

Modeling atmospheric aerosols in climate systems: Representation, dynamics, and radiative impacts

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Abstract

Aerosols, with their complex interactions and far-reaching effects, represent a cornerstone of atmospheric sciences and climate modeling. This article explores the representation of the various dynamic processes governing aerosols, including gas-particle partitioning, coagulation, condensation, and nucleation in Global Climate Models, emphasizing their significance in influencing radiative forcing. This article also highlights the challenges in representing microphysical processes and the uncertainties arising from simplifying assumptions.

1. Introduction

Atmospheric aerosols are defined as small, suspended liquid or solid particles that range in size from a few nanometers to approximately ten micrometers¹. These particles, comprised of mineral dust, sea salt, black and organic carbon, sulfate, nitrate, ammonium, and water, originate from diverse natural sources such as volcanic eruptions, sea spray, and desert dust, as well as anthropogenic activities like fuel combustion and biomass burning². While aerosols can be emitted directly into the atmosphere, they are also produced indirectly through the transformation of precursor gases³. The anthropogenic contribution to atmospheric aerosols has led to substantial increases in aerosol concentrations, impacting both air quality and climate. Aerosols influence air quality with direct implications for human health, and their complex interactions with solar and terrestrial radiation position them as critical agents in climate change¹.

Aerosols affect the Earth's radiative balance through scattering and absorbing radiation, quantified as aerosol radiative forcing (ARF)¹. Scattering aerosols, such as sulfates, typically reflect sunlight, contributing to a net cooling effect, while absorbing aerosols like black carbon retain heat, adding to atmospheric warming. Beyond these direct effects, aerosols also serve as cloud condensation nuclei (CCN) or ice-nucleating particles (INP), which alters cloud microphysical properties, longevity, and albedo¹. These interactions—collectively referred to as aerosol indirect effects—create feedbacks that influence cloud formation, precipitation, and overall climate sensitivity, particularly in the troposphere where aerosols are highly relevant to both pollution and climate.

The field of atmospheric science leverages global aerosol models to simulate the chemical and microphysical properties of aerosols and to analyze their spatial and temporal distributions, life cycles, and resulting climate effects. These models, including advanced global climate models (GCMs), help estimate how anthropogenic aerosols affect past, present, and potential future climate conditions. By integrating observational data with simulated outputs, these models provide spatially and temporally resolved information about aerosol-induced changes in clouds, radiation, and atmospheric chemistry. However, measurements remain indispensable for validating model quality and setting parameters that theoretical calculations cannot determine alone.² Despite significant advancements, global aerosol modeling continues to face challenges, primarily due to computational constraints and the complex interactions among aerosol processes and atmospheric components. Most climate models adopt simplified representations—such as size-resolved or mode-based approaches for aerosol size distribution—to balance accuracy and computational feasibility.³

The structure of this paper is as follows: Section 2 reviews the representation of aerosols in global climate models, focusing on aerosol types, size distributions, and the methods used to capture these variations. Section 3 delves into aerosol dynamics and microphysical processes, including emissions, chemical transformations, and the complex interactions that govern aerosol lifecycles. Section 4 is a summary of the various emissions inventory and chemical transformations of aerosol species. Furthermore, Section 5 addresses aerosol removal mechanisms, examining both wet and dry deposition processes critical to understanding aerosol persistence and distribution. Section 6 analyzes the simulation of aerosol effects on the atmosphere and climate, with an emphasis on radiative forcing and aerosol feedback mechanisms. Finally, Section 7 provides a summary and

conclusion, highlighting key insights from this review and suggesting areas for further research to advance aerosol modeling in climate systems.

2. Representation of aerosol in global models

2.1 Aerosol types and size distribution

Global atmospheric aerosol modeling requires comprehensive approaches to capture the intricacies of aerosol behavior and impact. These models must account for the atmospheric mass concentration (mass per unit air volume) of various aerosol components, as well as the particle number concentration (particles per unit air volume).² A critical aspect of these models is the representation of particle size distributions, as many aerosol lifecycle processes and climate impacts are size-dependent. Aerosol constituents often exhibit varying degrees of mixing; for example, black carbon particles may exist as pure particles or with a coating of other substances, influencing their radiative and cloud interaction properties. Thus, resolving different mixing states across size ranges is essential³.

Early global aerosol models focused solely on mass concentrations. However, advancements in computational power have enabled the inclusion of particle number concentrations, size distributions, and composition in many current models. To reduce computational demands, simplified aerosol representations are often utilized. One common approach is to approximate aerosol size distributions as a combination of size modes represented by log-normal functions. These functions, such as the one described below, are characterized by parameters like total particle number concentration N_t , geometric standard deviation σ , and median diameter D_{med} . This approach effectively represents the number size distribution, which describes the differential particle concentration within a size range.

$$n(\ln D) = \frac{dN}{d \ln D} = \frac{N_t}{\sqrt{2\pi} \ln \sigma} \exp \left[-\frac{(\ln D - \ln D_{med})^2}{2(\ln \sigma)^2} \right]$$

Previous studies have validated the modal approach by showing that the size distribution of tropospheric aerosols can be approximated by three log-normal modes: the Aitken mode (a few nanometers to ~ 100 nm), the accumulation mode (~ 100 nm to $1 \mu\text{m}$), and the coarse mode ($\sim 1 \mu\text{m}$ to several tens of μm). Many global models also include additional modes, such as a nucleation mode for ultra-fine particles or modes to distinguish between particle types in similar size ranges.²

Simplifications, like constant geometric standard deviations σ for the modes, reduce the number of free parameters required to define size distributions. Typically, the parameters N_t and D_{med} suffice to derive other quantities, such as total aerosol volume concentration.

$$V_t = \frac{\pi}{6} \int_{-\infty}^{+\infty} D^3 n(\ln D) d(\ln D) = N_t \frac{\pi}{6} D_{med}^3 \exp \left[\frac{9}{2} (\ln \sigma)^2 \right]$$

This derivation assumes spherical particles and incorporates specific densities of aerosol constituents. As a result, global aerosol models simulate mass concentrations of constituents and

particle number concentrations for each mode as primary prognostic variables, enabling the characterization of aerosol size distributions effectively and efficiently.²

2.2 Aerosol life cycle

The atmospheric lifecycle of aerosols is governed by a complex interplay of particle sources, transport mechanisms, chemical and microphysical transformations, and particle sinks. To accurately capture these processes, GCMs incorporate a variety of sub-models, each tailored to simulate specific aspects of the aerosol lifecycle.

Microphysical processes that influence aerosol size distributions at the microscale—particularly aerosol-aerosol interactions (e.g., coagulation) and aerosol-gas interactions (e.g., condensation and evaporation)—are often encapsulated within dedicated aerosol modules. These modules provide a focused framework for resolving the dynamic evolution of aerosol properties. Other critical processes, including particle emissions, chemical transformations, atmospheric transport, and particle deposition onto the Earth's surface, are typically addressed through distinct sub-models integrated into the larger framework of global aerosol simulations.²

The representation of aerosol lifecycle processes in global models often employs modal schemes, which approximate particle size distributions using log-normal modes. These schemes simplify the computational treatment of aerosols while retaining key information about their size-resolved behavior. The following sections provide a detailed examination of the processes commonly included in GCMs and their technical implementations in modal approaches.

3. Aerosol dynamics and microphysical processes

3.1 Condensation/Nucleation

The mass transfer flux J of an aerosol species from the gas phase to the aerosol phase is governed by the difference between its gas-phase concentration c_∞ and the equilibrium concentration at the particle surface c_{eq} , as described³:

$$J = 4\pi R_p D_i f(Kn, \alpha) (c_\infty - c_{eq})$$

Here, R_p represents the particle radius, D_i is the diffusion coefficient, and $f(Kn, \alpha)$ accounts for the effects of Knudsen number and accommodation coefficient. Organic vapors condense onto aerosol particles when $c_\infty > c_{eq}$. The equilibrium concentration depends on both the inherent properties of the organic species and their ability to interact with and form solutions with pre-existing aerosol-phase compounds. Smaller particles achieve equilibrium more rapidly than larger ones, with aerosol growth typically occurring on time scales ranging from minutes to hours³.

Under specific thermodynamic and chemical conditions, gas molecules aggregate to form molecular clusters, which, upon further growth, lead to the formation of ultrafine aerosol particles. This nucleation process is often parameterized in large-scale atmospheric models using empirical formulations that express the nucleation rate as a function of temperature, humidity, and the concentrations of trace gases. GCMs frequently focus on the nucleation of sulfuric acid (H_2SO_4) and water vapor, as these are critical precursors in the formation of new particles. The resulting

increase in particle number and mass concentrations is typically assigned to the smallest size mode in modal models.

The surfaces of liquid aerosol particles are sites of continuous molecular exchange with the surrounding gas phase. In GCMs, the evaporation of low-volatility compounds is often assumed to be negligible, simplifying the simulation of their condensation. These processes are typically described using particle-size-dependent physical formulations of gas fluxes to particle surfaces. GCMs, which typically operate on time steps of a few minutes to an hour, may therefore explicitly calculate condensation and evaporation fluxes for cases where equilibrium is not achieved instantaneously.

3.2 Coagulation

Coagulation refers to the process of collision and subsequent sticking of aerosol particles, resulting in the formation of larger particles. This process is primarily driven by the Brownian motion of particles in the atmosphere². In modal aerosol models, coagulation is categorized into two types: intramodal coagulation, involving particles within the same mode, and intermodal coagulation, involving particles from different modes. The total coagulation rates are typically calculated by integrating over the entire particle size distribution, accounting for coagulation efficiencies (also referred to as coagulation kernels) that depend on the sizes of the interacting particles.

Intramodal coagulation reduces the number concentration of particles within the respective mode. In contrast, intermodal coagulation results in a decrease in the number concentration in the smaller-sized mode, with a corresponding transfer of mass to the larger-sized mode³. The combined effects of condensation and coagulation influence particle size, composition, and overall aerosol dynamics. Such transformations may also occur during the uptake of aerosol particles into cloud droplets. Upon cloud droplet evaporation, the remaining aerosol particles reflect an aggregate of those taken up by the droplet, along with additional aerosol mass formed within the droplet. These dynamic changes may necessitate the reassignment of particles to different modes within the aerosol model, which is performed at each time step to maintain consistency in particle mass and number distributions.

The rate of coagulation is influenced by both particle number concentration and size³. The rate of change in particle number concentration due to Brownian coagulation is described by the coagulation coefficient $k_{1,2}$, expressed as³:

$$K_{1,2} = 4\pi(r_1 + r_2)(D_1 + D_2) \beta$$

Here, D_1 and D_2 are the diffusion coefficients of the particles, r_1 and r_2 are their radii, and β represents the Fuchs correction factor³. Coagulation significantly impacts particle number concentrations in environments with high particle concentrations, such as regions near emission sources. Furthermore, it can alter the mass and chemical composition of particles through internal mixing.

In urban and regional modeling studies that focus primarily on aerosol mass distributions, coagulation is often neglected due to its relatively minor impact on total mass³. However, in studies focusing on aerosol number concentrations, particularly for fine-mode aerosol distributions, coagulation is a crucial process that reduces particle numbers.

3.3 Gas-particle partitioning

Semi-volatile volatile organic compound (VOC) oxidation products exhibit partitioning behavior between the gas and particulate phases in the atmosphere. Ambient atmospheric particulates typically consist of a complex mixture of organic and inorganic components, along with water. These particulates may exist as single-phase or multiphase systems. Organics in the particulate phase can occur as liquids, solids, mixed phases, or dissolved in aqueous solutions. In the case of a multiphase aerosol, it is commonly assumed that the system comprises at least two distinct sub-phases: an organic phase and an aqueous phase. The equilibrium gas-particle partitioning of a condensable organic species into an aerosol phase containing both organic and aqueous sub-phases is governed by well-established thermodynamic principles³.

Let the aerosol-phase concentrations of partitioning secondary organic species be denoted as O_i ($\mu\text{g m}^{-3} \text{ air}$), where $i = 1, 2, \dots, n$, and n represents the number of partitioning species. The equilibrium partitioning of each species between the gas phase and the absorbing organic aerosol phase is characterized by the organic aerosol-phase partitioning coefficient, $K_{om,i}$. If G_i represents the gas-phase concentration of species i ($\mu\text{g m}^{-3} \text{ air}$) and M_o is the total organic aerosol mass concentration ($\mu\text{g m}^{-3} \text{ air}$) available for partitioning, the equilibrium can be expressed as:

$$K_{om,i} = \frac{O_i}{M_o G_i}$$

where M_o is the sum of all O_i values and the mass concentrations of primary organic aerosol species. The gas-phase concentration G_i for each species is computed using the relationship which relates $K_{om,i}$ to various thermodynamic properties:

$$K_{om,i} = \frac{RT}{10^6 MW_{om} \gamma_i p_{L,i}^o}$$

where R is the ideal gas constant ($8.2 \times 10^{-5} \text{ m}^3 \text{ atm mol}^{-1} \text{ K}^{-1}$), T is the temperature (K), MW_{om} is the average molecular weight of absorbing organics (g mol^{-1}), including both primary and secondary organic compounds, $p_{L,i}^o$ is the pure component vapor pressure of species i (atm), and γ_i is the activity coefficient of species i in the organic phase. The activity coefficient γ_i accounts for non-ideal interactions within the solution. For instance, $\gamma_i > 1$ indicates that species i has limited miscibility with the absorbing aerosol, favoring its persistence in the gas phase. The factor 10^6 ensures unit consistency by converting grams to micrograms. It is important to note that $K_{om,i}$ is inversely proportional to the pure component vapor pressure $p_{L,i}^o$. Consequently, uncertainties in vapor pressure values directly propagate to uncertainties in the predicted gas-particle partitioning behavior.

4. Emissions and chemical transformations of aerosol species

4.1 Emissions

Aerosol particles are introduced into the atmosphere either through direct emissions or via chemical transformations of trace gases, referred to as aerosol precursors². Direct emissions can occur naturally, such as wind-driven dust release from the Earth's surface, or anthropogenically, such as the production of black carbon particles during combustion processes. In contrast, secondary aerosol particles form through chemical reactions in the atmosphere involving precursor gases, like sulfur dioxide or volatile organic compounds³.

The representation of aerosol and precursor emissions in global climate models is tailored to the nature of the emission process. These emissions are typically incorporated as prescribed spatial and temporal distributions of emission fluxes. For example, anthropogenic emissions from industrial activities are often based on predefined inventories. However, emissions dependent on meteorological conditions, such as the release of mineral dust or sea salt driven by wind, require dynamic modeling. In these cases, meteorological parameters like wind speed are used to calculate emissions during simulations, ensuring a more accurate coupling between atmospheric conditions and emission rates. This dynamic approach enhances the consistency between emissions and meteorological states within the model.

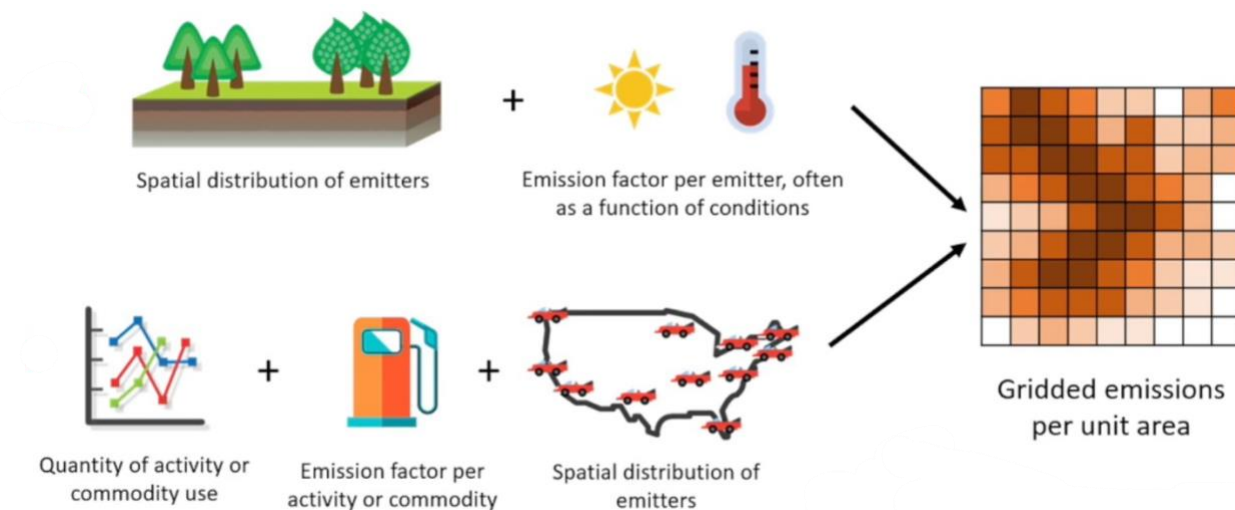


Fig. Framework for Gridded Emissions in Atmospheric Models. The image illustrates the construction of gridded emissions data for atmospheric models by combining various factors. These include the spatial distribution of emitters, the emission factor per emitter (often dependent on conditions such as temperature), and the activity or commodity-specific emission factors. For example, vehicle emissions are mapped using their spatial distribution, fuel use, and emission factors associated with fuel combustion. This approach integrates these components into a gridded representation of emissions per unit area, which is essential for high-resolution climate and air quality modeling.

4.2 Chemical transformations and aerosol precursors

Atmospheric chemical processes facilitate the transformation of highly volatile species into condensable vapors, which either contribute to the nucleation of new aerosol particles or enhance the condensation of aerosol mass onto existing particles. One notable example is the photochemical conversion of gaseous sulfur dioxide (SO_2) into gaseous sulfuric acid (H_2SO_4),

which subsequently forms aerosol sulfate. Similarly, SO_2 can undergo transformation into sulfate through liquid-phase chemistry within cloud droplets. When such cloud droplets evaporate, the sulfate remains as aerosol residues⁴.

These transformation processes are typically represented in GCMs using gas-phase and liquid-phase chemistry schemes. However, due to the high computational cost of detailed chemistry simulations, many GCMs simplify aerosol precursor chemistry by considering a limited number of chemical species and reactions. In some cases, models utilize pre-calculated datasets of trace gas concentrations to reduce computational demands. Key chemical pathways that must be included in these models involve the oxidation of SO_2 to H_2SO_4 and the conversion of nitrogen oxides ($\text{NO}_x = \text{NO} + \text{NO}_2$) into nitric acid (HNO_3). These processes are critical for accurately simulating aerosol formation and evolution in the atmosphere⁵.

5. Transport and removal of aerosols

5.1 Atmospheric transport

The transport of aerosols, their precursor gases, and other relevant atmospheric quantities is a fundamental aspect of global aerosol modeling. To accurately represent this process, global models employ specialized transport schemes. These schemes account for various scales of atmospheric motion, including large-scale wind systems (advection), small-scale vertical motions induced by the local heating of air masses (convection), and turbulent mixing on finer scales (diffusion)².

The meteorological inputs necessary for these transport calculations, such as wind fields and temperature profiles, can either be prescribed from external datasets or calculated dynamically within the model. Dynamically calculated meteorology allows for interactive simulations where aerosol feedback on weather and climate can be captured, providing a more comprehensive understanding of aerosol-climate interactions. In contrast, the use of prescribed meteorology simplifies the modeling framework but limits the simulation to passive aerosol transport, excluding the potential effects of aerosols on atmospheric processes.

5.2 Wet deposition

Wet deposition serves as a critical process for removing aerosols from the atmosphere, occurring through interactions between aerosols and cloud or precipitation particles. Aerosols may become incorporated into cloud droplets or ice crystals by acting as nuclei for cloud particle formation or through collision with pre-existing cloud particles. When these cloud particles form precipitation, such as rain or snow, the aerosols are either washed out of the atmosphere or re-released at lower altitudes if the precipitation evaporates or sublimates. Additionally, falling raindrops or snowflakes can scavenge more aerosols, further enhancing aerosol removal.

GCMs simulate wet deposition by coupling aerosol modules with cloud and precipitation schemes. These aerosol modules provide information on aerosol properties, while the cloud and precipitation schemes determine cloud coverage, microphysical characteristics, and precipitation rates. Together, they estimate the aerosol removal rate through wet deposition. The process is divided into in-cloud and below-cloud scavenging².

In-cloud scavenging involves aerosol particles acting as CCN or ice nuclei (IN), facilitating their incorporation into cloud particles. This process is influenced by aerosol microphysical properties, particularly their size, composition, and hygroscopicity. For instance, the Köhler equation⁶ describes the nucleation of aerosol particles as CCN. Polar soluble organic compounds, such as dicarboxylic acids, are particularly efficient at being scavenged due to their high hygroscopicity. Conversely, insoluble or hydrophobic aerosols, such as elemental carbon, are less efficiently scavenged.

Below-cloud scavenging, on the other hand, occurs when aerosols are captured by falling precipitation through processes like diffusion, impaction, or interception. The scavenging efficiency in this case is determined by the size, shape, and phase (solid or liquid) of the precipitation particles. For example, dendritic snowflakes or rimed snow are more effective at collecting aerosols than other snowflake³. However, the representation of aerosol scavenging by snow and ice remains highly uncertain in global models, particularly in high-latitude regions where mixed-phase clouds dominate.

5.3 Dry deposition

Aerosol particles can also be removed from the atmosphere in the absence of precipitation through a process known as dry deposition. This mechanism involves the direct transport and adherence of particles to the Earth's surface. The deposition occurs through various pathways such as turbulent diffusion, Brownian motion (for fine particles), or gravitational settling (for larger particles). In global climate models, dry deposition is typically quantified using a deposition velocity, which is determined based on particle size, composition, and local aerodynamic as well as surface properties.

Dry deposition is a significant removal process near the Earth's surface. The dry deposition flux F is modeled as being directly proportional to the concentration C of the deposited aerosol species³:

$$F = -u_d C$$

Here, u_d , the dry deposition velocity, is calculated using an electrical resistance analogy. In this framework, the deposition process is conceptualized as overcoming three resistances connected in series: *aerodynamic resistance* r_a , *quasi-laminar boundary layer resistance* r_b , and *canopy resistance* r_c .

1. Aerodynamic resistance r_a governs the turbulent transport of aerosols from the atmosphere close to the surface.
2. Quasi-laminar resistance r_b depends on the physical and chemical properties of the aerosols, as well as the surface characteristics.
3. Canopy resistance r_c , which is generally considered zero for aerosols, assumes that particles that reach the surface are effectively deposited and cannot escape back into the atmosphere.

For aerosol particles, the overall deposition velocity u_d is expressed as:

$$u_d = (r_a + r_b + r_a r_b u_g)^{-1} + u_g$$

where u_g is the gravitational settling velocity. The aerodynamic resistance is derived from the micrometeorological conditions in the lowest model layer, incorporating factors such as momentum flux, eddy diffusivity, and friction velocity. The quasi-laminar layer resistance is modeled based on Brownian diffusion, while u_g is calculated using Stokes' Law, which depends on the particle size and density.

Particle size is the most critical factor influencing the dry deposition velocity. For very small particles, turbulent transport and Brownian diffusion are dominant removal mechanisms, leading to efficient deposition. Larger particles, on the other hand, primarily settle to the surface due to gravity. Medium-sized particles present a challenge as neither turbulent transport nor gravitational settling is particularly effective, resulting in these particles having the longest atmospheric lifetimes in terms of dry deposition.

In GCMs, calculating the dry deposition velocity of organic aerosols involves a detailed integration of micrometeorological conditions and aerosol properties. However, the evaluation of dry deposition remains uncertain due to the lack of direct eddy-correlation measurements for carbonaceous aerosols. While dry deposition becomes increasingly important for coarse particles due to gravitational sedimentation, the removal of aerosols in the accumulation mode (particles smaller than $2.5\ \mu\text{m}$) is predominantly governed by wet deposition processes.

6. Feedback mechanisms and model integration

6.1 Radiative effects

Perturbations in atmospheric aerosol concentrations, whether anthropogenic or natural, can significantly influence the Earth's radiative budget, thereby impacting weather patterns and climate systems. GCMs account for these effects through submodels designed to simulate the transfer of shortwave (solar) and longwave (terrestrial) radiation through the atmosphere. These submodels incorporate processes such as the scattering and absorption of radiation by aerosol particles, which necessitate detailed information on the optical properties of aerosols. Optical properties, including parameters like the single scattering albedo, asymmetry parameter, and extinction coefficient, determine how aerosols interact with radiation across different wavelengths. These properties are derived from the microphysical and chemical characteristics of aerosols, which are calculated within dedicated aerosol modules in GCMs⁴.

Aerosols impact the Earth's radiative budget through two primary mechanisms¹:

Direct Radiative Effects (Aerosol-Radiation Interactions, ARI): Aerosols directly modify the radiative balance by scattering and absorbing incoming solar radiation and outgoing terrestrial radiation. For instance, sulfate aerosols predominantly scatter sunlight, exerting a cooling effect, whereas black carbon aerosols absorb solar radiation, contributing to localized atmospheric warming. The net effect of ARI depends on the aerosol type, size distribution, chemical composition, and altitude.

Indirect Radiative Effects (Aerosol-Cloud Interactions, ACI): Aerosols act as CCN, thereby influencing cloud microphysics. By altering cloud droplet number concentration (CDNC) and size,

aerosols increase cloud albedo (reflectivity), a phenomenon known as the first indirect effect or the Twomey effect. Furthermore, aerosols can enhance cloud lifetime and thickness by suppressing precipitation, known as the second indirect effect or the Albrecht effect. These modifications influence cloud radiative forcing and, consequently, the Earth's energy budget.

The quantification of radiative effects remains one of the major sources of uncertainty in climate projections. This is partly due to the complexity of aerosol-cloud-radiation interactions and the spatial and temporal variability of aerosol distributions. Advanced parameterizations in climate models, along with observational constraints from satellite and ground-based measurements, are essential for improving our understanding of aerosol radiative effects. Integrating these effects into GCMs enables researchers to better assess the role of aerosols in driving climate variability and change.

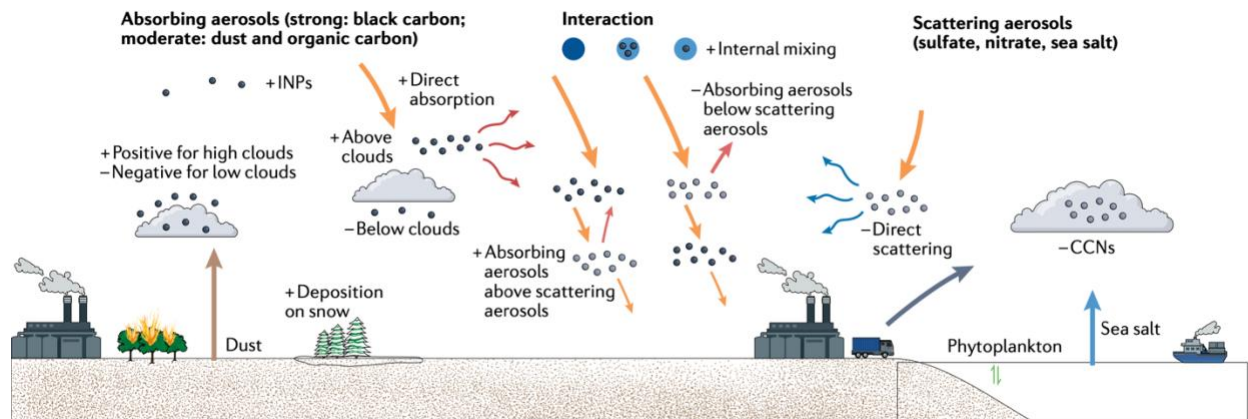


Fig. Schematic Representation of Aerosol Radiative Effects

The schematic illustrates the radiative impacts of absorbing aerosols (dark grey dots) and scattering aerosols (light grey dots), along with their interactive effects. Scattering aerosols exert a negative radiative forcing (–) by directly reflecting incoming solar radiation and interacting with clouds. Absorbing aerosols predominantly contribute to positive radiative forcing (+) by absorbing radiation, leading to warming, though interactions with clouds may induce localized cooling effects. Interactions between absorbing and scattering aerosols amplify absorption, further enhancing warming. The diagram uses light orange arrows for incoming solar radiation, dark orange arrows for scattered radiation by scattering aerosols, red arrows for re-emitted radiation from absorbing aerosols, and dark blue arrows for scattered sunlight. Abbreviations: CCN – Cloud Condensation Nuclei, INPs – Ice-Nucleating Particles.¹

6.1.1. Impact of absorption on aerosol-radiation interactions

Aerosol particles interact with solar (shortwave) and terrestrial (longwave) radiation through scattering and absorption processes – ARI⁴. The efficiency of these interactions is governed by several factors, including the radiation wavelength, particle size distribution, particle shape, and refractive index. The refractive index, in turn, is influenced by the chemical composition and mixing state of the aerosol particles⁴.

Both scattering and absorption contribute to the extinction of radiation by aerosol particles, and the relative contribution of scattering is quantified by the single-scattering albedo (SSA), denoted as ω_0 , which is expressed as:

$$\omega_0 = \frac{\sigma_{\text{sca}}}{\sigma_{\text{sca}} + \sigma_{\text{abs}}}$$

Here, σ_{sca} and σ_{abs} are the scattering and absorption cross-sections, respectively, measured in units of area. This parameter can also be applied to a population of aerosol particles and is a critical factor in determining the overall radiative effects of aerosols¹.

The net impact of aerosols on Earth's radiative budget depends on their optical properties and their interaction with the surface albedo. Scattering aerosols, such as sulfates, tend to increase local albedo by reflecting more sunlight back into space, particularly when situated over darker surfaces like oceans or forests. This results in a cooling effect. However, the same scattering aerosols have a diminished radiative impact over brighter surfaces, such as snow or desert regions, due to the already high albedo.

Conversely, absorbing aerosols, such as black carbon or mineral dust, reduce the albedo by absorbing solar radiation. Over bright surfaces, these aerosols significantly decrease the reflectivity, enhancing localized warming effects. However, their impact is less pronounced over darker surfaces where the contrast in albedo is lower. This dual role of aerosols highlights the complexity of aerosol-radiation interactions and underscores the importance of considering the surface-aerosol interaction when evaluating their climate effects.

7. Conclusion

Aerosols are essential to the Earth's climate system, significantly impacting radiative balance, cloud formation, and atmospheric processes in profound ways. This article emphasizes the complexities of modeling aerosols in GCMs, highlighting the importance of accurately representing aerosol properties, dynamics, and interactions. Current GCMs employ parameterizations to simulate aerosol processes, including their direct and indirect radiative effects, lifecycle (emission, transport, and deposition), and interactions with clouds and precipitation. While these approaches have evolved considerably, they are limited by computational constraints and uncertainties, particularly in representing aerosol-cloud-radiation feedbacks comprehensively.

One critical challenge lies in simplifying the diversity of aerosol characteristics—such as size distributions, chemical compositions, and mixing states—within GCMs. The intricacy and unpredictability of aerosols in the actual atmosphere are frequently not captured by current methods, such as modal schemes for particle size and assumptions of homogenous mixing states. In a similar vein, our capacity to precisely forecast the climate impacts of aerosols, clouds, and precipitation is hampered by ongoing uncertainties in their feedback mechanisms.

Future studies must give priority to developing aerosol modeling frameworks in order to overcome these constraints. A more realistic modeling of aerosol processes might be achieved by incorporating more intricate aerosol modules, like bin schemes, which would allow for greater resolution of particle size distributions². To fully capture the spectrum of aerosol interactions, models must contain a wider range of aerosol forms, chemical contents, and mixing states. The modeling of aerosol modules' transformations and effects will be further improved by coupling

them with more thorough gas- and liquid-phase chemistry methods. Another crucial topic for future research is improving the representation of aerosol-cloud interactions. Accurately simulating aerosols' functions as CCN and INPs, as well as their subsequent impacts on precipitation and cloud microphysics, requires more focus. Additionally, improving the quality of aerosol simulations through increased geographical and temporal resolution of climate models will provide more accurate and regionally relevant insights. These advancements will be made possible in large part by advances in computational capability.

In conclusion, simulating aerosols in GCMs is still a challenging but necessary task for comprehending their complicated effects on the climate system. This article attempts to contribute to a more thorough and reliable understanding of the representation of aerosols in climate models by addressing current constraints and utilizing technical improvements. As a result, we will be better equipped to anticipate and lessen the effects of aerosols, providing important information for climate research and policy.

8. References

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