

Large Dust: Scatter and absorb

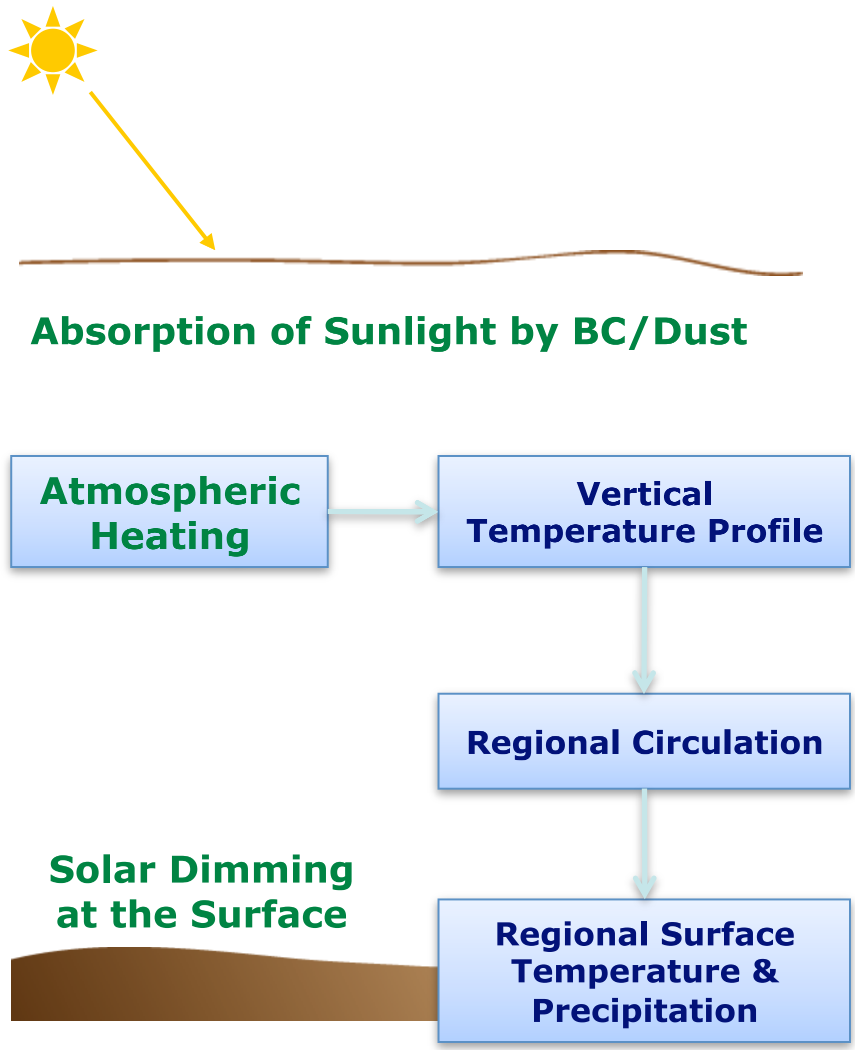
Effect of Aerosol Vertical Profile



(Zhang, Gu et al. 2013)

With the same column-integrated AOD, the use of different aerosol profiles produces substantial changes in estimated radiative fluxes

Aerosol Direct Radiative Forcing & Regional Climate



Aerosol Semi-direct Effect
- Aerosol Absorption & Cloud

Hansen et al (1997):

aerosol absorption of sunlight → **heats the lower troposphere** → **reduces large-scale cloud cover**

- **Could be comparable to the aerosol direct and/or indirect effects** (Gu et al. 2006)
- **Great complexity in the cloud response to the heated layer, depending on the relative position of the absorbing aerosol layer to the cloud** (Johnson et al. 2004; Fan et al. 2008), **underlying surface, etc.** (Allen and Sherwood 2010, Koch and Genio 2010)
- **At present, the semi-direct effect is poorly understood and its role in cloud distribution and regional climate change is unclear** (Koch and Del Genio 2010).

Aerosol Indirect Effect (Counter Greenhouse Effect)

- **Solar Albedo Effect (1st indirect):**

  Air pollution → more condensation nuclei (aerosols) → competing for water vapor → more smaller water droplets and more clouds → reflect more sunlight → cooling

- **Precipitation Reduction (2nd indirect) because of smaller droplets:**

  More aerosol → more droplets → less coalescence → less rain → higher IWP → higher cloud fraction → longer lifetime

Dust-Ice Clouds

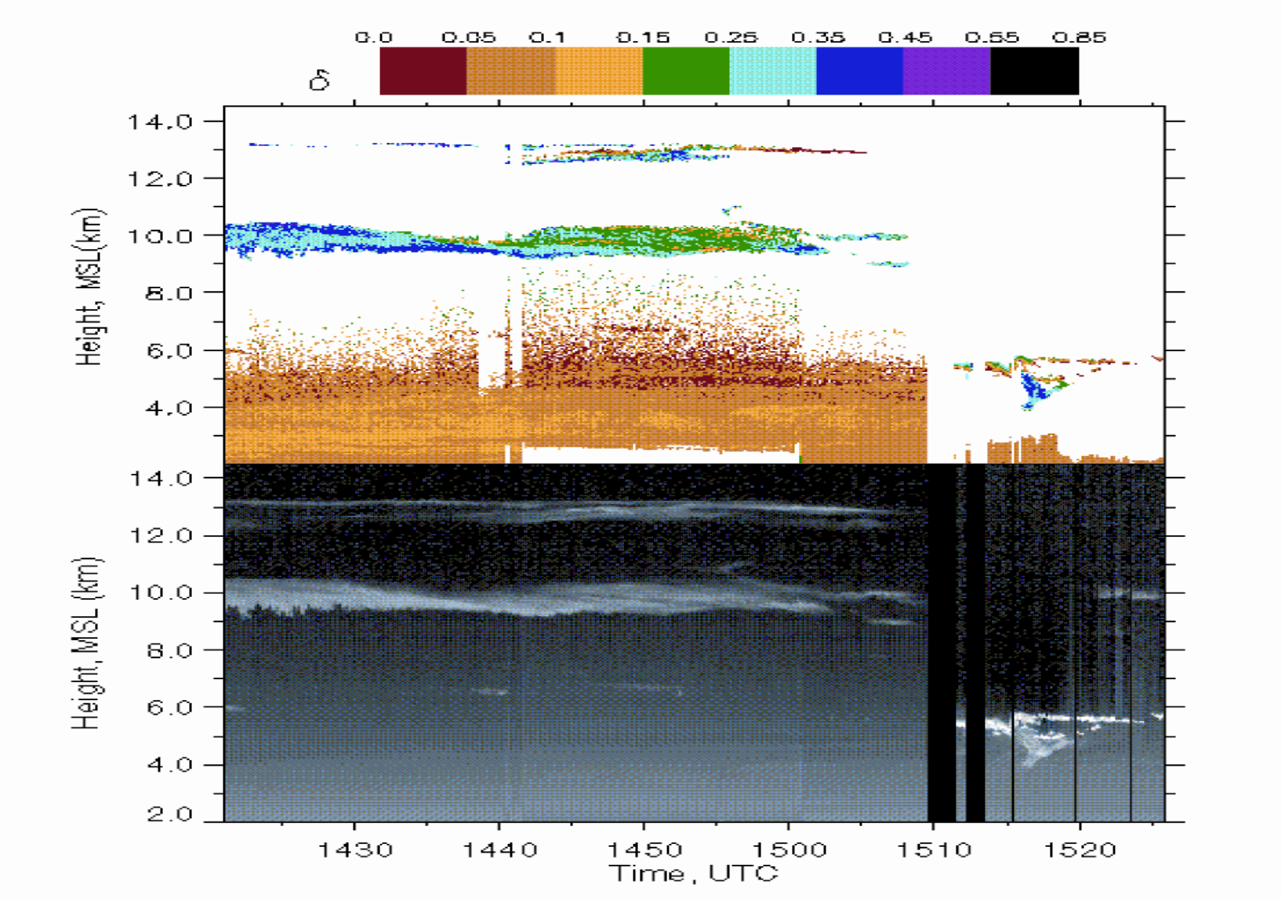
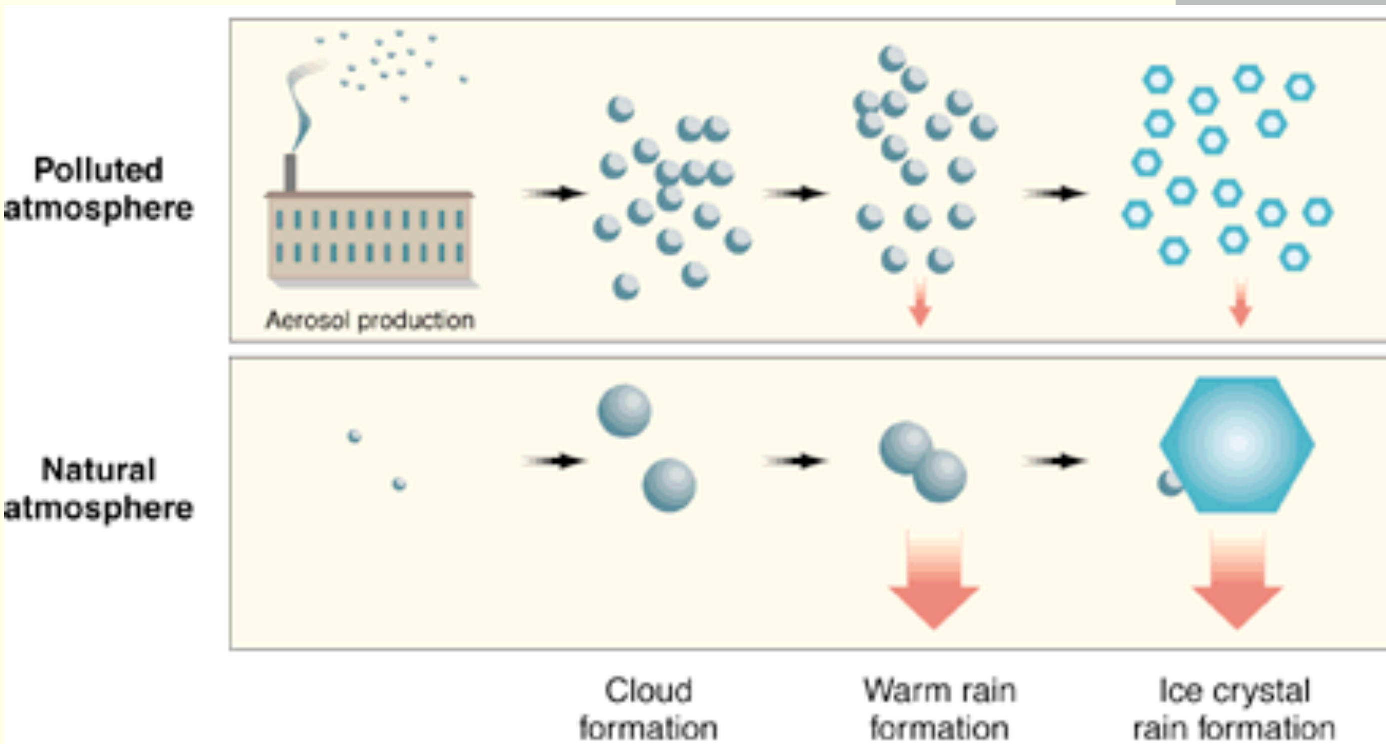


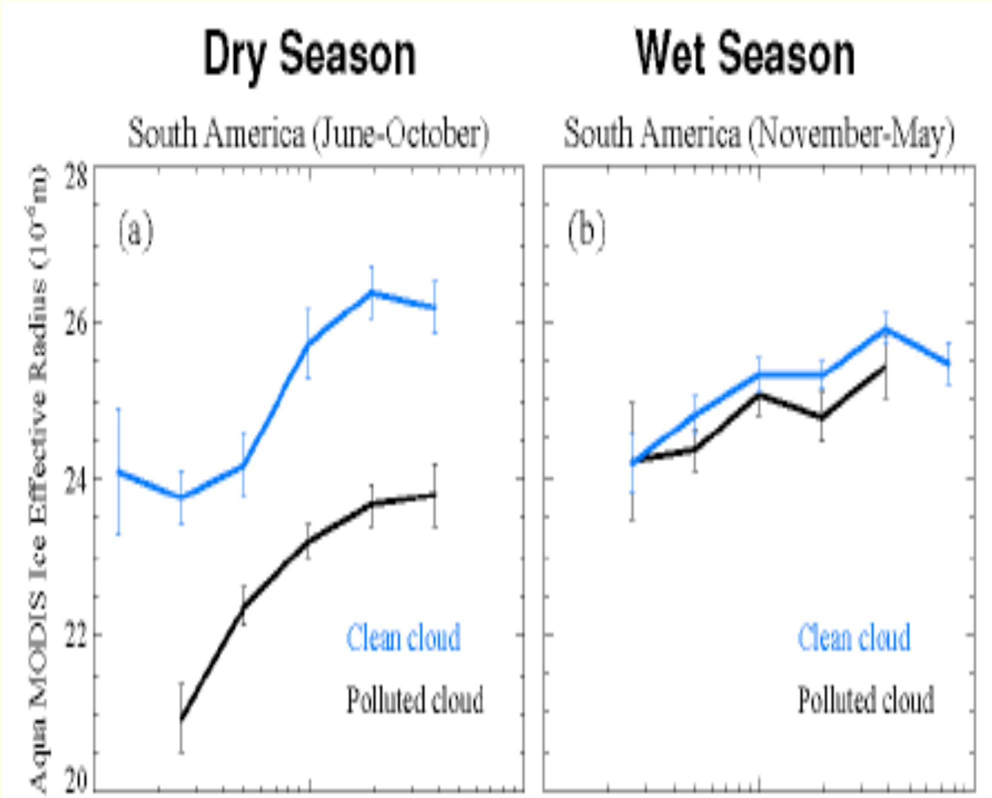
Figure 1. PDL linear depolarization ratio (see color δ scale at top) and relative retrured power (in a logarithmic gray scale) height-versus-time displays on the morning of 29 July 2002, from the Ochopee field site of the CRYSTAL-FACE program. Note that low-altitude signals often cannot be used to calculate δ -values because of strong off-scale signals. Depicted are strongly depolarizing (δ ~ 0.2 to 0.4) upper troposheric cirrus clouds, aerosols (δ ~ 0.10 to 0.15) extending up to ~5.5 km, and at far right a super cooled liquid altocumulus cloud (δ ≈ 0 at cloud base). Note the temporary glaciation of this cloud as it descended into the top of the dust layer (after Sassen et al. 2002).

Cloud Particle Size & Pollution



Processes by which aerosols affect clouds. The polluted cloud contains eight times as many droplets → half the size → twice the surface area → twice the optical depth → higher reflectivity than the natural cloud.

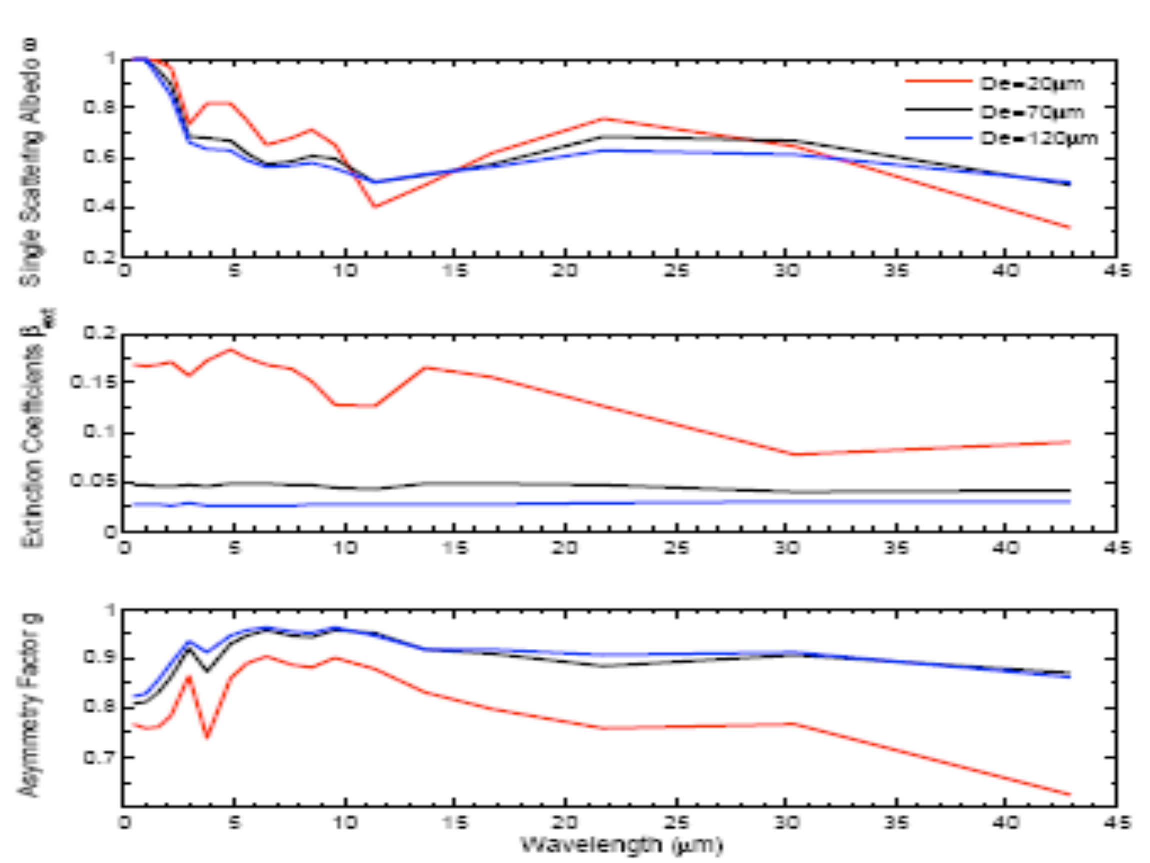
Ice Crystal Size & Clean/Polluted Clouds: Indications From Satellite Data Analysis



Aura MLS IWC (mg/m^3) Aura MLS IWC (mg/m^3)

From Jiang et al. (2008, GRL)

Single-Scattering Properties for Different D_e



Smaller D_e

→Larger single-scattering albedo & extinction coefficient

→Smaller asymmetry factor

→ More reflected solar and more trapped IR

Parameterizations for Ice Number or Mean Effective Ice Crystal Size (*De*)

- Relate ice nucleation and number to aerosol concentration on the basis of explicit microphysics modeling, laboratory studies, as well as theoretical considerations_
- Mean effective ice crystal size calculated from ice mass & number for radiation calculation

large uncertainties in the parameterization of ice microphysics processes and tremendous computational cost

- A common practice to prescribe a mean effective ice crystal size in numerical models (e.g., Köhler 1999; Ho et al. 1998; Gu et al. 2003)
- Used IWC and/or temperature produced from the model to determine mean effective ice crystal size (Kristjánsson et al. 2005; Gu and Liou 2006, Ou and Liou 1995, Ou et al. 1995, McFarquhar et al. 2003, Liou et al. 2008).

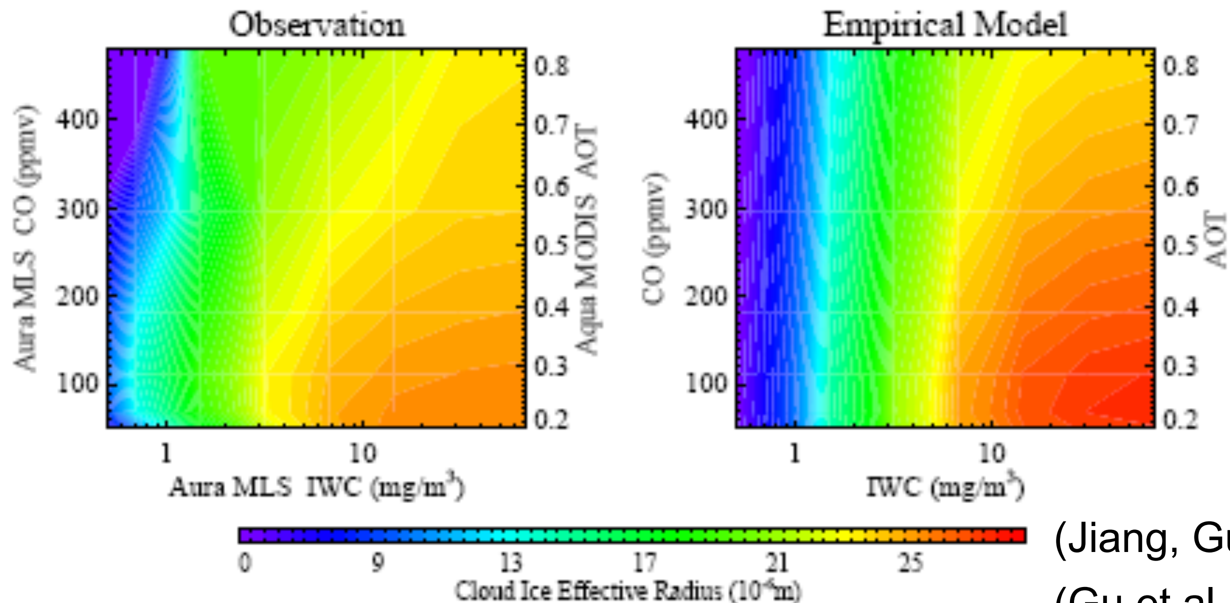
None of the De parameterizations accounted for the distinction between "polluted" and "clean" clouds

Empirical IWC-re-AOT Relations Based on Satellite Data

Using least-squares fitting, we obtained an empirical formula for *Re* as a function of *IWC* and *AOT*. This function broadly captures the variation of *Re* with *IWC* and *AOT*.

$$R_e = \frac{R_e^0}{(IWC / IWC^0)^{-\alpha} + 1} \cdot \frac{(\varepsilon \cdot AOT - \tau)^{\gamma}}{(\varepsilon \cdot AOT - \tau)^{\beta} + 1}$$

α=1.436, β=0.5858, γ=0.282, ε=8.09
R_e^0=56.0293, IWC^0=1.3838, τ=1.11



(Jiang, Gu et al. 2011)

(Gu et al. 2012)

This empirical relationship of R_e with *IWC* and *AOT* can serve as a 1st-order parameterization of the first indirect effect of aerosols on ice clouds for application to climate models.

Radiative Transfer Schemes in WRF

- All single column (one-dimensional in vertical) schemes – Good approximation if model layer thickness is much less than horizontal resolution
- Gaseous Absorption due to H_2O, O_3, and CO_2 are normally included; Traces gases such as CH_4, N_2O, O_2, N_2, CFCs, etc. are also included in many schemes
- Water cloud and ice clouds are normally included; Some include rain, snow, and graupel
- Multiple spectral bands
- Two-stream approximation is commonly used
- Aerosol direct radiative effect are allowed in several schemes

Comparison of Radiation Schemes in WRF

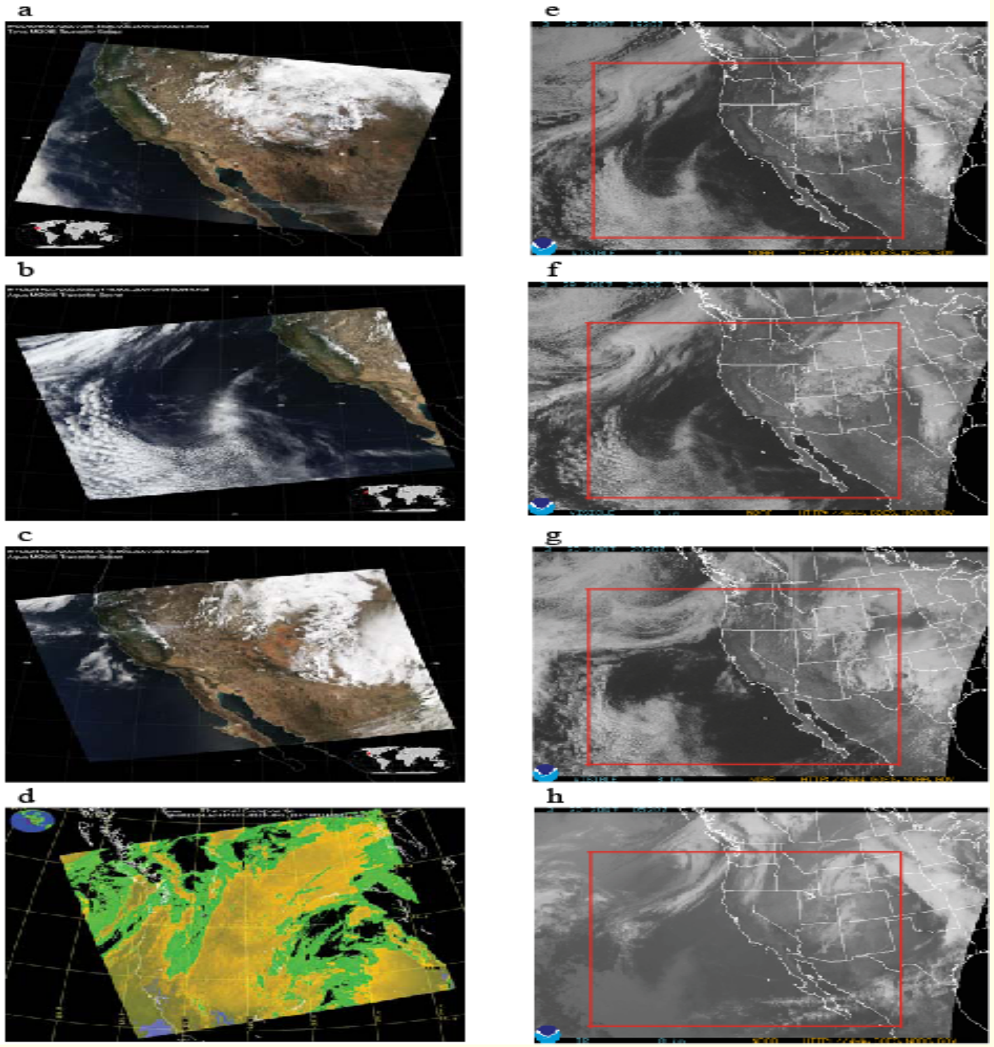
Scheme Name	Gases	Clouds	Aerosols	Band	Flux Solution
Dudhia-shortwave only (ra_sw_physics=1)	H2O (No O3, CO2)	Look-up tables: Cloud albedo and absorption	aerosol scattering		Downward integration **Sloping and shadowing**
RRTM -longwave only (ra_lw_physics=1)	H2O, O3, CO2 Trace gases K-distribution	Water & ice clouds: prescribed mass abs coef & LWP/IWP		Multiple	Two-stream
GSFC-shortwave only (ra_sw_physics=2)	H2O, O3, CO2	Water & ice, rain: Parameterized ~LWC/IWC & r_e/D_e	GOCART (WRF-Chem)	Multiple	Two-stream
CAM (3)	H2O, O3, CO2 Trace gases	Water & ice: Parameterized ~LWC/IWC & r_e/D_e re/De ~ temperature	Allow: monthly climatology 7 groups	Multiple	Solar: Two-stream IR: Absorptivity-emissivity
RRTMG (4) - New version of RRTM	H2O, O3, CO2 Trace gases k-distribution	Water & ice: Parameterized ~LWC/IWC & r_e/D_e McICA random overlap		Multiple	Two-stream
Goddard (5) New Goddard scheme	H2O, O3, CO2 Trace gases k-distribution	Water & ice, snow, rain, and graupel: Parameterized LWC/IWC & r_e/D_e De ~ temperature		Multiple	Two-stream
FLG (7)	H2O, O3, CO2, O2 Trace gases k-distribution	Water & ice, rain, graupel: Parameterized ~LWC/IWC & re/De De ~ IWC or IWC & AOD Maximum/random overlap Horizontal inhomogeneity	Allow: 18 aerosol types	Multiple	Solar: Four-stream IR: two/four combination
GFDL (99)	H2O, O3, CO2	Water & ice: prescribed abs coef; COD ~ Qc Maximum/random overlap		Multiple	Two-stream

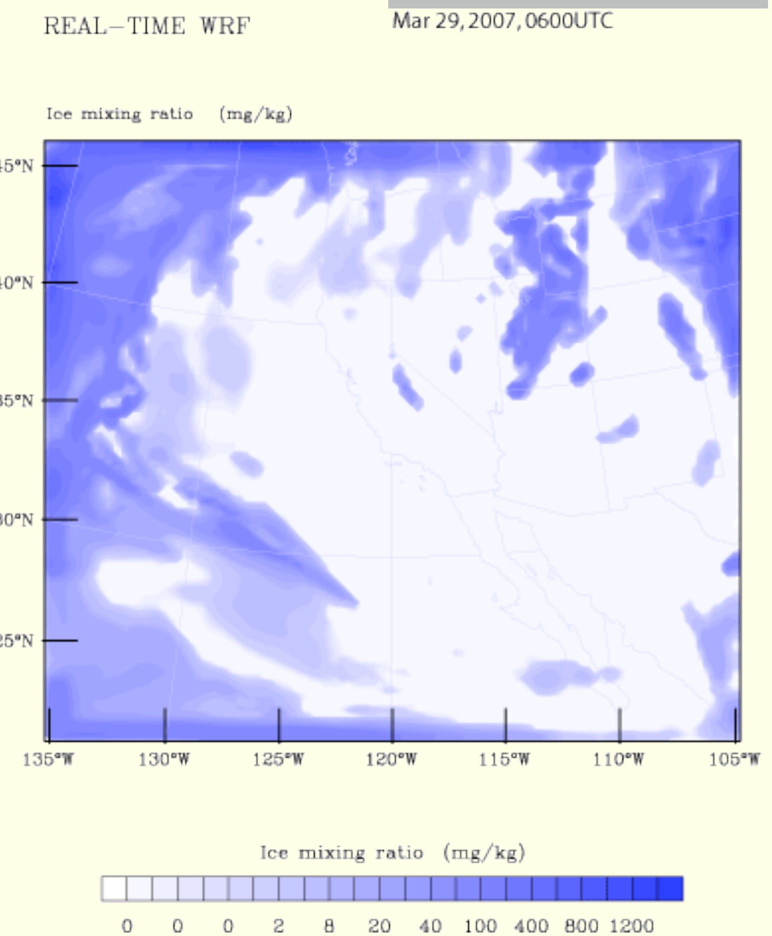
Does Choice of Radiation Scheme Matter?
- WRF Simulations for Real Case Cirrus clouds

- 28 Eta levels, 30 km horizontal resolution
- Same initial and boundary conditions
- Same physics options except for radiation scheme:
  - CTRL: LW: RRTM Scheme; SW: Dudhia Scheme
  - RADI: FLG for both SW and LW
- Observed cirrus cases over northeastern pacific Ocean and western United States during March 29 – 30, 2007
- 48 hour simulations starting March 29, 2007, 0000 UTC

WRF Simulations

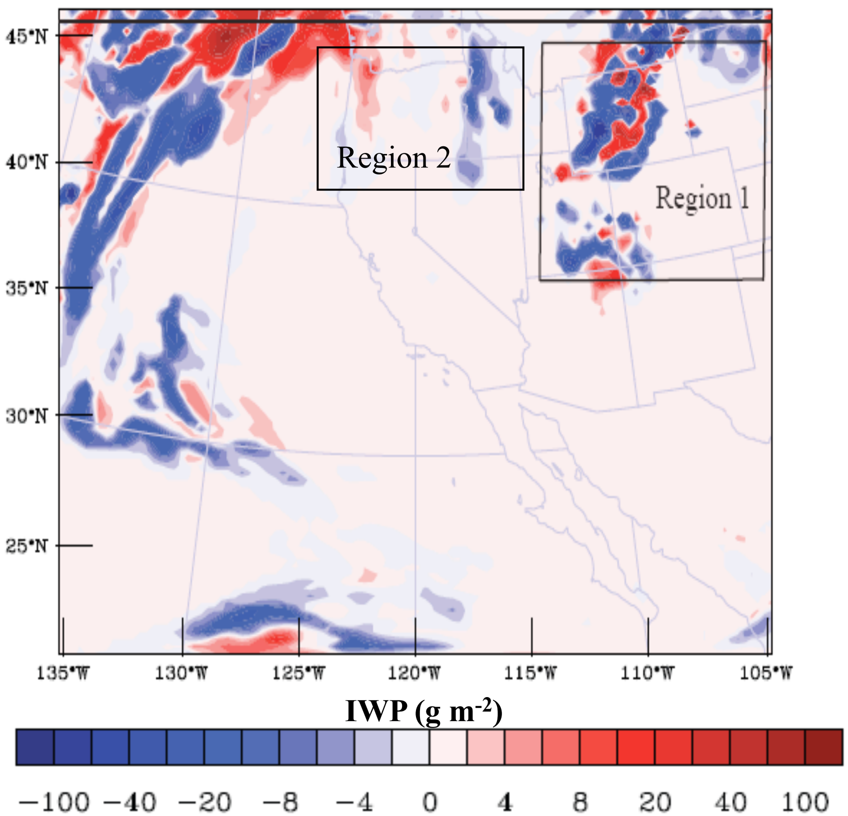
(48 hours simulations, starting at Mar. 29, 2007, 0000UTC)

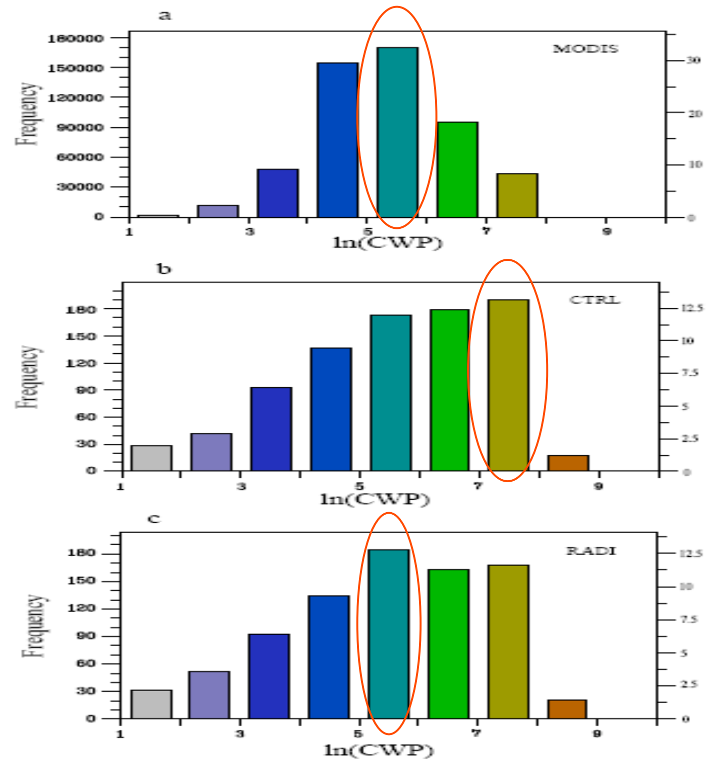




March 29, 1800 UTC

RADI – CTRL





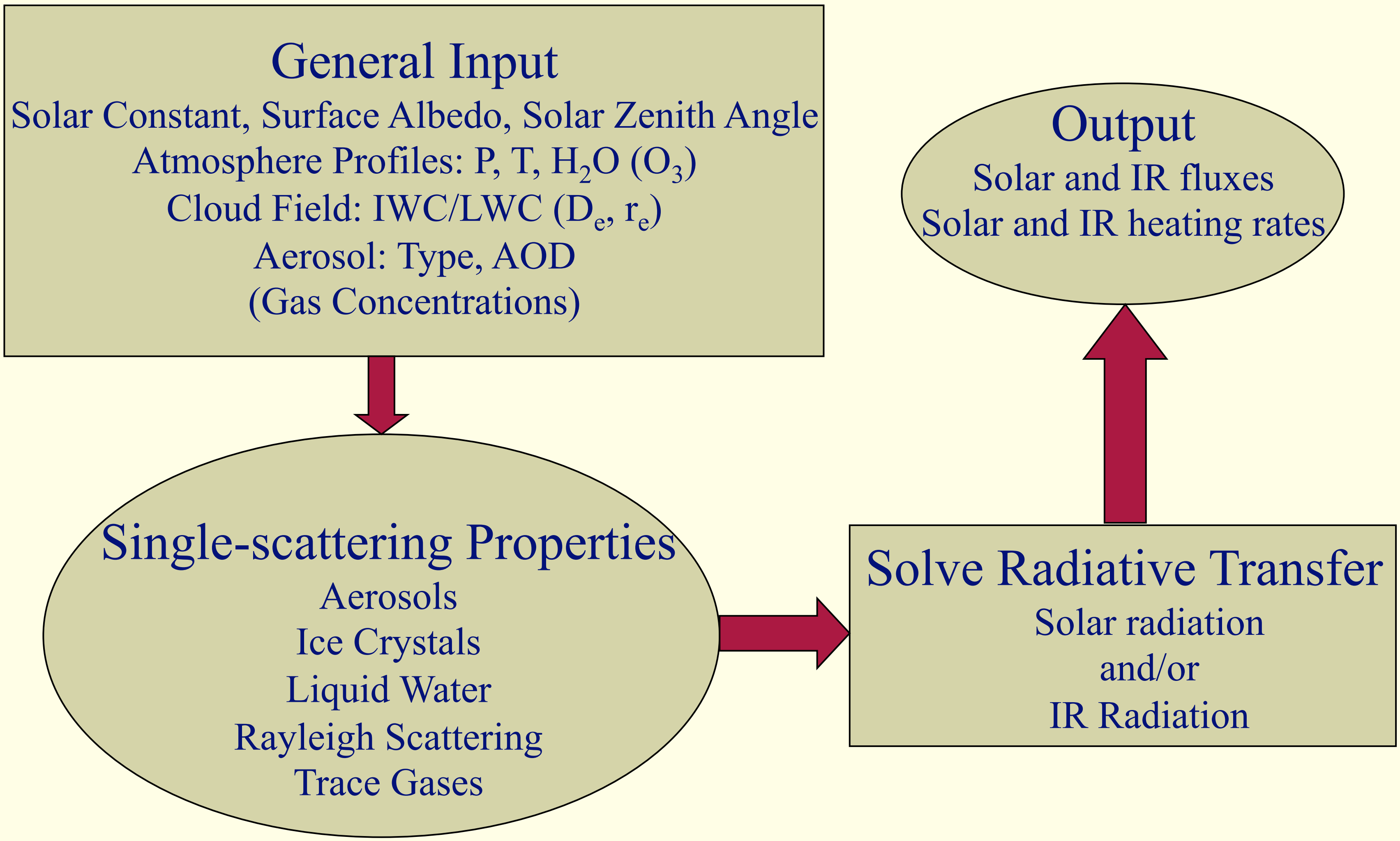
Region 1

Obs/Exp	Mean IWP (g m^{-2}) Region 1	Mean IWP (g m^{-2}) Region 2
MODIS	38.02	23.2
CTRL	43.18	30.47
RADI	41.26 (~5%)	25.76 (~20%)

Obs/Exp	Cirrus Fraction (%) Region 1	Cirrus Fraction (%) Region 2
MODIS	42.33	17.00
CTRL	49.67	17.70
RADI	46.97	17.97

Gu et al. 2011, JGR

Flow Chart of the Radiation Program in Models



Play Around with Radiation Scheme

What could be possibly tuned in a radiative transfer scheme?

- **For clouds**

  - Prescribed absorption coefficient: vary from 100. to 1000. (e.g., GFDL)

  - Mean effective radius/size: modified prescribe values (r_e~10 μm; D_e ~ 75 μm) or develop new parameterizations (e.g., CAM, Goddard, FLG)

  - Cloud optical depth adjustment: the adjustment factor could be tuned (e.g., GFDL)

Future Work

- 3D radiative transfer in climate models

- Cloud sub-grid scale structure: horizontal and vertical inhomogeneity

- Aerosol nonsphericity and inhomogeneous effects – Lorenz-Mie theory of light scattering by spheres cannot be applied to nonspherical particles; significantly affect the aerosol single-scattering properties and consequently produce important impact on climate forcing.


Thank you!