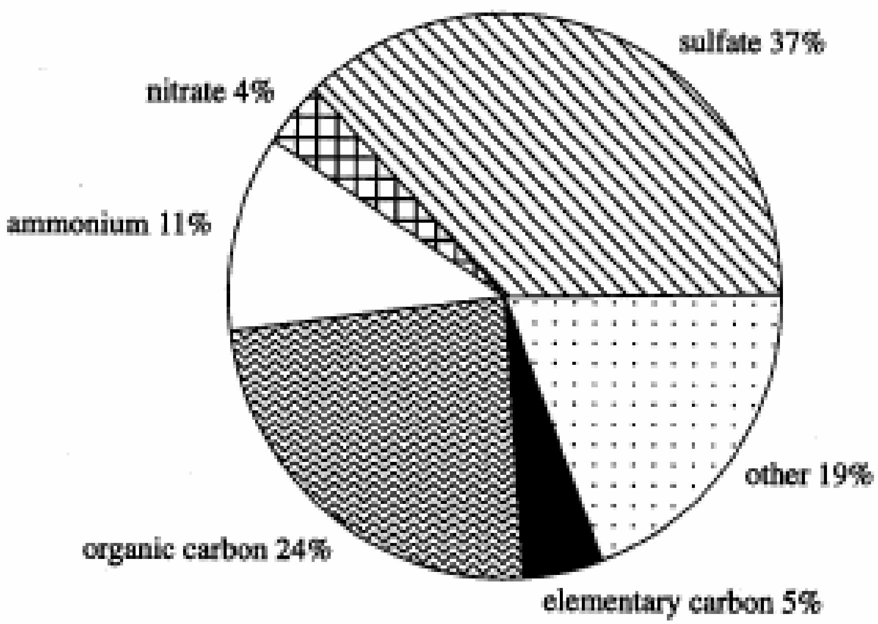
AOSC400-2015
October 8, Lecture # 11

Aerosols *(refresher and continuation):*

- Composition, size distribution, optical properties (e.g., optical depth, single scattering albedo, asymmetry parameter)
- Climate sensitivity
- Aerosol radiative forcing
- What is IPCC?
- IPCC estimates of aerosol radiative forcing

Ultrafine aerosols – nucleation mode (10^{-3}-10^{-2} μm)

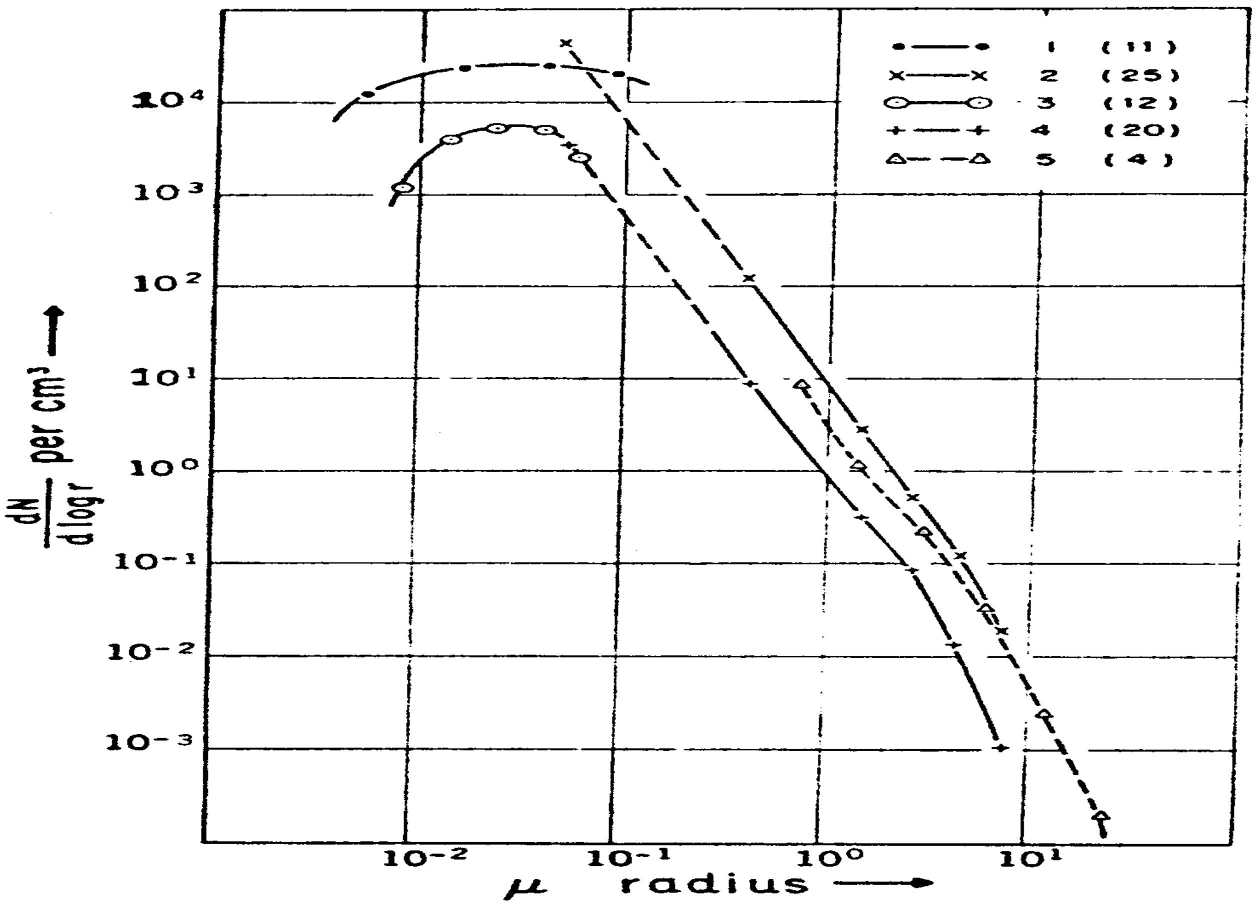
Fine aerosols – condensation and coagulation (0.01-1 μm)



Accumulation mode:

Too small to sediment and are lost mainly by scavenging by cloud droplets

Coarse mode – form by wind erosion (land, sea, vegetation).



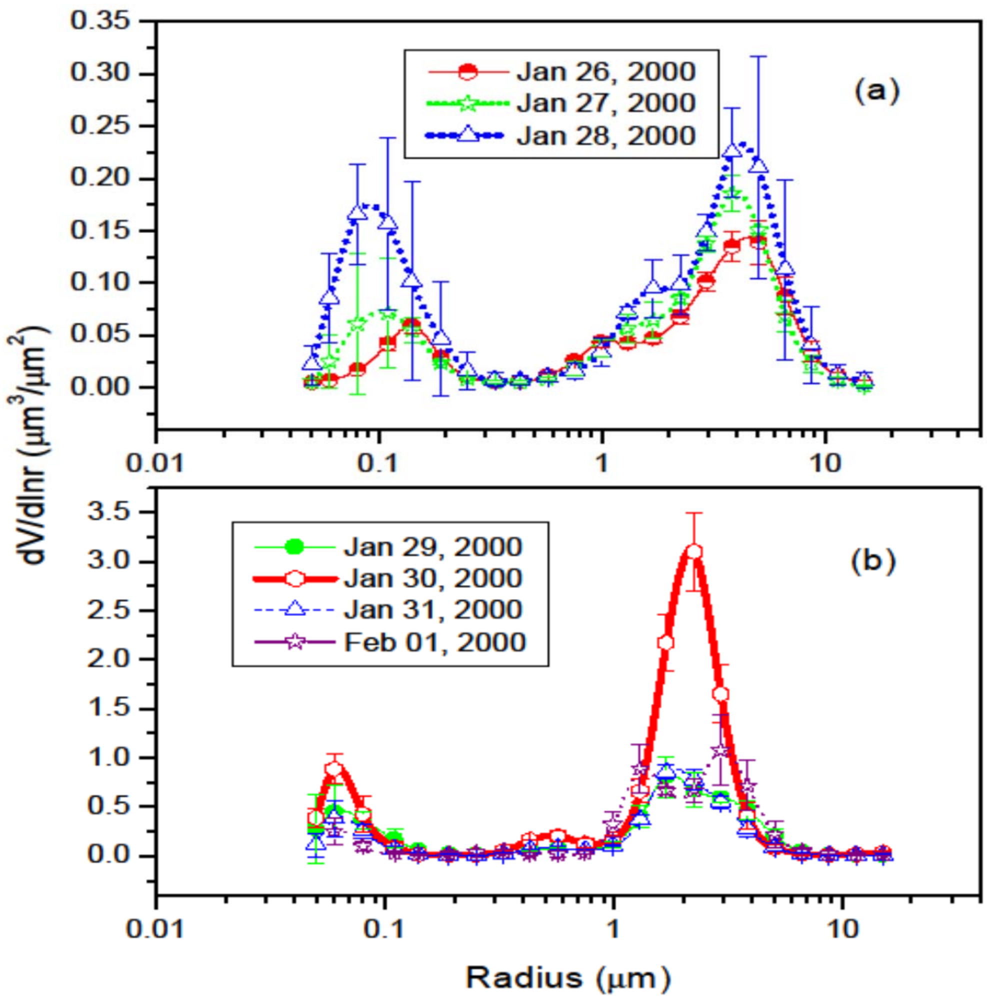
Junge distribution of aerosol sizes.

Power law distributions are often used to model the aerosol populations, but because of the peaks in the accumulation and coarse modes, such models are only useful over limited ranges.

Other observed distributions of aerosols.

Upper figure: Normal days

Lower figure: During dust outbreak



Bimodal Variability in size distribution during “normal” days and dust outbreak in Africa.

Aerosol Size Distribution

The total aerosol optical depth due to extinction may be written as

$$\tau_{\lambda}^{M} = \int_0^{\infty} \beta_e(\lambda, z)\,dz \qquad (2)$$

Let the height dependent aerosol size distributionbe defined by n(z, a) ($cm^{-3}\mu m^{-1}$)

Then, the extinction coefficient (cm^{-1}) is given as

$$\beta_e(\lambda, z) = \int_0^{\infty} \sigma_e(a, \lambda)\, n(z, a)\, da \qquad (3$$

where σ_e stands for the extinction cross section (cm^2) for an individual particle. For the retrieval purpose, the Junge size distribution can be assumed to a good approximation.

Aerosol Size Distribution

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$$\tau_{\lambda}^{M} = \int_0^\infty \beta_e(\lambda, z)\, dz \qquad (2)$$

Remember from Lecture 10:

$$\tau_{R\lambda} = \int_0^\infty \kappa_{R\lambda} \rho \, dz = \int_0^\infty \beta_{R\lambda} \, dz$$

where

$\kappa_{R\lambda}$ = Rayleigh mass scattering coefficient

$\beta_{R\lambda}$ = volume scattering coefficient

$$\kappa_{R\lambda} \rho = \beta_{R\lambda}$$

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Scattering coefficient σ_s

The scattering coefficient is defined as the fraction of radiant flux lost from a collimated beam per unit thickness of aerosol due to scattering (deflection of light into directions different from the direction of the transmitted beam) and is given in units of reciprocal length.

Absorption coefficient σ_a

The absorption coefficient is defined as the fraction of radiant flux lost from a collimated beam per unit thickness of aerosol due to absorption (transformation of light into other forms of energy) and is also given in units of reciprocal length.

$$\sigma_e = \sigma_s + \sigma_a$$

Single scattering albedo ($\overline{\omega}$)

The ratio of the scattering coefficient to the extinction coefficient is called the single scattering albedo (SSA). It describes the contribution of scattering to extinction.

$$\overline{\omega} = \frac{\sigma_s}{\sigma_e} = 1 - \frac{\sigma_a}{\sigma_e} = (1 - \alpha) \qquad (11)$$

Where the ratio of the absorption coefficient to the extinction coefficient

is termed as the absorption number $\alpha = \frac{\sigma_a}{\sigma_e}$

It is the fraction of scattered light with respect to the total light which interacts with the particles.

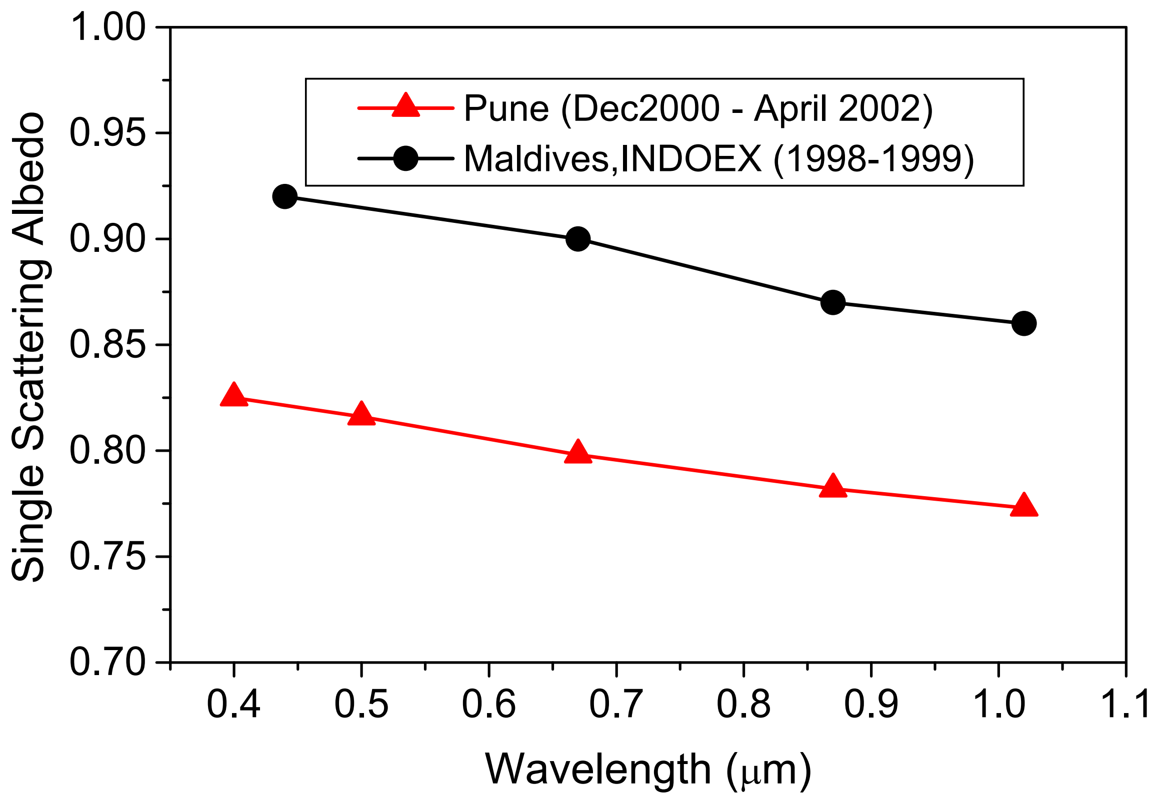
Single scattering albedo - ω_0

Ratio of scattering to extinction, describes aerosol absorption.

0.0----------0.5-----------0.8-----------0.9----------0.95---------1.0

black carbon	smoke from fire	urban pollution	sea salt
			volcanic
	dust in blue		sulfuric acid
		dust in green to NIR	

Comparison of single scattering albedo over Pune (inland station in India) with Maldives INDOEX Experiment (AERONET) measurements



The following 5 slides have been introduced for completeness of the discussion. May be a bit above the class level.

Angular scattering cross section $\beta(\theta)$

Angular scattering cross section is defined as that cross section of an incident wave, acted on by the particle, having an area such that the power flowing across it is equal **to** the scattered power per steradian at an observation angle σ.

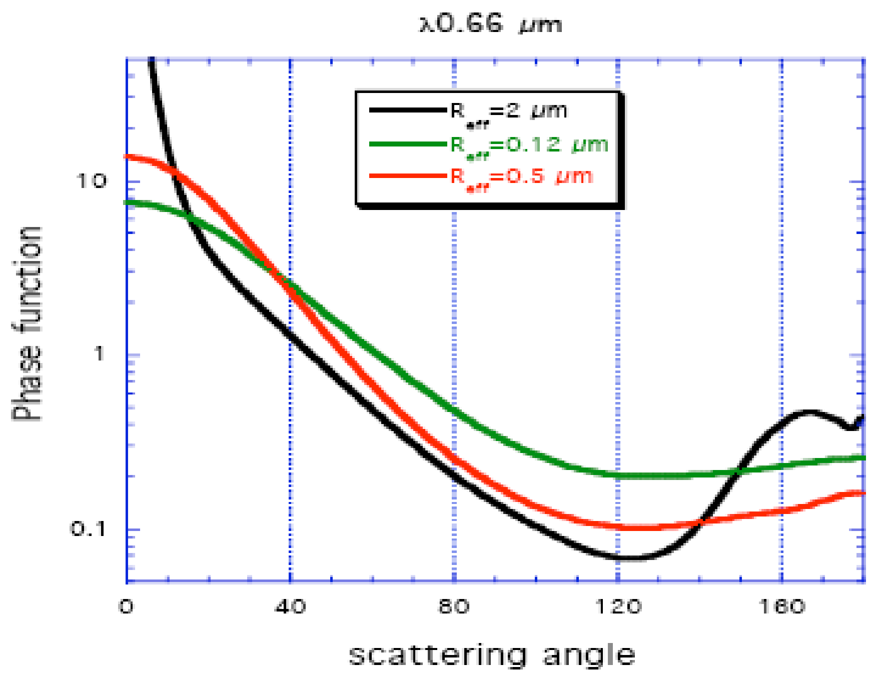
The total scattering cross section is the integral over the total solid angle $d\omega$ given by

$$\sigma_{sc} = \int_0^{4\pi} \sigma(\theta)\, d\omega \qquad (12)$$

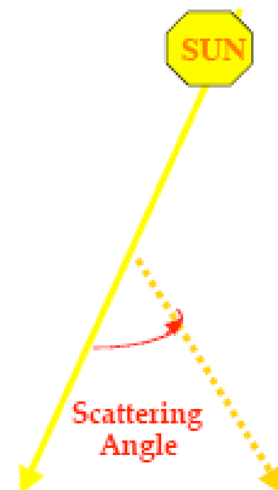
Angular scattering coefficient for a poly-dispersion is given by

$$\beta(\theta) = \int_{r_1}^{r_2} \sigma(\theta)\, n(r)$$

Angular distribution of the scattered light.



Scattering angle:
The angle of deviation from the direct beam –



Total scattering coefficient β

The total scattering coefficient β_{sc}

is the integral of $\beta(\theta)$

over $d\omega$

$$\beta_{sc} = \int_0^{4\pi} \beta(\theta) d\omega \qquad (13)$$

Phase function P(θ)

Phase function defines the angular scattering properties or the way scattered energy is redistributed as a function of the scattering angle. It is defined as the energy scattered per unit solid angle in a given direction to the average energy in all directions.

$$P(\theta) = \frac{\beta(\theta)}{\beta_{sc} / 4\pi} \quad (14)$$

where $\beta(\theta)$ is the angular scattering cross section and βsc is the total scattering cross section.

Asymmetry factor g (λ)

Asymmetry factor g is the mean value of cos *(θ)* (where θ is the scattering angle) over the total solid angle weighted by the phase function.

$$g(\lambda) = \frac{\int_0^{4\pi} \cos(\theta) P(\lambda, \theta) d(\cos \theta)}{\int_0^{4\pi} P(\lambda, \theta) d(\cos \theta)} \qquad (16)$$

Effect of aerosols on the solar radiation

Aerosol forcing

Aerosols modify incoming solar and outgoing IR radiation. Change in the radiation flux caused by aerosol is referred to as 'aerosol radiative forcing'. The effect of aerosols on the top of the atmosphere (TOA) radiative fluxes is called TOA radiative forcing and that on the surface radiative fluxes is called the surface radiative forcing. The difference between the two is the atmospheric radiative forcing. In all cases (i.e., surface, atmosphere and TOA), 'aerosol forcing' is the difference in radiative fluxes with and without aerosols. Aerosol Forcing:

$$\Delta F = Flux_{noaerosol} - Flux_{aerosol}$$

The change in forcing per unit aerosol optical depth is called

aerosol forcing efficiency

$$\text{Forcing Efficiency} = \Delta F/\Delta \tau$$

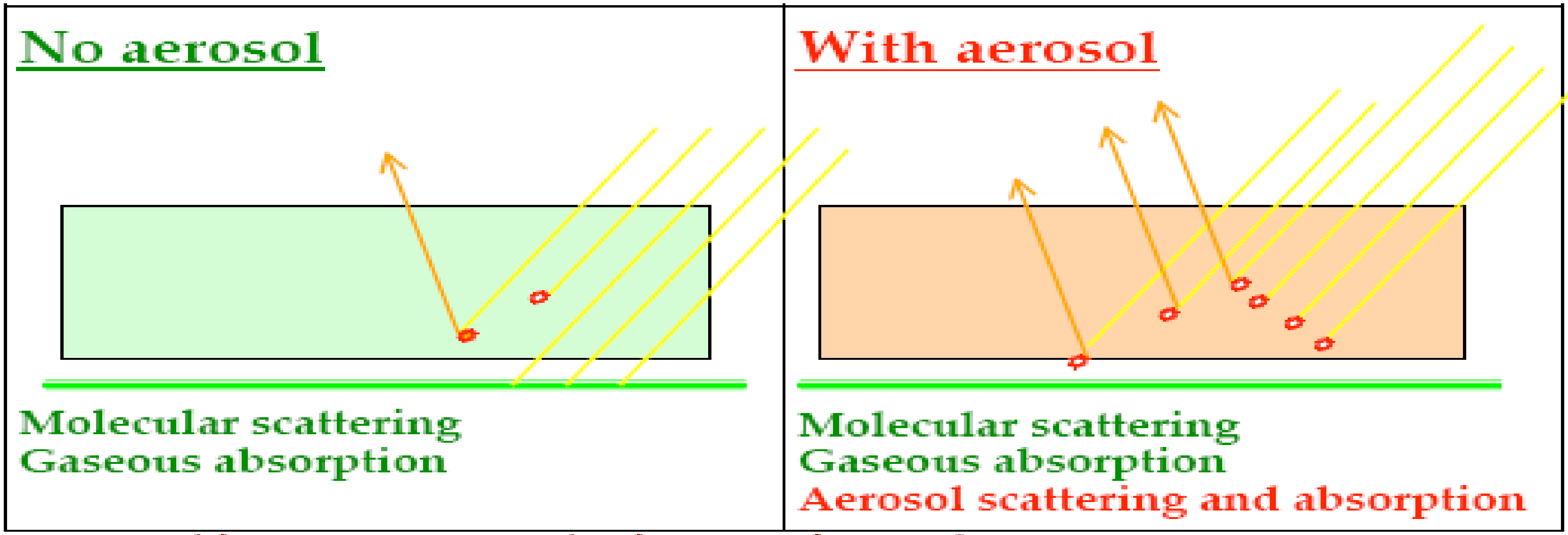
The actual forcing is the product of forcing efficiency and aerosol optical depth.

Effect of aerosols on the solar radiation (also called aerosol radiative forcing) can be broadly classified into two (1) direct forcing and (2) indirect forcing.

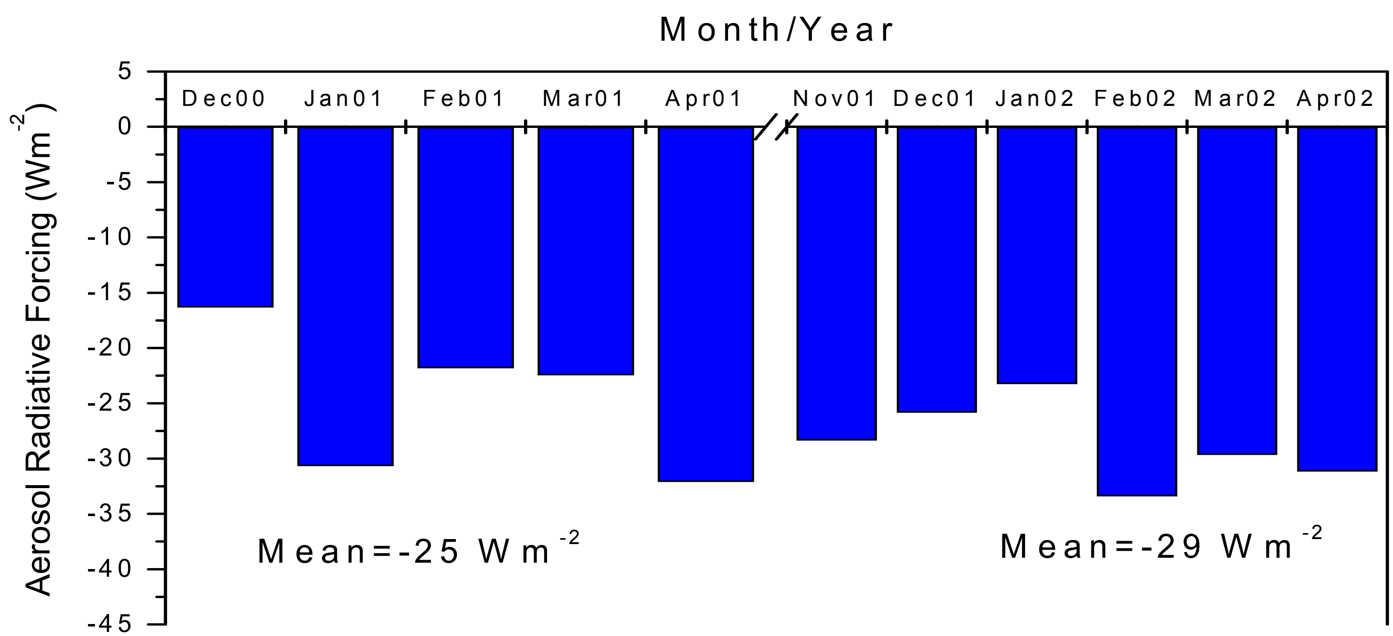
Direct forcing

Scattering and absorption of solar radiation is known as "direct effect" of aerosols on the global climate, which lead to either cooling or warming of the atmosphere depending on the proportion of light scattered to that absorbed.

What is radiative forcing of aerosol?



- Aerosol increases atmospheric scattering and absorption
- Aerosol increases reflection of sunlight to space and reduces sunlight reaching the surface
- Aerosol absorption changes the temperature profile in the atmosphere



Monthly mean aerosol forcing at the surface observed during dry seasons of years 2001 and 2002 in India

Aerosol radiative forcing over a tropical urban site in India , Pandithurai, G., Pinker, RT, Takamura, T., Devara, PCS , 2004.
GEOPHYSICAL RESEARCH LETTERS Volume: 31 Issue: 12

Indirect forcing

Aerosols have an "**indirect effect**" on climate by altering the properties of clouds. It is difficult to form cloud droplets without small aerosol particles acting as "condensation nuclei" to start the formation of cloud droplets. According to *Twomey,* if the aerosol concentration increases in a cloud,

There is competition for it among the droplets, each of which will be smaller. Consequences: clouds with smaller droplets reflect more sunlight, and last longer, because it takes more time for small droplets to coalesce and become large enough to fall to the ground. This effect is supported by observations showing that aerosols from forest fires and urban pollution can suppress rain and snow fall.

Radiative forcing of aerosols differs from that of greenhouse gases; unlike greenhouse gases, aerosol particles are relatively short lived in the troposphere, resulting in spatial and temporal non-uniformity. Aerosol forcing is greatest in daytime and in summer; greenhouse gas forcing acts over the full diurnal and seasonal cycles. Furthermore, unlike molecular absorption, which is invariant from one molecule to the next for a particular species, the optical properties of aerosols depend on particle size and composition. Such differences make a description of the aerosol influences on climate much more complex than a treatment of radiative influences of greenhouse gases.

Radiative forcing

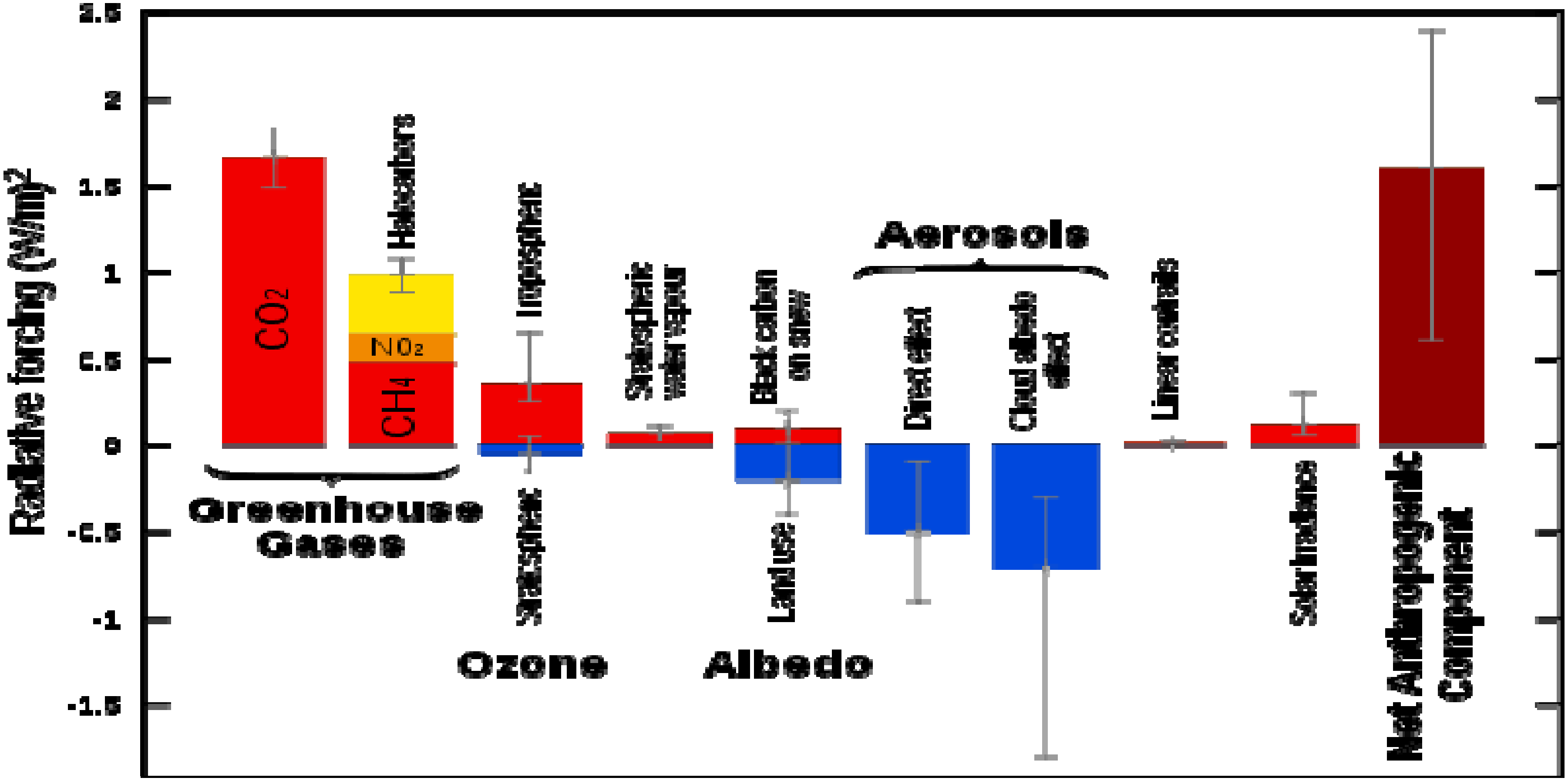
- A quantity that *measures the contribution of a greenhouse gas to the atmospheric greenhouse effect*
- Refers to the change in outgoing infrared flux due to the change in the concentration of a particular greenhouse gas
- Example: a doubling of atmospheric carbon dioxide produces a radiative forcing of ~ 4.4 W/m^2

Climate sensitivity

Climate sensitivity is defined as the amount of global average surface warming following a doubling of carbon dioxide concentrations.

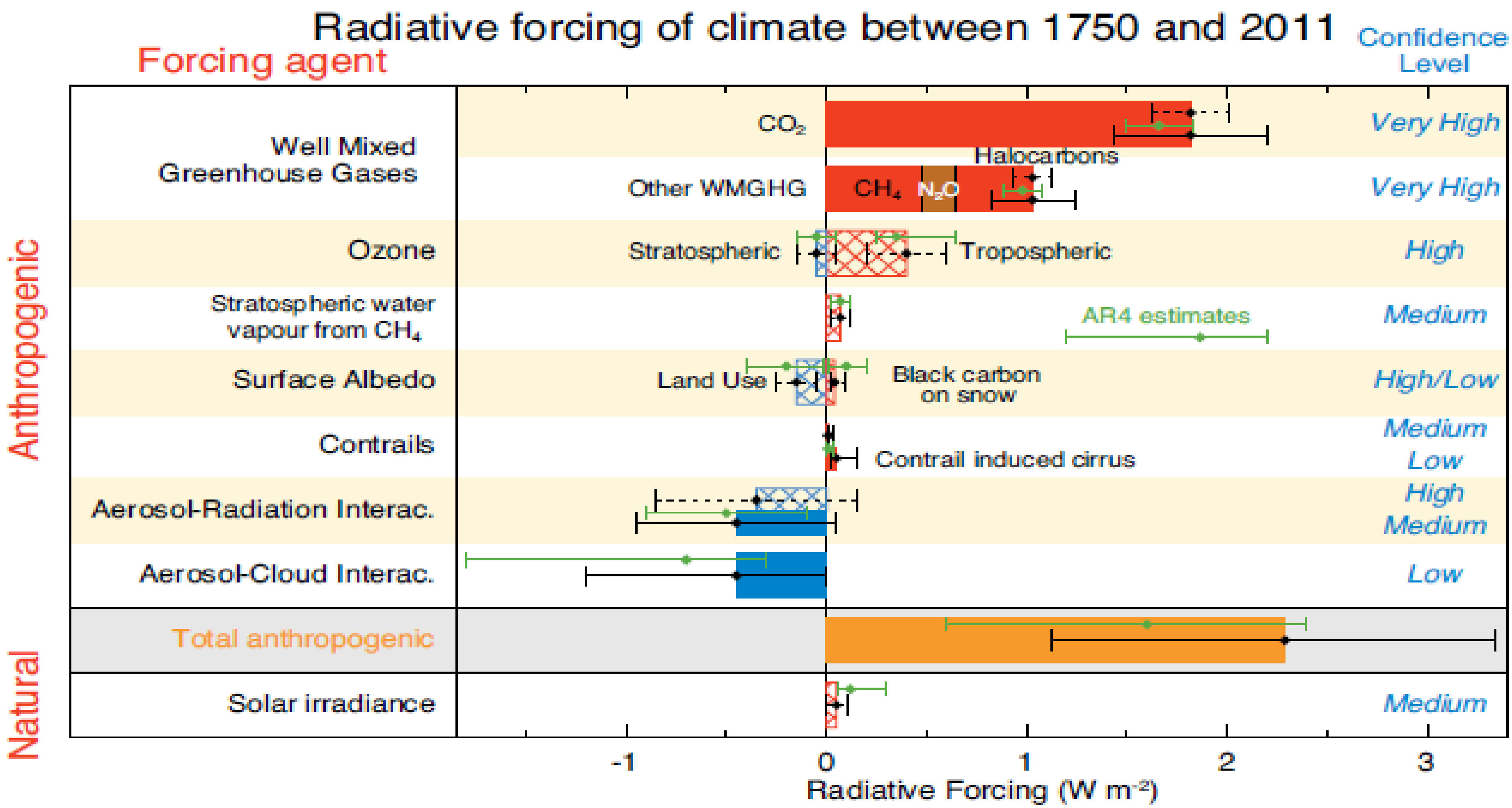
It is estimated to be in the range of *2 to 4.5 °C*, with a best estimate of about 3 °C.

Radiative Forcing Components

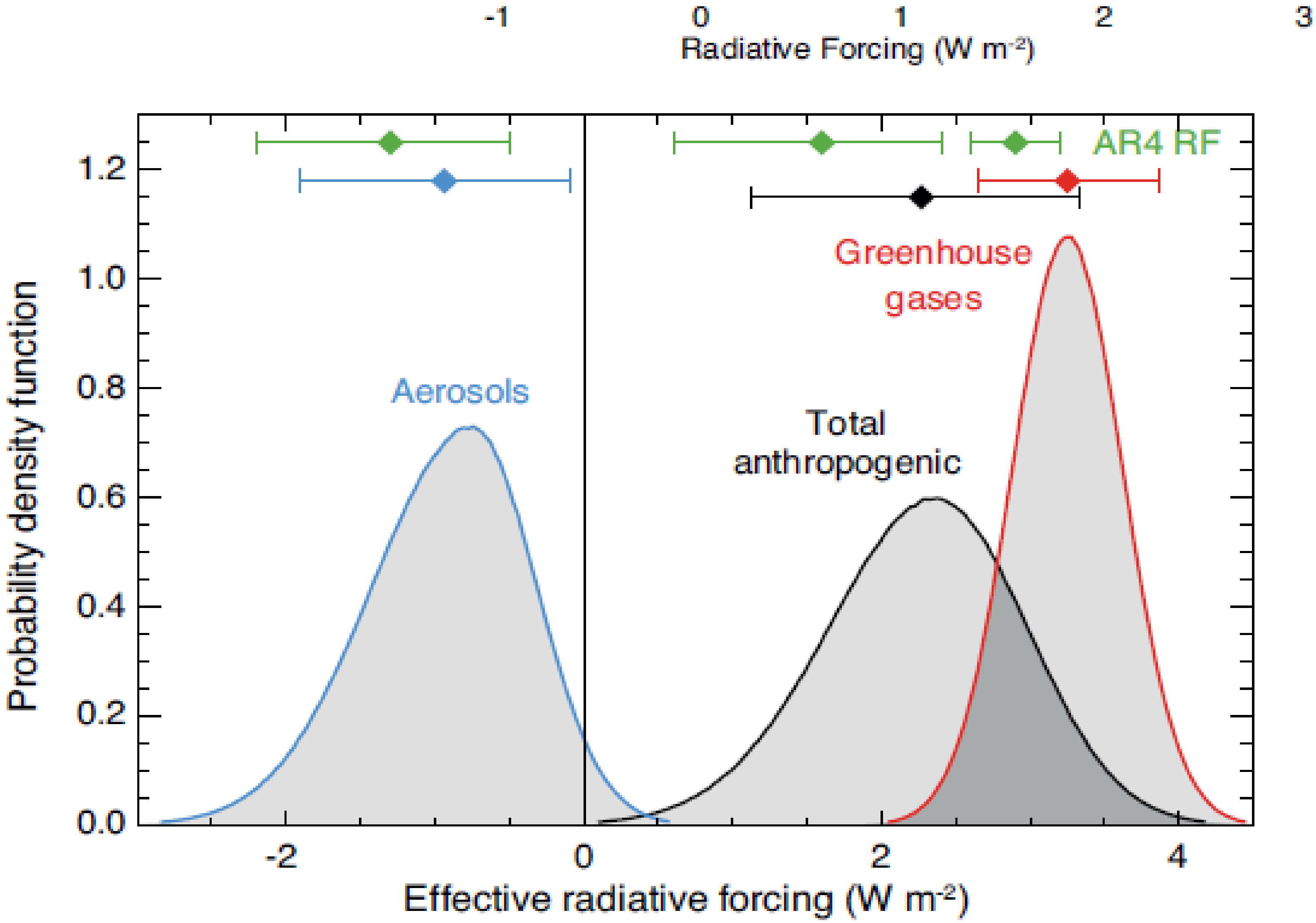


Changes in radiative forcing between 1750 and 2005 as estimated by the IPCC 2007

IPCC 2013

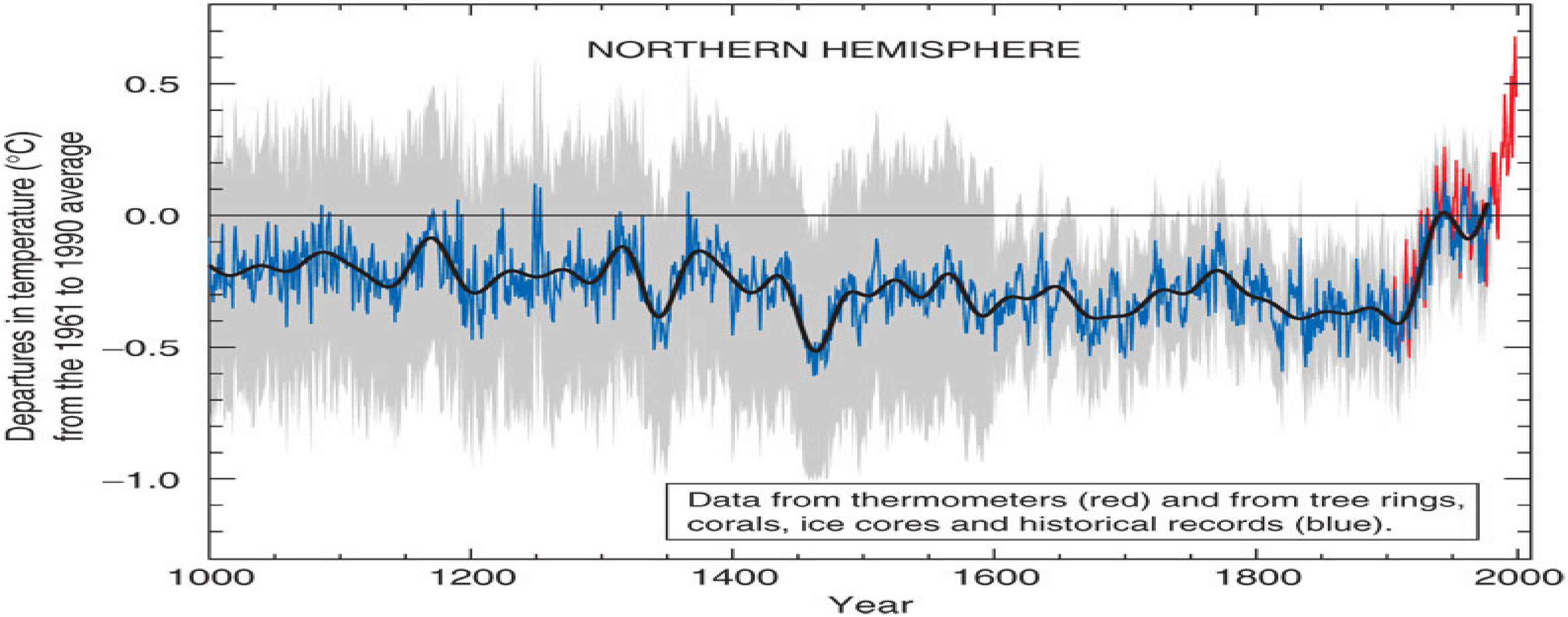


Radiative forcing of climate between 1750 and 2011
Forcing agent
Confidence Level
Well Mixed Greenhouse Gases
CO₂
Very High
Halocarbons
Other WMGHG
CH₄
N₂O
Very High
Ozone
Stratospheric
Tropospheric
High
Stratospheric water vapour from CH₄
AR4 estimates
Medium
Surface Albedo
Land Use
Black carbon on snow
High/Low
Contrails
Contrail induced cirrus
Medium
Low
Aerosol-Radiation Interac.
High
Medium
Aerosol-Cloud Interac.
Low
Total anthropogenic
Solar irradiance
Medium
Anthropogenic
Natural
-1
0
1
2
3
Radiative Forcing (W m⁻²)



-1
0
1
2
3
Radiative Forcing (W m-2)
AR4 RF
Greenhouse gases
Aerosols
Total anthropogenic
Probability density function
1.2
1.0
0.8
0.6
0.4
0.2
0.0
-2
0
2
4
Effective radiative forcing (W m-2)

"Hockey stick controversy"



The "hockey stick" graph as shown in the 2001 IPCC report. This chart shows the data from Mann *et al.* 1999. The blue lines are temperatures estimated from proxy indicators, red lines are temperatures from thermometers, and the gray shaded region represents estimated error bar.

How can we predict change in climate?

- Used are the most sophisticated climate models developed by several institutions in various countries as a tool to do such predictions.
- These models are run with prescribed scenarios of possible changes in various atmospheric inputs.
- The Intergovernmental Panel on Climate Change (IPCC) is charged with overseeing the conduct of the various experiments.

The Intergovernmental Panel on Climate Change (IPCC) is the leading international body for the assessment of climate change.

It was established by the United Nations Environment Programme (UNEP) and the World Meteorological Organization (WMO) in 1988 to provide the world with a clear scientific view on the current state of knowledge in climate change and its potential environmental and socio-economic impacts.

In the same year, the UN General Assembly endorsed the action by WMO and UNEP in jointly establishing the IPCC.

- The IPCC is a scientific body under the auspices of the United Nations (UN). It reviews and assesses the most recent scientific, technical and socio-economic information produced worldwide relevant to the understanding of climate change. It does not conduct any research nor does it monitor climate related data or parameters.

- Thousands of scientists from all over the world contribute to the work of the IPCC on a voluntary basis. Review is an essential part of the IPCC process, to ensure an objective and complete assessment of current information. IPCC aims to reflect a range of views and expertise.

- The Secretariat coordinates all the IPCC work and liaises with Governments.

- It is supported by WMO and UNEP and hosted at WMO headquarters in Geneva

- The IPCC is the pre-eminent international scientific organization whose *purpose* is to provide a "*consensus*" scientific view on *global warming.*

- At the end of 2007 the IPCC was awarded the Nobel Peace Prize.

Climate Change 2013: Report Overview

- Working Group I Report: The Physical Science Basis.
- Working Group II Report: Impacts, Adaptation and Vulnerability.
- Working Group III Report : Mitigation of Climate Change .
- Task Force on Greenhouse Gas Inventories
- **The Synthesis Report (SYR):** Summary for Policymakers (SPM)

WORKING GROUP I CONTRIBUTION TO THE IPCC FIFTH ASSESSMENT REPORT
CLIMATE CHANGE 2013: THE PHYSICAL SCIENCE BASIS
Final Draft Underlying Scientific-Technical Assessment

2.3.3 Changes in Surface Radiation Budget

2.3.3.1 Surface Solar Radiation

Final Draft (7 June 2013) Chapter 2 IPCC WGI Fifth Assessment Report

Globally complete satellite estimates are available since the early 1980s (Hatzianastassiou et al., 2005; **Pinker et al., 2005**; Hinkelman et al., 2009). Since satellites do not directly measure the surface fluxes, they have to be inferred from measurable top-of-atmosphere signals using empirical or physical models to remove atmospheric perturbations. Available satellite-derived products qualitatively agree on a brightening from the mid-1980s to 2000 averaged globally as well as over oceans, on the order of 2–3 W m–2 per decade (Hatzianastassiou et al., 2005; **Pinker et al., 2005**; Hinkelman et al., 2009).

Review for Test

Atmosphere's role in heating of the Earth

Atmospheric composition

- Important atmospheric greenhouse gases (ozone, CO_2 - the Keeling curve)
- Probing vertical structure (as a function of height and pressure).
- Know basic concepts: Pressure, Density, Energy, temperature, heat, Latent Heat

Heat transfer in the atmosphere

Solar radiation - sun source of energy

- ❑ Electromagnetic spectrum
- ❑ Inverse-Square; Sun Angle

Seasons

- ➢ Solar zenith angle and how to compute it
- ➢ Solar time-equation of time
- ➢ Declination

How much energy received on a horizontal surface

- Reflectance/albedo
- how to compute length of day

Thermal radiation

- Basic long-wave (thermal) radiation laws:

  Plank; Stefan-Boltzmann; Wien
- Simple radiation balance climate model without greenhouse effect

o Radiative Properties of Non-black Materials

o Simple radiation balance climate model with greenhouse effect

- Relative optical path length
- Empirical expression for air mass
- Instruments to measure components of radiative fluxes

Physics of Scattering and Absorption and Emission

- ❑ Simple aspects of radiative transfer in the visible
- ❑ Beer-Bouguer-Lambert Law
- ❑ Langley Plots

Applications of the Beer's Law

- ❖ To study atmospheric transmissivity
- ❖ To study the impact of aerosols
- ❖ Introduction to AERONET
- ❖ **Rayleigh scattering**
- o Refractive Index; Snell Law

Aerosols:

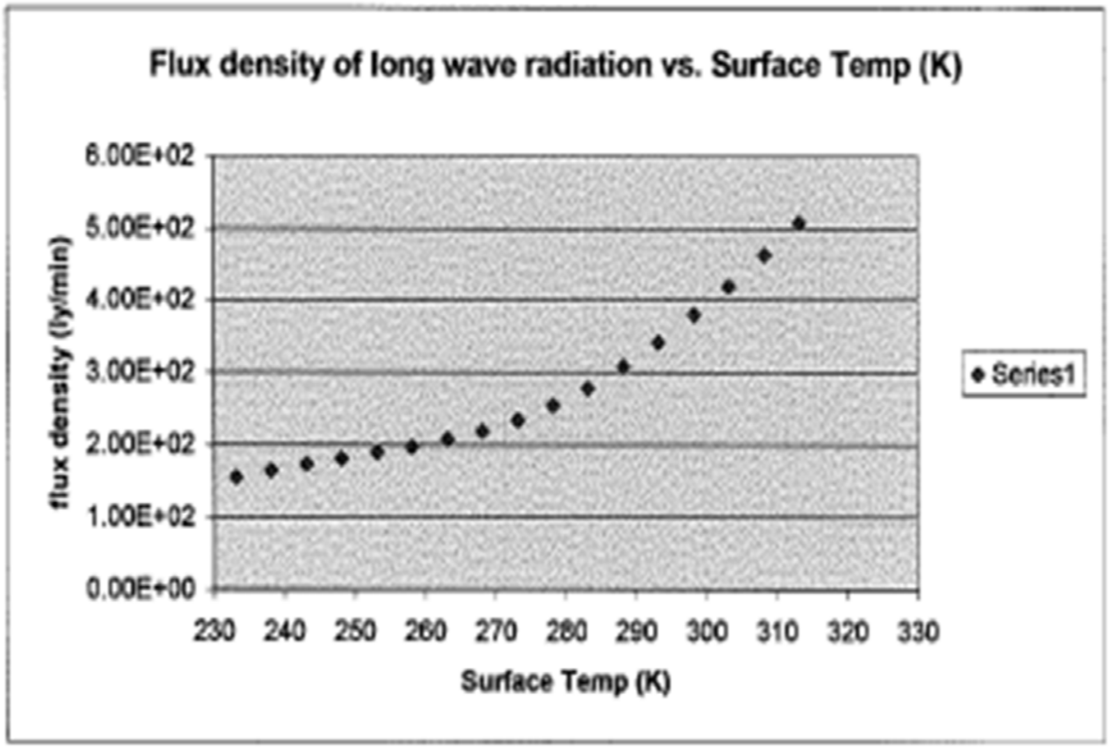
- o Composition, size distribution, optical properties (e.g., optical depth).

Review of Homework # 2

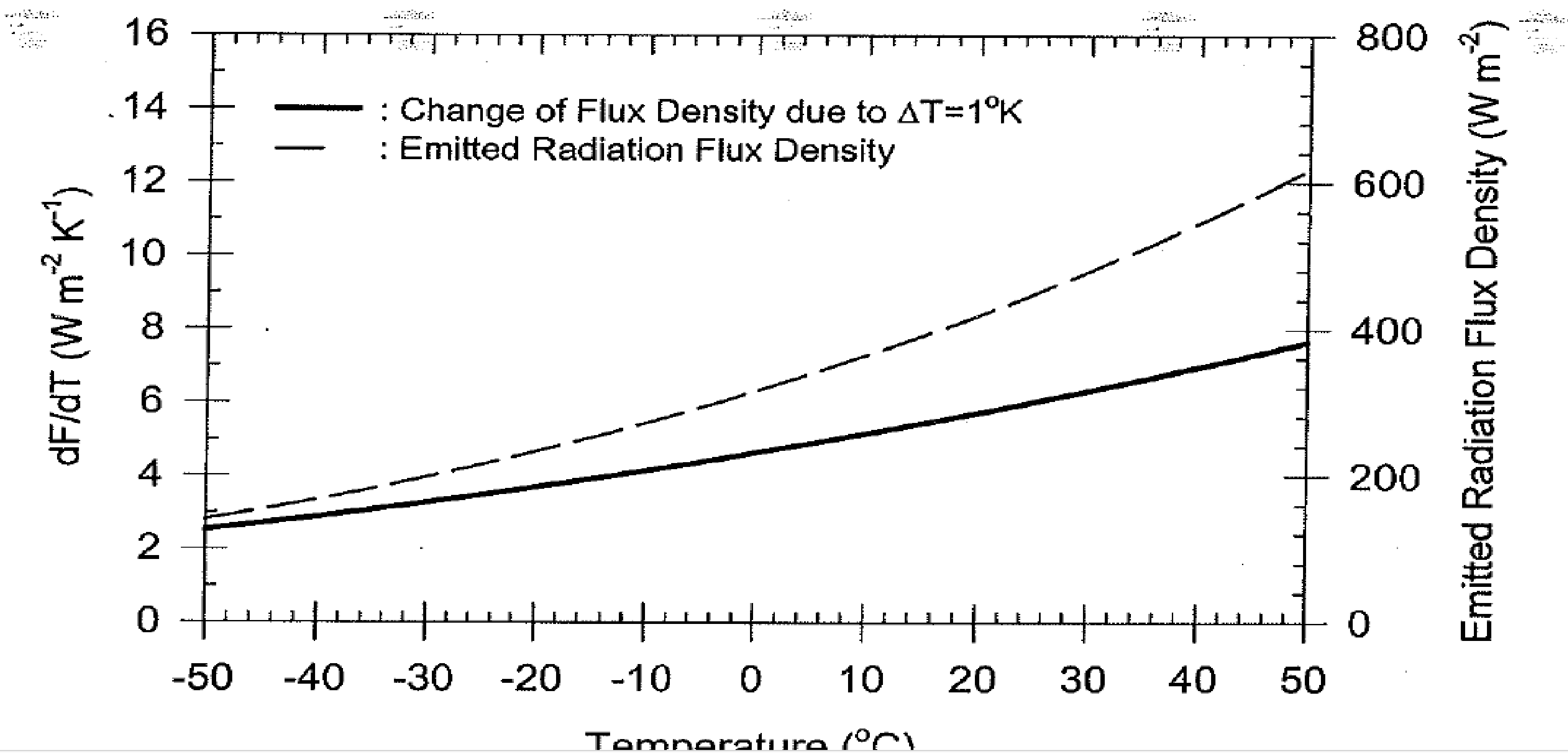
Problem # 7

T=0 C	273.16	
increment	5	

Temperature (C)	T (K)	L
-40	233.16	1.56E+02
-35	238.16	1.65E+02
-30	243.16	1.73E+02
-25	248.16	1.81E+02
-20	253.16	1.89E+02
-15	258.16	1.97E+02
-10	263.16	2.07E+02
-5	268.16	2.19E+02
0	273.16	2.35E+02
5	278.16	2.54E+02
10	283.16	2.78E+02
15	288.16	3.07E+02
20	293.16	3.41E+02
25	298.16	3.79E+02
30	303.16	4.19E+02
35	308.16	4.63E+02
40	313.16	5.07E+02



Problem # 3



: Change of Flux Density due to ΔT=1°K
: Emitted Radiation Flux Density
dF/dT (W m-2 K-1)
Emitted Radiation Flux Density (W m-2)
16
14
12
10
8
6
4
2
0
800
600
400
200
0
-50
-40
-30
-20
-10
0
10
20
30
40
50