Lecture 5: Aerosol Optical Properties

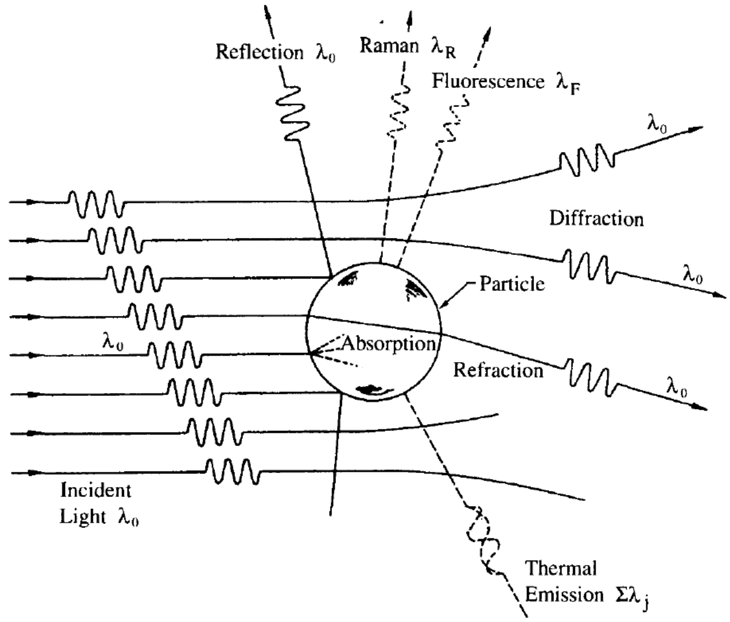
and Aerosol Cloud Interactions

1

Aerosol Optical Properties

2

Interaction of Radiation and Aerosol



3

3

Scattering of light:

$$\tilde{F}_{scat} = C_{scat} F_0$$

F_0 (w/m^2): incident intensity of radiation
C_{scat} (m^2) is the single-particle scattering cross section

Absorption:

$$\tilde{F}_{abs} = C_{abs} F_0$$

C_{abs} (m^2) is the single-particle absorption cross section

4

4

Extinction cross section:

$$C_{\text{ext}} = C_{\text{scat}} + C_{\text{abs}}$$

C_{ext} has the units of area; in the language of geometric optics, one would say that the particle casts a "shadow" of area C_{ext} on the radiative energy passing the particle. This shadow, however, can be considerably greater, or much less, than the particle's geometric shadow. The dimensionless *scattering efficiency* of a particle Q_{scat}, is C_{scat}/A, where A is the cross-sectional area of the particle. Defining Q_{abs} and Q_{ext} in the same way, we obtain

$$Q_{\text{ext}} = Q_{\text{scat}} + Q_{\text{abs}}$$

Extinction, scattering, and absorption efficiency

5

5

Single-scatter albedo

The ratio of Q_{scat} to Q_{ext} is called the *single-scattering albedo*,

$$\omega = \frac{Q_{\text{scat}}}{Q_{\text{ext}}} = \frac{C_{\text{scat}}}{C_{\text{ext}}}$$

Thus the fraction of light extinction that is scattered by a particle is ω, and the fraction absorbed is 1 - ω.

6

6

The key parameters that govern the scattering and absorption of light by a particle are

(1) the wavelength of the incident radiation;

(2) the size of the particle, and

(3) the particle optical property relative to the surrounding medium, the *complex refractive index:* $N = n + ik$

7

7

Size parameter:

$$\alpha = \frac{\pi D_p}{\lambda} = \frac{circumference}{wavelength}$$

Based on the α value, light scattering can be divided into three domains:

***α* « 1 *Rayleigh* scattering**

***α* ~ 1 *Mie* scattering**

***α* » 1 *Geometric* scattering**

8

Mie Scattering :

Scattering efficiency is maximum when particle radius = λ
⇨particles in 0.1-1 μm size range are efficient scatterers of solar radiation

By scattering solar radiation, aerosols decrease visibility and increase the Earth's albedo

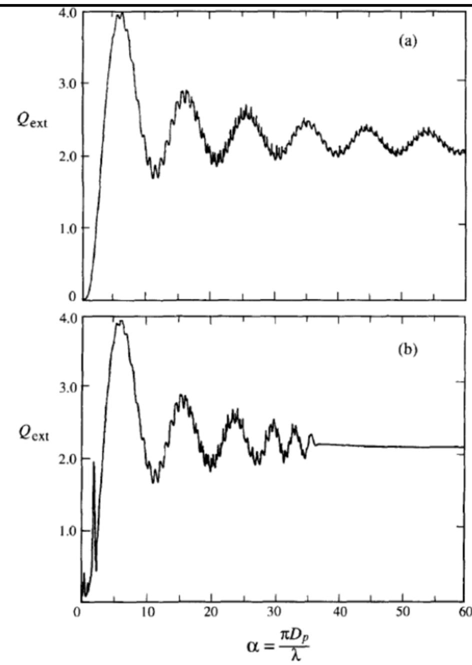


FIGURE 15.3 Extinction efficiency Q_{ext} for a water droplet: (a) wavelength of the radiation is constant at $\lambda = 0.5\,\mu m$ and diameter is varied; (b) diameter $= 2\,\mu m$ and wavelength is varied (Bohren and Huffman 1983).

9

Extinction by an Ensemble of Particles

Consider the solar beam traversing an atmospheric layer containing aerosol particles. Light extinction occurs by the attenuation of the incident light by scattering and absorption as it traverses the layer.

$$\frac{dF}{dz} = -b_{\text{ext}}F$$

b_{ext} is the extinction coefficient, having units of inverse length (m^{-1})

If F_0 is the incident flux at $z = 0$, taken to represent, say, the top of the layer, then intensity at any distance z into the layer is

$$\frac{F}{F_0} = \exp(-b_{\text{ext}}z) = \exp(-\tau)$$

$\tau = b_{\text{ext}}\, z$ is called the ***optical depth*** of the layer

10

For a population of monodisperse, spherical particles at a number concentration of *N,* the extinction coefficient is related to the dimensionless extinction efficiency by

$$b_{\text{ext}} = \frac{\pi D_p^2}{4} N Q_{\text{ext}}$$

With a population of different sized particles of identical refractive index *m* with a number size distribution function of *n(Dp),* the extinction coefficient is given by

$$b_{\text{ext}}(\lambda) = \int_0^{D_p^{\max}} \frac{\pi D_p^2}{4} Q_{\text{ext}}(m, \alpha) n(D_p) dD_p$$

11

$$\tau = \int_{\Delta z} b_{ext}(z) dz$$

$$b_{\text{ext}} = b_{\text{scat}} + b_{\text{abs}}$$

$$b_{\text{abs}} = b_{ag} + b_{ap}$$

$$b_{\text{scat}} = b_{sg} + b_{sp}$$

b_{sg} = scattering coefficient due to gases (the so-called Rayleigh scattering coefficient)
b_{sp} = scattering coefficient due to particles
b_{ag} = absorption coefficient due to gases
b_{ap} = absorption coefficient due to particles

12

$$\tau_p = \int_{\Delta z} b_{sp}(z) + \int_{\Delta z} b_{ap}(z) dz = \tau_{sp} + \tau_{ap}$$

13

Angstrom Exponent/Coefficient

$$\tau(AOD) = \beta\lambda^{-\alpha}$$

$$\alpha = -\frac{ln\frac{AOD_{\lambda 1}}{AOD_{\lambda 2}}}{ln\frac{\lambda 1}{\lambda 2}}$$

AOD = aerosol optical depth

Describes how AOD changes with wavelength

This is highly dependent upon which wavelengths you choose to calculate the Angstrom exponent

14

Angstrom Exponent/Coefficient

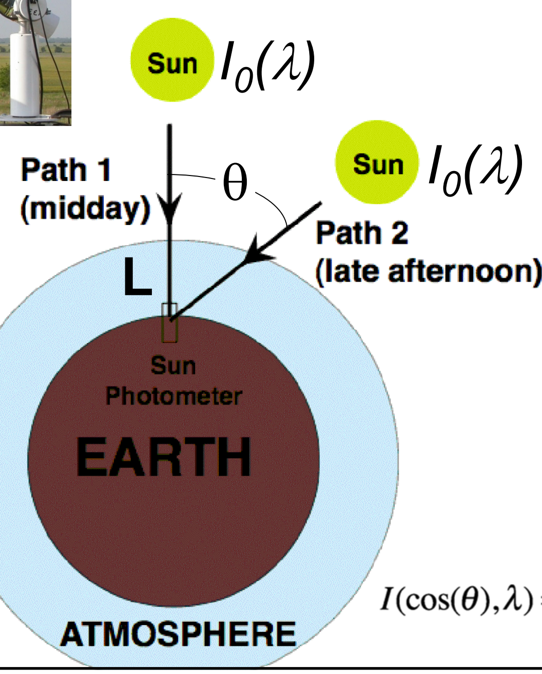
$$\tau(AOD) = \beta\lambda^{-\alpha}$$

Some typical values:

- Very coarse mode (large) aerosols: α is small (much less than 1) - e.g. dust and sea salt
- Accumulation mode (very small) aerosols: α is large (more than 1, and sometimes more than 2 - e.g. fresh smoke, urban aerosols
- Clouds: α=0

15

HOW TO OBSERVE AOD at SURFACE?



Simple Idea of Sun Photometers:

1. The sun's output arriving at the top of the atmosphere doesn't change from mid day to late afternoon.
2. The path length of sunlight through the atmosphere changes during the day.

$$I(\cos(\theta),\lambda) = I_0(\lambda)\exp\left(-\frac{\beta_{ext}L}{\cos(\theta)}\right)$$

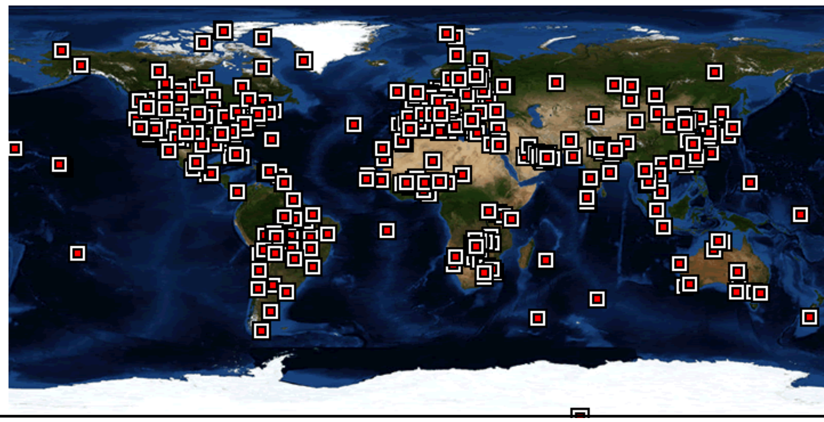
16

AERONET

http://aeronet.gsfc.nasa.gov/

AErosol **RO**botic **NET**work

Worldwide collection of sun photometers



17

Aerosol Cloud Interactions

18

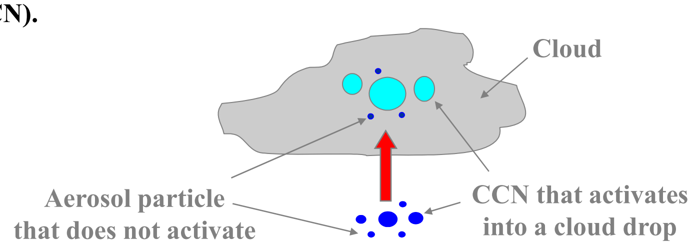
How do clouds form?

Clouds form in regions of the atmosphere where water vapor is supersaturated. We focus on liquid water clouds.

Water vapor supersaturation is generated by cooling (primarily through expansion in updraft regions and radiative cooling)

Cloud droplets form from pre-existing particles found in the atmosphere (aerosols). This process is known as *activation*.

Aerosols that can become droplets are called cloud condensation nuclei (CCN).



19

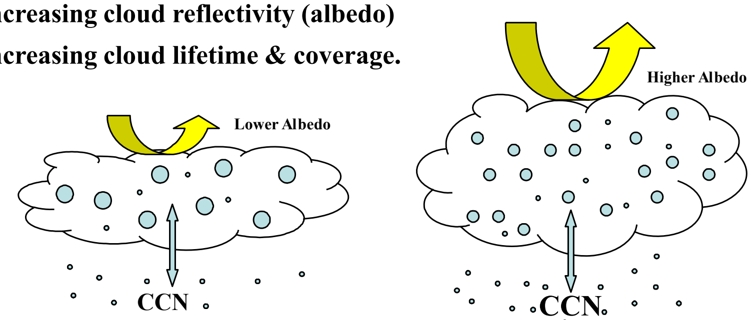
How can humans affect clouds?

By changing CCN; cloud properties are a *strong* function of their concentration.

This phenomenon is known as aerosol indirect effect.

The aerosol indirect effect can lead to *climatic cooling* by:

- Increasing cloud reflectivity (albedo)
- Increasing cloud lifetime & coverage.



Clean Environment (few CCN) | Polluted Environment (more CCN)

20

Aerosol Direct and Indirect Climate Effects

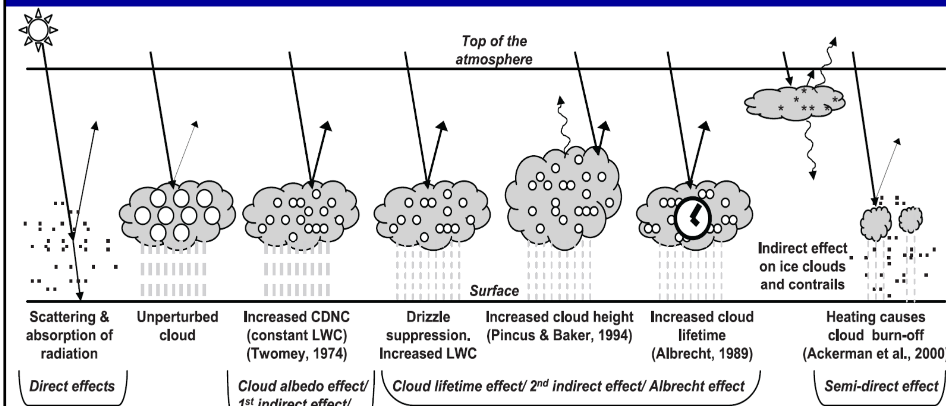


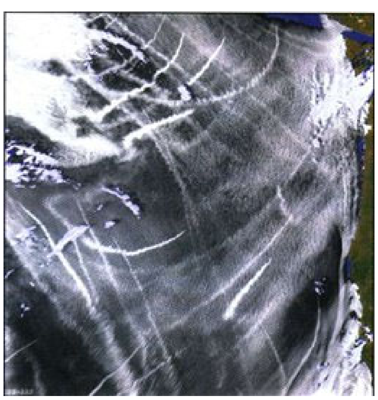
Figure 2.10. *Schematic diagram showing the various radiative mechanisms associated with cloud effects that have been identified as significant in relation to aerosols (modified from Haywood and Boucher, 2000). The small black dots represent aerosol particles; the larger open circles cloud droplets. Straight lines represent the incident and reflected solar radiation, and wavy lines represent terrestrial radiation. The filled white circles indicate cloud droplet number concentration (CDNC). The unperturbed cloud contains larger cloud drops as only natural aerosols are available as cloud condensation nuclei, while the perturbed cloud contains a greater number of smaller cloud drops as both natural and anthropogenic aerosols are available as cloud condensation nuclei (CCN). The vertical grey dashes represent rainfall, and LWC refers to the liquid water content.*

IPCC, 2007

21

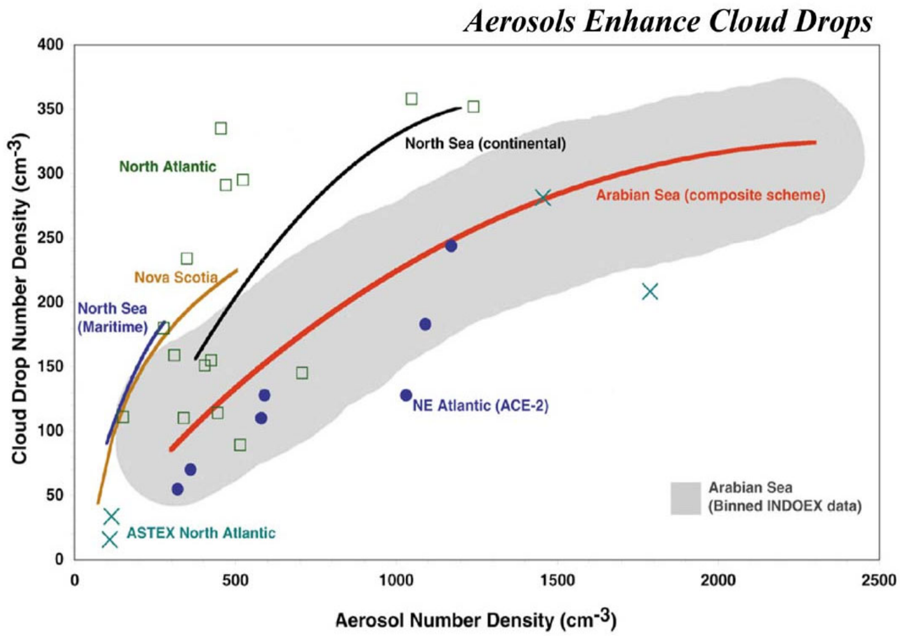
Aerosols and climate: indirect effect

“Ship Tracks” off the coast of Washington



- aerosols are the “seeds” upon which water vapor condenses to form a cloud (these are called “cloud condensation nuclei, or CCN).
- If people make more aerosols, we make more cloud droplets, but because there is a fixed amount of water vapor in the air these droplets will be smaller.
- smaller droplets scatter light more efficiently!
- smaller cloud droplets may also impact rain from these clouds.
- very difficult effect to observe and model!

22



Ref: Ramanathan, Crutzen, Kiehl and Rosenfeld, *Science* 2001.

23

Formation of Cloud droplets

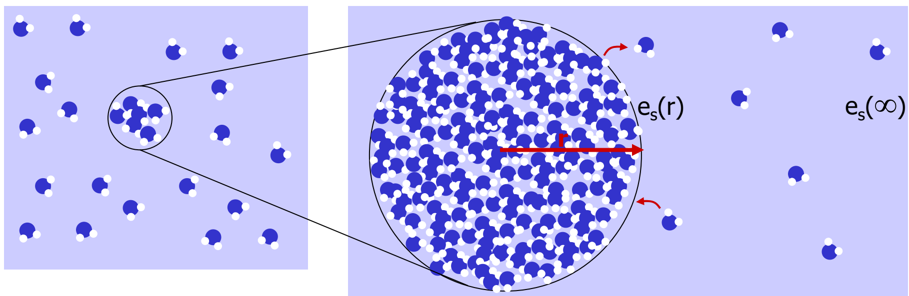
- Cloud drops form on aerosols
  - condensation nuclei or hygroscopic nuclei
- Rate of formation is determined by the number of these nuclei present.
- Nuclei keep supersaturation from exceeding a few percent.

24

Formation of Cloud Droplets

Why do cloud droplets form almost immediately upon reaching supersaturation?

In air containing water vapor above the saturation pressure, can chance collisions form a stable droplet of pure water?



$$e_s(r) = e_s(\infty)e^{\frac{2\sigma}{rR_v\rho_L T}}$$

Smaller drops require higher e_s for equilibrium

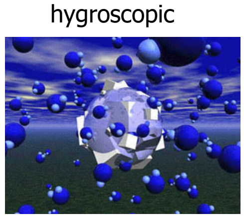
σ is the surface tension (energy/area or force/length)

25

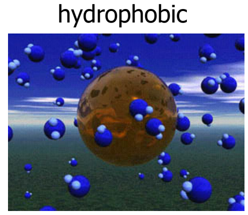
Nucleation (phase transition across a free energy barrier) of droplets requires a particle (condensation nucleus).

Hygroscopic nuclei are soluble in water and decrease $e_s(r)$ significantly.

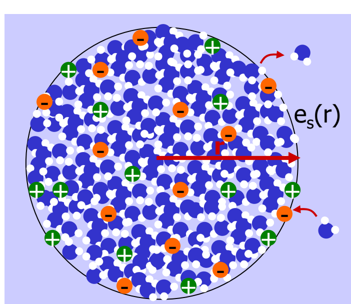
hygroscopic



hydrophobic



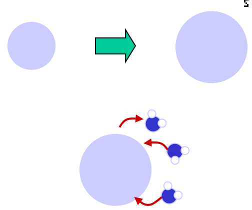
http://terra.nasa.gov/FactSheets/Clouds/



With non-water molecules on the surface, the equilibrium (equal transfer across the interface) occurs at lower pressure

26

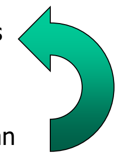
Curvature effect $\frac{e_s(r)}{e_s(\infty)} = e^{\frac{2\sigma}{r R_v \rho_L T}} = e^{\frac{a}{r}}$ (note: saturation vapor pressure is the vapor pressure required for equilibrium)



Increased r, decreases equilibrium/saturation vapor pressure over the drop (fewer molecules required outside the drop at equilibrium)

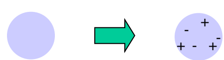
To attain this new equilibrium, vapor molecules will want to enter the drop at a higher rate than they leave (growth)

(Why does this process require supersaturation?)



$\boxed{\text{But this positive feedback can't get started at typical atmospheric saturation ratios.}}$

Solution effect $\frac{e'_s(r)}{e_s(\infty)} = \left[1 - \frac{b}{r^3}\right] e^{a/r} \approx 1 + \frac{a}{r} - \frac{b}{r^3}$

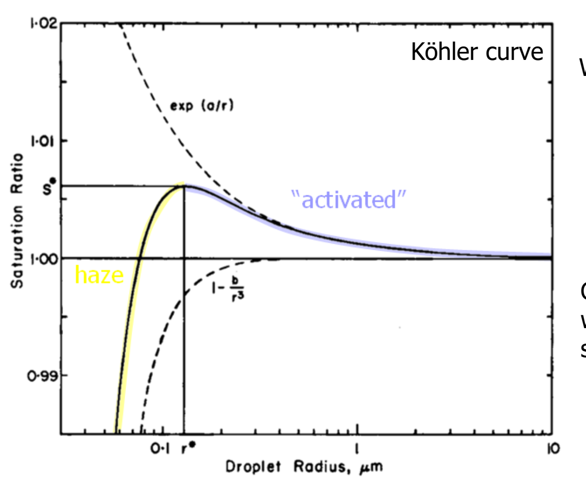


Adding solute, decreases equilibrium vapor pressure over the drop since fewer liquid molecules are available to escape (fewer molecules required outside at equilibrium)

To attain this new equilibrium, vapor molecules will want enter the drop at a higher rate than they leave (growth)

27

Now for a solution droplet (compared to a pure water plane surface) the equilibrium vapor pressure is increased due to curvature effects and decreased due to solution effects: $\frac{e'_s(r)}{e_s(\infty)} = \left[1 - \frac{b}{r^3}\right] e^{a/r} \approx 1 + \frac{a}{r} - \frac{b}{r^3}$



Equilibrium saturation ratio of a solution droplet formed on an ammonium sulfate condensation nucleus of mass 10^{-16} g.

Why does the Köhler curve approach 1.0 for large r?

Growth does not continue without bound since drops start to compete for vapor

28

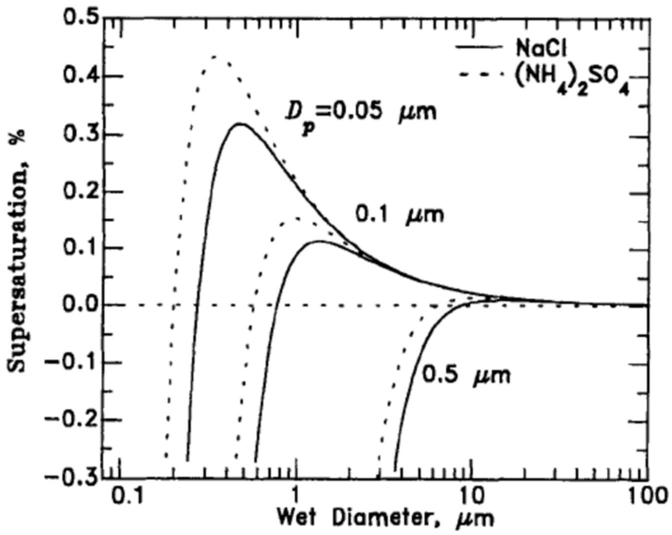


FIGURE 17.5 Köhler curves for NaCl and $(NH_4)_2SO_4$ particles with dry diameters 0.05,0.1, and 0.5 μm at 293 K (assuming spherical dry particles). The supersaturation is defined as the saturation minus one. For example, a supersaturation of 1% corresponds to a relative humidity of 101%.

29

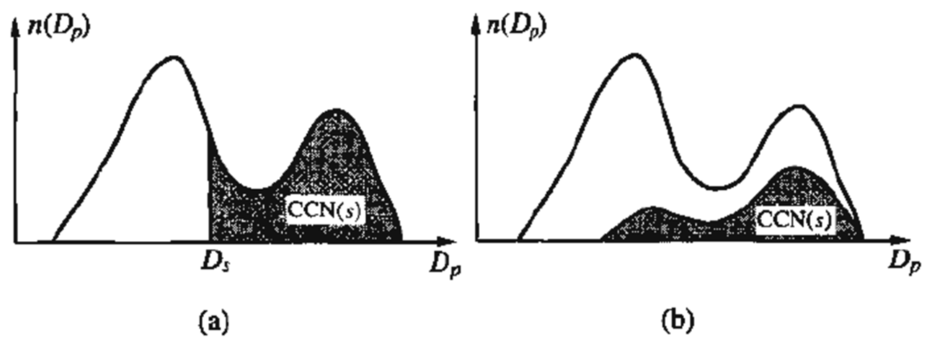


FIGURE 17.16 Schematic of an aerosol size distribution with the shaded area indicating the CCN for (a) uniform chemical composition and (b) a typical multicomponent aerosol population that is neither externally nor internally mixed and has a size-dependent composition.

30

Effect of aerosol on cloud optical thickness

$$\tau = \Delta z \int_0^\infty Q_e \pi\, r^2 n(r)\, dr \approx 2\pi\, \Delta z\, R_{eff}^2 N$$

$$(Q_e \sim 2)$$

$$\mathrm{LWC} = \rho \int_0^\infty \frac{4}{3}\pi\, r^3 n(r)\, dr \approx \frac{4}{3}\pi\rho R_{eff}^3 N$$

$$\tau \approx \frac{3}{2}\frac{\mathrm{LWC}\,\Delta z}{\rho R_{eff}} = 2.4\Delta z\left(\frac{\mathrm{LWC}}{\rho}\right)^{2/3} N^{1/3}$$

$$\frac{\Delta\tau}{\tau} = \frac{1}{3}\frac{\Delta N}{N}$$

31

Effect of aerosol on cloud albedo

The albedo A of a cloud is given to a good approximation by

$$A = \frac{(1-g)\tau}{1+(1-g)\tau}$$

For the scattering of solar radiation by clouds, $g \approx 0.85$.

$$A \approx \frac{\tau}{6.7+\tau}$$

Provided the LWC and cloud depth are held constant,

$$\frac{\Delta A}{\Delta N} = \frac{A(1-A)}{3N}$$

32

Effect of aerosol on cloud albedo

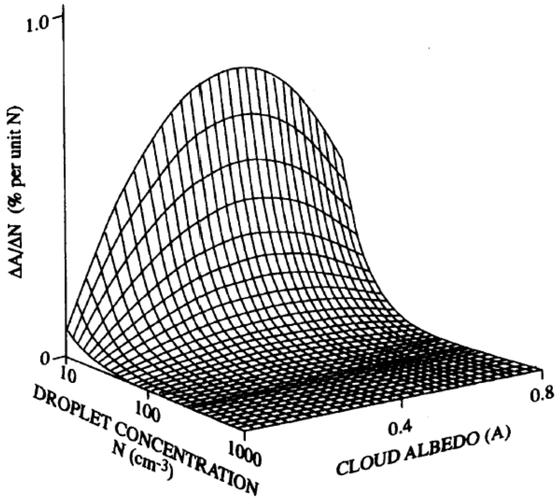


Figure 6 The change in the albedo of a cloud per unit change in the droplet concentration ($\Delta A/\Delta N$) as a function of the cloud albedo (A) and the droplet concentration (N), for a cloud with a constant liquid-water content. (Adapted from Twomey, 1991.)

From Hobbs, 1993.

33