1 Clouds

Clouds are fascinating to watch for their myriad of shapes. They are scientifically challenging because aspects about their formation, evolution, distribution, and impact on Earth's climate and the hydrological cycle are still not fully understood. Understanding them requires knowledge about both the large-scale meteorology and microscale cloud processes involved in the formation of cloud droplets, ice crystals, raindrops, snowflakes, graupel, and hailstones.

Clouds are an integral part of Earth's atmosphere and are a major factor in Earth's radiation budget (Wielicki et al., 1995; Kiehl and Trenberth, 1997; Wild et al., 2013). Their influence on the radiation budget differs for the solar (shortwave) and terrestrial (longwave) radiation and depends on the cloud type and altitude (Chapter 2).

Clouds play an important role in Earth's hydrological cycle, by transporting water vapor through the atmosphere and from the oceans to the continents, eventually delivering the water from the atmosphere to Earth's surface as rain or snow (Chahine, 1992; Trenberth et al., 2007) (Section 10.6). Clouds are important for air quality because precipitation removes soluble gases and aerosol particles (Chapter 6) from the atmosphere and deposits them on the surface. Clouds provide a medium for multiphase chemical reactions, meaning reactions that take place inside cloud droplets or on ice crystal surfaces (e.g., Ravishankara, 1997). Furthermore, vertical motions associated with clouds, called updrafts and downdrafts (Chapter 5), largely determine the vertical redistribution of trace species, temperature, and moisture.

1.1 Formation, Distribution, and Importance of Clouds

1.1.1 Definition and Formation of Clouds

A cloud is a visible aggregate of cloud droplets or ice crystals, or a combination of both, suspended in air. For a cloud to be visible, the cloud particles, also called *hydrometeors*, need to exist in a sufficiently large number concentration. Historically, for operational weather forecast definitions, clouds were defined based on visibility only, where observers indicated the fraction of the sky covered with clouds. Today, cloud cover can also be defined based on a minimum number concentration of hydrometeors or a minimum water content from in situ or remote sensing observations.

Clouds form when air is lifted. Lifting cools the air, and increases the relative humidity, eventually allowing the water vapor to condense on liquid or deposit on solid aerosol particles, leading to the formation of cloud droplets or ice crystals. Air masses often need a trigger in order to start rising. Broadly, such triggers include four different mechanisms: (i) Air masses are lifted along fronts of low-pressure systems, that is, along a warm front or cold front in

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

an extratropical cyclone. (ii) Air masses can be lifted along a mountain ridge, that is, due to orography. (iii) Air parcels (smaller volumes of air than air masses) rise in response to local (radiative) surface heating and associated convection. A prominent example of cloud formation due to solar heating of Earth's surface in summer is deep convective clouds that may be associated with thunder and lightning. (iv) Large-scale convergence of air parcels or air masses at the surface can also cause their ascent and cloud formation. These four major mechanisms associated with cloud formation in the troposphere operate on different spatial scales. Depending on the mechanism that is causing the air masses to rise, the clouds that form differ in their properties, as we will see throughout this book.

1.1.2 Global Cloud Coverage

Satellites detect and define clouds based on their optical depth. The dimensionless optical depth corresponds to the amount of solar radiation that is removed by scattering and absorption of cloud particles, as detailed in Chapters 2 and 12. There are several global cloud climatologies, most of them retrieved from satellite data (e.g., Stubenrauch et al., 2013). Some of the oldest cloud satellite retrievals are from the International Satellite Cloud Climatology Project (ISCCP) (Schiffer and Rossow, 1983, 1985; Rossow and Schiffer, 1999), which contains cloud information dating back to 1983. These early climatologies used passive observations such as infrared and visible radiances captured from satellite-borne radiometers to document coverages of distinct cloud types and estimate associated radiative properties. Satellite data are best for information on the highest clouds and their cloud top heights. Since 2006, some satellites have been equipped with cloud radars and/or cloud-aerosol lidar sensors (Sassen and Wang, 2008). Radar and lidar instruments are active sensors that emit electromagnetic radiation and detect cloud properties from the backward scattered electromagnetic radiation. The principle of cloud radars will be discussed in more detail in Chapter 10. Combined, cloud radar and a cloud-aerosol lidar provide detailed information about the vertical structure of clouds in terms of their frequency of occurrence, as well as their liquid and ice water contents, that is, their thermodynamic phase, providing the most accurate information on clouds currently available (Stephens et al., 2002; Winker et al., 2003; Mace et al., 2009). With the launch of the EarthCARE satellite in 2024[1], even more detailed information on cloud properties and updraft speeds has and will become available in the years to come. Figure 1.1 shows the global distribution of the annual mean total cloud fraction. Note that we use the terms cloud fraction, cloud amount, cloud cover, or cloud coverage interchangeably. An average cloud fraction of, for example, 50 % can be interpreted as having a 100 % cloud cover 50 % of the time, or as having a 50 % cloud cover over a given region 100 % of the time, or a mixture of both. Figure 1.1 is based on satellite data: the CloudSat cloud radar (Stephens et al., 2002) and the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) (Winker et al., 2003) data sets. Note that satellite retrievals still have problems in identifying clouds over ice-covered surfaces. Therefore, cloud fractions over polar regions are rather uncertain. In addition, the area

[1] www.esa.int/Applications/Observing_the_Earth/FutureEO/EarthCARE

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

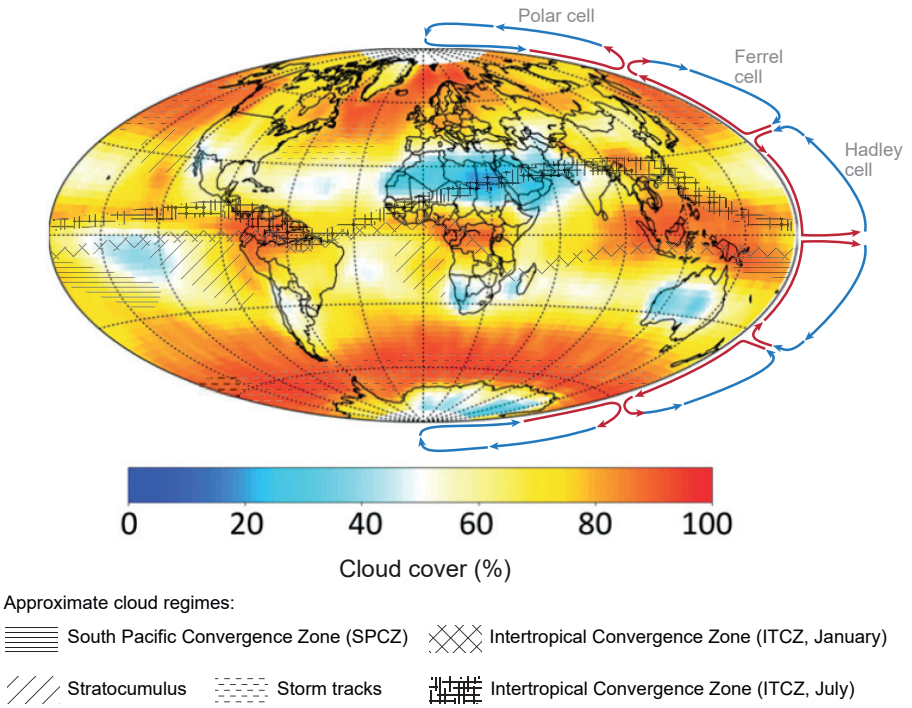


Figure 1.1 Annual mean total cloud cover (%) averaged over the years 2007–2010, based on CloudSat and CALIPSO satellite data sets. The approximate regions where clouds occur most often are also indicated, as defined in the legend. Figure adapted from L'Ecuyer et al. (2019). ©American Meteorological Society. Used with permission.

of lower cloud fraction over the Indian Ocean seen in Figure 1.1 is an artifact due to the poorer satellite coverage in this region.

In the global annual average, 70.8 % of Earth's surface is covered with clouds. The cloud cover is 5–15 % higher over oceans than over land (Table 1.1). The occurrence of clouds is closely related to atmospheric dynamics, from the planetary scale circulation patterns to synoptic and mesoscale dynamics (e.g., Muller, 2020).

A closer inspection of Figure 1.1 reveals that clouds occur most commonly in the *Intertropical Convergence Zone* (ITCZ). The ITCZ forms where the air masses of the south and north trade winds converge at the surface and ascend, forming the rising branch of the *Hadley*[2] *circulation* (Hadley, 1735), see Section 1.3.3. These (tropical) clouds are closely coupled to planetary scale circulation. The location of the ITCZ has an annual cycle that follows the position of the Sun's *zenith*. Its exact location is not constant along one latitude but varies with longitude because of the different heating of land and ocean. In the annual average, the ITCZ is located between the equator and 10–15 °N. Moreover, high cloud fractions are observed in the *South Pacific Convergence Zone* (SPCZ), which is a northwest- to southeast-oriented band starting at 120 °E at the equator and extending eastward to 120 °W and 30 °S. In both

[2] www.britannica.com/biography/George-Hadley

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

Table 1.1 The 10 cloud types, plus fog, and their acronyms, typical altitude ranges of their bases in polar regions (pr), midlatitudes (ml), and tropical regions (tr), and their characteristic features according to the definition of the World Meteorological Organization (WMO, 1975). Global annual fractions of the different cloud types are obtained from satellite observations (Raschke et al., 2005) and from surface observations over land (1971–2009) and ocean (1954–2008), with their average base heights in km above the surface (Eastman and Warren, 2013; Khvorostyanov and Curry, 2014). Cloud base of continental Cu is obtained from Zhang and Klein (2013), and cirrus cloud fractions from Stubenrauch et al. (2017).

		Global fraction (%)		Base height (km)	
Cloud type/genera	Description	Land	Ocean	Land	Ocean
Fog	A "cloud" that touches the ground	1	1	0	0
Low, layered clouds: 0–2 km (pr, ml, tr)		17	35		
Stratus (St)	Very low, gray, uniform layer; Sun outline very distinct when visible	2–5	2–13	0.5	0.4
Stratocumulus (Sc)	Low, gray-white, patch or layer with elements, rolls, or rounded masses	8–13	14-22	1	0.6
Low clouds with vertical extent: 0–2 km (pr), 0–3 km (ml, tr)		14	24		
Cumulus (Cu)	White, detached, dense elements with sharp outlines and vertical growth	5–8	13–15	1–3	0.6
Cumulonimbus (Cb)	Very deep, dense and precipitating, with flattened top (anvil)	3–5	2–6	1	0.5
Nimbostratus (Ns)	Gray, dark, diffuse, uniform cloud with steady precipitation	2–4	3–5	0.1–3	
Mid-level clouds: 2–4 km (pr), 2–7 km (ml), 2–8 km (tr)		21	24		
Altostratus (As)	Uniform or striated gray/blue sheet; Sun can be seen as through translucent glass, that is , there is no clear outline	4–12	6–7	3–5 (ml)	
Altocumulus (Ac)	Gray or white broken sheets, elements, bands, rounded masses	5–17	3–18	2–6 (ml)	
High clouds: 3–8 km (pr), 5–13 km (ml), 6–18 km (tr)		25	20		
Cirrus (Ci)	Detached white filaments or patches with fibrous appearance or silky glimmer			7–10 (ml)	
Cirrostratus (Cs)	Thin white translucent veil, either fibrous or smooth in appearance (halo)			6–8 (ml)	
Cirrocumulus (Cc)	Thin white sheet or patch without shading, composed of very small ripples, grains			6–8 (ml)	
Total cloud cover		54	68		

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

regions, the ITCZ and SPCZ clouds form as a result of air mass convergence at the surface and local heating causing convection. Consequently, the characteristic cloud types associated with the ITCZ and SPCZ are organized deep convective clouds with large vertical extent (Section 1.2.2 and Chapters 5 and 11), usually spanning the whole depth from the surface to the tropopause.

High cloud fractions can also be seen in the midlatitude storm tracks in both hemispheres, over the Southern Ocean and the semi-permanent Aleutian and Icelandic low-pressure regions in the north Pacific and north Atlantic, respectively. Here cloud cover can exceed 90 % in the annual mean (Figure 1.1). Clouds in the storm tracks are closely linked to synoptic scale dynamical processes, mostly frontal systems (Chapter 10).

The extensive cloud cover off the west coasts of North and South America and Africa (Figure 1.1) is associated with low clouds that form in subtropical high pressure systems over cold ocean currents, where large-scale subsidence in the free troposphere prevails. In regions of large-scale air mass descent such as those of the descending branches of the Hadley and Walker[3] circulations (Bjerknes, 1969), the air masses are compressed and warm adiabatically (Chapter 3), often leading to a temperature inversion, where warmer air masses reside on top of colder air masses. In these situations, deep convection is suppressed, and clouds are confined to the *planetary boundary layer* (PBL), the lowest layer of the troposphere. The PBL or atmospheric boundary layer is defined as the part of the atmosphere, which is affected by Earth's surface in terms of friction and is often capped by a temperature inversion. Depending on the season and local meteorology, the PBL depth varies from less than 100 m to several kilometers (Ao et al., 2012; Chan and Wood, 2013; Zhou et al., 2021). The PBL is typically less than 1 km in polar regions (Ao et al., 2012) and over oceans, while it can exceed 5 km over deserts (Cuesta et al., 2008). The PBL also exhibits a pronounced diurnal cycle over land with maxima in early afternoon and minima during night (Liu and Liang, 2010).

In subtropical high-pressure systems over cold ocean currents, the moisture flux from the ocean leads to the formation of stratus (Section 1.2.1) or stratocumulus clouds in the shallow PBLs over the oceans. As the air masses follow the trade winds toward the equator, where the (sea) surface temperatures are higher, shallow convection can take place, and more convective cloud types such as shallow cumulus form (Section 1.2.2). Apart from these regions, the subtropics are characterized by low cloud fractions, in particular over the main deserts, such as the Sahara and the Kalahari, as well as the desert and arid regions of Australia and the Arabian Peninsula.

1.2 Macroscopic Cloud Properties and Cloud Types

Clouds can be grouped into three basic categories, following the French naturalist Jean Baptiste Lamarck (Lamarck, 1802) and English chemist Luke Howard (Howard, 1803): cumulus, stratus, and cirrus. The Latin term *cumulus* means heap or pile and denotes dense isolated clouds, which appear as white tuberose flowers, composed of individual cloud elements. In

[3] www.britannica.com/biography/Gilbert-Walker

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

a group of cumulus clouds, the individual clouds are disconnected from each other. The tower-like structure of cumulus clouds indicates that they usually extend more in the vertical (height) than in the horizontal (width). Cumulus clouds form due to convection, that is, the rising of air masses in locally unstable air (Chapter 4). Therefore, cumulus clouds are often referred to as convective clouds, where their vertical extent is given by the depth of the atmospheric instability. Vertical extent refers to the distance between cloud base and cloud top, also referred to as cloud depth.

In contrast, stratus clouds represent clouds with dimensions that are much larger in the horizontal than in the vertical, as indicated by the term *stratus*, which means flat. Consequently, they appear as uniform cloud layers when seen from Earth's surface. When stratus clouds cover the whole sky, the sky appears gray and may not have any structure. In contrast with cumulus clouds, stratus clouds are usually formed by large-scale vertical air motions in statically stable air (Klein and Hartmann, 1993; Wood and Bretherton, 2006), that is, there is little internal vertical motion.

Lastly, the word *cirrus* means hair or curl and is used to describe clouds that appear wispy and fibrous, composed of delicate filaments. By definition, cirrus clouds consist purely of ice crystals, which give them their characteristic hair-like appearance (Heymsfield et al., 2017). Cloud phase is not a defining factor for the other cloud types. Thus, all cumulus and stratus clouds can consist entirely of cloud droplets, ice crystals, or both, depending on their temperature and other parameters (Chapters 8 and 9).

Cirrus clouds have horizontal dimensions that are much larger than their vertical dimension. Because of their low ice water content, they appear almost transparent. Ice crystals can leave cirrus clouds as precipitation, which causes the cloud edges to become optically thin and to appear diffuse (Penner et al., 1999). In contrast, cloud droplets are much smaller than ice crystals and do not leave the cloud as precipitation until they have grown to raindrop size. Therefore, the edges of cumulus clouds appear quite sharp as long as they consist only of cloud droplets.

The World Meteorological Organization (WMO) further distinguishes clouds within these three categories and specifies them first in terms of the cloud type they belong to, also called *genera* (WMO, 1975). On the basis of the cloud shape, that is, the arrangement of the elements that compose it, a cloud type can be further divided into subtypes, called *species*. Finally, another subcategory can be added, called *varieties*. Varieties of cloud types/species allow the characterization of the spatial arrangement of the macroscopic cloud elements and the associated degree of transparency. Here we confine ourselves to a description of clouds only in terms of genus and species. For example, a cirrus cloud can be subdivided into the species *uncinus* or *floccus*. Cirrus uncinus (Ci unc) describes a cloud that has a comma-like shape (Figures 1.2 and 1.9c) and cirrus floccus describes small tufts of cirrus.

In total, 10 distinct cloud types exist according to the WMO. They are shown in Figure 1.2. It has proven useful to divide these cloud types into subgroups on the basis of the height at which they form, which reflects the synoptic situations in which they form. The subgroups contain low, mid-level, and high clouds, where *level* refers to the altitude of the cloud base above mean sea level. Cloud base heights differ for clouds in polar regions, midlatitudes, and tropical regions.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

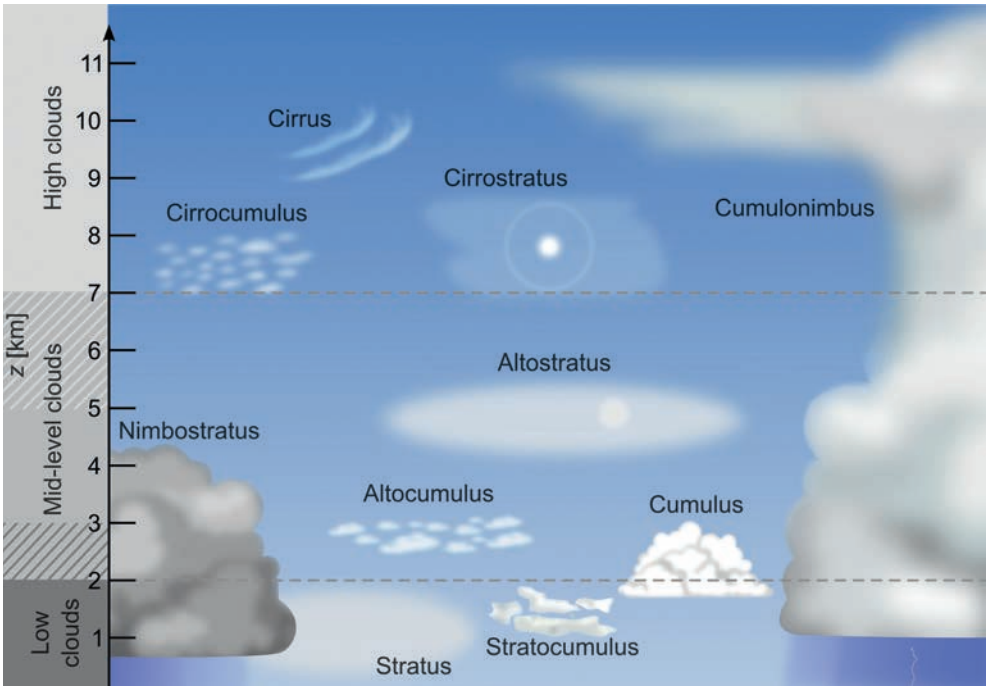


Figure 1.2 Schematic of the 10 cloud types according to WMO and as described in Table 1.1. Figure courtesy of Ditte Thomsen.

In order to distinguish the various cloud types, it helps to ask the following questions: How much of the sky is covered by this type of cloud? Is the Sun visible and, if so, is the Sun's disk sharply defined or diffuse? Is rain or snow falling from the cloud? If so, is the precipitation widespread or concentrated in narrow shafts? Are particular patterns, such as small cloud elements, rolls, or undulations visible, or does the cloud base appear uniformly gray? Table 1.1 summarizes the 10 main cloud types along with their abbreviations, their appearances, their typical base heights above Earth's surface as well as their global annual coverages. The range of cloud fractions and cloud base heights is illustrated in Figure 1.3.

In general, the annual mean cloud coverage is higher over the ocean, 68 %, as compared with that over land, 54 % (Eastman and Warren, 2013). The coverage of stratus, stratocumulus, and cumulus clouds is 8–10 % higher over the oceans than over land (Table 1.1), and their cloud bases are at lower altitudes because of the higher relative humidity in the marine PBL. Cirrus clouds, however, have a larger coverage over land, which could be due to a real physical difference or to difficulties in observing cirrus clouds over the oceans from ground (Eastman et al., 2011). One physical explanation for a higher cirrus coverage over land is cirrus formation caused by gravity waves triggered when air flows over mountains (Section 4.3).

1.2.1 Low Layered Clouds

Stratus (St) and *stratocumulus* (Sc) clouds (Figure 1.4d,e) comprise the low layered cloud types. These clouds are shallow stratiform clouds, usually less than 1 km in vertical extent

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

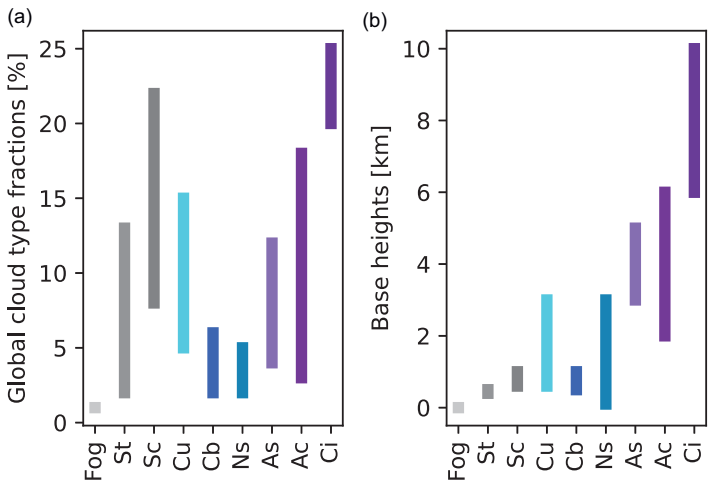


Figure 1.3 Visualization of the ranges of the (a) fractions and (b) base heights of the different cloud types and fog, as listed in Table 1.1. Differences between ocean and land are neglected here. Figure courtesy of Nadja Omanovic.

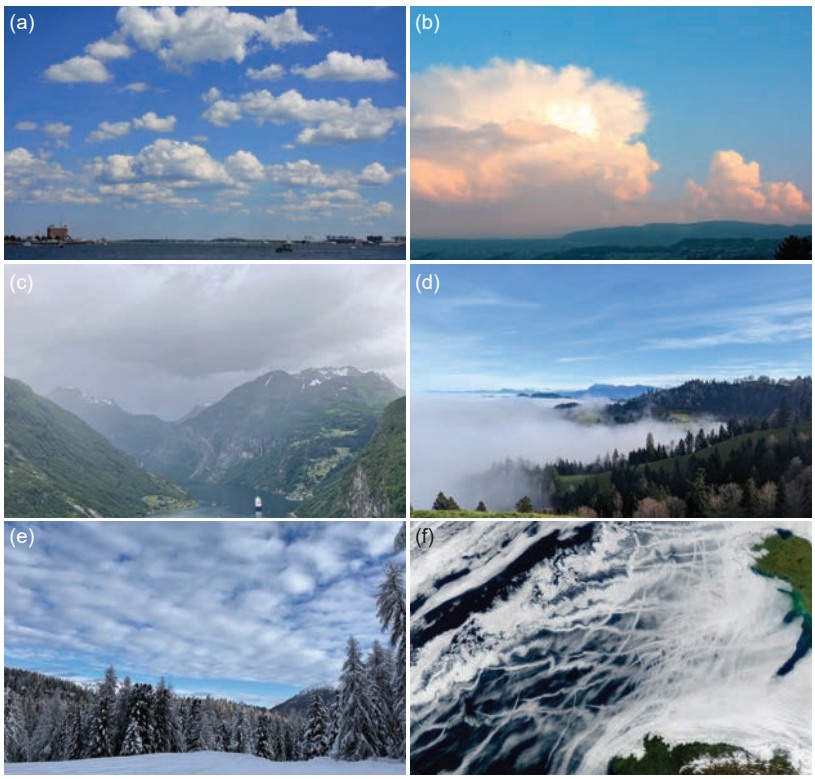


Figure 1.4 Photographs of (a) cumulus humilis (Cu hum), (b) cumulonimbus (Cb), (c) nimbostratus (Ns), (d) stratus (St), (e) stratocumulus (Sc), as viewed from above, and (f) satellite image of ship tracks off Europe's Atlantic coast. Photographs taken by Fabian Mahrt (a), (b), Jan Henneberger (d), and Ulrike Lohmann (c). Photo (e) is taken from the NASA Visible Earth Image Gallery (https://modis.gsfc.nasa.gov/gallery/) and was taken by the moderate resolution imaging spectroradiometer (MODIS) Land Rapid Response Team.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

(Rangno, 2002). Low layered clouds are often prevalent in areas that are characterized by subsiding air motions that correspond to areas with increased lower tropospheric stability (Klein and Hartmann, 1993). For example, stratus formation prevails in regions with strong subsidence. By contrast, if the subsidence is weak, the lower tropospheric stability is lower, and mostly cumulus clouds form.

Stratocumulus clouds show some degree of patchiness, which is why cumulus enters the name. While neighboring cumulus clouds can be seen as clearly separate from each other, individual stratocumulus cloud elements are connected, forming a cloud layer and have flattened cloud tops. In contrast with stratus clouds, which are purely gray, stratocumulus clouds consist of a pattern of gray and white colors, where the rounded air masses of individual cloud elements can be recognized.

Stratus and stratocumulus clouds tend to produce drizzle in clean air masses in which the concentration of aerosol particles is low. These are primarily found over the oceans in the absence of anthropogenic pollution. *Drizzle* denotes light rain and has drop radii between 25 μm and 0.25 mm (Table 1.2 and Figure 1.5). Drizzle drops are large enough to fall by the action of gravity but are smaller than raindrops, with radii between 0.25 and 5 mm (Houze, 1993). Rain and drizzle formation will be discussed in Chapter 8. Sometimes so-called *ship tracks*, which owe their existence to exhaust from ships, can be observed over the oceans. Ship tracks are visible as bright lines in Figure 1.4f. Ship exhaust creates a much higher number concentration of aerosol particles than normally found over the oceans. As will be discussed in Chapter 7, a higher number concentration of aerosol particles causes the cloud to consist of more cloud droplets. Generally this implies that the droplets will be smaller, as more of them have to compete for the water vapor available for condensation, which has implications for the cloud radiative effect. In Chapter 8 we will see that it takes much longer for small droplets to form precipitation than for large droplets. The time it takes for precipitation formation in

Table 1.2 Typical sizes and densities of hydrometeors. Values in parentheses indicate typical ranges.

Hydrometeor	Size (mm)	Density ($kg\,m^{-3}$)	References
Cloud droplet radius	0.005 (0.002–0.025)	1000	
Drizzle drop radius	0.05 (0.025–0.1)	1000	Peng et al. (2002); Sant et al. (2013)
Raindrop radius	0.5 (0.1–5)	1000	Pruppacher and Klett (1997); Sant et al. (2013)
Ice crystal length	0.1 (0.001–1.5)	900 (solid ice), else 400–900	Straka and Mansell (2005); Lasher-Trapp (2025)
Snowflake maximum dimension	10 (0.1–100)	100–300	Straka and Mansell (2005); Lasher-Trapp (2025)
Graupel radius	0.25–2.5	400 (200–700)	Milbrandt and Morrison (2013); Heymsfield et al. (2018)
Hailstone radius	>2.5–9.2	800–900	Straka and Mansell (2005)

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

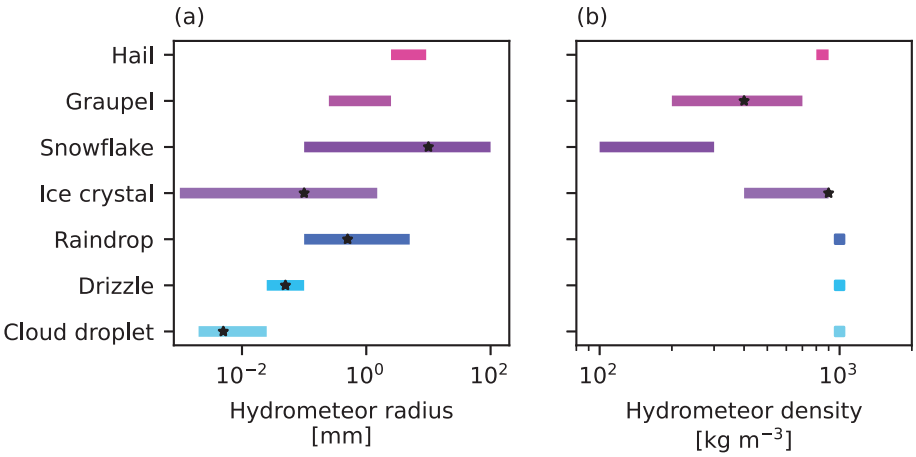


Figure 1.5 Visualization of the hydrometeor characteristics, as listed in Table 1.2. Figure courtesy of Nadja Omanovic.

a cloud with a high cloud droplet number concentration, such as in a ship track, can be longer than the lifetime of the cloud, which means that drizzle formation is inhibited and the cloud may dissipate before any precipitation has been formed. Ship tracks broaden with increasing distance from the moving ship due to the turbulent mixing of environmental air masses with the ship track. This causes the dilution of the ship track until it either evaporates or cannot be distinguished from the background cloud any longer.

1.2.2 Low Clouds with Vertical Extent

Nimbostratus (Ns) and convective clouds (Cu and Cb) are characterized by a low base height and a large vertical extent. Though nimbus is the Latin word for a cloud, the term is normally used to denote precipitating clouds. Nimbostratus is always found in the mid-troposphere. Ns clouds typically extend to low levels, but can also reach high levels. Since its base height is usually around 2 km, we regard nimbostratus as having a low cloud base. Nimbostratus is a formless cloud that is almost uniformly dark gray because it is optically thick enough to block most of the sunlight (Section 2.3). Nimbostratus is typically associated with long-lasting stratiform precipitation (Figure 1.4c and Section 10.4.1) and forms when air masses are lifted along a warm front in a low-pressure system.

Convective clouds, on the contrary, develop in unstable air due to buoyancy. In the atmosphere, stability or instability is mainly determined by the vertical temperature gradient. In an unstable atmosphere, the environmental temperature decreases strongly with increasing altitude such that a rising air parcel remains warmer than the surrounding air, in spite of the cooling due to adiabatic expansion (Chapter 3). If the air is sufficiently moist, buoyantly rising air parcels can cause the formation of convective clouds (Section 5.2.3). Upon cloud formation, the air parcel gains additional buoyancy from the latent heat released during condensational growth of cloud droplets.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

An example of a cumulus cloud, a cumulus humilis (Cu hum), is shown in Figure 1.4a. The designation *Cu hum* is a combination of the *cumulus* genus and the *humilis* species to which the cloud belongs. Cu hum is the smallest cloud of the cumulus genus, also called *fair weather cumulus*, with a vertical extent of less than 1 km (Zhang and Klein, 2013). Zhang and Klein (2013) classified continental Cu hum into thin ones, with an average cloud depth (vertical extent) of 220 m and an aspect ratio (depth/width) of 0.5, and thick ones, with an average cloud depth of 720 m and an aspect ratio of 2. They found that cloud bases vary generally between 2 and 3 km for the thin ones and between 1 and 2 km for the thick ones. The vertical growth of Cu hum can be restricted by a temperature inversion, which terminates the upward motion of an air parcel. However, a temperature inversion is not required to stop the growth of cumuli. An air parcel stops rising when its temperature is colder than that of the environment. This height can be predicted by comparing how the environmental (ambient) temperature changes with height, the so-called *environmental lapse rate*, with the *air parcel's lapse rate* (Chapter 4).

Capping inversions or smaller environmental lapse rates cause the uniform cloud top heights of Cu hum; they are a feature of this cumulus species only. A capping inversion often coincides with the top of the PBL. Its vertical extent typically ranges from 50 m during wintertime in the Arctic to over a few hundred meters in mid-latitudes at night and up to 2 km in the tropics.

A cumulus cloud that has a slightly higher buoyancy than the surrounding cumuli is able to penetrate through the capping inversion. There it may develop into the slightly taller *cumulus mediocris* (Cu med) if the atmosphere above the inversion is unstable, allowing the cloud to extend further vertically. Taller (and hence bigger) cumuli also develop if the atmosphere is unstable throughout. Like Cu hum, Cu med clouds do not precipitate and frequently develop into *cumulus congestus* (Cu con) (Figure 1.6). This growth occurs over the course of the day if the Sun continues to heat the surface, leading to thermally induced upward motions. Moreover, Cu con can form behind a cold front where the atmosphere becomes increasingly unstable and rising air parcels acquire more buoyancy. These clouds can develop vertical extents of 6 km. They appear as cauliflowers, where the cauliflower elements are indicative of the exchange between cloudy and environmental air in the form of entrainment and detrainment (Section 5.2.6). They may produce abundant precipitation, especially in the tropics (Johnson et al., 1999). Their sharp outlines suggest that they still consist mainly of cloud droplets even at temperatures below 0 °C.

Cu con can further develop into *cumulonimbus* (Cb, Figures 1.4b and 1.6) if the atmosphere becomes even more unstable and the cloud extends high enough that ice crystals form. Part of the buoyancy in cumulonimbus originates from the freezing of cloud droplets into ice crystals and the associated release of latent heat of fusion, which further warms the cloud relative to its environment and hence causes further upward motion. When cumulonimbus reach the temperature inversion of the tropopause, further ascent is stopped by the stable stratification of the stratosphere. This causes the tops to flatten and the cloud to spread out horizontally, leading to the formation of an anvil. Cumulonimbus that have developed an anvil are characterized by a region of active convection on the upwind side and a stratiform anvil on the downwind side (Chapter 11). The smooth fibrous anvils of cumulonimbus are indicative of ice crystals falling with different speeds and sublimating below the cloud.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

On a given day, it is common to observe cumuli at different stages of development (Figure 1.6) because of slight differences in buoyancy. After some time, the seemingly random cumuli can develop some organization, enabling the growth of taller cumuli (Grabowski et al., 2006; Schlemmer and Hohenegger, 2014). The taller cumuli in turn develop downdrafts, which can either cause their dissipation or lead to the formation of larger complexes such as multicells and supercells (Chapter 11). Only cumulonimbus can be accompanied by lightning, thunder, or hail. The common prerequisite for these hazards is the coexistence of water, ice, and strong updrafts.

1.2.3 Mid-Level Clouds

Mid-level clouds are optically thinner than their corresponding low clouds. This is because of the exponential decrease of air density, including water vapor with increasing altitude, such that less water vapor condenses to form the cloud. In addition, the number concentration of cloud active aerosol particles decreases with increasing altitude. Mid-level clouds comprise *altocumulus* (Ac) and *altostratus* (As). They typically owe their existence to slow upward lifting of air masses (Table 1.3) with vertical velocities (= updraft speeds) on the order of centimeters to tens of centimeters per second in the mid-troposphere (Rangno, 2002). Ac are usually 200–700 m thick, whereas As clouds are usually 1–3 km thick (Khvorostyanov and Curry, 2014).

Like St clouds, As clouds are layered with a uniform appearance. Their cloud bases are usually in the mid levels of the troposphere (2–7 km), while St clouds have low base heights around 0.5 km (Figure 1.3b). Even though As clouds are usually found in mid levels, they often extend higher. Moreover, they have smaller optical depths than St clouds, that is, they attenuate radiation less effectively and consequently appear less gray, but their optical depths are larger than those of cirrus clouds. The optical depth of As is small enough that the Sun's appearance is as if one were looking through translucent glass, meaning that the Sun does not have a clear outline (Figure 1.7a). If the large-scale cloud coverage of the sky changes from Cs to As, this is often an indicator of an approaching warm front or occlusion (Chapter 10).

Like other cumulus clouds, Ac clouds are associated with local instabilities and convection. They consist of distinct cloud elements and can exist as either detached clouds (altocumulus

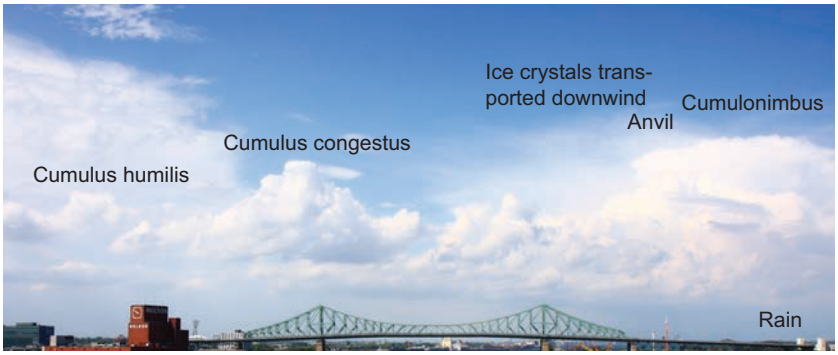


Figure 1.6 Photograph of cumulus clouds in different stages of development, taken by Fabian Mahrt.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press



Figure 1.7 Photographs of (a) altostratus (As) and (b) altocumulus (Ac), taken by Ulrike Lohmann.

castellanus [Ac cas]) or as rolls in layers and patches (as shown in Figure 1.7b). Ac castellanus clouds observed in the morning are a good indicator of afternoon convection[4]. Altocumulus lenticularis (Ac len) are caused by an oscillatory motion of the air in a statically stable atmosphere (Section 4.3).

Sometimes precipitation falling from As and Ac clouds is visible in the form of fallstreaks, also called *virga*[5], see Figure 1.8. Virga is the term for precipitation that evaporates or sublimates before it reaches the ground (Section 8.3).

1.2.4 High Clouds

The three cloud genera comprising in situ-formed high clouds are Cc, Cs , and Ci, as shown in Figure 1.9a–c. Cc is a finely granulated cloud, consisting of many small, similar-looking cloud elements. These small granular structures are often arranged in a regular pattern, where the individual cloud elements are identifiable and intersected by small areas of blue sky. Cs appears as a whitish veil, having a waveless structure where no individual cloud elements can be identified. A distinct characteristic of Cs is the appearance of a halo (Figure 1.9b) if the conditions are right. A halo is an optical phenomenon that appears as a bright ring at angular radii of 22 ° or 46 ° around the Sun (Section 2.4.4). The Ci genera consists of delicate filaments that belong to neither the Cc or Cs genera.

These three cloud genera and their associated species are solely comprised of ice crystals. Because of the decrease of water vapor with height in the troposphere, cirrus clouds have a much lower water content than mid-level or low clouds (Table 1.3). Thus, cirrus clouds are generally not optically dense enough to produce shading, except when the Sun is near the horizon (Rangno, 2015).

Small ice crystals can be advected by the wind, causing cirrus clouds to become vertically sheared, as is the case for cirrus uncinus (Ci unc) shown in Figure 1.9a, with characteristic hooks at their upwind ends. However, not all high clouds are characterized by this fibrous

[4] https://forecast.weather.gov/glossary.php?letter=a
[5] https://cloudatlas.wmo.int/en/clouds-supplementary-features-virga.html

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

Table 1.3 Typical macroscopic and microphysical properties for various cloud types: cloud droplet number concentration (N_c in cm^{-3}), cloud droplet radius (r_d in μm), liquid water content (LWC in $g\,m^{-3}$), ice crystal number concentration (N_i in cm^{-3}), ice crystal length (in μm), volume-mean ice crystal radius ($\bar{r}_i$ in μm), ice water content (IWC in $g\,m^{-3}$), vertical velocity (in $m\,s^{-1}$), vertical extent (in km), and lifetime (either in min or in h). Note that the lifetimes refer to the cloud as a whole, not the time air parcels spend in them, which can be considerably shorter. "oc" refers to clouds over oceans and "la" to clouds over land. Data have been compiled from the WMO cloud atlas (WMO, 1975; Pruppacher and Jaenicke, 1995; Pruppacher and Klett, 1997; Miles et al., 2000; Lehmiller et al., 2001; Quante, 2004; Deierling and Petersen, 2008; Cotton et al., 2011; Khvorostyanov and Curry, 2014). References for the individual cloud types are included in the table.

Parameter	Typical value	Typical range	References
Liquid fog			Mazoyer et al. (2019)
N_c	60	5 to 160	
r_d	7.5	1 to 15	
LWC	< 0.2	0.05 to 0.2	
Vertical velocity	0.01		
Vertical extent	0.1	0.05 to 0.6	
Lifetime (h)	3	2 to 6	
St, Sc: drizzle and snow possible			Wood (2005, 2012)
N_c	75 (oc), 290 (la)	<10 to 500 (oc), 40 to 1000 (la)	
r_d	7 (oc), 4 (la)	2 to 12.5	
LWC	0.2	0.1 to 0.6 (oc), 0.05 to 0.5 (la)	
N_i		3×10^{-3} to 0.75	Hobbs and Rangno (1996); Lohmann et al. (2016)
IWC		0.1 to 0.2	Hobbs and Rangno (1996); Lohmann et al. (2016)
Vertical velocity	0.1 (St), 0.5 (Sc)		
Vertical extent	0.5 (St), 0.3 (Sc)	0.2 to 0.7 (oc), 0.2 to 0.9 (la)	
Lifetime (h)	6	2 to 12	
Cu hum, Cu med			Squires (1958); Zhang and Klein (2013)
N_c	50 (oc), 250 (la)	20 to 200 (oc), 100 to 1000 (la)	
r_d	14 (oc), 6 (la)	2 to 25	
LWC	0.3	0.2 to 1	
Vertical velocity	0.7	0.5 to 3	
Vertical extent	0.5 (Cu hum, la), 1.5 (med)	0.2 to 1.2 (Cu hum, la)	
Lifetime (min)	20	10 to 40	
Cu con: rain, snow and graupel possible			
N_c	60	20 to 100 (oc), 50 to 900 (la)	
r_d		4 to 50	
LWC	2	0.5 to 2.5	

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

Table 1.3 Cont.

Parameter	Typical value	Typical range	References
N_i		0.01 to 1600	Heymsfield et al. (2013); Lawson et al. (2015, 2022)
IWC	1	0 to 2.8	Heymsfield et al. (2013); Lawson et al. (2015, 2022)
Vertical velocity	4	3 to 10	
Vertical extent	5	>3	
Lifetime (min)	30	20 to 45	
Cb: rain (always), snow, graupel and hail possible			
N_c		1 to 100 (oc)	
LWC		1.5 to 14	
N_i		10^{-3} to 1.6	Heymsfield et al. (2013); Lawson et al. (2015, 2022)
Ice crystal length		10 to 10^4	
IWC		10^{-3} to 8	Heymsfield et al. (2013); Lawson et al. (2015, 2022)
Vertical velocity	10	5 to 50	
Vertical extent	10	6 to 12	
Lifetime (min)	60	45 to 180	
Ns: rain (always), snow and ice pellets possible			
LWC		0.2 to 0.5	
Vertical velocity	0.1	0.1 to 2	
Vertical extent	3-4	2 to 10	
Lifetime (h)	8	6 to 12	
Ac, As: rain, snow, ice pellets			
N_c	100	30 to 1000	
r_d	4	2 to 10	
LWC, IWC	0.2	0.01 to 0.5	
Vertical velocity	0.1	0.01 to <0.5 (As), 0.01 to <1 (Ac)	
Vertical extent	0.5 (Ac), 1 (As)	0.2 to 0.7 (Ac), 1 to 3 (As)	
Lifetime (h)		6 to 12	
Cc, Ci, Cs			Krämer et al. (2009, 2020); Heymsfield et al. (2017)
N_i	0.03	2×10^{-3} to 0.3	
Ice crystal length	100	1 to 1500	
$\bar{r}_i$	20	2 to 80	
IWC		10^{-5} to 1	
Vertical velocity	0.1	0.01 to 1.5	
Vertical extent		0.2 to 0.4 (Cc), 0.1 to 3 (Cs)	
Lifetime (h)		0.2 to 12	

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press



Figure 1.8 Photograph of virga, taken by Laurent Barbe.

pattern. Like cumulus clouds, the cirrocumuli shown in Figure 1.9c consist of individual cloud cells rather than thin ice particle streamers. When the cloud edges and the ice crystals are advected away from their initial cell, a fibrous appearance develops (Houze, 2014).

In addition to in situ-formed high clouds, cumulonimbus leave stratiform *anvil cirrus* behind in their dissipating stage (Section 11.1). Deep convection transports moisture and condensate from the lower troposphere to the upper troposphere, causing anvil cirrus to have a higher ice water content than in situ-formed cirrus (Krämer et al., 2016) and a larger vertical extent (Stubenrauch et al., 2023). During transport, they had time to grow to larger sizes than ice crystals in in situ-formed cirrus (Heymsfield et al., 2013; Krämer et al., 2016). Anvil cirrus also have a larger optical depth than in situ-formed cirrus (Stubenrauch et al., 2023).

Water vapor associated with aircraft emissions can cause anthropogenic clouds to form, so-called *contrails* or long condensation trails (Schumann and Heymsfield, 2017; Kärcher, 2018), which are also called *cirrus homogenitus*[6] (Figure 1.9d). For contrails to form, the relative humidity conditions need to be just right: If the upper troposphere is too dry, no contrail is visible. If it is a bit moister, contrails are only visible for a short time, that is, they form but quickly dissipate again. If it is even moister, contrails form and are long-lived (e.g., Schumann, 2012). Over time, turbulence causes them to widen, causing the initial homogeneous, line-shaped contrail to evolve into contrail cirrus (Voigt et al., 2017).

[6] https://cloudatlas.wmo.int/en/aircraft-condensation-trails.html

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

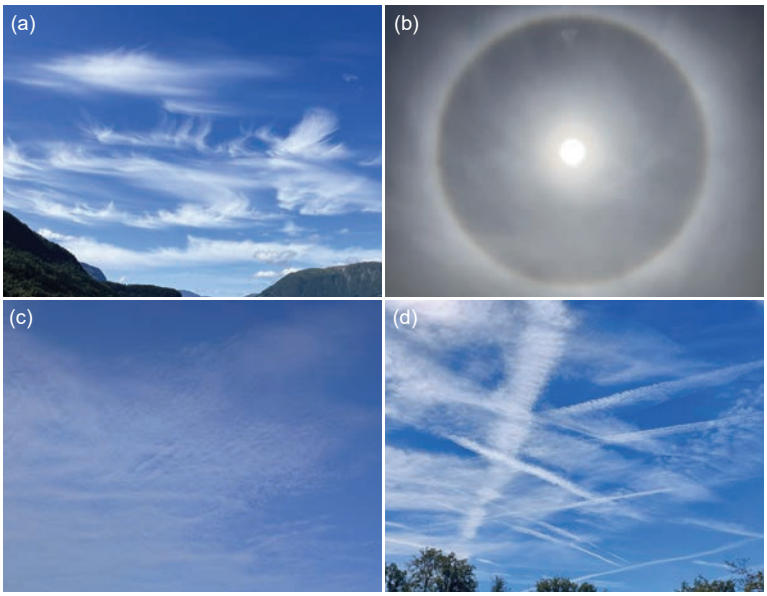


Figure 1.9 Photographs of (a) Cirrus uncinus (Ci unc), (b) cirrostratus (Cs) with a 22° halo, (c) cirrocumulus (Cc), and (d) contrail cirrus (cirrus homogenitus). Photographs taken by Ulrike Lohmann (a), (b) and by Fabian Mahrt (c), (d).

1.2.5 Other Cloud Types

There are some additional cloud types that are not included in the WMO cloud atlas. A cloud type that is restricted to isolated and sharp mountain peaks, such as the Matterhorn in Switzerland or Mount Everest in Nepal, is the *banner cloud* shown in Figures 1.10 and 1.11a. It is a cloud plume that extends downwind from an isolated obstacle even on otherwise cloud-free days. Banner clouds form because an obstacle in a horizontal flow causes the air masses to split: part of the air may be forced vertically upward over the obstacle, while the remaining air diverges horizontally around it. The diverging air masses cause a vortex in the lee of the mountain that rotates around a horizontal axis. This lee vortex is associated with pronounced upwelling close to the leeward face of the mountain (Voigt and Wirth, 2014; Prestel and Wirth, 2016). If this air is sufficiently moist, the contained water vapor will condense before the air reaches the top of the mountain and thus forms a banner cloud. This mechanism implies that the air forming the banner cloud on the leeward side of the mountain originates from lower levels than the air on the windward side. If this is not the case, a cloud would also form on the windward side (Schween et al., 2007). After the cloud has been formed, it experiences the main wind flow around the mountain and hence is advected by it, leading to the formation of an elongated banner.

Another rarely observed cloud type is the so-called *roll cloud* that can be seen in Figure 1.11b. It is characterized by strong updrafts in front of it and downdrafts at its rear causing it to appear as if rolling. This cloud type occurs in connection with *land-sea breezes* (Section 10.6) and convective storm activity. Roll clouds are frequently observed in Australia, where they

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

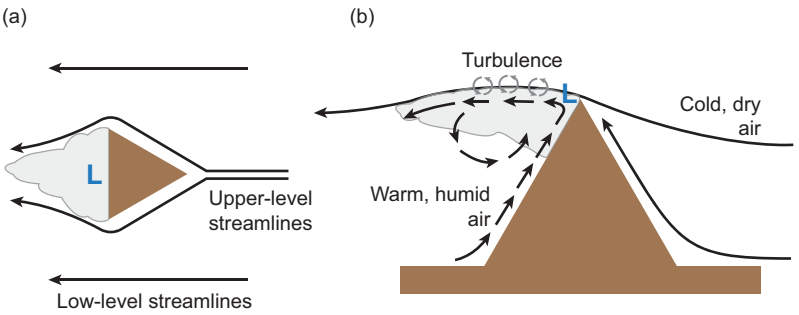


Figure 1.10 Schematic illustration of the formation of a banner cloud viewed from top (a) and from the side (b).



Figure 1.11 Photographs of (a) a banner cloud, (b) roll cloud (the dark shape above a light horizon), (c) mammatus cloud, (d) asperatus cloud, (e) Kelvin–Helmholtz wave cloud, and (f) Ac lenticularis clouds. Photographs taken by Christopher Fuchs (a), Donovan Reese/Photodisc/Getty Images (b), Robert David (c), Christian Grams (d), Kate Calder/Moment Open/Getty Images (e), and Don Smith/Photodisc/Getty Images (f).

are associated with the land-sea breeze. The only place where they occur regularly is over the southern part of the Gulf of Carpentaria in northern Australia in the Austral spring. As these roll clouds typically arrive in the morning at the coast of Queensland, they are also called *morning glory clouds* (Porter and Smyth, 2002; Goler et al., 2006). Roll clouds can also indicate the leading edge of convective storm outflow (Chapter 11). They form horizontally because of the circulating (rolling) air masses. Along the direction in which the convective storm is moving, cold air outflow at Earth's surface precedes the storm and marks its leading edge. The cold outflow forces the warm environmental air to be lifted and, if it is sufficiently moist, a roll cloud forms.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

Mammatus clouds got their name from "mamma", the Latin name for breast. They are a rare example of clouds in sinking air (Figure 1.11c) (e.g., Schultz et al., 2006; Ravichandran et al., 2020). The term mammatus refers to the cellular pattern of pouches hanging underneath the cloud base, caused by gravity. Mammatus clouds typically develop below the anvil of a convective storm. The high concentration of heavy hydrometeors (Section 1.4.2) in anvils exerts a gravitational force, so that the air in the anvil starts sinking. The warming of the subsiding air and the associated decrease in relative humidity lead to evaporation/sublimation of the hydrometeors. This, in turn, causes cooling of the air due to the latent heat consumption of evaporation/sublimation and counteracts the warming due to subsidence. If the latent heat consumption exceeds the adiabatic warming, the air will continue to sink. The subsiding air with the evaporating/sublimating hydrometeors eventually appears as rounded pouch-like structures below cloud base.

A similar feature is seen in the so-called *asperatus cloud* (Figure 1.11d). The term asperatus means roughened or agitated and refers to their wavy cloud base. This unique feature distinguishes them from all other cloud types. Their wavy cloud base again indicates sinking cloudy air. They are particularly common in the Great Plains in North America during the morning or midday after convective storms. Even though asperatus clouds appear dark and storm-like, they are harmless leftovers of convective storms and will dissipate.

Another special cloud is the *Kelvin–Helmholtz wave cloud*. A Kelvin[7]–Helmholtz[8] instability arises from small perturbations at the interface of two fluids with different velocities, an example of which are waves on a lake or ocean. Such a shear instability can also occur within one medium. In the atmosphere, the resulting waves form along the interface between two atmospheric layers with a sufficiently large difference in horizontal velocity. The shear instability is manifested in the form of Kelvin–Helmholtz billows (Figure 1.11e) if the air that originates from the moister lower layer is sufficiently lifted for the air to reach saturation and to condense. Finally, another special cloud type are lens-shaped lenticularis clouds. An example of an Ac lenticularis clouds is shown in Figure 1.11. These cloud types often form due to oscilllatory motion of the air, as we will see in Chapter 4.

1.2.6 Fog and High Fog

While clouds can form due to adiabatic expansion and cooling in rising air parcels or air masses, that is, vertical motions (Section 3.2), fog forms as a result of isobaric cooling, that is, cooling at a constant pressure in the absence of vertical motions (Section 5.1), typically occuring at or near the ground. Similar to the definition of a cloud, fog is defined according to visibility. Fog is defined as when the horizontal visibility is reduced to <1 km, whereas if the visibility is larger, the term *mist* or *light fog* is used.[9]

Several types of fog can be distinguished, as shown in Figure 1.12 (e.g., Eugster, 2008). Most commonly, fog forms by radiative cooling at night and it occurs most often in the early mornings during autumn and winter. This kind of fog is called *radiation fog*. At night, the

[7] www.britannica.com/biography/William-Thomson-Baron-Kelvin
[8] www.britannica.com/biography/Hermann-von-Helmholtz
[9] https://cloudatlas.wmo.int/en/fog-compared-with-mist.html

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

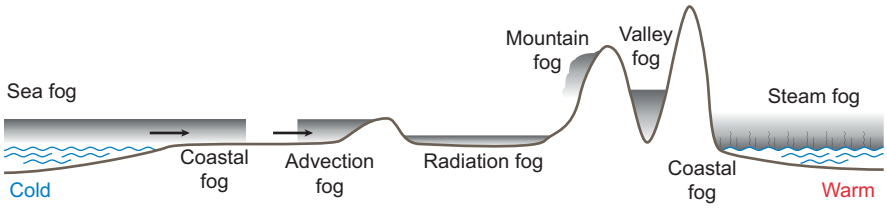


Figure 1.12 Schematic illustration of the different fog types after Eugster (2008).

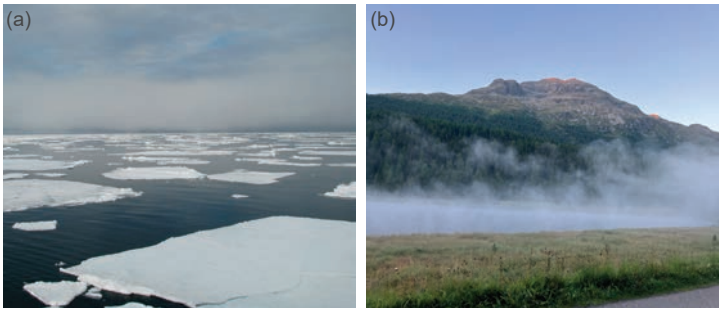


Figure 1.13 Photographs of (a) radiation fog and (b) steam fog, taken by Blaz Gasparini and Ulrike Lohmann, respectively.

longwave radiation emitted from Earth's surface causes the air to cool. If it is sufficiently moist, the relative humidity can reach 100 % with respect to water, allowing condensation to set in. Usually, fog does not extend much in the vertical (a few hundred meters at most) so that heating by solar radiation during the day causes the fog droplets to evaporate and the fog to disappear. A photograph of a radiation fog is shown in Figure 1.13a. Other types of fog are advection fog, which can have its origin in radiation fog near the coast over land, from where it is advected over the ocean by the nocturnal land breeze (Pilié et al., 1979). Fog can also form when moist and warm air passes over a cooler surface and cools at its surface. This then leads to mixing fog, because the mixing of the warmer, moister air above with the colder air below could lead to a supersaturated mixed air mass out of which the mixing fog forms (Section 5.1). The opposite case of mixing fog is often observed in fall, when colder air passes over warmer water (lakes, rivers, ocean). This fog is termed *steam fog* (Figure 1.13b). Typically, fog consists of cloud droplets. If fog is solely composed of ice crystals, it is termed *ice fog*[10] (Gultepe et al., 2015). Ice fog usually occurs at temperatures below −30 °C but can already occur at temperatures below −15 °C (Gultepe et al., 2015). Ice fog consists of small ice crystals with lengths (maximum dimensions) of less than 200 μm (Gultepe et al., 2020). If the horizontal visibility in ice fog drops below 50 m, this suggests that the ice crystals are even smaller, with diameters between 2 and 30 μm.[11]

In German-speaking countries, the term *high fog* is used to characterize low stratus trapped between mountains. In terms of appearance, high fog cannot be distinguished from stratus clouds. High fog can either result from isobaric mixing at the altitude where it occurs (typically at the top of the PBL, well separated from Earth's surface) or as a result of ground fog being heated near the surface by surface heat fluxes and lifted to higher altitudes. During winter,

[10] https://glossary.ametsoc.org/wiki/Ice_fog
[11] https://cloudatlas.wmo.int/en/ice-fog.html

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

when the solar radiation reaching the fog layer is not strong enough to dissipate it, high fog may persist for days. As such, high fog is a frequent occurrence in Alpine valleys and over the Swiss Plateau in fall and winter (Scherrer and Appenzeller, 2014; Henneberger et al., 2023).

1.3 Global Coverage of Individual Cloud Types and Cloud Regimes

1.3.1 Clouds at Different Altitudes

The categorization of clouds into different altitude levels is important for their radiative properties (Section 2.3). Figure 1.14 provides an overview of the cloud cover in different levels of the atmosphere, based on satellite data (Mace et al., 2009; Chepfer et al., 2010) for the period 2006–2011 for December, January, February (DJF) and June, July, August (JJA) (Boucher et al., 2013). Note that, from the satellite perspective, low clouds with vertical extents, which are considered as a separate class in the WMO cloud type definition (Table 1.1), are included in the high cloud category because of their high cloud tops. In regions with high clouds, low clouds are often underestimated by satellite data, because of shadowing by overlying clouds (Chepfer et al., 2008).

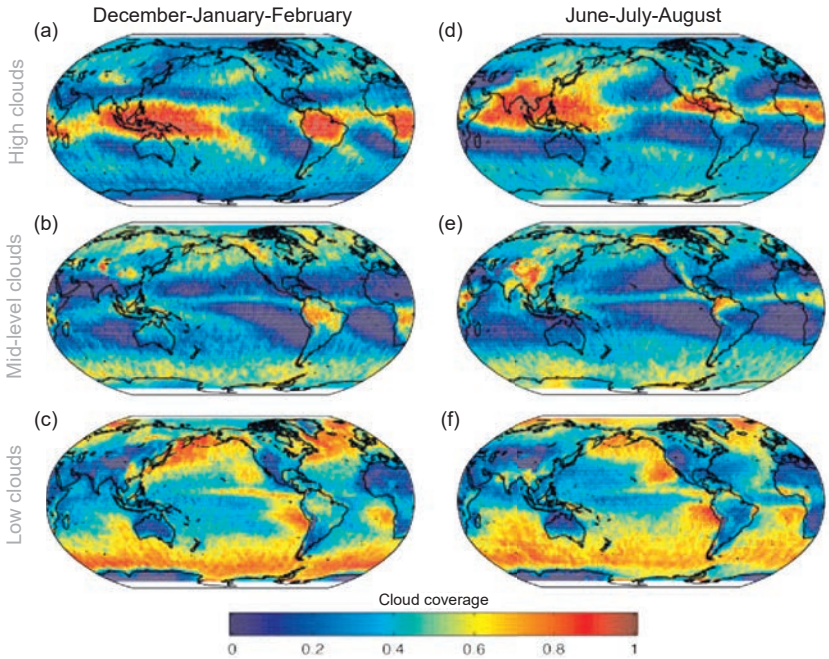


Figure 1.14 (a)–(c) DJF and (d)–(f) JJA averaged high, mid-level, and low cloud coverage based on data from CloudSat and CALIPSO for the period 2006–2011 (Mace et al., 2009). For low clouds, only the CALIPSO data set for 2007–2010 (Chepfer et al., 2010) is used at locations where it indicates a larger fractional cloud cover than the CloudSat data set, because the latter removes some clouds with tops at altitudes below 750 m. Figure adapted from Figure 7.6 of IPCC AR5 (Boucher et al., 2013).

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

The areas with the largest fraction of high clouds are found in the tropics (Figure 1.14a, d). These regions are associated with large-scale water vapor transport into upper levels by persistent deep convection and encompass the West Pacific Warm Pool, the Amazon, and central Africa. The ITCZ and SPCZ are also visible. As a consequence of the annual cycle of Sun's zenith, the ITCZ is on average at 5 °S in DJF but moves to, on average, 15 °N in JJA because of the larger area of land masses in the northern hemisphere which warm faster than the oceans. The largest area of high cloud cover in JJA is associated with the deep convective clouds and torrential rainfalls that characterize the Indian summer monsoon. Here, the shift in the ITCZ is most pronounced. In DJF, maxima in mid-latitude high clouds are associated with extratropical cyclones in the southern ocean, in the north Pacific, and in the north Atlantic along the extratropical storm tracks, while in JJA mid-latitude high clouds in the northern hemisphere are mainly associated with summertime convection.

Mid-level clouds are found least frequently of all clouds (Figure 1.14b, e) because the mid-troposphere is on average rather dry. Summing up the cloud cover per cloud type in Figure 1.15 yields a coverage of mid-level clouds (As, Ac, and Ns) of 11.5 %, compared to that of high clouds (Ci, multi-layer clouds (MLC) and deep convective clouds (D.C.)) of 38.6 % and that of low clouds (St, Sc, and Cu) of 20.8 %.

Usually, deep convective clouds are accompanied by smaller convective clouds in their vicinity whose tops reach only into the mid-troposphere, giving rise to the maxima in mid-level clouds seen in the ITCZ and SPCZ. At mid and high latitudes, mid-level clouds form in connection with extratropical cyclones. Unlike deep convective clouds, which are predominantly caused by static instability (Section 4.4), mid-level clouds can also form by shear instability. In the atmosphere, shear instability is commonly caused by a vertical gradient in the horizontal wind, for which the Kelvin–Helmholtz instability is an extreme example (Fig. 1.11e).

Low clouds form under the subsiding branch of the tropical Hadley circulation cells (Figure 1.1) and over cold ocean currents, such as the California current off the coast of California, the Humboldt[12] current off the west coast of South America, the Benguela current off the west coast of southern Africa, the Canary current off the west coast of northern Africa, or the West Australian current. The vertical extent of low clouds is limited by the temperature inversion at the top of the shallow PBL, as discussed earlier. In addition, low clouds are found in the storm tracks of both hemispheres, where their coverage can exceed 80 % (Figure 1.14c, f). They also form in the warm sector of extratropical cyclones as well as behind cold fronts (see Chapter 10). The Arctic ocean has a high coverage of stratus clouds during summer when the sea-ice fraction is small and the Arctic ocean serves as a moisture source for cloud formation (Kay and Gettelman, 2009).

1.3.2 Individual Cloud Types

To understand the role of clouds for precipitation formation and climate in more detail, the partitioning of cloud cover into low, mid-level, and high clouds may not suffice. Thus, in the following, we discuss the distribution of the individual cloud types. Figure 1.15 shows

[12] www.britannica.com/biography/Alexander-von-Humboldt

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

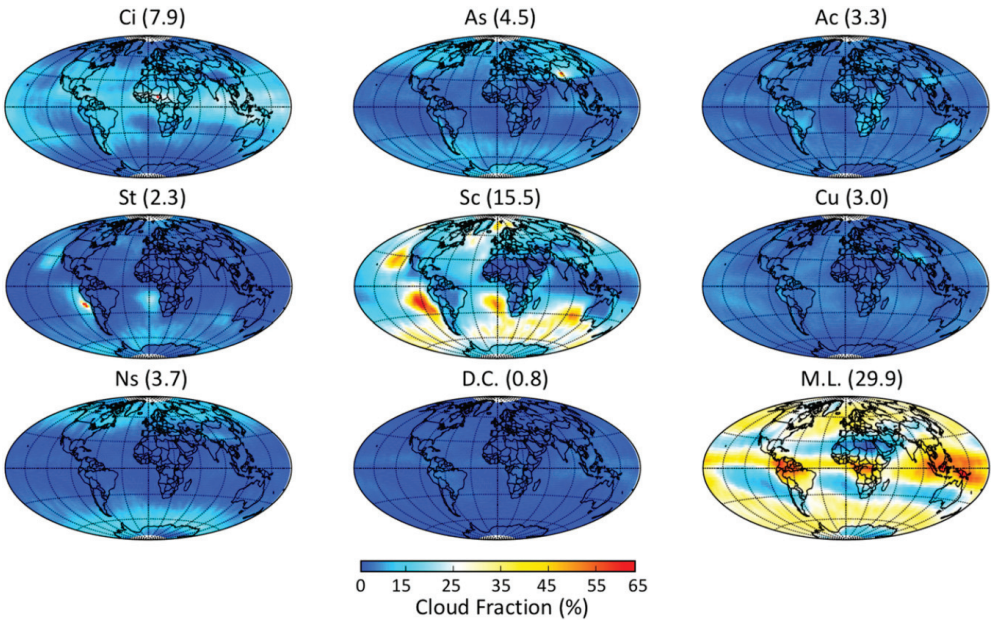


Figure 1.15 Percentage contribution of individual cloud types to the total cloud cover (Fig. 1.1) over the years 2007–2010, as obtained from the CloudSat and CALIPSO satellite data. Cloud types as classified in Table 1.1. D.C. denotes deep convective clouds. M.L. denotes multi-layer clouds, defined as when multiple distinct cloud layers are identified at different heights (vertical positions) at a given location. The values in parenthesis indicate the area-weighted global average cloud cover for each cloud type. Figure reproduced from L'Ecuyer et al. (2019). ©American Meteorological Society. Used with permission.

the breakdown of the global, annual mean cloud fraction by cloud type. Specifically, the cloud fractions shown correspond to the contribution of each cloud type to the total cloud cover over the period 2007–2010 shown in Figure 1.1 (L'Ecuyer et al., 2019), derived by applying classification schemes (Sassen and Wang, 2008) to the CloudSat and CALIPSO satellite data.

The low cloud types (St, Sc, and Cu) show the largest coverages over oceans, in particular off the west coasts of North and South America, Africa, and Australia, as well as over the storm track regions. The mid-level cloud types (As and Ac) are predominantly found over the storm tracks and the extratropics. Cirrus clouds are rather ubiquitous in the tropics and midlatitudes.

Most importantly, Figure 1.15 reveals that often more than one cloud type is present at a given location. In fact, 29.9% of the globe is covered by multi-layer clouds (L'Ecuyer et al., 2019), termed M.L. in Figure 1.15. They consist of multiple cloud layers that are vertically separated. In this satellite climatology, they are attributed to the high clouds category (L'Ecuyer et al., 2019). Note that earlier cloud categorizations often lack this multilayer cloud category and define the coverage in terms of the cloud type that is seen first (high clouds from satellite, low clouds from surface observations). As a comparison, the total percentage of the globe that is covered by single cloud layers or types amounts to 41%. Sc clouds are the WMO cloud type that occurs most common, with an area-weighted global average of 15.5%, whereas the

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press



Figure 1.16 Photograph of a pileus cloud. Photograph taken by Sylke Boyd, obtained from the WMO International Cloud Atlas, with permission from the WMO.

percentage coverage of the other WMO cloud types is below 8%. Deep convective (D.C.) clouds, equivalent to our definition of Cb clouds, make up a relatively small fraction, because they are horizontally small and short lived as compared to stratus clouds. D.C. clouds only represent 0.8% of the cloudy atmosphere and also count toward the high clouds category (L'Ecuyer et al., 2019). Most of the deep convective clouds in the tropics are multilayer clouds, because they have a clearly separated cirrus or *pileus* cloud on top (L'Ecuyer et al., 2019), an image of which is shown in Figure 1.16.[13]

1.3.3 Cloud Regimes in the Framework of the Global Circulation

As discussed earlier there are preferred geographical locations for some cloud types because their formation is associated with typical synoptic conditions. Instead of referring to geographical regions, it has been found useful to define cloud regimes in which similar cloud scenes are grouped together.

An example for a cloud regime are the different marine stratocumulus regions shown in Figure 1.1 (Section 1.1.2). These PBL clouds are found over the cold ocean currents in the subtropics, where large-scale subsidence in the free troposphere of the subtropical high pressure systems prevails. Geographically, they are found off the coasts of Africa, as well as North and South America, that is, in different geographical regions. However, their formation mechanisms are the same, hence they can be regarded as one popular cloud regime.

These marine stratocumulus organize into distinct patterns (e.g., Glassmeier and Feingold, 2017) and accordingly these regions can be further subdivided in terms of their appearance. The marine PBL has two stable states, a clean state that tends to remain clean and a polluted state that tends to remain polluted (Baker and Charlson, 1990). Evidence of these two states of the marine PBL is visible in *open cells* versus *closed cells* (Wood and Bretherton, 2006; Watson-Parris et al., 2021) as shown in Figure 1.17.[14] Open cells refer to rings of clouds in which air rises at the cell boundaries and subsides in the cell interior. The rising motion is generally triggered by the warm temperatures of the ocean. Clouds in these open cells produce drizzle, removing a fraction of the aerosol particles, and thus remain clean (Stevens et al., 2005, see Figure 1.17c). Closed cells, on the contrary, have a circulation opposite to that in open cells,

[13] https://cloudatlas.wmo.int/en/clouds-accessory-pileus.html
[14] https://earthobservatory.nasa.gov/images/43795/open-and-closed-cell-clouds-over-the-pacific-ocean

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

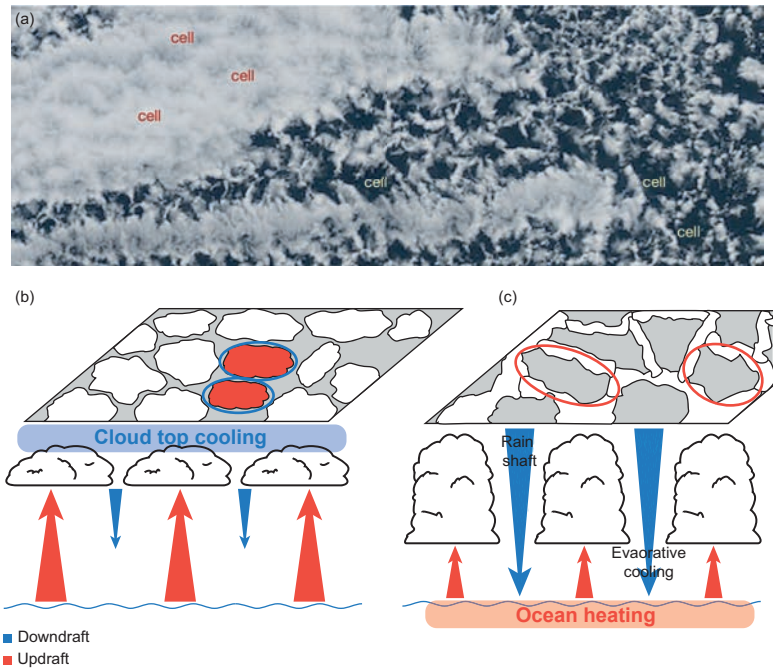


Figure 1.17 (a) Satellite images of closed-cell (marked in red) vs. open-cell (marked in yellow) marine PBL clouds as observed from the MODIS satellite on April 17, 2010 over the Pacific ocean off the coast of Peru. Schematic of the formation of (b) closed cells and (c) open cells over the ocean. Satellite image taken from the NASA Visible Earth Image Gallery (https://modis.gsfc.nasa.gov/gallery/).

with air rising in the middle and subsiding at the cell boundaries (Figure 1.17b). They tend to be characterized by the absence of drizzle, allowing aerosol particles to accumulate (Rosenfeld et al., 2006). Closed cells tend to predominantly form by cooling from top, that is, by emissions of longwave radiation. As shown in Figure 1.17, closed and open cells can also form in close proximity to each other due to local differences in ocean temperature or aerosol concentration (Rosenfeld et al., 2006).

Other methods to distinguish cloud regimes are according to cloud top properties retrieved from satellite data (cloud top pressure and optical depth) or according to large-scale subsidence and/or lower tropospheric static stability. The latter acknowledges that cloud regimes found in areas with little instability are different from those in areas with large instability. Examples of cloud regimes that form in subsidence regions are polar stratus clouds, closed cells over the cold ocean currents (Eastman et al., 2021) (Figure 1.17 in red) or associated with cold air outbreaks (Fletcher et al., 2016), and marine stratocumulus (Wood, 2012). Cloud regimes that require a more unstable environment are marine shallow cumulus and open cells (Stevens et al., 2005) (Figure 1.17). Deep convective clouds as found in the ITCZ and SPCZ (Figure 1.1) mark a cloud regime characterized by the highest instability. Within a given dynamically defined regime, further geographic refinements are possible. The stratus cloud regime can be

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

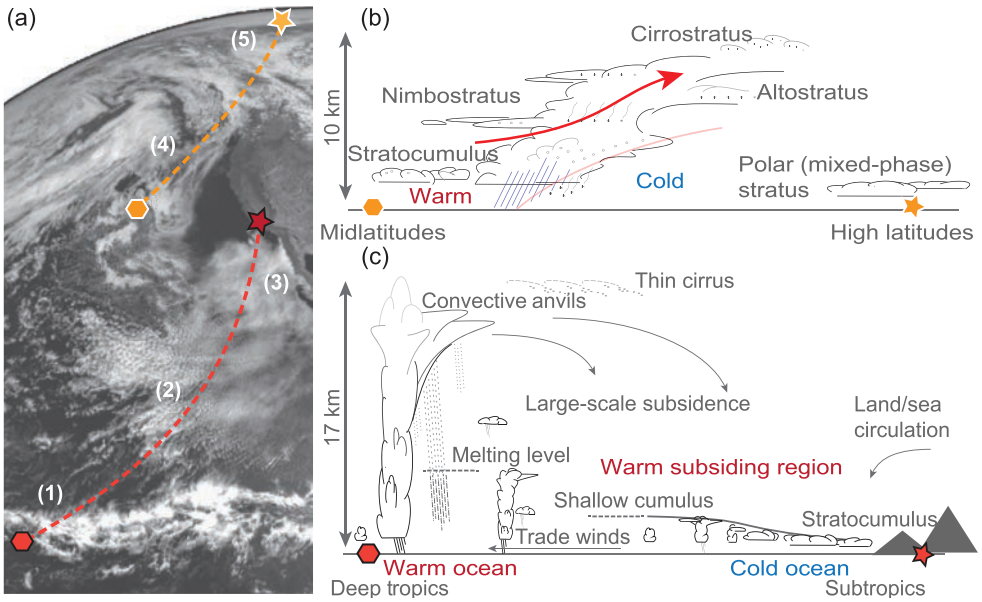


Figure 1.18 (a) A visible-wavelength geostationary satellite image over the northeastern Pacific. The numbers denote characteristic cloud regimes: (1) organized tropical convection, (2) trade wind cumulus, (3) marine stratocumulus, (4) stratiform clouds of a warm front, and (5) polar stratus. (b) A schematic of the different cloud types associated with extratropical cyclones along the orange black line moving from mid to high latitudes in (a). (c) A schematic of the different cloud types found along the dashed red line in (a) going from the subtropics off the coast of California to the tropics. Figure adapted from Figure 7.11 of IPCC AR5 (Boucher et al., 2013).

distinguished into warm-phase stratus decks in low- and midlatitudes, which cool the climate, and polar mixed-phase stratus decks, which warm the climate in DJF (Villanueva et al., 2022).

Figure 1.18a shows a geostationary satellite image in which some important cloud regimes can be identified. Going from south to north, these are: (1) organized tropical convection, (2) trade wind cumulus, (3) marine stratocumulus, (4) stratiform clouds of a warm front, and (5) polar stratus. The typical cloud types associated with frontal lifting along a warm front of an extratropical cyclone are shown in Figure 1.18b. For an observer, an approaching warm front is visible by its shield of cirrus clouds. These cirrus clouds turn into altostratus and then into an extensive region of nimbostratus as the front moves from mid to high latitudes.

Figure 1.18c schematically shows the Hadley circulation (Hadley, 1735) with its typical cloud types. The term *Hadley circulation* describes a thermally driven tropical circulation cell resulting from the latitudinal energy distribution. In the tropics, more shortwave radiation is absorbed than longwave radiation is emitted back to space. In contrast, the polar regions receive less shortwave radiation than the longwave radiation they emit to space. This results in an energy surplus in the tropics and an energy deficit in the polar regions. Thus, energy needs to be transported from the tropics to the polar regions. Part of this poleward energy transport occurs via ocean currents, and the other part occurs within the atmosphere. The atmospheric energy transport of the global circulation starts with the Hadley cell, which is the largest overturning circulation cell in the atmosphere. Heating from the Sun causes the air in

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

the ITCZ to rise and to form precipitating deep convective clouds with extensive cirrus anvils. They mark the ascending branch of the Hadley cell in the tropics (Figure 1.18a, cloud regime (1), and Figure 1.1). Due to their low cloud top temperature, anvil cirrus emit the least amount of longwave radiation to space and appear white in the satellite image.

The deep convective clouds in the ITCZ can occur in isolation or can form mesoscale convective complexes (MCCs), which are an example of mesoscale convective systems (MCSs) (Section 11.3). The outflow from the deep convective anvils toward the subtropics is marked by thin cirrus clouds. The Hadley circulation extends approximately up to 30 ° latitude in the annual mean. Here, due to the Coriolis force (Section 4.1), the winds at high altitudes no longer have a poleward component but only flow eastward. At these latitudes, the air masses from the Hadley cell converge with those from the mid-latitude Ferrel[15] cell (e.g., Holton and Hakim, 2012), which is the forced (not thermally driven) mid-latitude overturning circulation in the atmosphere. Because of mass continuity, the air needs to sink here, causing the subtropics to be characterized by large-scale subsidence. Adiabatic compression in the subsiding, cloud-free mid-troposphere air warms this air and leads to a marked temperature inversion at the top of the PBL over the cold oceans.

In the shallow PBL beneath the inversion, evaporation from the ocean leads to a high relative humidity, so stratus and stratocumulus clouds with cloud bases of only 400–600 m above the ocean can form (Eastman and Warren, 2013). In Figure 1.18a, they are visible as stratus and stratocumulus toward the northern end of the dashed red line (cloud regime (3)). The Hadley circulation is closed by the trade winds near the surface, which advect the subsided air masses equatorwards. As the air gets closer to the equator, it encounters warmer ocean temperatures; thus, the PBL depth increases and the stratus clouds break up first into stratocumulus and then into shallow cumulus, as shown in Figure 1.18a (cloud regime (2)) and Figure 1.18c. Close to the ITCZ, the rising air becomes so buoyant that cumulus clouds with higher tops form, such as Cu med, Cu con, and Cb.

1.4 Microphysical Cloud Properties

Clouds can also be characterized in terms of their microphysical properties. Examples that can readily be observed and have been reported for different cloud types are the number concentrations, sizes, and types of the hydrometeors within a cloud and their phase (liquid, ice, or a mixture of both). These microphysical properties strongly impact cloud processes and the radiative properties of clouds; thus, their climate impact strongly depends on whether the cloud phase is liquid, ice, or a mixture thereof (Lohmann, 2002b; Matus and L'Ecuyer, 2017; Tan and Storelvmo, 2019).

1.4.1 Cloud Phase

Warm clouds consist purely of liquid water (cloud droplets, drizzle drops, and raindrops). We thus use the terms warm and liquid clouds synonymously. Their temperature can be above or

[15] www.britannica.com/biography/William-Ferrel

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

below 0 °C as long as ice particles are absent. Warm clouds can exist down to temperatures close to −38 °C, the temperature at which small, supercooled water droplets freeze homogeneously and form ice particles (Chapter 9). Cold clouds are defined by having ice particles: they can consist purely of ice hydrometeors (ice clouds), or they can contain a mixture of liquid and ice hydrometeors (mixed-phase clouds). While the mixed-phase is limited to temperatures between −38 and 0 °C, cold clouds are not constrained in temperatures as long as they contain ice crystals, that is, their cloud bases can be above or below 0 °C.

Observations of cloud phase as a function of temperatures can be obtained from in situ observations and remote sensing, as shown in Figure 1.19. Note that satellite-based remote sensing cannot distinguish between liquid and mixed-phase clouds when detecting liquid water at cloud top and thus group all clouds containing liquid water into one liquid cloud category (e.g., Villanueva et al., 2021). On the contrary, ground-based remote sensing observations (Bühl et al., 2013; Radenz et al., 2021) cannot distinguish between mixed-phase and ice clouds when they detect ice at cloud base, so they group all clouds containing ice particles into the mixed-phase cloud category. In situ measurements are most precise in this regard but suffer from limited spatial coverage.

The occurrence of liquid clouds generally decreases with decreasing temperature: Starting with occurrences between 40 and 100% of the time at temperatures above −5 °C, the liquid cloud fraction decreases to below 30 % at temperatures below −15 °C and to much less below −30 °C, according to the in situ data shown in Figure 1.19. We note, however, that the percentage of liquid clouds at temperatures below −15 °C in Villanueva et al. (2021) is higher in the satellite data, because it includes mixed-phase clouds. Overall, the transition temperature where MPCs become more frequent than liquids clouds varies depending on the region and season, and depends on the availability of ice nucleating particles (e.g., Carlsen and David, 2022).

Mixed-phase clouds have been observed over the entire temperature range of −35 to 0 °C. The trend of their occurrence frequency is less clear compared to warm clouds. For instance, a study of orographic clouds in the Swiss Alps found more than 80 % to be mixed-phase, and reported most mixed-phase clouds to occur between −25 and −20 °C (Lohmann et al., 2016). As another example, mixed-phase cloud occurrence exceeded 50 % between −15 and −10 °C in the clouds sampled over Israel but remained around 25 % or less at all temperatures in the frontal clouds sampled over Canada by Korolev et al. (2003) and in the Swiss Alps for air masses coming from the southeast (L16-SE, Lohmann et al., 2016). Note that the mixed-phase occurrences from ground-based remote sensing observations (Radenz et al., 2021) include pure ice clouds and are thus higher below −10 °C.

The frequency of occurrence of ice clouds generally increases with decreasing temperature opposite to liquid clouds. Starting from values between 0 and 30 % at temperatures above −5 °C, the frequency was observed to increase to more than 80 % at temperatures lower than −30 °C in the data sets obtained over Canada and Israel. However, their frequency of occurrence was observed to be around ∼ 10 % when below −20 °C in the northwest wind cases of the orographic clouds (L16-NW) studied in the Swiss Alps (Lohmann et al., 2016). These differences in occurrence can be explained by differences in their formation mechanisms. The sampled orographic clouds in the northwest wind cases encountered higher updraft speeds and

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

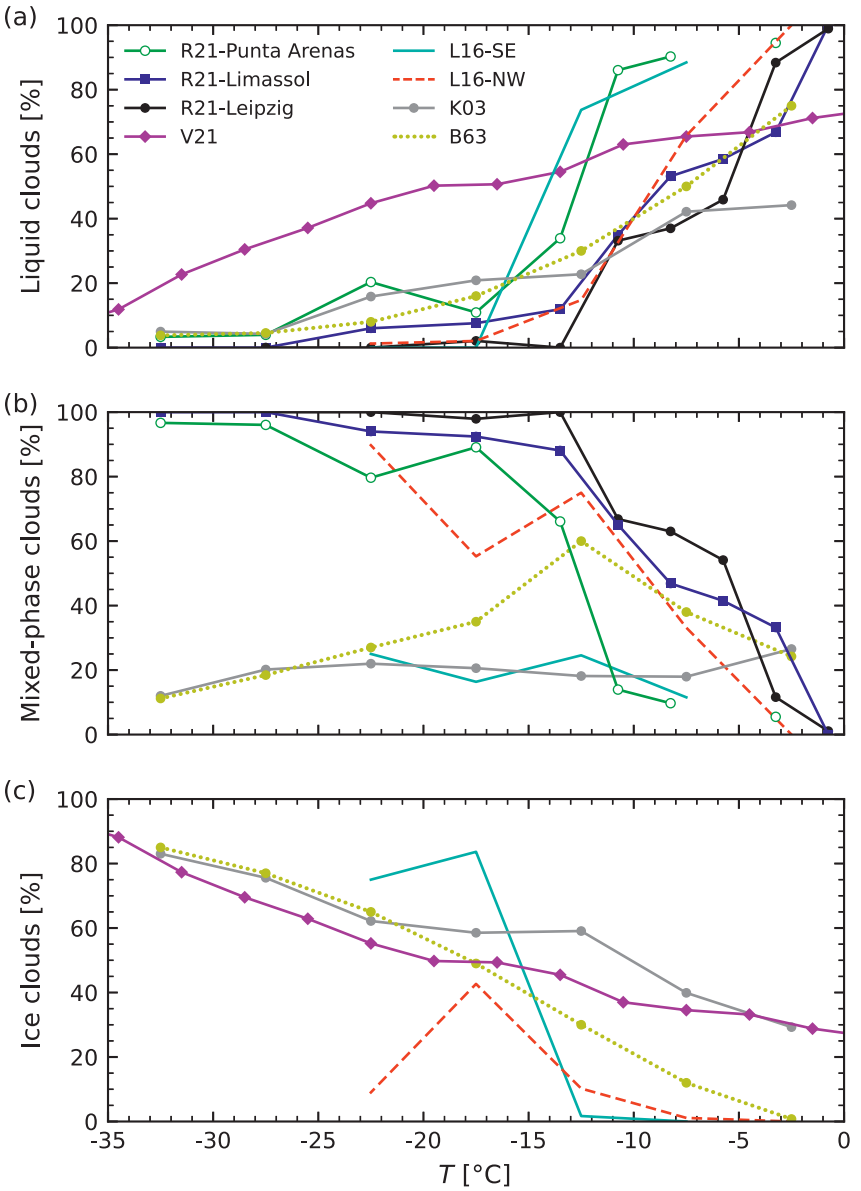


Figure 1.19 Percentage of (a) liquid clouds, (b) mixed-phase, and (c) ice clouds between −35 and 0 °C from different in situ studies: frontal clouds over Canada (K03, updated from Korolev et al. (2003)), various clouds over Israel (B63, Borovikov et al., 1963), and orographic clouds sampled at the Jungfraujoch in the Swiss Alps (L16, Lohmann et al., 2016). No clouds with temperature below −25 °C were sampled in the Swiss Alps. Also shown are data from remote-sensing studies: ground-based remote sensing over Leipzig (Germany), Limmasol (Cyprus), and Punta Arenas (Chile) (R21, Radenz et al., 2021), and obtained from the DARDAR satellite product between 30 and 60 °N/S (V21, Villanueva et al., 2021).

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

hence higher supersaturation than for the southeast wind cases, which permit the simultaneous existence of cloud droplets and ice crystals (Sections 8.1 and 9.3).

Detailed characterization of the cloud phase is important to correctly predict their microphysical and radiative properties. For example, if clouds with temperatures −38 to 0 °C only contain supercooled liquid cloud droplets, they reflect considerably more shortwave radiation back to space, that is, cooling the Earth-atmosphere system, compared to the same cloud consisted purely of ice crystals (Lohmann, 2002b). This mainly results from ice crystals having generally larger sizes than cloud droplets (Table 1.2), causing ice clouds to consist of fewer cloud particles and to be optically thinner compared to liquid clouds with the same water content. In addition, larger ice crystals can sediment more rapidly (Section 9.3.2), which reduce the cloud's water content and lifetime (Chapter 12).

1.4.2 Size Distributions of Hydrometeors

Cloud particles are also called *hydrometeors*, meaning particles comprised of water. The term hydrometeor is typically used more broadly and includes precipitation particles that leave the cloud. Therefore, cloud particles are sometimes also referred to as cloud hydrometeors. Typical sizes and size ranges of the various hydrometeors are summarized in Table 1.2. Cloud droplets are the smallest cloud particles, with a typical radius of 5 μm and a range in radius between 2 and 25 μm. Drops between 25 and 250 μm in radius are called *drizzle drops*. By contrast, raindrops are of millimeter size.

When it comes to the ice phase, some hydrometeors, such as ice crystals and snowflakes, are better described in terms of a maximum dimension because their shapes can deviate significantly from a spheroid (9.2). Thus, ice particles cannot be categorized by radius as neatly as liquid hydrometeors. Ice crystal lengths typically range between 1 μm and 1.5 mm, which overlaps with the maximum dimension of snowflakes, between 0.1 mm and 10 cm. Only rimed hydrometeors can be separated by radius, with graupel, also called *ice pellets* or snow pellets, being smaller than 2.5 mm and hailstones being larger than 2.5 mm (Table 1.2). Hailstones may be spheroidal, conical, or irregular in shape.

The various cloud types contain different hydrometeors. As such, graupel and hail are limited to convective clouds of a considerable vertical extent (Cu con and Cb; Table 1.3 and Section 11.4.1). The liquid hydrometeors (cloud droplets and raindrops) require temperatures above −38 °C and can be observed in all cloud types, except for cirrus, which by definition are composed of ice crystals only. Ice crystals and snowflakes can, in principle, be observed in all cloud types at sub-zero temperatures but the conditions for ice formation must have been met.

The size distributions of hydrometeors can be described by a range of functional forms, including exponential (Marshall and Palmer, 1948), log-normal (Feingold and Levin, 1986), gamma (Borovikov et al., 1963), or generalized gamma distributions (Wu and McFarquhar, 2018). In the following, we use the gamma distribution as a general example to describe cloud particle size distributions, because both the exponential and the log-normal distributions are special cases of the gamma distribution. The number size distribution per unit length (also referred to as number distributions or size distributions for brevity) of cloud droplets and ice

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

crystals can be described in terms of the three-parameter gamma distribution function $n_N(r_h)$, given as

$$n_N(r_h) = N_{0,\gamma} r_h^{\mu} \exp(-\Lambda r_h). \tag{1.1}$$

Here, r_h is the radius of the hydrometeor, and $N_{0,\gamma}$, Λ, and μ are the intercept, slope, and shape parameters, respectively, and $n_N(r_h)$ is the number of hydrometeors per unit volume of air within a size range of unit length around radius r_h. Frozen hydrometeors are described either in terms of their equivalent radius, r_{eq}, that is, the radius of the water sphere that results when the frozen hydrometeor is melted, or in terms of their maximum dimension.

By integrating over the entire hydrometeor size distribution, the number concentration N_h per unit volume and the mass mixing ratio q_h in kg per kg of air (from now on referred to as $\mathrm{kg\,kg^{-1}}$ for simplicity) can be obtained as

$$N_h = \int_0^{\infty} n_N(r_h) dr_h = \int_0^{\infty} N_{0,\gamma} r_h^{\mu} \exp(-\Lambda r_h)\, dr_h, \tag{1.2}$$

$$q_h = \int_0^{\infty} m(r_h) n_N(r_h) dr_h = \int_0^{\infty} m(r_h) N_{0,\gamma} r_h^{\mu} \exp(-\Lambda r_h)\, dr_h, \tag{1.3}$$

where $m(r_h)$ is the hydrometeor mass as a function of the hydrometeor radius r_h. Because r_h enters to the zeroth power in the expression of N_h (Eq. 1.2) and to the third power in the expression of q_h (Eq. 1.3, because $m \propto r_h^3$), N_h and q_h are also referred to as zeroth and third *moments* of the hydrometeor size distribution. Raindrops and snowflakes tend to follow an exponential distribution (Section 10.1), which is a special case of the gamma distribution with shape parameter zero (i.e., $\mu = 0$).

Instead of mass mixing ratios, in situ observations often report water content, for example, the liquid water content (LWC) or ice water content (IWC). These refer to the amount of water or ice per cubic meter of air and are usually given in $\mathrm{g\,m^{-3}}$. LWC and IWC are defined as follows:

$$LWC = \frac{4}{3}\pi N_c \bar{r}_d^3 \rho_l, \tag{1.4}$$

$$IWC = \frac{4}{3}\pi N_i \bar{r}_i^3 \rho_i, \tag{1.5}$$

where N_c and N_i are the cloud droplet and ice crystal number concentration, $\bar{r}_d$ and $\bar{r}_i$ denote the volume mean cloud droplet and ice crystal radii, and ρ_l and ρ_i are the liquid water and ice densities. Integrating LWC and IWC over the depth of cloud, yields the liquid water path (LWP) and ice water path (IWP), respectively.

LWP and IWP obtained from satellite data (Li et al., 2012; Elsaesser et al., 2016, 2017) are shown in Figure 1.20. The geographical distribution of LWP mimics the geographical distribution of low and mid-level clouds in Figure 1.14 with maxima in the ITCZ, SPCZ, and in the extratropical storm tracks, as these clouds predominantly consist of liquid water. The geographical distribution of IWP, on the other hand, mimics the geographical distribution of high clouds in Figure 1.14 with maxima in regions of persistent deep convective clouds, which contain a lot of ice. These regions include the West Pacific Warm Pool, the Amazon, central Africa, the ITCZ, the SPCZ, and the extratropical storm tracks.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

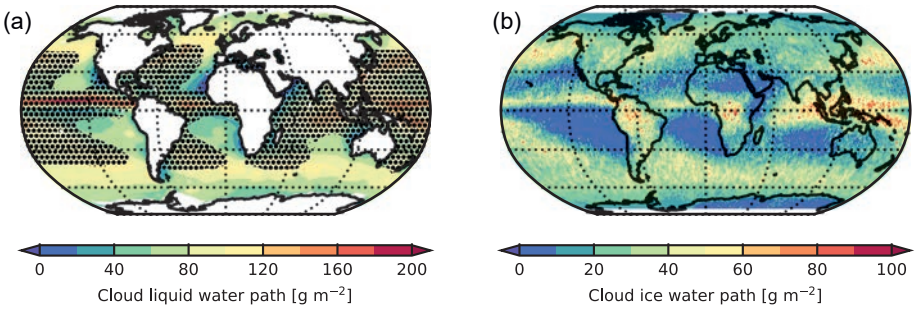


Figure 1.20 (a) Annual mean vertically integrated cloud liquid water content (liquid water path, LWP) in (g m^{-2}) over the oceans from the Multi-Sensor Advanced Climatology averaged over the years 1988–2016 (Elsaesser et al., 2016, 2017). Hatching denotes areas in which the LWP could be influenced by rain, that is, where rain contributes more than 20 % to the LWP. (b) Annual mean vertically integrated ice water content (ice water path, IWP) in (g m^{-2}) from CloudSat and CALIPSO satellite observations averaged over the years 2007–2010 compiled by Li et al. (2012). Figure adapted from Neubauer et al. (2019).

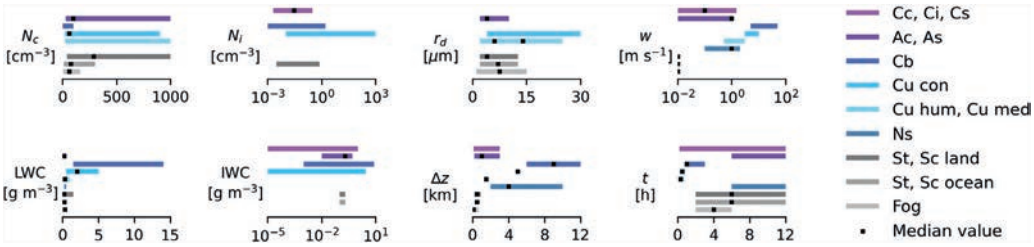


Figure 1.21 Visualization of typical cloud properties listed in Table 1.3. Figure courtesy of Nadja Omanovic.

The formation of precipitation depends on the water content of a cloud as well as on the cloud droplet or ice crystal number concentration, which determines the average sizes of the cloud droplets and ice crystals. In addition to being important for the formation of precipitation-sized particles, the sizes of hydrometeors also influence the optical properties of the cloud, for instance, the fraction of the sunlight that is reflected back to space (Chapter 12).

Typical values and ranges of these microphysical properties for the cloud types introduced above are summarized in Table 1.3 and visually summarized in Figure 1.21. Generally, cloud droplets are smaller and more numerous in clouds over land than over oceans (e.g., Han et al., 1998). An example is shown in Figure 1.22, contrasting N_c in marine and continental cumuli (Squires, 1958). The value of N_c is determined by the number concentration of aerosol particles that serve as *cloud condensation nuclei* (CCN) (Section 7.6). Generally, CCN concentrations are higher over land due to more aerosol sources than over oceans. In addition, N_c depends on how much the relative humidity with respect to liquid water exceeds 100 %, the so-called *supersaturation with respect to liquid water*, or supersaturation in short.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

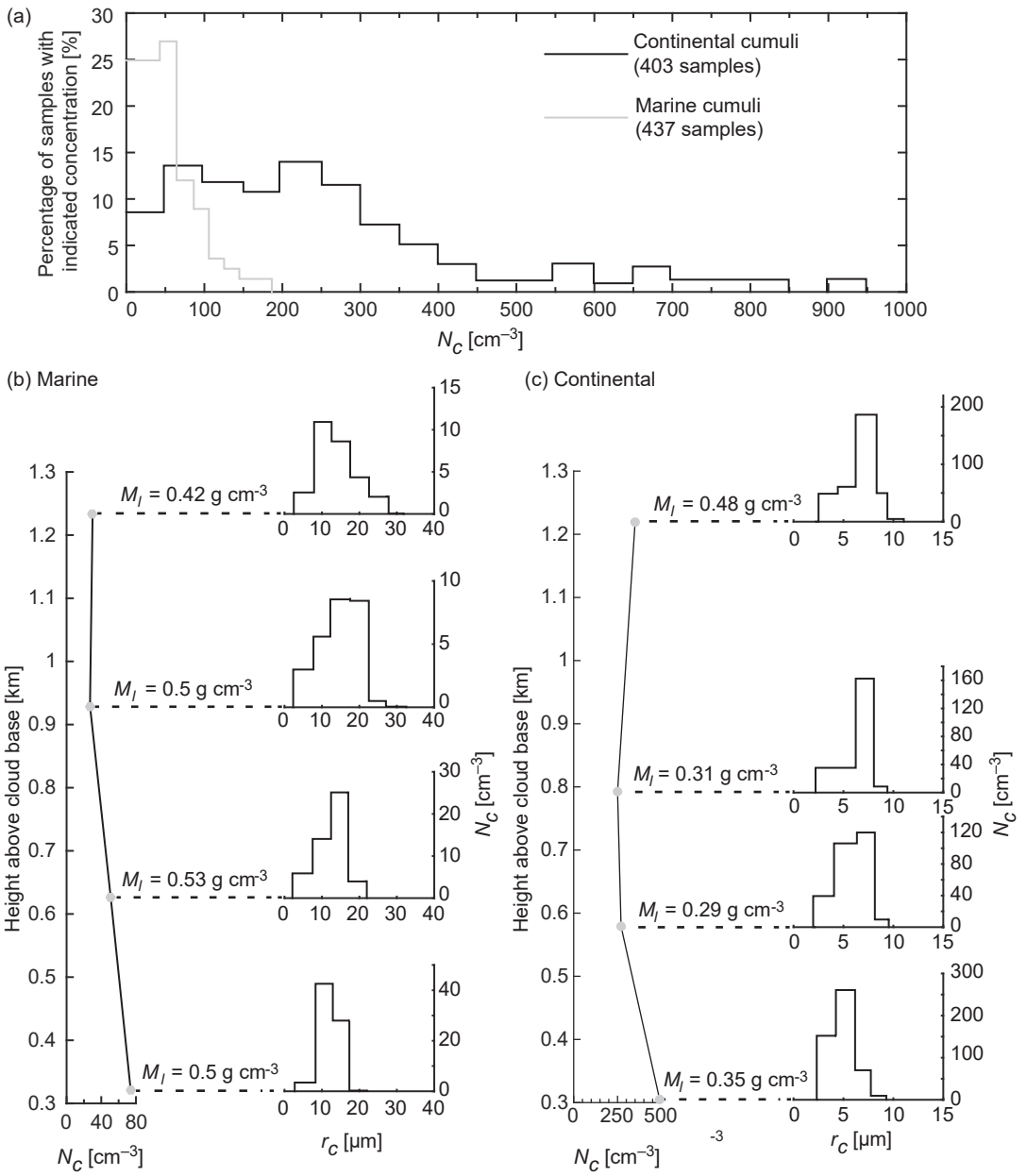


Figure 1.22 Observed cloud droplet number concentrations in marine and continental cumuli. (a) Percentages of marine and continental cumuli with indicated droplet concentrations N_c. Change in the droplet concentration and size distribution with height above cloud base in (b) marine and (c) continental cumuli; M_l denotes the liquid water content. Note the different x-axis scales in (b) and (c). Figure adapted from Squires (1958).

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

Supersaturation is produced by adiabatic expansion and cooling in rising air parcels and thus is determined by the updraft speed (Chapter 8). Updraft speeds are generally higher over land than over ocean (Table 1.3) because the Sun heats up the land faster than the ocean because of its smaller heat capacity. This heating produces buoyancy (Chapter 4) and rising air. With higher updraft speeds over land, more and smaller aerosol particles are activated (Chapter 7). As a result, N_c rarely exceeds 200 cm^{-3} in marine cumuli, while N_c reaches up to 950 cm^{-3} in continental cumuli (Figure 1.22a). Due to the lower N_c values in marine clouds, cloud droplets experience less competition for the available water vapor. Droplets can thus grow to larger sizes more quickly in marine clouds than in continental clouds, as indicated in the evolution of the cloud droplet size distribution with increasing altitude (Figure 1.22b,c). In the marine cumuli shown in Figure 1.22b, cloud droplets grow up to 30 μm in radius at altitudes of 1.2 km above cloud base, whereas they remain at or below 10 μm in radius in the continental cumuli at all altitudes, shown in Figure 1.22c.

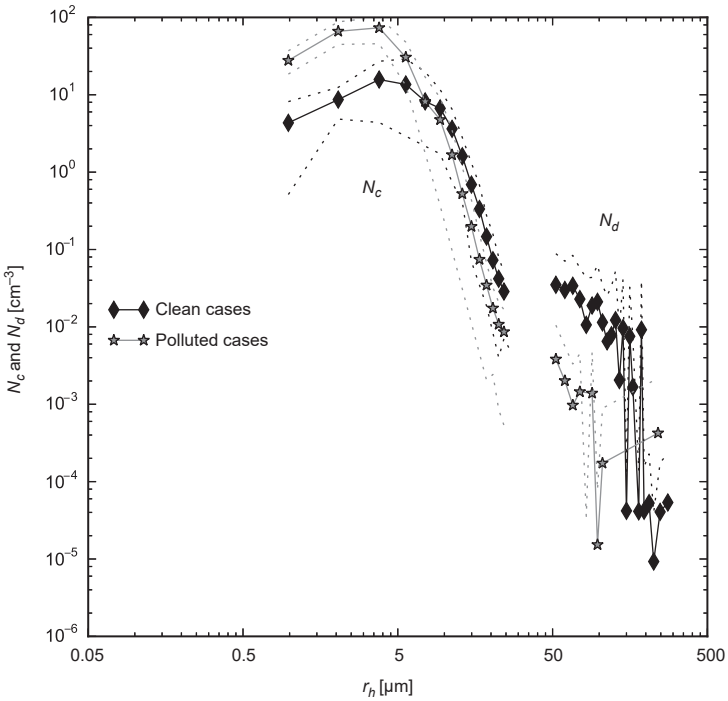


Figure 1.23 Cloud droplet N_c and drizzle drop N_d size distributions for clean clouds (diamonds) and polluted clouds (asterisks) as a function of the hydrometeor radius r_h. Each point refers to the averaged concentration over all clean or all polluted clouds. The solid and dotted lines give the averages and the standard deviations for the clean and polluted cases, respectively. The values of N_d minus one standard deviation are not shown because they extend below the lowest values shown. Figure adapted from Peng et al. (2002).

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

Supersaturations observed in fog are normally less than those observed in clouds, especially in convective clouds. While relative humidity in convective clouds and clean stratus clouds can exceed 101 % (Hudson et al., 2010, 2015), it typically remains below 100.1 % in fog (Duplessis et al., 2021).

Similar differences, in terms of size distribution and number concentration, to those between marine and continental clouds are found between polluted and clean clouds. An example of an observed size distribution of cloud droplets (N_c) and drizzle drops (N_d) in marine clouds off the coast of Nova Scotia, Canada, is shown in Figure 1.23 (Peng et al., 2002). These data are sorted according to their aerosol concentration (N_a), measured with an instrument which sampled aerosol radii larger than 0.07 μm. In this study, clouds with $N_a < 300\,\text{cm}^{-3}$ were considered clean and those with $N_a > 300\,\text{cm}^{-3}$ were considered polluted. The polluted clouds have average concentrations $N_a = 810\,\text{cm}^{-3}$, $N_c = 260\,\text{cm}^{-3}$, and $N_d = 8 \times 10^{-4}\,\text{cm}^{-3}$ and the clean clouds have average concentrations $N_a = 160\,\text{cm}^{-3}$, $N_c = 70\,\text{cm}^{-3}$, and $N_d = 9.4 \times 10^{-3}\,\text{cm}^{-3}$.

Figure 1.23 shows that the concentration of small cloud droplets (with $r_c < 5$ μm) is larger in the polluted clouds than in the clean clouds. At a radius of about 7.5 μm, the curves intersect, and the concentration of larger drops becomes higher in the clean clouds. The gap in the data between radii 25 μm and 55 μm is caused by the use of different instruments to obtain the size distributions and their respective measurable size ranges.

The differences between marine versus continental or clean versus polluted clouds show that typical cloud droplet size distributions for a given cloud type do not exist. In addition to differences in aerosol concentration and vertical velocity at the cloud base, clouds of the same cloud type can have different cloud base temperatures, which determines the growth rate of the cloud droplets (Chapter 8). They can also experience different levels of mixing with the environmental air, so called *entrainment* (Chapter 5). Lastly, they could have been sampled at different stages of their life cycle and at different heights above cloud base. Therefore, large variations exist between the individual size distributions of the various cloud types.

A few general differences between certain cloud types can be stated, with caution, keeping the above points in mind. As shown in Figure 1.22, cloud droplets are larger higher up in a cloud. Thus, one can expect that as a fair weather cumulus continues to develop into a Cu con, the cloud droplets become larger and fewer in number, mainly due to growth by collision-coalescence (Section 8.2). Furthermore, a comparison of St and Sc with Cu hum and Cu med shows that cumuli have higher updraft speeds than stratus clouds, causing N_c to typically be higher in cumulus than in St and Sc. Clouds which can produce precipitation (mainly Ns, Cb, and Cu con) typically have broader drop size distributions, with drop radii exceeding 15 μm (Chapter 8), and/or they contain an ice phase (Chapter 9).

Aerosol number concentrations generally decrease with altitude, as the distance to the emission sources at the Earth's surface increases. As a result, the CCN number concentration also decreases with altitude, which results in a lower N_c in mid-level clouds (Ac and As) than in the respective low clouds (Sc and St). The water vapor mixing ratio also decreases with altitude in the troposphere due to the decrease in temperature with increasing altitude and the exponential relationship between air temperature and the amount of water it can contain (Chapter 3). If less water vapor is available, less can condense or deposit and form cloud droplets or ice crystals. Therefore, the liquid and ice water contents also decrease with increasing height

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

(decreasing temperature) from the cloud base. The exponential decrease of water vapor with increasing altitude generally exceeds the decrease in CCN concentration, because of the longer residence times and additional sources of CCN in the free troposphere, as will be discussed in Chapter 6. Therefore, mid-level clouds typically have smaller cloud droplets than low clouds.

Cirrus clouds are characterized by the lowest number concentrations and the largest average ice crystal size. However, there is a large spread in the microphysical properties of cirrus clouds (Tables 1.2 and 1.3). First, the ice water content in cirrus is a strong function of temperature and decreases drastically as the environmental water vapor concentration decreases at colder temperatures (Heymsfield et al., 2017; Krämer et al., 2020). Second, there are different mechanisms by which cirrus clouds form and which determine the ice crystal number concentration. The variety of ice crystal shapes, number concentrations, and sizes (Chapter 9) additionally complicates the characterization of ice clouds.

1.5 Temporal and Spatial Scales of Cloud Systems and Atmospheric Phenomena

Atmospheric processes and phenomena including clouds occur on very different spatial and temporal scales. Three spatial scales are distinguished, as shown in Figure 1.24: the local or

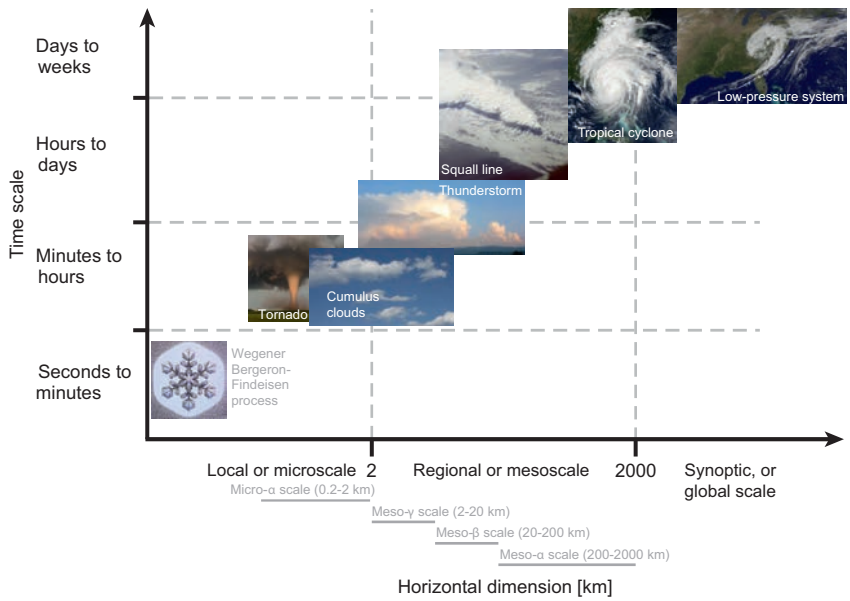


Figure 1.24 Examples of atmospheric processes or phenomena that occur at different temporal and spatial scales. The scale separations were defined by Orlanski (1975). Image sources from left to right: Kenneth Libbrecht, NOAA image gallery, Fabian Mahrt (two images), NASA (two images), and NOAA image gallery.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press

microscale, with horizontal dimensions less than 2 km; the regional or mesoscale, between 2 and 2000 km; and the synoptic, global, or macroscale, with weather systems larger than 2000 km (Orlanski, 1975). All of these scales can be further subdivided (Orlanski, 1975; Hobbs, 1981).

Examples for the micro-α scale (0.2–2 km) are thermals (Section 5.2), tornados (Section 11.2.5), or Cu hum (Section 1.2.2). The meso-γ scale (2–20 km) includes ordinary-cell convective storms (Section 11.1) and thunderstorms. The meso-β scale (20–200 km) encompasses land-sea breezes and most squall lines (Section 11.3.1). Typical meso-α scale (200–2000 km) phenomena are fronts and MCSs, including large quasi-linear convective systems (QLCSs) (Section 11.3). Tropical cyclones (Section 11.5) also mainly belong to the meso-α scale, but they can extend into the lower end of the synoptic scale, which also includes low and high pressure systems.

Next to the spatial scales, the temporal scales of these phenomena differ and can range from seconds to weeks. Generally, the duration increases for phenomena on the microscale to the synoptic scale.

https://doi.org/10.1017/9781009485654.002 Published online by Cambridge University Press