

Chapter 2

Atmospheric Aerosols

Abstract This chapter provides an introduction to atmospheric aerosols by offering a definition and discussing various ways of categorizing and characterizing aerosols. The natural and anthropogenic sources of atmospheric aerosols are reviewed and discussed, leading to a description of their spatial distribution in the atmosphere. The chapter also introduces qualitatively some aerosol concepts, such as their chemical composition and size distribution, paving the way for a more complete description in the following chapters. Some examples are provided to illustrate how aerosols, despite being microscopic particles, can manifest themselves in the atmosphere. Finally, the various pathways by which aerosols impact the climate system are reviewed, introducing a revised categorization of aerosol–radiation, aerosol–cloud, and aerosol–surface interactions.

Keywords Aerosol definition · Aerosol properties · Aerosol types · Aerosol–radiation interactions · Aerosol–cloud interactions

2.1 Definitions

An aerosol is by definition a collection of solid or liquid particles in suspension in a gas. Strictly speaking the term aerosol includes both the particles and the suspending gas. In atmospheric sciences, it is usual however to use the term aerosol in its plural form to refer to the “aerosol particles” without the atmosphere as the suspending gas. We follow that convention in this book, but occasionally refer to “the aerosol” when we refer to atmospheric aerosols in a generic way. Atmospheric scientists also like to differentiate cloud particles from other types of particles in the atmosphere. For this reason we define aerosols as solid or liquid particles in suspension in the atmosphere to the exception of all hydrometeors (cloud droplets, ice crystals, raindrops, snowflakes, and graupel).

Aerosols are always present in the atmosphere but in extremely variable concentrations. This is due to the very large heterogeneity in aerosol sources and their relatively short residence time in the atmosphere (of the order of hours to weeks). The vast majority of aerosols are not visible to the naked eye because of their microscopic size but one can easily see the collective effect of aerosols in the atmosphere as soon as their concentrations are large enough. For example, a haze that reduces

Fig. 2.1 Vegetation fire in the Alps on a cloudy day. One can notice the *blueish* colour of the smoke plume which contrasts with the *white* colour of clouds. The *blueish* colour is due to the small aerosol size in the smoke plume and their larger effectiveness at scattering radiation in the shorter wavelengths of the visible spectrum. (Photograph by the author)



Fig. 2.2 Saharan dust event over the Alps. The snowpack has taken an *orange* colour due to the deposition of dust. (Photograph taken in April 2002 by the author)



the atmospheric visibility and whitens the sky is nothing else than a collection of aerosol particles that interact with solar radiation. A smoke plume is composed of microscopic particles that stem from incomplete combustion of carbonated fuels; these particles collectively darken the sky (see Fig. 2.1). Aerosols can also be visible when they get deposited in great quantity on the Earth's surface, as it is occasionally the case for Saharan dust particles (Fig. 2.2). Furthermore, one can "see" aerosol particles one by one with the help of an electron microscope as shown on Fig. 2.3. The large variety of sizes and shapes among atmospheric aerosols is already striking from these photographs.

The amount and properties of aerosols are extremely variable in space and time. This is why one is usually interested in characterizing a population of aerosols rather than individual particles. The most important characteristics of an aerosol population are the size distribution, chemical composition, and shape of the particles. It is useful to classify aerosols in different categories according to their properties. There are several possible classifications:

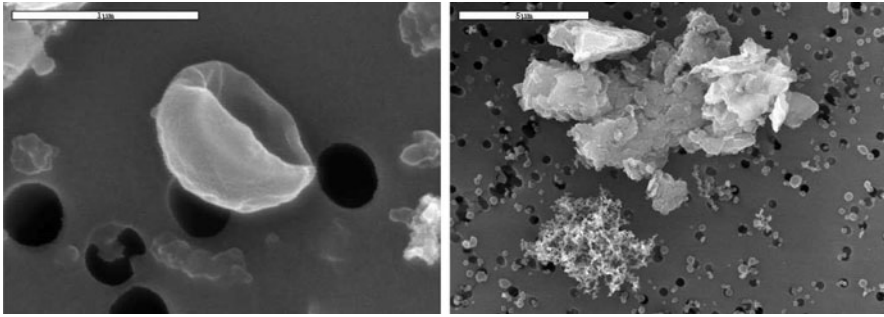


Fig. 2.3 Photographs of atmospheric aerosols collected on a filter: vegetal debris (*left*) and soot (black carbon) particle next to a mineral aggregate (*right*). The *white segments* represent 1 and 5 μm , respectively (© LISA/CNRS)

1. *Primary aerosols* have been emitted into the atmosphere as particles. This is the case of aerosols produced by the effect of the wind friction on an oceanic or terrestrial surface and aerosols produced during an incomplete combustion. *Secondary aerosols* designate those particles that have not been emitted directly in the particulate phase but come instead from the condensation of atmospheric gas-phase species. These gas-phase species, which can undergo a number of chemical transformations before they condense, are called *aerosol precursors*. The primary or secondary origin of aerosols offers a first way to categorize the atmospheric aerosol. The chemical composition of the aerosol usually provides a first idea as to whether the aerosol is primary or secondary.
2. Aerosol properties vary spatially and some of these properties can vary more or less systematically with the type of environment. One can thus speak of *urban aerosols*, *semiurban aerosols*, *continental aerosols*, *desertic aerosols*, *marine aerosols*, *volcanic aerosols* or *stratospheric aerosols*. This is an imperfect categorization in that aerosols can be transported a long way and are therefore not necessarily representative of the location where they can be found. It is for example possible to observe marine aerosols above the continents and continental aerosols above the ocean. When local effects dominate, it may be useful though to refer to such aerosol types in a broad sense.
3. Aerosols can also be classified according to their origin. One can distinguish natural from anthropogenic sources. Natural sources consist of emissions from the ocean, soils, vegetation, fires, and volcanoes. Anthropogenic sources are largely dominated by emissions from the combustion of fossil fuels (i.e. coal and oil), biofuels (plant biomass including wood, vegetable oils, animal waste), other fuels (e.g. peat), or vegetation fires caused by humans. Industrial activities, transportation, heating, or even domestic activities related to cooking in developing countries, are important sources of aerosols. Some industrial and agricultural activities can also emit primary aerosols referred to as industrial dust and arable dust, respectively.

None of these classifications can categorize fully and systematically the aerosol. The different aerosol populations mix and interact with each other in the atmosphere so that some of the terms and aerosol classes that we have just introduced are somehow misuses of language.

2.2 Sources of Aerosols and Aerosol Precursors

Let us now go through the different sources of aerosols in a more systematic way. Table 2.1 summarizes the amount of aerosols and aerosol precursors emitted by different sources.

2.2.1 *Marine Aerosols*

The wind friction at the ocean surface ejects fine particles of salty marine water into the atmosphere. A fraction of the water evaporates, so that the concentration of salt in the particle increases. This gives rise to sea salt particles that are more or less hydrated according to the ambient humidity. Although these particles are often called sea salt aerosols, this is yet another misuse of language because these particles may also contain biological material and other impurities. It is therefore more appropriate to refer to *sea spray aerosols*. Sea spray aerosols cover sizes that range from typically 100 nanometres (nm) to several tens of micrometres (μm). The largest particles fall back fairly quickly to the ocean surface and are therefore of lesser climatic importance.

2.2.2 *Desert Dust*

The wind friction on continental surfaces can detach soil particles and suspend them in the atmosphere. This is particularly the case in desertic, arid, and semiarid regions where the wind is not slowed down by the vegetation that is either completely absent or fairly sparse. The emission of soil particles to the atmosphere also requires a reduced soil humidity so that cohesive forces between soil particles are also reduced. As for sea spray aerosols, *desert dust* particles span sizes that range from typically 100 nm to tens of μm . Larger particles can also be lifted but they fall down quickly. Desert dust aerosols are also called *mineral dust* or mineral aerosols. Emissions of desert dust depend very much on environmental and meteorological conditions. They are sporadic in nature. Figure 2.4 shows a photograph of a dust storm generated by an atmospheric gravity current known as a *haboob*. Figure 6.3 shows an episode of Saharan dust aerosols advected above the Atlantic Ocean over a period of three days.

Table 2.1 Approximative emission fluxes from different types of primary aerosols and gaseous precursors of secondary aerosols. The climate importance of aerosols depends not only on the strength of their emissions, but also on their physical and chemical properties. Estimates are compiled from Penner et al. (2001), Dentener et al. (2006), Guenther et al. (1995), Jaenicke (2005), Burrows et al. (2009), Heald and Spracklen (2009). Tg = 10^{12} g = 1 million of tons. Gg = 10^9 g = 1 thousand of tons

Aerosol type	Emission flux (per year)
<i>Natural primary aerosols</i>	
Desert dust	1000–3000 Tg
Sea spray	1000–6000 Tg
Biomass burning aerosols	20–35 Tg
Terrestrial primary biogenic aerosols	Order of 1000 Tg
Including bacteria	40–1800 Gg
Including spores	30 Tg
<i>Precursors of natural secondary aerosols</i>	
Dimethylsulphide (DMS)	20–40 Tg S
Volcanic SO ₂	6–20 Tg S
Terpenes	40–400 Tg
<i>Anthropogenic primary aerosols</i>	
Industrial dust	40–130 Tg
Biomass burning aerosols	50–90 Tg
Black carbon (from fossil fuel)	6–10 Tg
Organic carbon (from fossil fuel)	20–30 Tg
<i>Anthropogenic secondary aerosols</i>	
SO ₂	70–90 Tg S
Volatile organic compounds (VOCs)	100–560 Tg C
NH ₃	20–50 Tg N
NO _x	30–40 Tg N

C carbon, S sulphur, N nitrogen

2.2.3 Volcanic Aerosols

Volcanoes can emit fragments of pulverized rocks and minerals, usually called *volcanic ash*, during explosive eruptions. These particles have sizes typically ranging from a micrometre to millimetres. Volcanic ash can be transported over distances of a few hundreds to a few thousand kilometres but being micronic particles they tend to fall down rapidly. Hence their climate effect is limited. Volcanoes also emit sulphur-rich gases (in the form of sulphur dioxide, SO₂, and hydrogen sulphide, H₂S) that get oxidized in the atmosphere to form submicronic sulphate aerosols. If these sulphur-containing gases are emitted in the troposphere, the residence time of

Fig. 2.4 Desert dust storm in 2004 at Agoufou in Mali.

These dust storms are known as haboobs in Africa but their names vary according to the country. (© Patricia de Rosnay, LEGOS/CNRS)



the subsequent aerosols will be short, a few weeks at most. However, if the eruption is powerful enough to inject the sulphur gases in the stratosphere, then the volcanic aerosols have a much longer residence time, of the order of a few months to more than a year, depending on the region and altitude of injection.

2.2.4 Biogenic Aerosols

The terrestrial biosphere is a source of *primary biogenic aerosol particles* (abbreviated PBAP). They comprise plant and insect debris, pollen (a fine powder produced by seed plants that contains the male gametes that serve for reproduction), spores (a reproduction cell or organ from many plants and fungi), bacteria and viruses. Once airborne, these particles can be transported by the wind on varying distances depending on their size. Debris are usually larger than 100 μm , pollen, spores and large bacteria are generally in the range of 1–100 μm , while small bacteria and viruses are generally smaller than 1 μm .

Seawater also can contain biological material, some of which is transferred to sea spray aerosols during the emission process. This primary organic matter is found preferentially in particles smaller than 200 nm in diameter (Leck and Bigg 2008) and its amount has been found to depend on the biological activity in ocean waters (Facchini et al. 2008).

Terrestrial and marine ecosystems are also an important source of aerosol precursors. Some species of phytoplankton produce dimethylsulphide (DMS), a gaseous compound that is oxidized in the atmosphere to form sulphur-containing aerosols. Plants and algae emit volatile organic compounds (VOCs) that are oxidized in the atmosphere and condense and contribute organic material to the atmospheric aerosol. These aerosols are referred to as *secondary biogenic aerosols*. Their sizes are typically of the order of a few tenths of a micrometre.

2.2.5 Biomass Burning Aerosols

In biology, biomass refers to material produced by living organisms, but we adjust the definition here to mean all biological (organic) material that comes from the living world and can potentially burn (i.e. vegetation, dead wood, animal dung, peat) while excluding so-called fossil fuels (coal, gas, and oil) that are formed on geological timescales. The burning of biomass generates primary aerosols that stem from the incomplete combustion of the organic matter. Biomass burning aerosols include *organic carbon*, that is associated with hydrogen and oxygen atoms, and *black carbon*, where the carbon content is very high. These aerosols are generally submicronic and are clearly visible in smoke plumes (Fig. 2.1). The sources of biomass burning aerosols are both natural and anthropogenic. The combustion of biomass also emits gaseous compounds, such as volatile organic compounds and sulphur dioxide, which are aerosol precursors.

2.2.6 Aerosols from Fossil Fuel Combustion

The combustion of coal and oil derivatives also produces black carbon and organic carbon, as well as sulphur dioxide that converts into sulphate aerosols. These are essentially submicronic particles, that are also a source of air pollution in developing and industrialized countries (Fig. 2.5). Air pollution due to particles and gas-phase pollutants, such as ozone and nitrogen oxides, is responsible for a wide range of adverse health and environmental effects. Effects on human health include increased respiratory and cardio-vascular diseases and associated mortality. Aerosols and acidic deposition are responsible for damages on historical buildings.

2.3 Spatial and Temporal Aerosol Distributions

Once they are in the atmosphere, aerosols are transported by winds but are also subject to removal pathways, which are called *atmospheric sinks*. Different aerosol types can stay for very different times in the atmosphere. It is useful to introduce the concept of atmospheric *residence time*¹, which is the average amount of time an aerosol spends in the atmosphere, and is defined as

$$\tau^* = B/S = B/R \quad (2.1)$$

¹ Residence time is a property that refers to the global system or a region (reservoir) big enough for which advective terms in and out of the considered region can be considered negligible in comparison to the source and sink terms integrated over the region. It should not be confused with the concept of *lifetime* which is defined in Chap. 4.

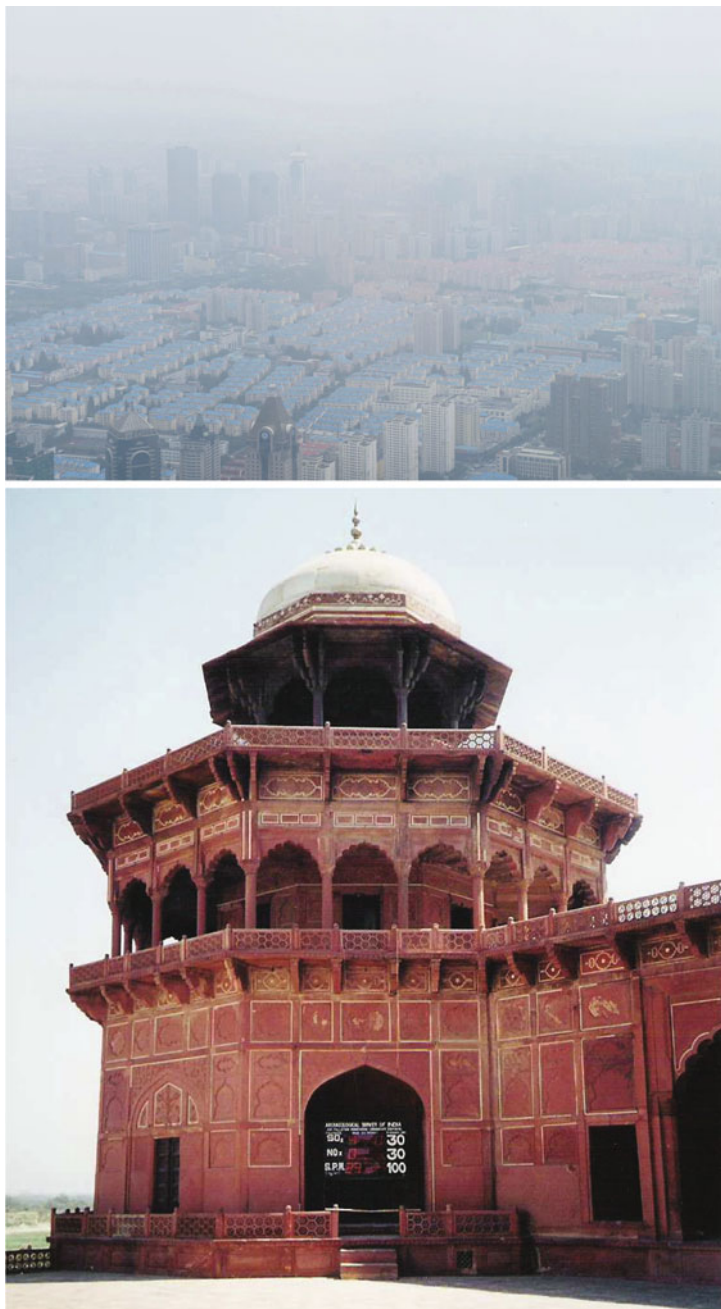


Fig. 2.5 Atmospheric pollution over the city of Shanghai in China (*top panel*, photograph by Eulalie Boucher). Air quality monitoring screen at the Taj Mahal entrance in India (*bottom panel*, photograph by the author). Atmospheric pollution has harmful effects on human health and terrestrial ecosystems. Acid deposition can also have deleterious effects on historical buildings

where B is the vertically-integrated aerosol concentration (also called the aerosol *burden* or *column*), S is the source flux and R is the removal or sink flux. The quantities are averaged over the globe and can be expressed as averages per unit surface (e.g. in units of kg m^{-2} for B and $\text{kg m}^{-2} \text{s}^{-1}$ for fluxes) or as integrated quantities over the globe (e.g. in units of kg for B and kg s^{-1} for fluxes). If the source and sink terms are averaged over a long enough period, they should be balanced and the residence time can be computed from either one or the other quantity. Schematically, in the troposphere, the smallest and largest aerosols have residence times of hours to days, while the bulk of the aerosols in the intermediate size range has residence times of days to 1–2 weeks.

The spatial and temporal distribution of aerosols thus strongly depends on the distribution of sources and sinks and how these interact with transport. The most important sinks are *dry deposition* at the surface and *wet deposition* from precipitations. Figure 2.6 shows spatial aerosol distributions as predicted by an aerosol model constrained by satellite observations. One recognizes plumes of industrial aerosols over Asia and to a lesser extent over Europe and North America, plumes of biomass burning aerosols over South America and Africa during the dry season, and a plume of desert dust originating from the Saharan region. The size of the aerosol plumes reflects the strength of the sources and the time spent in the atmosphere before removal. One can note seasonal variations in the aerosol quantities (in Fig. 2.6 between the months of January and July) but also interannual variations (here between years 2004 and 2007) that are due to variations in the intensity of sources and sinks in response to variations in the meteorology and human activity.

Aerosol properties evolve during atmospheric transport. As an example, we show on Fig. 2.7 the aerosol size distribution (expressed in both aerosol number and aerosol volume concentrations) in three different locations across the Atlantic Ocean. It can be observed that the average size of the aerosols increases as the aerosols are advected off the coast. The concentration of aerosols simultaneously decreases because dry and wet depositions act as sink terms. More details on aerosol size distributions are provided in Chap. 3 and appendix B.

The concentrations and properties of aerosols also vary in the vertical. Most aerosol sources being located at the Earth's surface, one expects aerosol concentrations to decrease with height. Concentrations are generally larger in the atmospheric boundary layer than in the free troposphere. However it is not uncommon for aerosol plumes to be lifted into the free troposphere (as shown for instance in Fig. 2.8). There exists also a local maximum of concentrations in the stratosphere where aerosols are trapped by much less efficient sink terms (see Chap. 12).

2.4 Aerosol–Cloud–Radiation Interactions

Interactions between aerosols, clouds, and electromagnetic radiation are schematized in Fig. 2.9. Aerosols interact with electromagnetic radiation that propagates in the atmosphere in several ways. Solar (shortwave) radiation that interacts with aerosols is deflected in all directions but anisotropically. This process is called *scattering*. Some

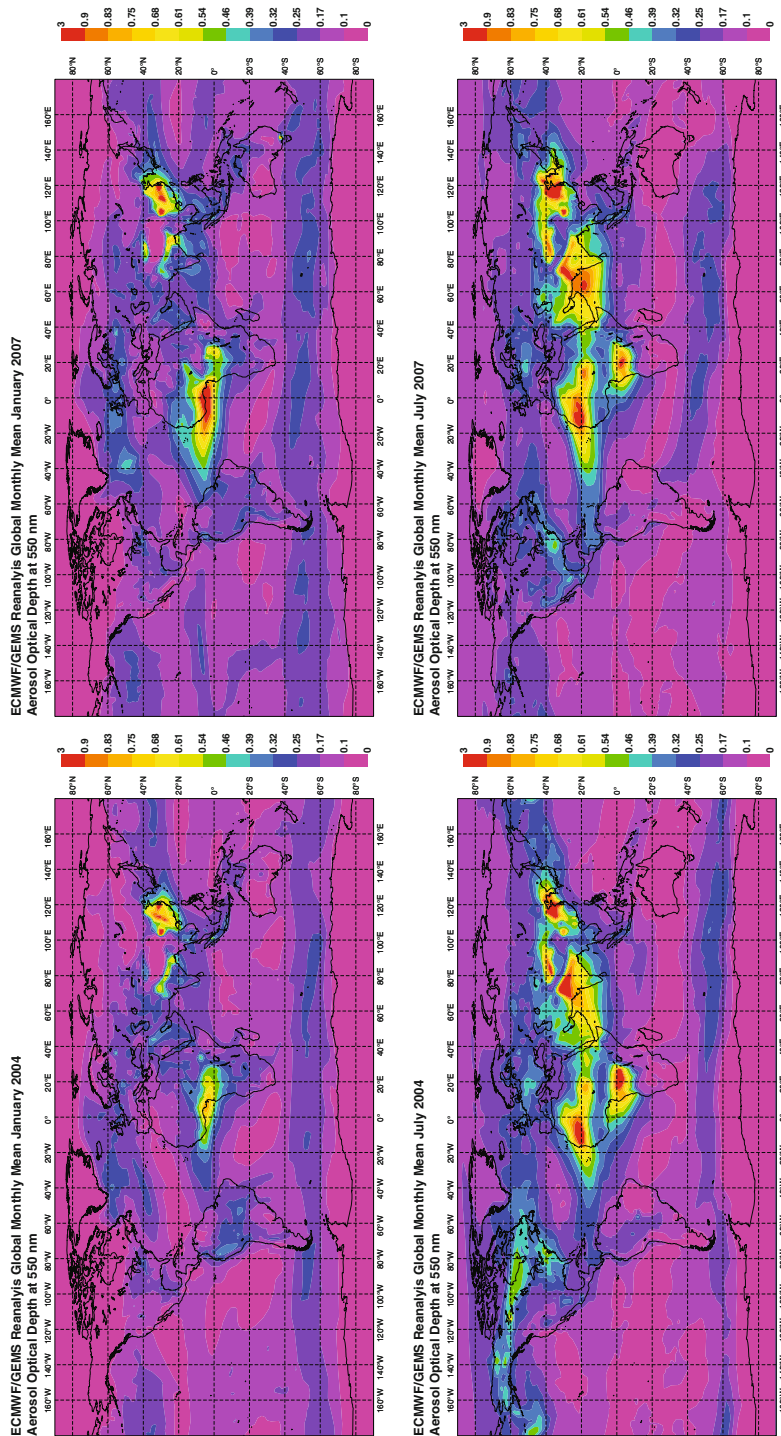


Fig. 2.6 Aerosol optical depth as simulated in a reanalysis where daily MODIS observations are assimilated in a global aerosol model for the months of January and July in 2004 and 2007. From the modelling system described in Morcrette et al. (2009) and Benedetti et al. (2009). (© ECMWF)

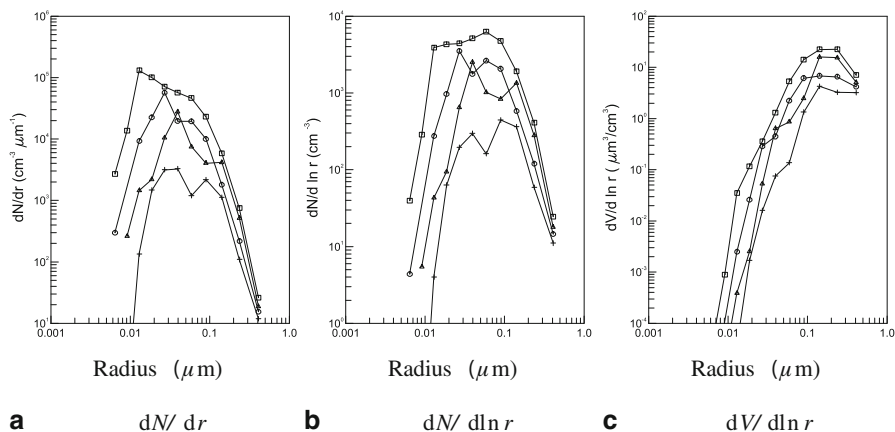


Fig. 2.7 Examples of aerosol number and volume size distributions as measured by Hoppel et al. (1990) during a field campaign across the Atlantic Ocean. These distributions illustrate the transition between a continental environment (with large concentrations of aerosols) to a maritime environment (with smaller concentrations of aerosols). The $\square$, $\circ$, $\triangle$, and $+$ symbols show four different locations ranging from the East American coast to the middle of the Atlantic Ocean. (Data provided by Glen Frick. See also Fig. 3.1)

aerosols also absorb solar radiation, thereby transforming the electromagnetic energy into heat, a process called *absorption*. Aerosol also scatter and absorb terrestrial (longwave) radiation emitted by the Earth's surface and atmosphere, and can emit such radiation. These interactions are described in more detail in Chaps. 3, 5, and 8.

Aerosols influence the microphysical properties of liquid clouds, in particular through their role as *cloud condensation nuclei* upon which water vapour can condense, which controls in part the concentration and size of the population of cloud water droplets. The chemical composition of the aerosol can also determine the degree of acidity of the cloud droplets and control the rate of some of the chemical reactions that take place in the droplets. Aerosols also modify the microphysical properties of ice clouds because a small fraction of the aerosols act as *ice nuclei* that favour the formation of the ice phase. For similar reasons, aerosols influence the properties of mixed-phased clouds (i.e. those clouds that contain both liquid droplets and ice crystals). These aerosol–cloud interactions are described in more detail in Chap. 9.

Conversely clouds influence aerosol populations. Precipitating clouds contribute to remove aerosols from the atmosphere. One differentiates *in-cloud scavenging*, whereby aerosols enter water droplets within the clouds before precipitation, from *below-cloud scavenging*, whereby aerosols are captured by precipitating raindrops themselves. The same holds for ice clouds although scavenging by ice crystals and falling snowflakes is much less efficient. Clouds therefore constitutes an important sink for aerosols through wet deposition. Non-precipitating clouds will eventually evaporate and release cloud droplet residues in the atmosphere, which turn into aerosols and can serve as cloud condensation nuclei in a future cloud. Because sulphur

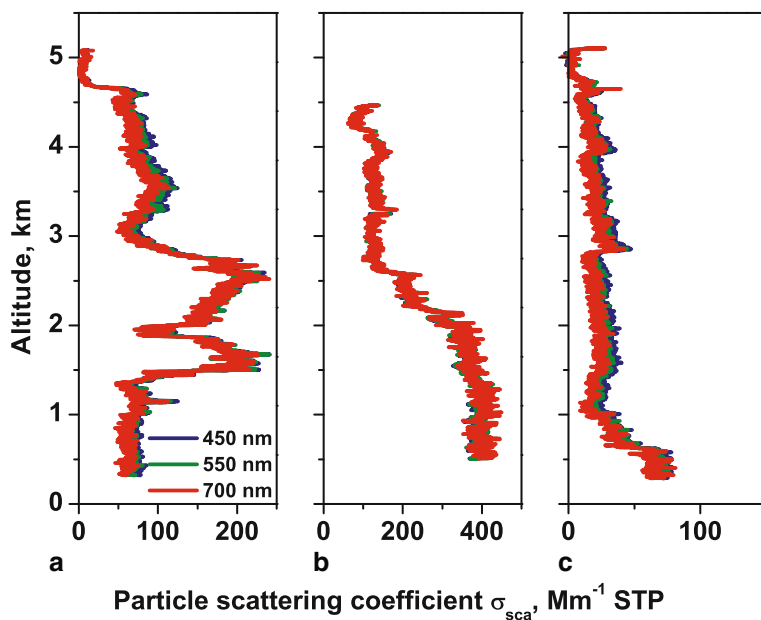


Fig. 2.8 Vertical profiles of the scattering coefficient, $\sigma^{\text{scattering}}(z)$, at 450, 550, and 700 nm (blue, green, and red lines, respectively) from desert dust aerosols during the AMMA field campaign in West Africa. The three profiles correspond to **a** dust aerosols transported at 14°N to the south of the intertropical discontinuity in the Saharan air layer between 1.5 and 5.5 km altitude, **b** dust aerosols transported in Saharan boundary layer north of the intertropical discontinuity, and **c** dust aerosols emitted from the Sahel region following the passage of a mesoscale convective system. The scattering coefficient is expressed in Mm^{-1} at standard temperature and pressure conditions ($1 \text{ Mm}^{-1} = 10^{-6} \text{ m}^{-1}$). (© LISA/CNRS)

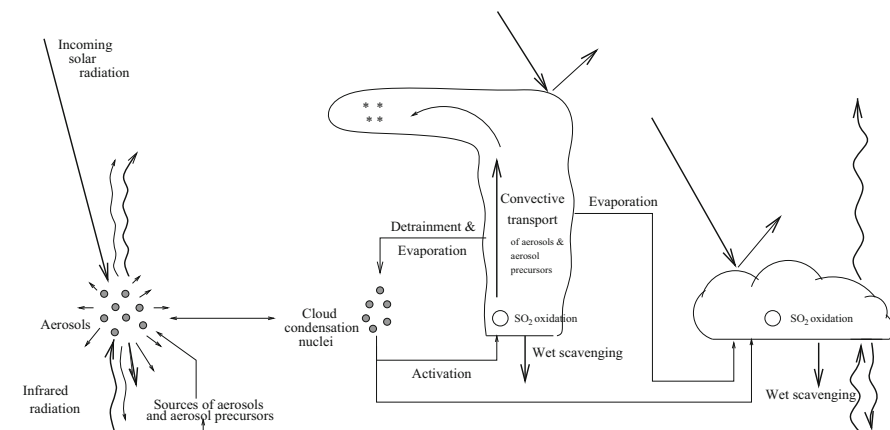


Fig. 2.9 Schematic of aerosol–cloud–radiation interactions. Aerosols interact with both solar (short-wave) and terrestrial (longwave) radiation. There are also two-way interactions with clouds as described in the text

dioxide (SO₂) is readily oxidized in cloud droplets, other interstitial aerosol material can be incorporated into cloud droplets, and coalescence of cloud droplets acts to form larger droplets, the chemical and physical properties of an aerosol population that has gone through a condensation/evaporation cycle can change substantially. The size distribution shifts to larger sizes, the aerosols get more mixed, with more soluble material in the larger size range. This cloud processing effect is particularly important in convective clouds, which are capable of activating small aerosols and release larger aerosols that can serve as cloud condensation nuclei in other clouds.

Clouds scatter solar radiation much more than they absorb it, and this results in clouds reflecting a significant fraction of solar radiation back to space. This leads to a cooling of the climate system. Clouds also absorb and emit terrestrial radiation. They are therefore responsible for a greenhouse effect that leads to a warming of the climate system. One traditionally defines the *cloud radiative effect*² as the difference in net³ radiative fluxes at the top of atmosphere with and without clouds:

$$\text{CRE} = F^{\text{all sky}} - F^{\text{clear sky}} \approx f (F^{\text{cloudy sky}} - F^{\text{clear sky}}) \quad (2.2)$$

where f is the fractional cloud cover. The cloud radiative effect can be decomposed in a (negative) shortwave component and a (positive) longwave component. For low clouds, the shortwave component wins over the longwave component and the net effect of such clouds is to cool the climate system. Inversely, for high clouds that are not too thick, the longwave components wins and the net effect of such clouds is to warm the climate system. Overall the net effect between cooling and warming depends on the thickness and height of the cloud, but also on the cloud microphysical and optical properties.

Conversely, radiation is an important factor that governs the evolution of a cloud. Heating, through absorption of radiation, and cooling, through emission of terrestrial radiation, modify the vertical temperature profile in the cloud, and hence its stability.

In conclusion there is a complex set of interactions between aerosols, clouds, and radiation. It is important to have a good grasp of these processes to understand the physics and chemistry of the atmosphere in general, and the role of aerosols in the climate system in particular.

2.5 Climate Effects of Aerosols

As shown in Fig. 2.10, aerosols influence climate in a number of ways:

1. They scatter and absorb solar radiation. Backscattering of solar radiation towards space results in a reduction of incoming solar radiation at the Earth's surface, a loss of energy and a cooling of the climate system. Absorption of solar radiation

² The term *cloud radiative effect* is preferred to the more usual term of *cloud radiative forcing* because clouds do not exert a climate forcing as defined in Chap. 1.

³ The term “net” refers here to energy gained by the surface–atmosphere system.

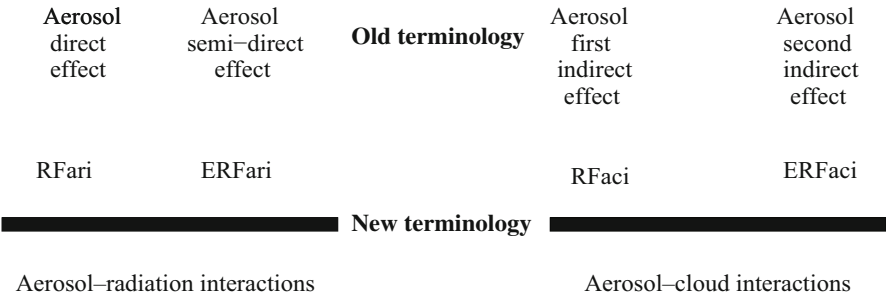


Fig. 2.10 Aerosol–radiation interactions and aerosol–cloud–radiation interactions and how they relate to the original terminology

- is accompanied by a heating within the aerosol layer, but also by a reduction of incoming solar radiation at the Earth’s surface. Such effects occur preferentially, but not uniquely, in clear sky conditions. These processes have traditionally been called the *aerosol direct effect* but the more precise term of *aerosol–radiation interactions* is preferred in this book.
2. The absorption of solar radiation by aerosols modifies the vertical temperature profile. This impacts the relative humidity, atmospheric stability, and therefore cloud formation. This effect has traditionally been called the *aerosol semi-direct effect*, but it can also be seen as a *rapid adjustment* of the atmospheric state that follows aerosol–radiation interactions.
 3. Aerosols serve as cloud condensation nuclei in liquid water clouds, thus exerting a partial control of cloud microphysical and optical properties. An increase in the concentration of aerosols leads to an increase in the concentration of cloud condensation nuclei, and generally to an increase in the concentration of cloud droplets. For a fixed cloud liquid water content, this is accompanied by a reduction in the cloud droplet size and an increase in the cloud reflectivity. Altogether this leads to less solar energy absorbed and a cooling of the climate system. This effect has been traditionally called the *aerosol first indirect effect*, but we use here the more general term of *aerosol–cloud interactions* that includes this and related effects.
 4. The modification of cloud microphysical properties is expected to have an impact on cloud evolution, in particular in terms of the ability of clouds to generate droplets that are large enough to initiate precipitation. This effect is traditionally called the *aerosol second indirect effect*, but is viewed here as a rapid adjustment (or a series of rapid adjustments) that stem from the role of aerosols as cloud condensation nuclei.
 5. Aerosols can impact the properties of mixed-phased and ice clouds, in particular through their role as ice nuclei. This has been referred to in the past as yet another indirect effect of aerosols, or a *glaciation effect*. It is also part of *aerosol–cloud interactions*.
 6. Absorbing aerosols may deposit onto snow and ice surfaces, thus making these surfaces less reflective (i.e. more absorbing). This contributes to warm the surface

and thus the climate system. This effect is known as the *aerosol on snow effect* but can be categorized as part of *aerosol–radiation interactions* or *aerosol–surface interactions* in our new terminology.

7. Finally, aerosols also interact with vegetation through changes in incoming solar radiation, fraction of diffuse radiation and as a source of nutrients. These are other contributions to *aerosol–surface interactions*.

As shown in Fig. 2.10, these multiple effects can be grouped into three main types of interactions: aerosol–radiation interactions (including the direct effect, the semi-direct effect and the aerosol on snow effect), aerosol–cloud interactions (including the first and second aerosol effect and the glaciation effect), and aerosol–surface interactions (effects on snow and vegetation). We follow this classification and discuss aerosol radiative effects in Chap. 8 and aerosol–cloud interactions in Chap. 9. However it is now necessary to introduce a number of concepts and tools to study and quantify the effects of aerosols on climate, which is the purpose of the next chapter.

Exercises

1. The flux of desert dust aerosols to the atmosphere is estimated to be 1000 Tg per year. Their residence time in the atmosphere is 2 days. Compute the burden of desert dust aerosols in units of Tg and mg m^{-2} . The radius of the Earth is taken equal to 6370 km.
2. The flux of SO_2 to the atmosphere is equal to 80 Tg S per year. Only 75% of the emitted SO_2 is oxidized into sulphate aerosols, the rest gets deposited to the surface by dry and wet deposition with a 4:1 ratio. Compute the wet deposition flux of SO_2 in units of $\text{mol S m}^{-2} \text{ year}^{-1}$ and $\text{mg SO}_2 \text{ m}^{-2} \text{ year}^{-1}$?
3. An aerosol is eliminated from the atmosphere at a rate of 10 % day^{-1} due to dry scavenging and 20 % day^{-1} due to wet scavenging. Other aerosol sinks are neglected. Compute the aerosol lifetime with respect to dry scavenging, τ_{dry} , and its lifetime with respect to wet scavenging, τ_{wet} , and its residence time, τ^* .

Solutions

1. $B = 5.5 \text{ Tg} = 11 \text{ mg m}^{-2}$.
2. $F_{\text{wet deposition}}^{\text{SO}_2} = 2.5 \cdot 10^{-4} \text{ mol S m}^{-2} \text{ year}^{-1} = 16 \text{ mg SO}_2 \text{ m}^{-2} \text{ year}^{-1}$.
3. $\tau_{\text{dry}} = C / [\partial C / \partial t]_{\text{dry}} = 1/0.1 = 10 \text{ days}$.
 $\tau_{\text{wet}} = C / [\partial C / \partial t]_{\text{wet}} = 1/0.2 = 5 \text{ days}$.
 $1/\tau^* = 1/\tau_{\text{dry}} + 1/\tau_{\text{wet}}$, i.e. $\tau^* = 3.3 \text{ days}$.

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