

CHAPTER II

AEROSOLS



2.1 Introduction

The atmosphere is composed of molecules of gases, small solid and liquid particles suspended in the air, called aerosols. Aerosols have both natural and anthropogenic origin. The size of these particles are ranged in radius from 0.001 to 100 μm and divided into 2 modes; fine mode (radius equal or less than 1 μm) and coarse mode (radius more than 1 μm). Also fine mode is divided into the nucleation mode (or Aitken particles, radius less than 0.1 μm) and accumulation mode (radius 0.1 – 1 μm). Aerosols could be classified according to location and chemical type as follows;

- Location: marine, continental, rural, industrial, polar, desert, stratospheric, tropospheric aerosols, etc.
- Chemical type: sulphate (SO_4^{2-}), nitrate (NO_3^-), soot (elemental carbon), sea-salt (NaCl), etc.

Most aerosols are in the troposphere, but large volcanic eruptions inject aerosols and gases into the stratosphere (Hobbs, 1993).

2.2 How Aerosols are Produced?

Aerosols are divided into two broad categories based on their method of creation: primary aerosols and secondary aerosols.

Primary aerosols are particles that are emitted directly into the atmosphere. The aerosols are produced by processes that occur in both natural and anthropogenic. Processes that introduce particles into the atmosphere, including wind lifting, breaking wave and bubble, burning of biomass and fossil fuel, volcano, agriculture and industry.

Secondary aerosols are particles that are produced in the atmosphere by chemical or physical transformations of precursor gases. Precursor gases are created by natural and anthropogenic processes. Processes that introduce precursor gas into the atmosphere, including burning of biomass and fossil fuel, emission from marine, soil and vegetation, volcano.

Due to the wide variety of sources, properties of atmospheric aerosol particle such as shape, chemical composition and size may be very heterogeneous (Voutilainen, 2001). Aerosol concentrations vary with location and time. The high concentrations are usually found in urban areas. The events such as dust storms and volcanic eruption also enhance aerosol concentrations. Once aerosols are in the atmosphere, their size and composition can be changed due to transformation processes, being transported in the atmosphere and being removed by dry and wet deposition. These processes depend on many factors including aerosol size, concentration, chemical composition, location and meteorological condition.

2.3 Major Physical Properties of Atmospheric Aerosol

2.3.1 Scatter, Absorb and Emit Electromagnetic Radiation

Aerosol particles can scatter and absorb electromagnetic radiation at different wavelengths. Scattering is a process which conserves the total amount of energy, but change direction of radiation. Absorption is a process that energy is converted to another form inside the particles, or re-radiated at different wavelength that is called emitting. The sum of scattering and absorption is called extinction or attenuation.

All aerosol particles scatter radiation. Some particles that made of absorbing material can also absorb radiation. The scattering and absorption of radiation by single aerosol particle is expressed by the refractive index as a complex number of the form $m=n-ik$, where the real part (n) represents scattering and the imaginary part (k) represents absorption (Reist, 1993). Refractive indices of various materials were given in Appendix B.

Consider the solar radiation traversing through atmospheric layer that contains aerosol particles; the extinction occurs by scattering and absorption as the radiation traverses the layer. The loss of intensity (F) over a depth of layer dz can be express as

$$dF = -b_{ext} F dz \quad (2.1)$$

where b_{ext} is the extinction coefficient (metre⁻¹).

The aerosol extinction coefficient can be expressed as a function of particle size, mass concentration, and chemical composition as follows:

$$b_{ext} = \int_{D_p, \min}^{D_p, \max} N_m(D_p) E_{ext}(D_p, \lambda, m) dD_p \quad (2.2)$$

where $N_m(D_p)$ is the mass size distribution function ($\text{gm}^{-3} \mu\text{m}^{-1}$), $E_{ext}(D_p, \lambda, m)$ is the mass extinction efficiency ($\text{m}^2 \text{g}^{-1}$) which is a function of particle diameter, D_p , wavelength, λ , and refractive index, m . The extinction coefficient can be expressed as the sum of scattering coefficient, b_{scat} and an absorption coefficient, b_{abs} . Equation 2.2 can be rewritten as follows:

$$b_{ext} = \int_{D_p, \min}^{D_p, \max} N_m(D_p) E_{scat}(D_p, \lambda, m) dD_p + \int_{D_p, \min}^{D_p, \max} N_m(D_p) E_{abs}(D_p, \lambda, m) dD_p \quad (2.3)$$

where $E_{scat}(D_p, \lambda, m)$ and $E_{abs}(D_p, \lambda, m)$ are the mass scattering and absorption efficiencies, respectively.

The mass scattering and absorption efficiencies at a wavelength of 550 nm for ammonium sulfate and soot aerosols are shown as function of particles diameter in Figure 2.1.

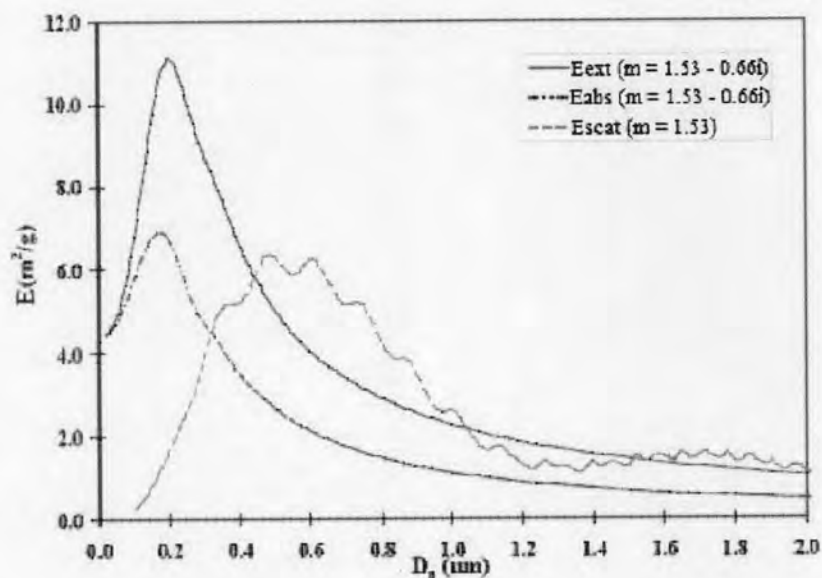


Figure 2.1 Mass scattering efficiency (550 nm) for a purely scattering aerosol ($m = 1.53$), and the mass absorption and scattering efficiency for an aerosol that has both scattering and absorbing components ($m = 1.53 - 0.66i$)

(From Aerosol Radiative Properties and Their Impacts, Bergin, 2006)

For the scattering aerosol ($m = 1.53$) the peak in the mass scattering efficiency occurs roughly where the aerosol diameter is equivalent to the wavelength of light (i.e. 550 nm). The mass scattering efficiency rapidly decreases for particles having diameters less than or greater than the wavelength of light. For the absorbing aerosol, the shape of the mass extinction curve is similar to that of the scattering aerosol, although there is an absorbing component in addition to light scattering. The peak in the mass extinction efficiency tends to be at a smaller diameter than for the scattering aerosol (Bergin, 2006). So the scattering is maximum for a scattering particle diameter corresponding to the wavelength of radiation. The location of the peak for the mass extinction efficiency at a given wavelength depends on refractive index. At any given value of n (real part), increasing k (imaginary part), the diameter at which the mass extinction efficiency peak occurrence will decrease (Seinfeld and Pandis, 1997). Atmospheric aerosols in the

accumulation mode are efficient scatters of solar radiation because their size is of the same order of wavelength of radiation.

2.3.2 Cloud Condensation Nuclei

Aerosol particles are necessary for cloud formation. Aerosols that are comprised of water soluble material are efficient cloud condensation nuclei. For examples water soluble aerosols containing sulphate, nitrate and sea salt. Water soluble aerosols are hygroscopic that are capable of attracting water vapor from the air.

2.4 Transformation Processes

Processes of precursor gases are converted to aerosol and evolution of aerosol in the atmosphere include:

2.4.1 Gas to Particle Conversion (GPC): Besides the aerosols that are directly emitted into the air, aerosol particles are produced by GPC, which occurs in the atmosphere. GPC consist of two types; homogeneous and heterogeneous nucleation. Homogeneous nucleation takes place when new particles are produced by condensation of precursor gases. The process generally requires that the involved gas vapors occur at high supersaturation. Heterogeneous nucleation takes place when gas molecules condense on pre-existing particles. The process occurrence requires only minor supersaturation of gases (Hobbs, 1993).

2.4.2 Cloud Processing: Aerosols play an important role in the formation of clouds. Clouds were formed when water vapor condenses onto water soluble aerosol knows as cloud condensation nuclei (CCN). When cloud droplets are evaporated, aerosol particles are left behind and suspended in the air. The particles are different from the original CCN, both chemically and physically properties (Seinfeld and Pandis, 1997).

2.4.3 Coagulation: The process that aerosol particles collide with one another and stick together to form larger particles. Coagulation leads to a removal of small particles replaced by larger ones.

2.5 Removal of Aerosol from the Atmosphere

The main aerosols removal processes in the atmosphere are dry and wet deposition. Deposition decreases the number concentration of aerosol particles in the air.

Dry deposition is the transfer of particles from the atmosphere to the earth's surface without the aid of precipitation. Dry deposition process depends on atmospheric stability, physical and chemical properties of the depositing particle and the specific properties of the contacting surface. Particle dry deposition processes such as gravitational settling (particles fall down due to gravitation), impaction, interception and Brownian diffusion (aerosol particles move randomly due to collisions with gas molecules; this may lead to collision with obstacles or surface).

Wet deposition combines all natural processes that the aerosol particles are scavenged by atmospheric hydrometeors (cloud and fog drops, rain, snow) and transferred to the earth's surface. Seinfeld and Pandis (1997) refer to wet deposition process by hydrometeors as follows:

- 1) Precipitation scavenging
 - 1.1) In-cloud scavenging: aerosols are removed via nucleation scavenging (growth of CCN to cloud drops) and then collection of a fraction of the remaining aerosol by cloud or rain droplets.
 - 1.2) Below-cloud scavenging: As a raindrop fall through air, it collides with aerosol particles and collects them.
- 2) Cloud interception: the impaction of cloud droplets on the terrain usually occurs at the top of tall mountains.
- 3) Fog deposition: aerosols are removed by settling of fog droplets.

The residence time of aerosol particles depends on their sizes, chemistry and height in the atmosphere. Particle residence times range from minutes to hundreds of days. The accumulation mode of aerosols remains in the atmosphere longer than the two size categories and are removed primarily by precipitation scavenging processes. The nucleation mode of aerosols is high rates of particle collision and coagulation, increases the size of individual particles and removes them from the nucleation mode. The coarser particles are mainly removed by sedimentation and have shorter atmospheric residence time than the other two size ranges.

2.6 Categories of Aerosols

Aerosols can be categorized according to source as follows:

2.6.1 Soil Dust

Dust particles enter the atmosphere directly by the action of wind. Dust source regions such as deserts, dry lake beds, and areas in drier regions where vegetation has been reduced or soil surfaces have been disturbed by human activities. Major dust sources are found in the desert regions of the Northern Hemisphere, while dust emissions in the Southern Hemisphere are relatively small. The atmospheric lifetime of dust depend on particle size: large particles are quickly removed from the atmosphere by the gravitational settling, while sub-micron sized particles can have atmospheric life time of several weeks (Intergovernmental Panel on Climate Change, 2001). Although dust particles are generally large size (more than 1 μm), they can be transported to long distances from their sources (The National science digital library, 2005).

2.6.2 Industrial Dust, Primary Anthropogenic Aerosol

Industrial and technical activities produce primary aerosol particles such as transportation, coal combustion, cement manufacturing, metallurgy, and waste incineration. Considering the source strength and fact that much industrial dust is present in a size fraction that is not optically very active (diameter more than 1 μm), it is

probable not of climatic importance. On the other hand, the growing industrialization without stringent emission controls, especially in Asia, may lead to increase in this source in next few decades (Wolf and Hidy, 1997).

2.6.3 Sea Salt

Sea salt aerosols are generated by an evaporation of sea spray produced by bursting bubble during whitecap formation and wave breaking. The strength of the ocean as a source of salt particles depends on wind speed (Hobbs, 1993). These aerosol particles are very efficient cloud condensation nuclei. Sea salt particles cover wide size range (about 0.05 to 10 μm diameter), and have a corresponding wide range of atmospheric lifetime (Intergovernmental Panel on Climate Change, 2001).

2.6.4 Aerosol from Fossil Fuel Burning

Aerosol particles and precursor gases are produced by fossil fuel burning from both stationary sources, such as factories and power plants, and mobile source, such as motor vehicles. The primary aerosol particles that are released from fossil fuel burning are soot and fly ash. Soot is carbonaceous that includes both elemental (black carbon) and organic compounds. Soot particles generally belong to the fine particle mode. The mass median diameter of black carbon in the atmosphere between 0.1 and 0.5 μm . Fry ash is non-organic and generally belongs to the coarse particle mode (The National science digital library, 2005).

2.6.5 Aerosol from Biomass Burning

Biomass burning is a globally importance source of aerosols, which consists of burning of savanna or forest during a shifting agriculture and permanent land clearance as well as the burning of fuel wood and agricultural waste.

Biomass burning produces both primary aerosol particles and precursor gases for secondary aerosol formation. The primary aerosol particles that are injected into the atmosphere during burning consist of soot particles (carbonaceous particles) and fly ash. Organics are the largest component of biomass burning aerosols (Andreae et al., 1988).

2.6.6 Biogenic Aerosols

Primary biogenic aerosols are particles that are produced from biosphere or microbes that are dispersed into the atmosphere. The primary biogenic particles can be categorized as follows:

1. Living/viable particle: bacteria, viruses, protozoa, fungi, pollen, spores, algae
2. Nonliving particle: dead microbes, dead skin, and leaf abrasions

Pollen, spores, and fragments of animals and plants are usually in the micrometer size range, while bacteria, algae, protozoa, fungi, and viruses are smaller (Hobbs, 1993)

Secondary biogenic aerosols are formed in the atmosphere by the gas to particle conversion of volatile organic compounds (VOC) that are released from terrestrial and marine biosphere. The biogenically produced compound can result in the formation of secondary aerosol particles such as isoprene and terpenes that are the most common released from vegetation, dimethylsulphine (DSM) from phytoplankton, Nitrous oxide (N_2O) from microbiological process in soil and natural water.

2.6.7 Sulphate

Sulphate aerosols are produced by chemical reactions in the atmosphere from gaseous precursors (with the exception of sea salt sulphate and gypsum dust particles). The two main sulphate precursors are SO_2 from anthropogenic sources and volcanoes, and dimethylsulphide (DMS) from biogenic sources, especially marine plankton. SO_2 emissions are mostly related to fossil fuel burning. Most SO_2 is converted to sulphate either in gas phase or in cloud droplets that later evaporate. Both processes produce sulphate mostly in sub-micron aerosols (Intergovernmental Panel on Climate Change, 2001).

2.6.8 Nitrate

Most nitrate aerosols in the troposphere are resulted from cloud droplets evaporate. The nitrate particles have radius much larger than $1 \mu m$ (Hobbs, 1993). Sub-micron nitrate is predominantly present in the form of ammonium nitrate (NH_4NO_3). If ammonia is available in excess of the amount required to neutralize sulphuric acid

(H_2SO_4) to form ammonium sulphate ($(\text{NH}_4)_2\text{SO}_4$), the remaining ammonia can combine with nitric acid (HNO_3) to form ammonium nitrate (Schaap et al., 2003). Nitric acid also deposits on existing aerosol, such as alkaline mineral or sea salt aerosols.

Nitric acid derives from Nitrous oxide (N_2O) that is emitted by biological sources in soil and water, biomass burning and fossil fuel use, etc. N_2O decomposes chemically to nitrogen (N_2) and nitric oxide (NO). NO is oxidized to nitrogen dioxide (NO_2), next NO_2 reacts with OH radical to produce nitric acid (Hobbs, 1993).

2.6.9. Volcanoes

Volcanic eruptions usually provide two components of most significance for aerosols: primary dust and gaseous sulphur. Volcanic eruptions can lead to an increase in the amount of the stratospheric aerosol. The stratospheric lifetime of these coarse particles is only about 1 to 2 months due to the efficient removal by settling (Intergovernmental Panel on Climate Change, 2001). Sulphur emission from volcanoes has a longer live effect on stratospheric aerosol loads.

2.7 Effect of Aerosol

Aerosols influence the atmospheric energy budget through direct and indirect effects. Direct effects refer to scattering and absorption of radiation. Back scattering of solar radiation by aerosols enhance the planetary albedo, causing cooling of the atmosphere, while absorption by aerosol such as carbon cause heating. Indirect effects refer to the increase in available cloud condensation nuclei due to an increase in anthropogenic aerosol concentration. This is believed to change the number of cloud droplet concentration for constant cloud liquid water content, and the resulting increase in cloud albedo influences the earth's radiation budget. Cloud lifetimes and precipitation frequencies are also thought to be effected (World Meteorological Organization, 2001).

Airborne aerosols also effect human health. The toxicity of particles depends on their chemical compositions and their sizes. The size of airborne particles is significant

as this determines where is in the respiratory tract. The particles are deposited when we breathe in. Larger particles are generally filtered by small hair in the nose and throat and do not cause problems. But particulate matter smaller than $10\ \mu\text{m}$ in size, referred to as PM10 can settle in the bronchia and lung and cause health problem. Particulate matter smaller than $2.5\ \mu\text{m}$ in size, PM2.5, can penetrate directly into the lung. They can cause breathing and respiratory problem, irritation, inflammation, cancer and cardio-vascular problems (Wikipedia, 2001; International Max Plance Research School, 2005). The composition determines either how the respiratory tract reacts or how the body responds. Toxic air pollutants attach themselves to airborne particle. This particle is breathed into the lungs and the toxic species are absorbed into blood and body tissues.

Aerosols are also linked to problems of visibility reduction, acid rain, and urban smog in many locations of the world.

2.8 Aerosol Optical Properties

This study focuses on the variation of aerosol in Bangkok, the urban area with a lot aerosol sources, by analyzed the time variation of aerosol optical data with consist of aerosol optical thickness, angstrom exponent and aerosol volume size distribution. Therefore, it is necessary to discuss the terms here.

2.8.1 Aerosol Optical Thickness (AOT)

When solar radiation enters the earth's atmosphere, a part of the incident energy is removed by scattering and a part is absorbed. Both influence the extraterrestrial spectrum by considerably modifying the spectral energy passing through the atmosphere. The scattered radiation is called diffuse radiation. The scattering by particles much smaller than the wavelength of incident radiation is Rayleigh scattering, such as air molecules. The scattering by the particle size of the order of the incident radiation is Mie scattering, such as aerosol particles. A portion of this diffuse radiation goes back to space while the other portion reaches the ground. The radiation arriving on

the ground directly in line with the sun is called direct or beam radiation. Figure 2.2 illustrates distribution of direct, diffuse and absorbed solar radiation. When passing through the earth's atmosphere, solar radiation is attenuated due to scattering and absorption by air molecules and aerosols.

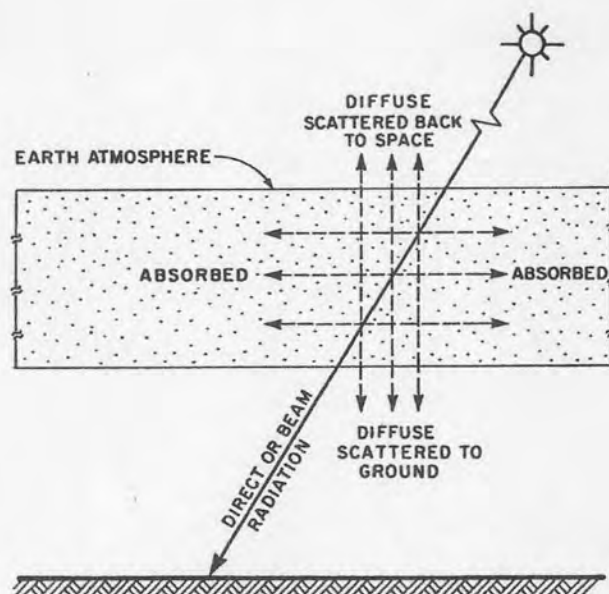


Figure 2.2 Distribution of direct, diffuse and absorbed solar radiation.

(From An Introduction to Solar radiation, Iqbal, 1983)

Absorption of solar radiation by aerosol is a continuous function of the wavelength. Absorption of solar radiation by gas is selective process. The most significant absorbing gases in the atmosphere are O_2 , O_3 , H_2O and CO_2 . O_2 and O_3 effectively absorb all solar radiation that is below 290 nm. As a result, solar radiation of wavelength shorter than 290 nm does not reach the earth surface. O_3 has a weaker absorption band from 300 to 350 nm and 450 to 770 nm. However, atmospheric absorption is not strong from 300 to 800 nm, forming a "window" in the spectrum. H_2O and CO_2 absorb much of the long-wave terrestrial radiation. The absorption spectra for various molecules were shown on Appendix C.

Aerosol optical thickness (AOT) describes the attenuation due to scattering and absorption of solar radiation by a column of aerosol at certain wavelengths. In general, scattering is much greater than absorption (Iqbal, 1983). The interaction of aerosols with the incoming solar radiation is strongly dependent on size of aerosol and the wavelength of radiation (Lata et al., 2003). The aerosol extinction is generally maxima for a scattering particle diameter corresponding to the wavelength of radiation. Measurement of AOT can provide important information about the concentration and variability of aerosol in the atmosphere.

AOT is determined from spectral extinction of solar radiation in the atmosphere at each wavelength (total optical thickness) by Lambert-Beer-Bouguer law (Iqbal, 1983) as shown:

$$I_{\lambda} = I_{0\lambda} \exp(-\tau_{\lambda} m) \quad (2.1)$$

where I_{λ} is the solar irradiance at wavelength λ

$I_{0\lambda}$ is the extraterrestrial irradiance at wavelength λ

τ_{λ} is the total optical thickness at wavelength λ

$m = \sec\theta_z$ where θ_z is the angle between the local zenith and the solar position, m is equals 1 when the sun is in the vertical direction.

The total optical thickness is composed of extinction optical thickness from several sources, such as scattering by air molecules, absorption of gases and extinction by aerosol particles. Total optical thickness is derived as follows;

$$\tau_{\lambda} = \tau_{R\lambda} + \tau_{G\lambda} + \tau_{A\lambda} \quad (2.2)$$

where $\tau_{R\lambda}$ is the Rayleigh optical thickness (for scattering by air molecules)

$\tau_{G\lambda}$ is the absorption optical thickness due to gases in the atmosphere

$\tau_{A\lambda}$ is the aerosol optical thickness

Then the aerosol optical thickness is obtained by subtraction of the other optical thicknesses from the total optical thickness as:

$$\tau_{A\lambda} = \tau_{\lambda} - \tau_{R\lambda} - \tau_{G\lambda} \quad (2.3)$$

The spectral ranges use for the AOT estimation do not include gas absorption spectral regions, except for the ozone and its broad semi-continuum absorption band extending from about 0.45 to 0.70 μm . (Esposito et al., 2004)

2.8.2 Angstrom Exponent

Angstrom exponent (α) is commonly used to characterize the wavelength dependence of AOT and to provide some basic information on aerosol size. Angstrom exponent is related to size distribution of the aerosol particles. Large value of α indicate a relatively high ratio of small particles to large particles. It appears obvious that α should vary from 4 to 0; when the aerosol particles are very small, of the order of air molecules, α should approach 4, and it should approach 0 for very large particles. Generally, α has a value of between 0.5 and 2.5 (Iqbal, 1983).

Angstrom exponent is derived using Angstrom turbidity formula (Angstrom, 1929) as shown:

$$\tau_{A\lambda} = \beta\lambda^{-\alpha} \quad (2.4)$$

where β is Angstrom turbidity coefficient, an index representing the amount of aerosols present in the atmosphere in the vertical direction

λ is wavelength in micrometers

From Equation (2.4) it can be facilitated by use of a linearized form i.e., $\ln \tau_{A\lambda} = \ln \beta - \alpha \ln \lambda$.

Then α is derives from slope of $\ln \tau_{A\lambda}$ versus $\ln \lambda$

2.8.3 Size Distribution

It is valuable to be able to express the number of particles of a given size for all the particle sizes in the atmospheric aerosol. The size distribution of particles can be described by the size distribution function $n(r)$ as follows:

$$n(r) = dN / dr \quad (2.5)$$

where $n(r)$ is number of particles per unit volume of air having radii in the range r and $r + dr$

The total number of particle per unit volume (the total number concentration), N , is

$$N = \int_0^{\infty} n(r) dr \quad (2.6)$$

Similarly, the mass of particles per unit volume in the range r and $r + dr$ is

$$m(r) = dM / dr \quad (2.7)$$

where $m(r)$ is the mass distribution function.

The total mass concentration is

$$M = \int_0^{\infty} m(r) dr \quad (2.8)$$

In addition, the dispersion of aerosol is sometimes expressed in term of surface area and volume distribution.

Since the size range of aerosol particles may cover several orders of magnitude, the size distribution function is usually expressed as a function of logarithm of particle radius, as shown:

$$n(r) = \frac{dN}{d \ln r} \quad (2.9)$$

The surface area and volume distribution can be express as function of logarithm of particle radius, as follows:

$$\frac{dS}{d \ln r} = S \frac{dN}{d \ln r} = 2\pi r^2 \frac{dN}{d \ln r} \quad (2.10)$$

$$\frac{dV}{d \ln r} = V \frac{dN}{d \ln r} = \frac{4}{3} \pi r^3 \frac{dN}{d \ln r} \quad (2.11)$$

The size distribution itself cannot be measured directly, an estimate for the size distribution is determined from indirect observations using computational methods. The problem of making inferences about a quantity that can not be observed directly is called an inverse problem.

2.9 Aerosol Measurement

The aerosol optical data in this study derive from measurement through the Aerosol Robotic Network (AERONET) by automatic sun and sky scanning radiometer (CIMEL). AERONET is an optical ground base aerosol monitoring network. There are 5 AERONET sites in Thailand which are:

- Omkoi, Chiang Mai (17.8 °N, 94.8 °E, Elevation 1120 m)
- Mukdahan (16.6 °N, 104.7 °E, Elevation 166 m)
- Pimai, Nakorn Ratchasima (15.2 °N, 102.6 °E, Elevation 220 m)
- Chulalongkorn , Bangkok (13.7 °N, 100.5 °E, Elevation 115 m)
- Hua_Hin, Prachuab Khirikhan (12.6 °N, 99.9 °E, Elevation 1 m)

The location of the AERONET sites in Thailand as shown on figure 2.3.

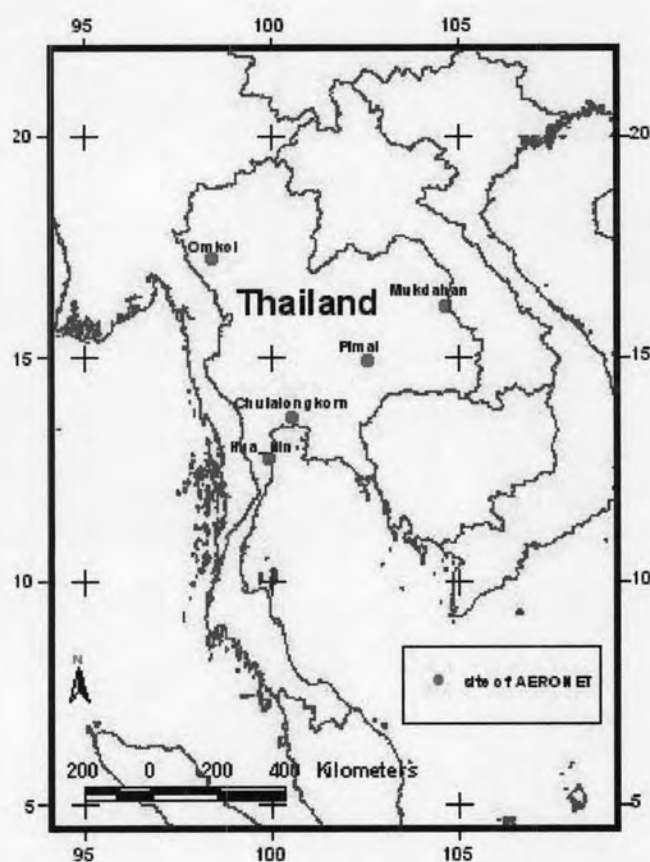


Figure 2.3 The AERONET sites in Thailand.

The CIMEL radiometer made the measurements of direct sun every 15 minutes in eight spectral bands (340, 380, 440, 500, 670, 870, 940 and 1020 nm) and sky radiance in four spectral bands (440, 670, 870 and 1020 nm). Seven of the eight direct spectral bands (excluding 940 nm) are used to acquire aerosol optical thickness data. The band at 940 nm is used to estimate total precipitable water content. Sky radiance measurements are used to retrieve aerosol volume size distributions over a range of size from 0.05 – 15 μm with the Dubovik inversion (Dubovik and King, 2000 and Dubovik et al., 2000).

Beside the aerosol optical data, the data of aerosol particle mass concentration have been used in this study. The measurements of aerosol particle mass concentration have been done by Pollution Control Department in Bangkok and other urban areas. They measure the total suspended particulates (TSP), the particle with an aerodynamic diameter of less than 10 microns (PM₁₀) and 2.5 microns (PM_{2.5}) at the surface.

The technique for measuring particle mass concentration involves filtration. An air pump draws ambient air flow through the filter. Each Filter is weighed in controlled temperature and humidity conditions before and after sampling. The mass concentrations are determined from an increase in filter mass and the volume of air sampled. This measurement is the manual gravimetric method.

The other method for measurement of the mass concentration is automated instrument such as beta attenuator method and Tapered Element Oscillating Microbalance. For the beta attenuator method, the instrument uses a paper filter tape that advances automatically at the end of the sample period. First measuring the attenuation of beta radiation is passed through the unexposed filter paper, and then measuring the attenuation after the tape has passed through the ambient air flow and the particulates are caught by filter paper. The attenuation of the beta radiation is proportional to the total sample mass. The attenuation measurement converts to a

measurement of the mass on the filter, so the filters are not required later laboratory analysis for mass variable. For Tapered Element Oscillating Microbalance, air is drawn through a tapered glass element with a filter attached. The element oscillates according to a characteristic frequency that decreases as mass accumulate on the attached filter. Measurement of the change in frequency converts to measurement of the accumulated mass.

2.10 Literature Review

Cachorro et al. (2000) have studied turbidity atmospheric condition during March to November 1995, at a rural station of the "Castilla y Leon" region in Spain of continental climate type. The solar direct irradiance data that were measured with a spectroradiometer in the 300-1100 nm range were used to determine the AOT as turbidity index and the Angstrom turbidity parameters (α and β) as a simple preliminary aerosol characterization. The data set of AOT at defined wavelength, 550 and 1000 nm, as well as Angstrom turbidity parameter were analyzed by means of instantaneous, monthly mean values and frequency histograms that gave the range and variability of turbidity due to background (non-industrial local source) atmospheric aerosols. The AOT values at 550 nm showed the high variability, ranging from 0.04 to 0.65 with the highest values in summer and the average value of 0.21 (S.D. = 0.11) for the 550 nm wavelength and 0.08 (S.D. = 0.06) for 1000 nm wavelength. The spectral features of AOT were shown by means of representative spectra for different turbidity condition. The AOT values showed a general decreasing tendency with wavelength. The α parameter values ranged from -0.5 to 2.6 while β parameter values ranged from 0.05 to 0.8. The Angstrom turbidity parameters seemed to indicate small-sized particles as components of background and continental boundary layer that had been polluted by sources of non-local origin. The obtained values of these turbidity parameters served as a reference data to estimate the climatological behavior of atmospheric aerosols in the area of study.

Esposito et al. (2004) used solar irradiance data measuring by spectroradiometer in the visible range (400-800 nm) during June-October 2001 in Tito Scalo, South Italy, the characteristic of continental climate, to estimate AOT, Angstrom turbidity parameter (α and β) and aerosol size distribution. They found that the lowest value of the AOT have been recorded in July, while the highest correspond to August. The back-trajectories revealed that in July air-masses came from the Atlantic Ocean, while in August air-masses mostly came from North-Eastern Europe. The injection of air particles arriving from different site caused the variation of the aerosol optical properties. The mean value of AOT at wavelength of 480 and 780 nm were 0.27 (S.D. 0.09) and 0.15 (S.D. 0.06), respectively. The Angstrom turbidity parameter α and β were high value in August, in general, high α values correspond to low β value. The α and β variation ranges were 0.61-2.73 and 0.03-0.21, respectively. The aerosol size distributions revealed a prevailing two-modal distribution.

Latha and Badarinath (2004) have studied the characterization of aerosols over urban and rural environments, Hyderabad and Srisailam in India, using AOT measurements from MICROTOP-II sunphotometer during February 2003. AOT was observed to be high at the urban environment compared to the rural environment. Aerosol studies for different wavelengths suggested a typical pattern of increase in AOT with decrease in wavelength as a result of dominance of accumulation mode particles and greater sensitivity of AOT at shorter wavelength. The turbidity (β) and wavelength exponent (α) values suggested high turbid atmosphere in association with dominant accumulation mode particle loading at the urban environment compared to the rural environment. Size distribution showed maximum particle loading observed at 0.2 μm both the urban and rural environments.

Smirnov et al. (2002) have studied characterization of ambient aerosol in 5 island locations of the AERONET of sun and sky radiometer (3 sites in the tropical Pacific Ocean and 2 sites in the tropical to subtropical Atlantic), spanning over 2-5 years .

Aerosol over the Pacific Ocean still resembled mostly clean maritime aerosol mainly free of continental influences. The AOT (500 nm) was stable with daily mean value of 0.07, mode value at 0.06. Angstrom parameter α ranged from 0.3 to 0.7. Over the Atlantic Ocean, the AOT was significantly higher with daily mean value of 0.14 and mode value at 0.10 due to the frequent presence of dust, smoke and urban/industrial aerosol.

Eck et al. (2004) have studied aerosol optical properties in Southeast Asia from AERONET observations during February to May 2003 with investigative data of AOT, α , single scattering albedo and volume size distribution, also studied of back trajectory. The study has being worked at urban and rural sites in Southeast Asia and other regions. They concluded that during the 2nd half of the dry season the aerosols were quite strongly absorbing in both urban and rural sites in Southeast Asia, the aerosol size distribution was dominated by fine mode particles (radius < 0.6 μm) and the monitoring in the beginning of the dry season would characterize the regional aerosol before biomass burning aerosols were added to the mixture.

Jin et al. (2005) have studied diurnal, weekly, seasonal and interannual variations of urban aerosols in summer months using 4 years of AOT and cloud properties measurement from satellite (MODIS) and AERONET, as well as PM_{2.5} data. They analyzed the similarities and differences of urban aerosols in difference climate regions, for one midlatitude city (New York) and one subtropical city (Houston). Interannual variations of monthly mean AOT (0.56 μm) values changed by 10% (τ_A 0.4–0.5) for the New York region in three consecutive June, little change ($\tau_A \sim 0.3$) occurred in Houston. They suggested such differences were partly a result of local weather and climate conditions, and the subsequent transport of aerosols. Seasonality was evident in AOT, with a minimum in winter (January) and a maximum in April to early summer, both New York and Houston. The diurnal variations of aerosols (PM_{2.5} and AOT), however, were detectable but largely affected by local and regional weather conditions, such as surface and upper-level winds. On clam clear days, aerosols peak during the two rush hours in

morning and evening caused by car combustion. The weekly cycle of AOT has been observed in New York with high value during weekdays and low value on weekends. This cycle might be interpreted as a signal of human activities. Nevertheless, this cycle was detectable for Huston but this signal was relatively weak. This was partly because the surface transport over Huston was generally stronger than New York, which distributed urban aerosol to other regions rapidly. No clear relationship was observed between monthly mean AOT and rainfall, implying that in summer the aerosol impact might not be the primary reason for change of urban rainfall amount.