Smoke above clouds

Aerosols in the atmosphere alter the radiative balance of the Earth by reflecting or absorbing solar radiation. Space-borne measurements of clouds and aerosols advected over the southeastern Atlantic Ocean indicate that the greater the cloud cover below the aerosols, the more likely the aerosols are to heat the planet.

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Many of the processes that emit carbon dioxide and other greenhouse gases into the atmosphere, such as the burning of fossil fuels and agricultural waste, also emit smoke and particles. Like greenhouse gases, these particle pollutants—collectively known as aerosols—alter the radiative energy balance of the Earth. However, greenhouse gases heat the planet by trapping outgoing terrestrial radiation, whereas aerosols can both heat and cool the planet by absorbing solar radiation, shading the Earth’s surface and altering the reflectivity of the Earth as seen from space. Whether aerosols exert a net warming or a net cooling effect depends, in part, on the reflectivity of the underlying surface. On page 181 of this issue, Chand and colleagues show that the radiative forcing potential of aerosols, that is their ability to heat or cool the planet, is primarily controlled by the fractional coverage of underlying clouds.

Clouds are an important determinant of the Earth’s radiative balance. Looking down at Earth from space, bright white clouds dominate our view. These brilliant surfaces reflect light back to space (Fig. 1). However, when dark smoky aerosols overlay these cloud decks, they darken the scene. This increases the amount of incoming solar radiation absorbed by the atmosphere and decreases the amount of solar radiation reflected back to space. In this scenario, aerosols will warm the planet. In contrast, on clear, cloudless days, the presence of aerosols in the atmosphere will often lighten the view from space, increasing the amount of solar radiation reflected out of the Earth system and cooling the planet. In this way, low-level clouds can exert a critical control on the radiative effect of aerosols that are higher up in the atmosphere.

Despite its importance, quantifying the effect of low-level cloud cover on aerosol forcing has proved tricky. Until recently, satellites were only able to measure the concentration of aerosols in cloud-free conditions, owing to the use of passive aerosol sensors, which rely on the reflection of natural sunlight from aerosol surfaces and are unable to directly measure the altitude at which a signal (be it cloud or aerosol) is generated (for example, see ref. 7). This made it impossible to resolve the spatial position of aerosols with respect to clouds, and vice versa. Thus, measurements of aerosol forcing have largely been confined to cloud-free conditions, and quantification of the impact of low-level cloud cover on aerosol forcing has relied on model simulations. These simulation studies indicate that low-level clouds can increase the warming potential of overlying aerosols by as much as 2.5 W/m² (ref. 6), which is equivalent to the magnitude of the total global forcing generated by greenhouse gases.

Chand et al. provide long-awaited observational verification of these model results. Using a space-based lidar—Cloud Aerosol Lidar and Infrared Pathfinder Satellite Observation, or CALIPSO—for short—they measured the amount of radiation scattered by aerosols that were advected over the partly cloudy surface of the southeastern Atlantic Ocean during July–October of 2006 and 2007. Unlike passive aerosol sensors, the lidar sends out its own signal, allowing it to measure the time until the reflection returns, and thus the altitude of the cloud or aerosol that is in the way. This allows the lidar to measure the properties of aerosols even on a cloudy day, as long as the aerosol layer is situated above the clouds. Combining the lidar data with satellite-derived measurements of regional cloud cover, from the Moderate Resolution Imaging Spectroradiometer (MODIS), they show that the more cloudy it is below the aerosols, the greater the aerosol-induced warming.
The linearity of the relationship between cloud cover and aerosol forcing makes it possible to define a critical cloud fraction at which aerosols switch from exerting a net cooling to a net warming effect. The average cloud cover in the southeastern Atlantic Ocean between July and October exceeded this critical threshold, suggesting that the net effect of aerosols in this region, at least at this time, will be to warm the atmosphere. Indeed, using their cloud and aerosol retrieval data, they calculated that aerosols will exert a positive radiative forcing of roughly 2.4 W m\(^{-2}\). However, when they ran the calculations assuming that the spatial pattern of aerosols was independent of the clouds, regional warming was reduced threefold, to 0.8 W m\(^{-2}\). The data indicate that spatial co-variation between lofted aerosols and low-level cloud cover is a critical control on aerosol forcing.

The Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC)\(^4\) estimates that the direct radiative forcing associated with aerosols is \(-0.5 \pm 0.4\) W m\(^{-2}\). That is, globally aerosols are expected to cool the planet. This aerosol induced cooling is roughly 20% of the magnitude of present day greenhouse-gas-induced warming, providing a significant counterbalance to a mostly positive forcing. However, the results of Chand et al.\(^4\) indicate that aerosols may, in fact, warm the tropical Atlantic Ocean, despite the negative global mean. This suggests that IPCC global estimates are unable to account for the complexity of aerosol forcing, which will vary regionally, seasonally and with cloud cover\(^5\).

The work of Chand et al.\(^4\) would not have been possible without the CALIPSO lidar and the MODIS imager, demonstrating the need for continued and comprehensive measurements of aerosol and cloud properties using both passive and active satellite sensors, supplemented by focused suborbital campaigns and long-term surface stations. Without these measurements, determining the climatic influence of particulate matter in the atmosphere, in a shifting hydrologic regime, will not be possible.

References

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1. The optical depth of sunlight at sea level due to Rayleigh scattering from air molecules is about

\[ \tau_R = 0.0088 \lambda^{-4.15+0.21} \]

where \( \lambda \) is the wavelength in \( \mu m \).

a. Compute the clear sky transmissivity of the solar flux at the blue \( (\lambda = 0.47 \mu m) \) and the red \( (\lambda = 0.64 \mu m) \) wavelengths, at solar zenith angles of 30° and 85° both. Ignore absorption by ozone and water vapor (small at these wavelengths) and aerosol effects.

b. Use these results to explain why the sky is blue and the setting sun is red.

HW Problem #3:

Petty Problem 12.8, in handout

\[ \text{\textit{(not relevant to HW!}}} \)
12. Scattering and Absorption By Particles

In reality, the inevitable presence of aerosols, which are much larger than molecules and don't satisfy the Rayleigh criterion, reduces the polarization somewhat. Also, multiple scattering, which is weak but not negligible in this instance, further reduces the polarization slightly.

Nevertheless, wearing a pair of polarized sunglasses, you can easily verify the above effect by viewing a portion of the blue sky at right angles from the sun and rotating the sunglasses (or your head) about the line-of-sight. The sky will appear darker or lighter, depending on whether the sunglasses transmit or block the polarized radiation. The bluer the sky (and therefore the less haze present) the more pronounced the effect will be.

12.2.4 Scattering and Absorption Efficiencies

We were able to infer the scattering phase function for small particles based on relatively simple handwaving arguments. Let's now turn to the question of how much radiation a small particle scatters and/or absorbs. While it is possible to obtain this information directly based on the dipole model we developed above (see BH83, section 5.2), the complete derivation requires more space and explanation than seems warranted at this introductory level. Among other things, it would be necessary to explain the relationship between the (relative) complex index of refraction \( m \) of the particle and its polarizability \( \alpha \), as well as to show how the imaginary part of \( \alpha \) bears on absorption of the incident electromagnetic wave by the particle.

An alternate way of getting at the same information is to take the general Mie solutions for spheres of arbitrary size, which I will briefly discuss in section 12.3, and find limiting expressions for \( x \ll 1 \). Specifically, you rewrite the solutions as power series in \( x \) and discard all but the first few terms. Here, I will give you the essential results without going through the derivations (see BH83, section 5.1).
be significant for absorption by clouds. Therefore, for the shorter wavelengths where absorption is comparatively weak, we plot the scattering co-albedo, defined as $1 - \bar{\omega}$, on a logarithmic vertical axis. Here are the basic points you should take away from these plots:

- The visible band ($0.4 \mu m < \lambda < 0.7 \mu m$) coincides almost exactly with a surprisingly narrow portion of the EM spectrum for which absorption by cloud droplets is, for all practical purposes, zero. You can think of it as an astonishing coincidence that clouds (when viewed from the sunlit side) appear to our eyes as white rather than gray, black, or some other color! As soon as you move into either the UV or near-IR bands, $\bar{\omega}$ quickly decreases to well below $1$, settling into the range 0.5–0.8 for most of the IR band. For even $\bar{\omega} = 0.8$, the albedo of a thick cloud is only around 15%.

- At several wavelengths, there is a significant difference between the single scatter albedo of a spherical ice particle and that of a water droplet of the same size (top row). For some of these wavelengths, ice particles are less absorptive than the water droplets; for others, the reverse is true. These differences can be exploited by satellite sensors to distinguish ice phase clouds (cirrus) from water clouds.\(^9\)

- For most wavelengths, there is a significant dependence of the single scatter albedo on the droplet radius in liquid water clouds (bottom row). As a general rule (although there are exceptions), a larger droplet has lower $\bar{\omega}$ (i.e., is more absorptive) than a smaller droplet at the same wavelength. Once again, satellite remote sensing techniques can exploit this property to estimate the effective droplet radius $r_{\text{eff}}$ in water clouds.

12.5.2 Radar Observations of Precipitation

Radar has become one of the most important observational tools of operational meteorologists as well as hydrologists. Weather radar

\(^9\)The fact that ice particles in clouds are generally not spheres complicates the problem somewhat, but the principle is still valid.
Fig. 12.11: Radar backscatter efficiency \( Q_b \) for water and ice spheres at the wavelength of the WSR-88D operational weather radar.

We can then replace the summation in (12.32) with an integral involving \( n(D) \) and \( \sigma_b(D) \):

\[
\eta = \int_0^\infty \sigma_b(D) n(D) \, dD, \tag{12.34}
\]

or

\[
\eta = \int_0^\infty Q_b(D) \left[ \frac{\pi}{4} D^2 \right] n(D) \, dD, \tag{12.35}
\]

where the term in square brackets is just the cross-sectional area of a sphere with diameter \( D \), and \( Q_b \) is the backscatter efficiency.

If our spherical particles happen to have size parameters \( x \ll 1 \), then we're in the Rayleigh regime. This means that (a) the Rayleigh formula (12.13) for \( \sigma_b \) applies, and (b) the phase function is given by (12.10). Substituting these into (12.33) gives

\[
Q_b = 4x^4 \left| \frac{m^2 - 1}{m^2 + 2} \right|^2. \tag{12.36}
\]

If our particles are too large, then Rayleigh theory no longer applies, and we have to calculate \( \sigma_b \) using Mie theory. Fig. 12.11 shows...
where $Z$ is the reflectivity factor, defined as

$$Z = \int_0^\infty n(D)D^6 \, dD.$$  \hspace{1cm} (12.39)

In other words, the reflectivity factor is numerically equal to the sum of the sixth powers of the diameters of all of the drops in a unit volume of air. The standard units of $Z$ used by meteorologists are [mm$^6$m$^{-3}$]. An estimate of the reflectivity factor $Z$ at each range $d$ along the beam is what most weather radars record and display.

Because observed values of $Z$ span an enormous range, meteorologists prefer to work with a logarithmic representation of $Z$, defining a nondimensional unit dBZ, which means “decibels with respect to one standard unit of $Z$.” You convert the reflectivity factor from standard units to units of dBZ as follows:

$$Z \text{ [dBZ]} = 10 \log_{10}(Z),$$  \hspace{1cm} (12.40)

where $Z$ on the right hand side is the numerical value of $Z$ expressed in standard (dimensional) units of reflectivity. Thus, an increase in reflectivity by 10 dBZ corresponds to a factor ten increase in $Z$ expressed in standard units. An increase of 30 dBZ implies a thousand-fold increase in reflectivity.

**Problem 12.6**: Depending on range, a typical weather radar can measure reflectivities from as low as $-20$ dBZ to as high as 70 dBZ. In terms of physical units, what is the ratio of the two reflectivity factors?

In converting the received power $P_r$ to an estimate of the reflectivity factor $Z$, the radar processing software assumes a value of $m$ appropriate to liquid water in (12.38). The displayed quantity is therefore actually better regarded as an equivalent reflectivity factor $Z_e$ which may or may not be equal to the true reflectivity factor $Z$ defined by (12.39), depending on whether the targets are liquid water or something else, like ice. If the particles are in fact ice, then

$$Z_e \approx 0.20Z.$$  \hspace{1cm} (12.41)
with a greater number of large raindrops, whereas light rain is usually characterized by smaller drops. On average, therefore, we expect heavy rain rates to be associated with large $Z$ and light rain rates to give rise to correspondingly weaker radar echoes.

Field observations of raindrops have revealed that the dropsize distribution $n(D)$ for rain is often reasonably well approximated by

$$n(D) = N_0 \exp(-\Lambda D),$$

(12.42)

where $N_0$ and $\Lambda$ are parameters that are functions of the rain rate $R$. In fact, the most widely used model of the above form is known as the Marshall-Palmer size distribution, after the researchers who developed it. In the Marshall-Palmer distribution, $N_0$ is a constant and $\Lambda = a R^b$, where the parameters $a$ and $b$ were chosen so as to maximize the agreement between the above size distribution function and a large number of actual observations of drop sizes at various rain rates.

It is beyond the scope of this text to discuss the M.-P. distribution in detail, except to note that, when it is substituted into (12.39) and combined with suitable assumptions about raindrop fall speed as a function of $D$, it is possible to obtain the following $Z$–$R$ relationship:

$$Z = 200R^{1.5},$$

(12.43)

where $R$ is assumed to be given in mm hr$^{-1}$, and $Z$ is in standard units of mm$^3$ m$^{-3}$. Other assumed (or measured) drop size distributions usually lead to $Z$–$R$ relationships having a similar form, but with different values for the two numerical coefficients.

Problem 12.9: Use the Marshall-Palmer $Z$–$R$ relationship above to estimate the rain rates $R$ associated with displayed radar reflectivities of (a) 10 dBZ, (b) 30 dBZ, and (c) 50 dBZ.

12.5.3 Microwave Remote Sensing and Clouds

Microwave radiometers operating at various frequencies from 3 to 183 GHz are assuming an increasingly prominent role in the satellite remote sensing of the atmosphere. One of the main attractions of