Chapter 7: Clouds and Aerosols

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Executive Summary

- Clouds cool the Earth on average, by about 17 W m^{-2}. This is the net result of a greenhouse (infrared) warming due mainly to high clouds (~30 W m^{-2}) and a cooling effect from reflecting solar radiation contributed by all cloud types (~47 W m^{-2}).

- New satellite observations and advances to models have given us global simulations that can explicitly resolve some types of clouds. Comparison to detailed observations with such models has led to improved understanding of cloud interactions with the meteorology and the climate. However observations alone do not provide a robust constraint on the sign and magnitude of cloud feedbacks.

- Evidence for a net positive feedback from water vapour and lapse rate changes has increased robustness. The net long-term feedback parameter is very likely positive with an interquartile range of 0.90 to 1.06 W m^{-2} K^{-1} from CMIP3 model [PLACEHOLDER FOR SECOND ORDER DRAFT: CMIP5 models].

- Cloud feedbacks on long-term greenhouse-gas induced surface temperature change are likely positive. Robust positive feedback mechanisms have been established while no mechanism for strong negative global cloud feedback has convincing observational or model-based support. The robust mechanisms include a rise in the heights of cirrus cloud tops and a reduction in subtropical cloudiness. The range of the cloud feedback parameter in the CMIP5 models is [xx to yy] W m^{-2} K^{-1}. Inconsistent prediction of changes to low cloud remains the largest source of spread in the cloud feedback parameter and equilibrium climate sensitivity among climate models. Since all cloud types are crudely represented by climate models, values outside the current spread of climate models cannot be ruled out.

- Observations, theoretical considerations and models indicate that the strength of extreme precipitation events, which can cause flooding, tend to strongly increase as the climate warms and atmospheric humidity increases. This tendency is expected to vary significantly by region.

- Observational studies since AR4 suggest that contrails and contrail-cirrus from aircraft, at current levels of coverage, exert only a small adjusted forcing (AF) of 0.03 W m^{-2} (with a 5%–95% uncertainty range of 0.01 to 0.06 W m^{-2}), and are unlikely to have an observable effect on surface temperature and diurnal temperature range.

- There has been continuous progress since AR4 on observing and modelling climate-relevant aerosols properties (including their size distribution, hygroscopicity, chemical composition, mixing state, optical and cloud nucleation properties) and their atmospheric distribution. The representation of aerosol processes in the CMIP5 models remain more simplistic than in some of the more detailed aerosol models used to assess radiative forcing (RF).

- The aerosol direct effect was assessed for the year 2010 relative to 1750 with all ranges indicative of 5%-95% confidence intervals. The RF for the total direct aerosol effect is –0.3 ± 0.3 W m^{-2} (not accounting for possible anthropogenic changes in mineral dust), using evidence from aerosol models and some constraints from observations. The semi-direct effect of atmospheric heating on clouds can be significant and of either sign regionally but is thought to be small globally (~0.1 W m^{-2}) and not significantly different than zero. The direct and semi-direct effects from anthropogenic aerosols are assessed together as an AF of –0.3 ± 0.4 W m^{-2}.

- Sulphate aerosol is responsible for a RF of –0.3 W m^{-2} (–0.2 to –0.6 W m^{-2}). Black carbon (BC) aerosol has a RF +0.2 ± 0.2 W m^{-2} (fossil fuel sources only) and +0.4 ± 0.2 W m^{-2} (fossil fuel and biomass burning including a possible small fraction from vegetation feedbacks). The largest uncertainties relate to the vertical profile of BC. Organic carbon aerosol from fossil fuel sources has a RF –0.05 ± 0.05 W m^{-2}. Biomass burning aerosol has a RF –0.01 W m^{-2} (–0.15 to +0.1 W m^{-2}). Secondary organic aerosol has a RF –0.04 W m^{-2} (–0.03 to –0.07 W m^{-2}) and nitrate aerosol has a RF of –0.1 ± 0.08 W m^{-2}. Finally mineral aerosol has a RF of –0.1 ± 0.2 W m^{-2} but this may include part of a feedback.

- Anthropogenic absorbing aerosols (BC and brown carbon) on snow and ice are responsible for a positive RF of +0.04 W m^{-2}, with a 0.01–0.10 W m^{-2} (5%-95%) uncertainty range. This radiative forcing is 2-4

time more effective at causing global mean temperature changes than an equivalent radiative forcing from CO_2.

- A number of climate feedbacks have been identified which involve aerosols, either through a change in the source strength of natural aerosols or a change in sink processes. There is low agreement in model simulations and no solid evidence to suggest that such feedbacks could be significant during the 21st century although they may be important at the regional scale.

- There has been continuous progress in our understanding of aerosol-cloud interactions in liquid clouds, in particular from observations and large-eddy simulating models, which reveal some compensating effects. Progress has been made on our understanding of aerosol interactions with mixed phase and ice clouds and their representation in climate models. This progress has led to a reduction in the estimate for the magnitude of global aerosol indirect forcings.

- The indirect radiative forcing (iRF) is very likely between –1 and –0.1 W m^{-2}, the lower bound being based on estimates from climate models, and likely between –0.4 and –0.2 W m^{-2}, the lower bound being based on studies that take satellite data into account. Following the same line of argumentation, the indirect adjusted forcing (iAF) is very likely between –1.5 and 0 W m^{-2} and likely between –0.7 and –0.2 W m^{-2}.

- There is limited, if any, evidence and no agreement that the small-scale impact of aerosols on cloud microphysical structure translates into a significant regional impact in terms of precipitation amount (beyond orographic locations) but there is medium evidence and agreement for an effect on timing and intensity of precipitation.

- New studies provide robust evidence that cosmic rays influence new particle formation through changes in atmospheric ionization rate. However there is medium evidence and high agreement that any effects from variations in cosmic rays on CCN and cloud properties are insignificant climatically.

- From a physical-science assessment basis, model studies, observations of the effects of volcanic eruptions, and physical arguments suggest that some Solar Radiation Management (SRM) strategies for geoengineering may be effective in offsetting the global average surface temperature increase. However SRM would produce an inexact compensation for the RF by greenhouse gases and there would be residual regional differences in temperature and rainfall patterns. SRM will not compensate for ocean acidification from increasing CO_2 and may have other impacts on the climate system (e.g., stratospheric ozone depletion from stratospheric aerosol injection). Termination of SRM would produce a reappearance of most of the avoided global warming within about a decade.

- Evidence from past volcanic eruptions and modelling studies suggest that increasing the amount of aerosols in the stratosphere can increase the Earth's albedo enough to counteract the global RF of at least up to a doubling of CO_2 (within the limitations stated above). The effectiveness and potential of SRM through cloud brightening is more uncertain than through stratospheric aerosol injection because of our limited understanding of aerosol indirect effects on clouds.

7.1 Introduction

The atmosphere, although mostly composed of gases, is full of particles. It is usual to partition these particles into cloud particles, atmospheric aerosols, and falling hydrometeors according to their size, water content and sedimentation velocity.

Clouds usually form in rising air, which expands and cools until cloud formation occurs through nucleation or freezing of aerosol particles. Cloud particles are generally larger than aerosols and mostly composed of water; they are suspended in the atmosphere and collectively form a cloud which is usually a visible body. The evolution of a cloud is governed by the balance between a number of dynamical, radiative and microphysical processes. Cloud particles of sufficient size become falling hydrometeors, which are categorised as drizzle, raindrops, snow crystals and graupel. Clouds affect the climate system by regulating the flow of radiation at the top of the atmosphere, by controlling precipitation, and through additional mechanisms too numerous to list here. Precipitation processes merit special attention not only because the hydrological cycle is important in its own right, and intricately linked to the structure of regional circulation systems, but also because the frequency and distribution of precipitation is an important sink of aerosol particles.

Atmospheric aerosols are relatively small solid and liquid particles in suspension in the air that can be of natural or anthropogenic origin. They interact with solar radiation, through absorption and scattering, and to a lesser extent with terrestrial radiation, through absorption, scattering and emission. Aerosols can serve as cloud condensation nuclei and ice nuclei upon which cloud droplets and ice crystals form. They also play a wider role in biogeochemical cycles in the Earth system, for instance by carrying nutrients to ocean ecosystems.

Cloud and aerosol amounts and properties are extremely variable in space and time. The short lifetime of clouds in the atmosphere often creates relatively sharp cloud edges and rapid horizontal variations in cloud properties, which is much less typical of aerosol layers. While the dichotomy between aerosols and clouds is generally appropriate and useful, it should be appreciated that there can be a continuum in particle size and a continuum between clear and cloudy sky which makes the distinction between aerosols and clouds, or clouds and rain more difficult and sometimes less relevant (Charlson et al., 2007; Koren et al., 2007).

Both clouds and aerosols are a major source of uncertainties in the climate system. Clouds respond to climate forcing mechanisms in multiple ways and individual cloud feedbacks can be positive or negative. The representation of cloud processes in climate models has been recognised for decades as a continuing source of much of the uncertainty surrounding climate change (e.g., Arakawa 1975, 2004; Bony et al., 2006; Cess et al., 1989; Charney, 1979; Randall, 1989), but with the exception of the Charney report (Charney, 1979) clouds have not been a focal point of past assessment reports. Key issues include the representation of both deep and shallow cumulus convection, microphysical processes in ice clouds, and partial cloudiness that results from small-scale variations of cloud-producing and cloud-destroying processes. Inter-model differences in cloud feedbacks constitute by far the primary source of spread of both equilibrium and transient climate responses simulated by the CMIP3 climate models (Dufresne and Bony, 2008) despite the fact that, as discussed by Randall et al. (2007) and also later in this chapter, most models agree on a near-neutral or positive cloud feedback.

Anthropogenic aerosols are responsible for a radiative forcing of climate through their direct effect (the interaction of aerosols with radiation) and their indirect effects (the interaction of aerosols with clouds). Quantification of the direct and indirect effects of anthropogenic aerosols has proven difficult, and is fraught with uncertainties (Haywood and Boucher, 2000; Lohmann and Feichter, 2005). While previous attempts to quantify the probability distribution function for the net anthropogenic radiative forcing from bottom up approaches have found that it is very likely or virtually certain to be positive, the possibility of negative values could not completely be ruled out (Forster et al., 2007; Haywood and Schulz, 2007). Our inability to better quantify non-greenhouse gas radiative forcings, and primarily that associated with atmospheric aerosols, is partly responsible for the uncertainty in observationally-constrained climate sensitivity (Andreae et al., 2005). It was also found that the total anthropogenic forcing is inversely correlated to climate sensitivity in climate models used for CMIP3 (Kiehl, 2007). This is important because, for a given climate model, the uncertainty in aerosol radiative forcing has proportionally more impact on the simulation of the

20th century's climate than of future climate projections (Dufresne et al., 2005) because scenarios of anthropogenic aerosol emissions tend to stabilise or decrease in the future (Lamarque et al., 2010).

Research into cloud-climate interactions has also progressed significantly since the Fourth Assessment report (AR4, IPCC 2007). Examples include additional field programmes (e.g., the Tropical Warm Pool International Cloud Experiment, TWPICE, as described by May et al. (2008)), greatly improved satellite data (e.g., Stephens and Kummerow, 2007), improved cloud parameterisations (e.g., Park and Bretherton, 2009), studies with high-resolution global models (e.g., Sato et al., 2009), and very-high-resolution regional models (e.g., Khairoutdinov et al., 2009).

The Third Assessment Report (TAR, IPCC 2001) dedicated a chapter to atmospheric aerosols (Penner et al., 2001), while their radiative forcing was assessed in a separate chapter (Ramaswamy et al., 2001). The AR4 updated our understanding of the aerosol radiative forcing (Forster et al., 2007), but did not assess all aspects of aerosol influences on climate. Aerosol indirect effects on climate were discussed in Denman et al. (2007). Our capability to observe aerosols has increased substantially since the TAR and the AR4, and so has our understanding of aerosol processes and our modelling capability, from the fine to the large scale. Many climate models now include more complex parameterisations of aerosol and cloud microphysics, incorporating understanding from process-based models. Models are being more thoroughly evaluated against in-situ (e.g., Koch et al., 2009b) and remote-sensing observations (e.g., Huneeus et al., 2011b), and that evaluation feeds back into model development. However an accurate treatment of aerosol processes and their impact on climate at the global scale remains a challenge.

For the first time in the IPCC WGI assessment reports, clouds and aerosols are discussed together in a single chapter. This offers the opportunity to assess in a unified framework not only cloud feedbacks and aerosol forcings, but also the multiple interactions among aerosols, clouds and precipitation and their relevance for climate and climate change.

Figure 7.1 provides an overview of the chapter's approach to clouds and aerosols in the context of climate change. *Forcings* associated with forcing agents such as greenhouse gases and aerosols act on global mean surface temperature through the global radiation budget. *Rapid forcing adjustments* (or rapid responses) arise when forcing agents, by altering flows of energy internal to the system, affect cloud cover (or other components of the climate system) and thereby alter the global budget indirectly. Because these adjustments do not operate through changes to global mean surface temperature, which are slowed by the massive heat capacity of the oceans, they are generally rapid and most are thought to occur within a week. *Feedbacks* amplify or damp changes to the global mean surface temperature via its impact on atmospheric variables that affect the global budget. Such a framework offers a clear distinction between the traditional concept of radiative forcing (RF, defined as the instantaneous radiative forcing with stratospheric adjustment only) and the new concept of adjusted forcing (AF, which includes other atmospheric and surface rapid adjustments) as introduced in Chapter 1 and detailed in Section 8.1.1. The aerosol semi-direct and indirect effects, which have been recognised as important for some time, will be quantified here through the concept of adjusted forcing.

[INSERT FIGURE 7.1 HERE]

Figure 7.1: Overview of feedback and forcing pathways involving clouds and aerosols. Forcings are represented by dark arrows; forcing agents are boxes with grey shadows, rapid forcing adjustments (or rapid response) are red arrows and feedbacks are other-colored arrows. See text for further discussion.

The Chapter aims to discuss clouds and aerosols in a comprehensive but climate-focused way. Section 7.2 describes our understanding of the role of clouds in climate change. Section 7.3 discusses aerosol properties and how these link up with best estimates of the aerosol direct radiative forcing and their uncertainties. Section 7.4 covers aerosol-cloud interactions and derives estimates for the aerosol indirect forcing. Finally Section 7.5 assesses solar radiation management techniques aimed at cooling the planet as a number of these techniques rely on the modification of aerosols and clouds. Our radiative forcing estimates for the aerosol effects feed into Chapter 8.

7.2 Clouds

7.2.1 Clouds in the Present-Day Climate System

7.2.1.1 Cloud Formation, Cloud Types, and Cloud Climatology

To form a cloud, air must cool or moisten until it is sufficiently supersaturated with respect to water vapor to nucleate some of the available condensation or freezing nuclei. Clouds form in diverse ways, including large-scale or orographically-driven ascent, small-scale turbulent buoyant ascent, radiative or evaporative cooling, or turbulent mixing of a moist layer. These formation mechanisms support many important cloud types, e.g., cirrus, stratus, or cumulus, organized by large-scale circulations into different climatological cloud regimes (such as subtropical marine stratocumulus) or transient cloud systems (such as tropical or midlatitude cyclones). Each cloud regime or system has a characteristic assemblage of cloud types that contribute to its radiative properties and precipitation characteristics. Such assemblages can be objectively isolated from satellite data (Jakob and Schumacher, 2008; Jakob et al., 2005).

Figure 7.2 shows a selection of widely occurring tropical and extratropical cloud regimes, and how they might look on at visible wavelengths on a typical geostationary satellite image.

[INSERT FIGURE 7.2 HERE]
Figure 7.2: Diverse cloud regimes reflect diverse meteorology. (a) A visible-wavelength geostationary satellite image shows (from top to bottom) expanses and long arcs of cloud associated with extratropical cyclones, subtropical coastal stratocumulus near Baja California breaking up into shallow cumulus clouds in the central Pacific, and mesoscale convective systems outlining the Pacific ITCZ. (b) A schematic vertical section through a typical warm front of an extratropical cyclone shows multiple layers of upper-tropospheric ice (cirrus) and mid-tropospheric water (altostratus) cloud upwind of the frontal zone, an extensive region of nimbostratus associated with frontal uplift and turbulence-driven boundary layer cloud in the warm sector. (c) A schematic cross section along the low-level trade wind flow from a subtropical west coast of a continent to the ITCZ shows typical low-latitude cloud types, shallow stratocumulus in the cool waters of the oceanic upwelling zone near the coast, trapped under a strong subsidence inversion, shallow cumulus of warmer waters further offshore and a transition into precipitating cumulonimbus cloud systems with extensive cirrus anvils associated with rising air motions in the ITCZ.

Figure 7.3a shows a corresponding geographical annual-mean plot of the total fraction of atmospheric columns that contain cloud, thresholded to remove very thin cloud of little radiative significance. Figure 7.3b shows a latitude-height section of annual- zonal-mean cloud fractional occurrence from CloudSat and Calipso, a combination of active sensors that can detect even thin clouds (optical depth less than rougly 3–5) and see deep into thick cloud layers. Clouds appearing above roughly the 400 hPa level (which are nearly all ice) are typically considered “high clouds”, while those appearing below roughly 700 hPa (which are mostly liquid but often contain ice outside the tropics) are considered “low” (Zelinka et al., 2011a).

[INSERT FIGURE 7.3 HERE]
Figure 7.3: Annual-mean cloud fractional occurrence based on four years of satellite observations (June 2006-February 2011) from CloudSat and Calipso (Kay et al., 2011; COSP simulator). (a) Geographical mean, with thin cloud (SR < 5) removed; (b) latitude-height section of zonal mean cloud cover. [PLACEHOLDER FOR SECOND ORDER DRAFT: further graphical refinement.]

[INSERT FIGURE 7.4 HERE]
Figure 7.4: Distribution of annual-mean SWCRE, LWCRE, net CRE (from CERES-EBAF) and precipitation (from CMAP).

7.2.1.2 Effects of Clouds on Earth’s Radiation Budget

The effect of clouds on Earth’s radiation budget can be inferred by comparing satellite estimates of upwelling solar and longwave radiation with the corresponding fluxes in the absence of clouds, obtained by screening cloudy pixels or using clear-sky radiative transfer models. Using this method, Loeb et al., (2009) estimate that in annual global mean, clouds enhance the planetary albedo, changing the net shortwave radiative flux into the top of the atmosphere (TOA) by –47 W m^{-2} compared to a cloud-free atmosphere (the global shortwave cloud radiative effect or SWCRE), and that clouds enhance the planetary greenhouse effect, reducing net longwave radiative energy loss by 30 W m^{-2} (the global longwave cloud radiative effect or LWCRE). Other published estimates differ from these by 10% or less (Loeb et al., 2009). The combined annual global mean net CRE of –17 W m^{-2} can be regarded as a cooling effect of clouds on climate. Both

global mean SWCRE and LWCRE are large compared to the 4 W m^{-2} radiative forcing of doubling CO_2. Hence, even modest changes in the global distribution of clouds could produce substantial radiative feedback on climate change.

The regional patterns of annual-mean SWCRE and LWCRE, shown in Figure 7.4a-b, reflect typical cloud regimes averaged across the seasonal cycle. High clouds, which are cold compared to the underlying surface, dominate patterns of LWCRE, while the SWCRE is sensitive to optically thick clouds at all altitudes. Figure 7.2 shows that regions of deep, thick cloud with large LWCRE and large negative SWCRE tend to accompany precipitation, emphasizing their intimate connection with the hydrological cycle. The net CRE is negative over most of the globe and most negative in regions of very extensive low-lying reflective cloud such as the midlatitude and eastern subtropical oceans, where SWCRE is strong but LWCRE is weak.

7.2.1.3 Coupling Between Clouds, Precipitation, Large-Scale Dynamics and Stratification

Precipitation is almost exclusively initiated in updrafts inside clouds. Thus, it is inextricably linked to clouds and their dynamical drivers.

7.2.1.3.1 Deep precipitating cloud systems

Over the tropical oceans, where most rain falls from cumulus cloud systems, we observe particularly tight statistical correlations over daily or longer timescales and space scales of a few hundred kilometres or more. More rainfall associates with more deep convective cloud cover (Wyant et al., 2006), higher relative humidity throughout the depth of the free troposphere (Holloway and Neelin, 2009) and more mean mid-tropospheric upward motion (Bony et al., 2004; Lintner et al., 2011). Similar relationships have been documented in the midlatitude storm tracks (Norris and Iacobellis, 2005). These relationships may change somewhat if climate changes; for instance, they could be affected by temperature changes that affect saturation water vapour mixing ratio, as well as the depth of the tropopause and the freezing level. However, we do expect a close correspondence between regional changes in mean vertical motion, precipitation and cloud cover that accompany a climate change, a theme explored further in the discussion of cloud feedbacks in Section 7.2.4.

7.2.1.3.2 Large-scale controls on boundary-layer cloudiness

While cumulus clouds develop in conditionally unstable layers of the atmosphere, marine stratocumulus clouds persist where there is a strong capping inversion that traps moisture in the atmospheric boundary layer. Klein and Hartmann (1993) showed that on seasonal and longer timescales, increased subtropical marine stratocumulus cloud cover is strongly correlated with greater lower tropospheric stability (LTS) between the surface and 700 hPa. Variants of LTS such as estimated inversion strength (EIS) (Wood; Bretherton 2006) or others that also account for humidity variations (Williams et al., 2006; Zhang et al., 2010a) better predict low cloud cover over the mid-latitude oceans and may be more applicable to changed climates (Wood and Bretherton, 2006).

7.2.1.3.3 Mixed-phase arctic clouds

Arctic clouds have become a focus of recent interest because they may affect the sensitivity of the Arctic to climate change, and because they can involve complex, vertically layered interactions between ice, liquid and aerosol particles. These clouds can persist for days, in spite of the inherent instability of the ice-water mix (Fridlind et al., 2007); theory suggests that mixed phase clouds persist if their updraughts are strong enough and ice nucleus concentrations low enough to maintain liquid water and prevent complete glaciation of the cloud (Korolev and Field, 2008). Morrison et al. (2011) argue this leads to a self-regulating structure in which thin layers of liquid water are maintained by radiatively-driven turbulence but are depleted by formation of ice that falls away from the liquid layer. Slight changes in meteorological forcing can rapidly glaciate the cloud layer and greatly diminish its radiative impact (Stramler et al., 2011). This combination of processes challenges numerical models of all scales, which struggle to simulate the balance between liquid and ice particles in field observations of Arctic boundary-layer clouds (Klein et al., 2009).

The response of high-latitude boundary-layer cloud cover response to the fractional cover of underlying sea-ice could be an important climate feedback and is discussed (including relevant observations) in Section 7.2.4.3.5.

7.2.1.3.4 Small-scale cloud-precipitation-circulation interaction

Precipitation can feed back profoundly on the small-scale circulations that accompany most cloud systems, for example by generating downdrafts and gusts that can produce more clouds, and by removing condensed water and aerosol particles. This helps foster characteristic 'open-cell' patterns of marine boundary layer cloud in midlatitude cold-air outbreaks (e.g., Muller and Chlond, 1996) and the low-latitude trade-wind belts (e.g., Xue et al., 2008); interactions with aerosols may also be involved, allowing rapid transitions between regimes of much vs. little cloud cover (see Section 7.4.3.2). Similar interactions help organize deep cumulus convection into mesoscale convective systems, squall lines, and tornadic thunderstorms (Houze, 1993), and the tendency to organize at these and larger scales gives convection a leading role in driving weather and climate variability especially at low latitudes. The organization is affected by details at small scales such as the fall speed of ice and snow crystals (Houze 1993). These interactions typically occur below the grid scale of climate models, partly explaining why many climate models reproduce convective organization poorly (Mapes et al., 2009).

7.2.2 Process Modelling and Observation of Clouds

7.2.2.1 Challenges in Modelling and Measurement of Cloud Processes

Cloud formation processes span scales from the submicron scale of cloud condensation nuclei to cloud system scales of up to thousands of kilometres. This range of scales is impossible to cover with direct numerical simulations on computers, and is unlikely to become so for decades if ever.

High-resolution models of individual cloud systems have nonetheless contributed greatly to our appreciation of interactions of turbulence with various types of cloud, e.g., cumulus, stratocumulus and cirrus. The usual strategy is called large-eddy simulation (LES) when applied to boundary-layer turbulence and cloud-resolving modelling (CRM) when applied to deep cumulus convection. The grid spacing is chosen to be small enough to resolve the dominant turbulent eddies that drive cloud heterogeneity, and the effects of unresolved eddies are parameterized. CRMs of deep convective cloud systems with horizontal resolutions of 2 km or finer can skilfully characterize statistical characteristics of the cloud ensemble, including fractional area coverage of cloud, vertical thermodynamic structure, the distribution of updrafts and downdrafts, and organization into mesoscale convective systems. However, some cloud ensemble properties are still sensitive to CRM microphysical parameterization assumptions, particularly the vertical distribution and optical depth of ice and mixed-phase clouds. Also, this type of modelling approach does not capture interactions with large scales, which requires a larger-scale model in which most of the cloud behaviour is parameterized.

Shallow cumulus cloud fields with clouds of below 2 km thickness are widespread over low latitudes. LES of such cloud fields with horizontal grid spacing of ~100 m and vertical grid spacing of ~40 m produces vertical profiles of cloud fraction, temperature, moisture and turbulent fluxes that agree well with available observations, though the simulated precipitation efficiency still shows some sensitivity to microphysical parameterizations (vanZanten et al., 2011).

LES of stratocumulus-topped boundary layers have shown considerable skill in simulating turbulence statistics and vertical thermodynamic structure (e.g., Ackerman et al., 2009; Stevens et al., 2005), and have been used to study the sensitivity of stratocumulus cloud organization, cloud thickness and albedo to changes in cloud condensation nucleus concentration (e.g., Savic-Jovcic and Stevens, 2008; Xue et al., 2008), sea-surface temperature (SST) and free-tropospheric conditions. However, model intercomparisons show that the simulated entrainment rate and stratocumulus cloud thickness are sensitive to the underlying numerical algorithms, even with vertical grid spacings as small as 5 m, due to under-resolution of the sharp capping inversion (Stevens et al., 2005).

7.2.2.2 Current Observing Capacities Relevant to Cloud Processes and Global Effects

Observations useful for detecting long-term (at least 20 year) changes in clouds, along with the significant difficulties in interpreting such observations given natural variability and other factors, are discussed in Chapter 2, Section 2.3.8. Here we discuss current observing capabilities for assessing climate-relevant cloud properties. A variety of observing systems, some new since AR4, combine to give us unprecedented insight

into different aspects of clouds on different temporal and spatial scales and to test and improve climate model simulations.

Satellites observations have proved particularly useful in the above regard due to their frequent and global (or near-global) coverage and because they directly measure how cloud characteristics affect outgoing radiation and hence the Earth's energy budget. Instruments available since the late 1970's on a combination of polar orbiters and geostationary weather satellites measure in the visible, near-infrared, thermal and microwave bands of the spectrum, and infer the reflectance, greenhouse effect, and approximate characteristic particle size and liquid water content of clouds. They observe nearly every spot on Earth at intervals from under an hour to many days depending on the instrument. Such satellite observations have limitations - meteorologically different distributions of cloud can look identical to the satellite, particularly if there are multiple cloud layers or cloud properties vary substantially across a satellite pixel (Wolters et al., 2010). Low-lying clouds over ice-covered surfaces are hard to detect in visible channels that are only useful during daytime. Most polar-orbiting satellites overfly each location at two particular local times of day, an issue in many regions with significant diurnal cycles of cloud cover. Geostationary satellites observe throughout the day, but with less spatial resolution.

Since the late 1990's, other satellites (e.g., NASA's Terra and Aqua) have carried advanced instruments that sample more wavelengths, polarisation, and multiple viewing angles to more accurately estimate the above cloud properties, as well as cloud-top altitudes, and to improve the estimation of rain rates although this remains a challenge (Marchand et al., 2010; Stephens; Kummerow 2007). Since 1997 the TRMM spaceborne radar has also observed the vertical structure of precipitation at latitudes equatorwards of about 35°N/S.

The 2006 launch by NASA of two coordinated, downward-pointing active sensors, the cloud profiling radar (CPR) on the CloudSat satellite and the CALIOP lidar on board the CALIPSO satellite has given the first accurate, near-global picture of the vertical distribution of cloud water and precipitation. Their high sensitivity, small footprint (roughly 1.5 km for CPR and 0.33 km for CALIOP), and vertical resolution (500 m and 30 m below 8 km, respectively) enable them to detect more tenuous and smaller clouds than most previous instruments (Winker et al., 2009). Their small daily sampling area makes them most useful for long-term statistics and for improving the interpretation of other instruments with overlapping, but broader, spatial coverage. Since AR4, satellite simulators (Chapter 9) have become widely used for comparing clouds and precipitation simulated by climate models with observations, and new simulators have been developed to compare with modern generations of cloud-observing satellite instruments.

Surface or aircraft-based observations give detailed process information on clouds in limited regions, often combining many types of measurements. Over the last 15 years, sites in different climate regimes, such as northern Europe, the central USA, China, the tropical Pacific, the West Indies, and the Arctic have begun continuously measuring overlying cloud properties using lidars, radars, and radiometers within a rich context of other atmospheric measurements. Such observations are widely compared with climate model behaviour (e.g., Phillips et al., 2004). Many field programs have been designed to improve understanding of cloud processes important to climate models using multi-instrument, multiplatform observations over a period of a few weeks that sample relationships between many quantities, followed by organized comparisons of climate and process models with the gathered data to evaluate and improve specific aspects of the simulations. Two recent examples are TWP-ICE for tropical cumulus convection in the Australian monsoon and cirrus formation (May et al., 2008) and VOCALS for aerosol-cloud-precipitation interaction in subtropical southeast Pacific stratocumulus clouds (Wood et al., 2011).

7.2.3 Representation of Clouds in Climate Models

7.2.3.1 Challenges of Parameterization Interaction and Subgrid Variability

Clouds form where rising air supersaturates sufficiently with respect to water vapour. The cloud droplets or ice crystals can then evolve, collide and grow to form precipitation. If these processes occurred uniformly across model grid cells, representing clouds in climate models would reduce to parameterizing their microphysics. Especially for ice and mixed-phase clouds this is already a challenge due to the complexity of microphysical processes.

Unfortunately clouds are often thin and short-lived, typically form through turbulent processes not resolved by the grid of a general circulation model (GCM), and may vary considerably within a GCM grid cell especially in convective systems or over mountainous terrain. Most CMIP5 climate model simulations use horizontal resolutions of 100–200 km in the atmosphere, with vertical layers varying between 100 m near the surface to more than 1000 m aloft. Within regions of this size in the real world, there is often enormous subgrid variability in cloud properties, associated with variability in humidity, temperature and vertical motion.

Because of this, the simulation of clouds in most modern climate models involves many interacting parameterizations that must work together as a system. These include parameterization of turbulence and cumulus convection, cloud fraction at each level, vertical overlap of these quantities at different heights, as well as cloud microphysics and aerosol and chemical transport. Each parameterization makes simplifying mathematical assumptions about the nature of subgrid variability within each grid cell; for pragmatic and historical reasons, these assumptions are frequently not fully consistent across the parameterizations used in one model, and vary significantly from model to model. For example, clouds in a grid column may be assumed to be vertically stacked for the radiation calculation, but not for calculating evaporation of precipitation. In summary, realistic simulation of clouds and their response to climate change forms one of the greatest challenges of climate modelling.

7.2.3.2 Advances in Microphysical Representation of Liquid Clouds Since AR4

Most microphysics schemes used in CMIP5-class climate models no longer assume that clouds are uniform within a model cell, but account for anticipated heterogeneity through simplifying assumptions and concise representations of subgrid-scale variability and covariance of relevant variables. Climate modellers have generally decomposed liquid clouds through a variety of attributes: 1) by considering gross categories of drop size (small cloud drops versus larger precipitating raindrops); 2) by classifying whether the drops developed within convective cloud cores or stratiform clouds that flow out of convective cores or form independently of them; 3) by predicting one or two parameters of an assumed drop size distributions; and/or 4) by assuming particular forms (e.g., Gaussian, top-hat) for the spatial variability of fields within a model cell or a cloud.

Most AR4 era climate models used a bulk (single moment) formulation for stratiform clouds, predicting only the time evolution of the average cloud and rain water mass in each gridbox. Two of 23 CMIP3 models (Storelvmo et al., 2006) employed two-moment formulations for cloud droplets with explicit prognostic equations for cloud drop number concentration (CDNC); others diagnosed CDNC from prognosed aerosols to allow for aerosol indirect effects (Storelvmo et al., 2009), using empirical relationships connecting aerosol mass to CDNC (e.g., Boucher and Lohmann, 1995; Menon et al., 2002), or based on proximity to land and altitude. The two moment schemes employed evolution equations for CDNC that activate cloud drops based on a subgrid vertical velocity and the size spectra of hygroscopic aerosols (Abdul-Razzak and Ghan, 2000; Ghan et al., 2011b; Nenes and Seinfeld, 2003). Many AR4 era models were forced to employ an arbitrary lower bound on CDNC to reduce the AIE, which is undesirable (Hoose et al., 2009).

More models participating in CMIP5 will use two moment schemes for liquid stratiform cloud. with the following advances. Some models include a diagnostic treatment of rain and snow number concentration as well as mixing ratio (Morrison and Gettelman, 2008; Salzmann et al., 2010), allowing the treatment of aerosol-scavenging and inclusion of the radiative effect of snow. Some models include an explicit treatment of subgrid cloud water variability for calculation of the microphysical process rates (Morrison and Gettelman, 2008). Some models no longer have to specify a lower bound on CDNC. Cloud drop activation schemes are becoming more sophisticated as aerosol schemes are becoming more complex, including more realistic accounting for aerosol hygroscopicity and particle size.

7.2.3.3 Advances in Microphysical Representation in Mixed Phase and Ice Clouds Since AR4

7.2.3.3.1 Mixed-phase clouds

New representations of the Bergeron-Wegener-Findeisen process in mixed-phase clouds (Lohmann and Hoose, 2009; Storelvmo et al., 2008b) compare the rate at which the pre-existing ice crystals deplete the

water vapour (Korolev 2007) with the condensation rate for liquid water driven by vertical updraft speed. Climate models are increasingly representing detailed microphysics, including mixed phase processes, inside convective clouds (Fowler and Randall, 2002; Lohmann, 2008; Song and Zhang, 2011). Such processes can influence storm characteristics like strength and electrification; more studies are needed to assess their importance for climate simulations.

7.2.3.3.2 Ice clouds

Although supersaturation with respect to ice is commonly observed in cirrus clouds, only one AR4 GCM (ECHAM) allowed ice supersaturation (Lohmann and Kärcher, 2002). Several global models now predict ice supersaturation (Gettelman et al., 2010; Liu et al., 2007; Salzmann et al., 2010; Tompkins et al., 2007). Tompkins et al. (2007) assume that once an ice cloud forms, the deposition process is sufficiently rapid that supersaturation is removed within a GCM time step. All other global models predict ice supersaturation based on parameterizations of homogeneous and/or heterogeneous freezing rates, which are discussed in Section 7.4.4.

7.2.3.4 Advances in Parameterization of Moist Turbulence and Cumulus Convection

Since AR4, parameterizations of cumulus convection and moist turbulence in many numerical weather prediction and climate models have continued to advance, leading to substantial improvement in their simulation of tropical rainfall and boundary-layer cloud and new capabilities for simulating cloud-aerosol interaction.

New 'adaptive' treatments of lateral entrainment into deep cumulus updrafts sensitive to environmental humidity or updraft buoyancy and velocity have improved simulations of the Madden-Julian Oscillation (MJO), tropical convectively-coupled waves, and mean rainfall patterns in the ECMWF operational weather forecast model (Bechtold et al., 2008) and the MIROC4 GCM (Chikira and Sugiyama, 2010). An adaptive detrainment parameterization implemented in the Met Office weather forecast model permits a computationally efficient bulk plume to mimic detrainment from a more realistic ensemble of clouds, improving forecast skill (Derbyshire et al., 2011). Various incremental cumulus parameterization changes have improved other climate models, e.g., modification of the deep convective trigger to account for entrainment effects on updraft buoyancy (Neale et al., 2008), combined with a new parameterization of cumulus momentum fluxes (Richter and Rasch, 2008), improved both the tropical mean state and ENSO variability simulated by the CCSM4 climate model.

Since AR4, more climate models have adopted cumulus parameterizations that calculate the typical vertical velocity in cumulus updrafts (e.g., Donner et al., 2011; Park and Bretherton, 2009), allowing more realistic representations of cloud microphysics and cloud droplet activation, a key issue for global simulation of aerosol-cloud interaction.

Several global models have adopted new approaches that more closely couple the parameterization of shallow cumulus convection and moist boundary layer turbulence. The eddy-diffusion mass flux (EDMF) scheme of Siebesma et al. (2007), adopted by ECMWF, combines an eddy-diffusion approach for small-scale turbulence with a mass-flux representation for strong non-cloudy and cloudy turbulent updrafts, which can incorporate a cumulus parameterization (Neggers, 2009; Neggers et al., 2009). In the CAM5 GCM, the shallow cumulus scheme of Park and Bretherton (2009) coupled to the turbulence parameterization of Bretherton and Park (2009) determines the cumulus-base mass flux from boundary layer updraft properties rather than ad-hoc closure assumptions typical of AR4 climate models. Using approaches such as these, many climate models simulate boundary-layer cloud radiative properties and vertical structure more accurately than at the time of AR4 (e.g., Köhler et al., 2011; Park and Bretherton, 2009).

7.2.3.5 High-Resolution Global Modelling

Since AR4, increasing computer power has made it possible to simulate the global circulation of the atmosphere and associated clouds with greater resolution. There have been three types of developments. First, models have been run with resolution that is higher than in the past, but not so high that cumulus clouds can be resolved. Second, models have been run with resolution high enough to resolve (or "permit") large individual cumulus clouds over the entire globe. In a third approach, the parameterizations of global

models have been replaced by embedded cloud-resolving models. The first approach is a continuation of the broad evolution of climate models toward finer simulated scales, so it is assessed in Chapter 9. The other approaches discussed below, are computationally intensive and have only been applied to weather and short climate simulations. Because they overcome many of the parameterization challenges associated with clouds, they make an interesting complement to conventional global atmospheric models.

7.2.3.5.1 Global cloud-resolving models

A Japanese global cloud-resolving model (GCRM) called NICAM (Miura et al., 2005; Tomita et al., 2005) has been run with a grid spacing of as little as 3.5 km. At present it can only be used for relatively short simulations of a few simulated weeks or months on the fastest supercomputers. Even with a 3.5 km grid spacing, NICAM can only resolve large cumulus clouds, not the eddies within boundary-layer clouds. Parameterizations of cloud microphysics, radiation, and turbulence are still needed, but these benefit from the fine grid spacing.

NICAM simulates many features of deep convection that are very challenging for conventional GCMs. Sato et al. (2009) show that NICAM can simulate the diurnal cycles of precipitation associated with land-sea breezes and thermally induced topographic circulations; results improve for finer grid spacing. Oouchi et al. (2009) shows NICAM accurately simulates most features of the Asian summer monsoon. Inoue et al. (2010) showed the cloudiness simulated by NICAM is in good agreement with observations from CloudSat and CALIPSO, but the simulation of cloud ice is not satisfactory and requires an improved parameterization of ice microphysics. Iga et al. (2010) found that changes in the turbulence and cloud microphysics parameterizations of NICAM can strongly affect the upper-level cloudiness, total precipitation, and Hadley circulation. In particular, the Hadley circulation weakens when the simulated high cloud amount increases, due to reduced tropospheric radiative cooling. Even with such fine grid resolution, microphysical parameterization uncertainties can significantly affect simulated climate.

Within the next decade, it may become computationally feasible to use GCRMs for century-long climate change simulations. Meanwhile, GCRMs can be used in shorter numerical experiments that shed light on results from lower-resolution models.

7.2.3.5.2 Models that use embedded cloud-resolving models as "super-parameterizations"

Grabowski and Smolarkiewicz (1999) and Grabowski (2001) pioneered the use of CRMs as 'super-parameterizations', i.e., substitutes for the parameterizations in conventional GCMs. Khairoutdinov and Randall (2001) tested the idea in a version of the Community Atmosphere Model (CAM). Tao et al. (2009) developed a similar model using a different global model and different CRM. As with NICAM, cloud microphysics, radiation, and turbulence must still be parameterized; the computational cost is intermediate between GCRMs and conventional GCMs

The "super-parameterized" CAM gives realistic simulations of the diurnal cycle of precipitation (Khairoutdinov et al., 2005; Pritchard and Somerville, 2010) and the MJO (Benedict and Randall, 2009). Its climatological biases in precipitation and mean circulation are comparable to other climate models, including excessive rainfall in boreal summer over the western Pacific Ocean and southern Asia (Khairoutdinov et al., 2005) and under-prediction of marine stratocumulus clouds (Blossey et al., 2009; Wyant et al., 2009). Stan et al. (2010) coupled the super-parameterized atmosphere model with a global ocean model for 20 simulated years, giving an improved simulation of the Asian summer monsoon (DeMott et al., 2011) and mean rainfall biases compared to the uncoupled model version, and an encouraging simulation of El Niño - Southern Oscillation.

Super-parameterized global atmospheric models can be used for climate simulations spanning decades or centuries with present-day computers. Like GCRMs, these models give improved simulations of climate variability on a range of time scales, but will benefit from improved parameterizations of turbulence and microphysics.

7.2.4 Cloud and Water-Vapour Feedback

Climate feedbacks are a central concern for projecting the magnitude of climate change, because they determine the sensitivity of climate to external forcing agents. The overall climate sensitivity of current

climate models is assessed in Chapter 9, but the individual contributing feedbacks are assessed in the relevant process-oriented chapters. Water vapour, lapse rate and cloud feedbacks are all assessed in this chapter because they involve moist atmospheric processes closely linked to clouds. In combination they produce most of the simulated climate feedback and also most of its intermodel spread.

Feedbacks are often expressed as a TOA net downward radiative flux change per degree of global surface temperature increase; in this section they are converted to dimensionless feedback factors by further multiplying by 0.31 K W^{-1} m^2 (Roe and Baker, 2007). Such feedback factors from different processes add to give the total feedback factor *f*. The equilibrium climate sensitivity is inversely proportional to 1-*f*, so a total feedback factor of *f* = 0.6 (typical of GCMs) would amplify the (blackbody) climate sensitivity by a factor of 2.5. Regional feedbacks are sometimes estimated based on surface, rather than TOA, energy fluxes; however, one must then also consider turbulent latent and sensible heat fluxes, complicating the analysis and making it a less useful predictor of the overall coupled system response, so we do not adopt this approach.

7.2.4.1 Cloud Altitude Feedback Mechanisms Involving High-Level Clouds

High clouds exert little net TOA radiative effect in the current climate due to near-compensation between their longwave and shortwave cloud radiative effects (Kiehl, 1994). Nonetheless, systematic changes in their properties could produce a significant radiative feedback by altering this balance.

New studies confirm that in typical global warming scenarios, longwave cloud feedback is consistently positive across CMIP3 climate models, with an interquartile range (IQR) of 0.06–0.12 (i.e., 0.2–0.4 W m^{-2} K^{-1}) (Soden and Vecchi, 2011; Zelinka et al., 2011a), and is primarily due to high clouds (Zelinka et al., 2011a). The dominant driver of longwave cloud feedback appears to be a robust consequence of global warming - an increase in the heights of the tropopause and the main level at which the deepest convective clouds stop rising and cloudy air flows outward, tentatively attributed in AR4 to the so-called fixed anvil-temperature (or FAT) mechanism (Hartmann and Larson, 2002). According to this mechanism, the outflow level from deep convective systems is determined ultimately by the highest point at which water vapour amounts are sufficient to emit significant infrared radiation; this point tends to occur at the same temperature regardless of climate, provided that relative humidity does not change too much, and therefore occurs at a higher altitude in a warmer climate. A positive cloud altitude feedback results because an otherwise identical cloud located higher in the atmosphere exerts a stronger greenhouse effect. New research has confirmed that the hypothesised behaviour is simulated in more realistic models (Kuang and Hartmann, 2007; Kubar et al., 2007), and explains the systematic tendency toward positive cloud feedback in GCMs (Zelinka and Hartmann, 2010). Zelinka et al. (2011b) find that in the CMIP3 models, the IQR of cloud altitude feedback factor is 0.08–0.15, which is even stronger than the overall longwave cloud feedback.

The observational record allows us to verify various elements of the expected FAT response to global warming. The global tropopause is rising as expected (Chapter 2). Cloud height is correlated to regional and seasonal changes in near-tropopause temperature structure (Eitzen et al., 2009), although the response is affected by changes in stratospheric circulation (Chae and Sherwood, 2010; Eitzen et al., 2009).

7.2.4.2 Feedback Mechanisms Involving the Amount of Middle and High Cloud

Nearly all GCMs also simulate an overall reduction in middle and high cloud amount in warmer climates, especially in the subtropics (Trenberth and Fasullo 2009; Zelinka and Hartmann 2010). This reduction is geographically correlated with simulated subtropical drying (Meehl et al., 2007), suggesting that it is tied to large-scale circulation changes (Sherwood et al., 2010; Wetherald and Manabe, 1980). The upward mass flux in deep clouds also decreases in a warmer climate (Section 7.2.5.1) which might affect cloud cover in ways difficult to capture in current models. In global average, both middle and high-level simulated cloudiness reductions cause a positive shortwave feedback, but the high cloud reductions also compensate nearly half of the longwave cloud altitude feedback. This may explain why researchers did not identify the important role of cloud altitude feedbacks sooner. From Figure 8 of Zelinka and Hartmann (2011b) the IQR of the CMIP3 multimodel mean net cloud feedback factor from middle and high clouds is estimated to be 0.08–0.16. This is similar to the cloud altitude feedback alone, so we conclude that the additional net feedback due to mid/high cloud amount reductions is small.

Bender et al. (2011) find decreasing trends in satellite-derived subtropical cloud cover, but also note large uncertainties involved in correcting for satellite radiometer and orbital drift as well as cross-calibration between satellites (see Chapter 2). Tselioudis and Rossow (2006) predict reduced storm-track cloud cover in warmer climates based on observed present-day relationships with meteorological variables combined with model-simulated changes to those driving variables. In agreement with the above analysis, they found that this reduction contributed little to net cloud feedback, due to compensating longwave and shortwave contributions.

Thin cirrus clouds exert a net warming effect on climate, and cover a significant area. These clouds could therefore exert an important feedback on climate if their area changed relatively moderately with global temperature (e.g., Rondanelli and Lindzen, 2010). We find no compelling evidence from observations, process models, or GCMs suggesting such a feedback is important, and the CMIP3 multimodel mean change of thin high cloud fraction is smaller than for other cloud types (Zelinka et al., 2011a). A caveat is that these clouds are challenging to accurately simulate in climate models (see Chapter 9).

A positive contribution to net cloud feedback also comes from the anticipated poleward shift of storm tracks in a warmer climate. Even if storm track clouds remained unaltered, this shifts clouds to latitudes of weaker sunlight, decreasing the planetary albedo. Such a shift occurs in most models (Yin, 2005) and has appeared in recent cloud and other data (see Chapter 2); similar shifts are also seen in indicators of the edge of the tropics (Scheff and Frierson, 2011). The impact of the observed shift on clouds, particularly the reduction in subtropical cloud cover, appears to be significant and would imply a strong positive feedback if it were due to global warming (Bender et al., 2011). Recent studies call into question how much of the observed shifts are temperature-driven vs. ozone-driven [Chapter 10.3.x], so the true magnitude of this feedback contribution remains highly uncertain. As most GCMs produce too little storm-track cloud in the southern hemisphere, it has been suggested that they underestimate this feedback (Trenberth and Fasullo, 2010) even if the model-predicted shifts are correct.

7.2.4.3 Feedback Mechanisms Involving Low Clouds

Low clouds exert a strong net cooling effect on the Earth, such that if their coverage or water content were climate-sensitive a feedback would result. Feedback contributions from low clouds continue to differ significantly among models, and to cause most of the spread in global climate sensitivity among GCMs for both transient and equilibrium simulations (e.g., Dufresne and Bony, 2008). Zelinka et al. (2011a) show that the CMIP3 IQR for low cloud feedback factor is 0.09–0.15; all analysed models had positive low cloud feedback, but of widely differing magnitudes. The feedback derives from slight decreases in low cloud cover, which in multimodel mean occurs throughout low and mid-latitudes (Zelinka et al., 2011b), though the geographical pattern varies widely between models (Webb et al., 2012).

No robust feedback mechanisms involving tropical and mid-latitude low clouds have yet been established, though many possible mechanisms for both positive and negative feedbacks have been proposed. It has long been suggested that cloud water content could increase in a warmer climate simply due to the higher water vapour mixing ratio (WVMR) in sub-cloud air or more condensation per unit height in an adiabatic cloudy updrafts, but this argument ignores the physics of crucial cloud-regulating processes like precipitation formation and turbulence. Observational evidence discounting the suggested effects was reported in AR4.

As noted in Section 7.2.1.3.2, at each latitude, low cloud cover tends to favor the coldest oceans, which can naively be interpreted to imply a positive feedback. This relationship is thought to be regulated by lower-tropospheric stability, but different measures of stability that are similarly skilful in the current climate imply significantly different cloud changes in a warmer climate (Section 7.2.1.3). Likewise, interannual variations of ocean surface temperature in marine stratocumulus regions tend to anti-correlate with changes in cloud cover and water content, again suggesting a negative feedback if taken at face value (Eitzen et al., 2011). However, since these regional variations are accompanied by local atmospheric stability changes that would differ from those accompanying a global warming, this interannual relationship is unlikely to indicate the global feedback (see Section 7.2.4.3.4).

Studies since the AR4 have yielded some further insight into the diversity of low-cloud feedbacks in GCMs. Several climate models that have been run with land removed (a so called aquaplanet) show cloud feedback

comparable to their usual configuration, suggesting that neither land-surface feedbacks nor land-induced heterogeneity is crucial in determining the low-cloud response to warming in the full GCMs (Medeiros et al., 2008). Low-cloud feedback in GCMs arises from a mix of cumulus and stratocumulus cloud feedbacks (Williams and Tselioudis, 2007; Williams and Webb, 2009; Xu et al., 2010). Bony and Dufresne (2005) found that that the intermodel spread among tropical feedbacks derives mainly from the regimes of moderate subsidence that support both these cloud types. Webb et al. (2012) found that the feedback factor and its spread is largest in cool-ocean (stratocumulus) regimes and shallow cumulus regimes covering a larger fraction of the subtropics.

Other post-AR4 studies have probed how different parameterizations affect simulated low-cloud feedbacks. For example, an increased positive cloud feedback in two successive versions of the NCAR GCM is due mainly to changes in the representation of shallow cumulus convection, but with significant compensation from the boundary-layer parameterisation, and the biggest simulated cloudiness reductions occur in the Southern Ocean stratocumulus regime due to interactions between cumulus and stratocumulus (Gettelman et al., 2011b). A single-column diagnostic study of Zhang and Bretherton (2008) found that the interaction between boundary layer, shallow and deep convective parameterisations in transporting moisture upward to form clouds was crucial in determining the cloud albedo feedback, even in a stratocumulus regime.

Cloud feedbacks have been examined in first-generation global cloud-resolving models and "superparameterized" GCMs, (Section 7.2.3.5), which avoid cumulus parameterization. These models have shown increases in low-cloud cover in warmer climates (Wyant et al., 2009; Wyant et al., 2006), predicting climate sensitivities at or below the low end from GCMs. GCMs with superparameterisations do not, however, simulate low clouds any better than (or even as well as) traditional GCMs, apparently due to the very fine grid (~100 m or better) required for proper, explicit simulation of the cloudy boundary layer (Section 7.2.2.1, Blossey et al., 2009). Limited-area high-resolutions models have also been used to estimate cloud feedbacks (Lauer et al., 2010; Xu et al., 2010), and have yielded a range of estimates even broader than those of global models. These models and methodologies are less well tested than traditional GCMs, so we assign relatively little confidence to their feedback behaviour at this time.

7.2.4.4 Feedbacks Involving Changes in Cloud Optical Depth

Another possible cloud feedback mechanism involves changes in cloud phase. At mixed-phase temperatures of –40 to 0°C, cloud ice particles that contribute most to light scattering are typically several-fold larger than cloud water drops (e.g., Mitchell et al., 2010), so a given mass of cloud ice reflects less sunlight than the same mass of cloud water droplets in today's atmosphere. As climate warms, the shift from ice to liquid clouds would raise albedos if cloud water mass did not change. The resulting negative cloud feedback appears in GCMs (Senior and Mitchell, 1993), with a magnitude that correlates with the simulated amount of cloud ice in mixed phase clouds (Tsushima et al., 2006). It is unlikely however that cloud water content would remain fixed as clouds changed phase, and the key physics is not represented in climate models, so this feedback mechanism is highly uncertain.

Zelinka et al. (2011b) isolated the component of simulated cloud feedback due to changes in cloud optical depth in a set of CMIP3 GCMs. The global-mean net feedback scattered around zero, but, there was a tendency toward slightly reduced optical depths at low and middle latitudes, and increases at latitudes poleward of 50° yielding a negative local feedback that they attributed partly to phase changes and partly to the greater cloud water amounts expected from increased poleward moisture transport (Vavrus et al., 2009).

7.2.4.5 Feedback from Arctic Cloud Interactions with Sea Ice

Arctic clouds, despite their low altitude, have a net warming effect at the surface in the present climate because their downward emission of infrared radiation over the year outweighs their reflection of sunlight during the short summer season. However, they also cool the atmosphere, so their effect on the energy balance of the whole system is ambiguous and depends on the details of the vertical cloud distribution and the impact of cloud radiative interactions on ice cover.

Visual cloud reports (Eastman and Warren, 2010) and lidar observations available since AR4 (Kay and Gettelman, 2009; Palm et al., 2010) now agree that low cloud cover over Arctic oceans is inversely

correlated to sea ice amount, with open water producing more cloud. The observed effect is weak in boreal summer, when the melting sea-ice is at a similar temperature to open water and stable boundary layers with extensive low cloud are common over both surfaces, and strongest in boreal autumn when cold air flowing over regions of open water stimulates cloud formation by boundary-layer convection (Kay and Gettelman, 2009; Vavrus et al., 2011). Kay et al. (2011) show that a GCM can represent this seasonal sensitivity of low cloud to open water, but only with an appropriate boundary-layer cloud parameterization. Vavrus et al. (2009) show that in a global warming scenario, GCMs simulate more Arctic low-level cloud in all seasons, but especially during autumn and winter when open water and very thin sea ice increase considerably, decreasing low-level stability and increasing upward moisture transport from the surface to levels at which it forms clouds.

A negative Arctic cloud feedback was suggested by Liu et al. (2008) on the basis that observed surface warming in recent decades was greater under clear-sky than under cloudy conditions, but this argument was not tested in a climate model and does not control for the large correlated effects of weather variability on both clouds and surface temperature. Gagen et al. (2011) present tree-ring evidence that summertime Arctic cloud cover was negatively correlated with Arctic temperatures over the last millennium, which is consistent with the conclusions of the above studies assuming there was less ice during warmer periods. While Gagen et al. (2011) presented this as evidence of negative cloud feedback, they ignored the year-round greenhouse effect of clouds, which could change the imputed feedback to a positive one (Palm et al., 2010). Note that there are pitfalls to using natural climate variations to infer cloud feedbacks, described in Section 7.2.4.3.7, that apply here as well.

7.2.4.6 Fast Adjustment of Clouds and Precipitation to a CO_2 Change

Climate feedbacks are usually estimated by comparing steady-state simulations with control and doubled CO_2, and analysing the contribution of different processes to the simulated changes in radiation balance, normalized by the change in global-mean surface air temperature (e.g., Soden and Held, 2006).

Gregory and Webb (2008) partitioned the transient response of the radiation balance of GCMs to an instantaneous doubling of CO_2 into a 'fast' (sub-seasonal) adjustment in which the land surface, atmospheric circulations and clouds respond to the radiative effect of the CO_2 increase, and an 'SST-mediated' response that develops more slowly as the oceans warm. They found that in some climate models, fast adjustment of clouds can have comparable top-of-atmosphere radiative effects to the ensuing SST-mediated cloud changes. However, Andrews and Forster (2008) found that this behavior was exceptional and that on average, fast cloud adjustments in a suite of climate models causes less than 20% of their equilibrium radiative feedback; low cloud adjustments have the biggest global net radiative impact (Colman and McAvaney, 2011). Fast adjustment may cause clouds to respond slightly differently during a transient climate change (in which SST changes have not caught up to CO_2 changes) than after equilibrium is reached and SST changes have been fully expressed. Studies of fast adjustment of climate models to a step CO_2 increase have also shown reduced global-mean precipitation (Cao et al., 2011) and cloud cover (Andrews and Forster, 2008), a shift of tropical clouds and precipitation from ocean to land (Lambert et al., 2011; Wyant et al., 2011), and shallowing of marine stratocumulus cloud (Caldwell and Bretherton, 2009; Wyant et al., 2011). The ensuing response to warming SST reverses many of these trends, increasing global-mean precipitation and shifting precipitation and high clouds back over the tropical oceans (Cao et al., 2011).

7.2.4.7 Observational Constraints on Global Cloud Feedback

A number of studies since AR4 have attempted to constrain cloud feedback (or total climate sensitivity) from observations; here we discuss those using modern cloud, radiation or other measurements. Section [12.5] discusses those based on past temperature data and forcing proxies.

One approach is to seek observable aspects of present-day cloud behaviour that reveal cloud feedback. In at least two climate models, large sets of runs with nonstandard parameter settings produce feedback strengths that correlate with the amount or water content of cloud simulated for the present day (Williams and Webb, 2009), but in other models this does not happen (Yokohata et al., 2010), and the resulting relationships do not hold across multiple models e.g., CMIP3 (Gettelman et al., 2011a). Among the AR4 models, net cloud feedback is strongly correlated with mid-latitude relative humidity (Volodin, 2008) and with characteristics

of the southern-hemisphere storm track (Trenberth and Fasullo, 2010); if valid either regression relation would imply a relatively strong positive cloud feedback in reality, but no mechanism has been proposed to explain or validate these empirical relationships. Likewise, Clement et al. (2009) found realistic decadal variations of low cloud over the North Pacific in only one model (HadCM3) and argued that the relatively strong cloud feedback in this model should therefore be regarded as more likely, but provided no evidence for such a link. Chang and Coakley (2007) examined midlatitude maritime clouds and found cloud thinning with increasing temperature, consistent with a positive feedback, while Gordon and Norris (2010) found the opposite result in following a methodology that tried to isolate thermal and advective effects. In summary, there is no evidence of a robust link between any of the noted observables and the global feedback, though some apparent connexions are tantalising and are being further studied.

Several studies have attempted to derive long-term climate sensitivity from interannual relationships between global-mean observations of top-of-atmosphere radiation and surface temperature. One problem with this is the different spatial character of interannual and long-term warmings; another is that the methodology can be confounded by cloud variations not caused by those of surface temperature (Spencer and Braswell, 2008). A range of climate sensitivities has been inferred based on such analyses (Forster and Gregory 2006; Lindzen and Choi, 2011). Crucially, however, among different GCMs there is no correlation between the interannual and long-term cloud-temperature relationships (Dessler, 2010), contradicting the basic assumption of these methods. On the other hand the GCMs, on average, do predict a cloud radiative response to these high-frequency global temperature changes that is consistent with the global feedback response and with observations, increasing the credibility of their predictions at longer timescales (Dessler, 2010). More recently there is interest in relating the time-lagged correlations of cloud and temperature to feedback processes (Spencer and Braswell, 2010) but again these relationships appear to reveal only a model's ability to simulate ENSO or other modes of interannual variability properly, which cannot be translated directly into cloud feedback on long-term global warming (Dessler, 2011).

While a number of studies have proposed methods to infer the long-term cloud feedback from observed variability, for a method to be accepted it should have a sound physical basis and be shown to work consistently when applied to climate models. No method yet proposed passes both tests. Moreover, some model studies show that the response of global cloud radiative effect to a global warming is sensitive to relatively subtle details in the geographic warming pattern, such as the slight hemispheric asymmetry due to the lag of southern ocean warming relative to northern latitudes (Senior and Mitchell, 2000; Yokohata et al., 2008). Cloud responses to specified uniform ocean warming without CO_2 increases are not the same as those to CO_2-induced global warming simulated with more realistic oceans (Ringer et al., 2006), partly because of fast adjustment (Section 7.2.4.3.6) and because low clouds also feed back tightly to the underlying surface (Caldwell and Bretherton, 2009). Simulated cloud feedbacks also differ significantly between colder and warmer climates in some models (Crucifix, 2006; Yoshimori et al., 2009). These sensitivities highlight the challenges facing any attempt to infer long-term cloud feedbacks from simple data analyses.

[INSERT FIGURE 7.5 HERE]

Figure 7.5: CFMIP figure on cloud feedbacks in CMIP5 models. [PLACEHOLDER FOR SECOND ORDER DRAFT: CMIP3 version used as placeholder.]

7.2.4.8 Feedback Synthesis

Together, water vapour-lapse rate and cloud feedback are the principal determinants of climate sensitivity. The combined global water vapour-lapse rate feedback simulated by all GCMs is strongly positive (feedback factor IQR 0.28–0.33), similar to AR4, and has the magnitude expected from simple physical arguments that on global scales, absolute humidity changes much more than relative humidity in a perturbed climate.

Cloud feedbacks on CO_2-induced climate change are less certain, but the combined evidence suggests a substantial positive net feedback from clouds. Two estimates of the IQR of overall cloud feedback factor, based on two slightly different groups of CMIP3 GCM simulations are 0.11–0.33 (Table 1 of Soden and Held, 2006), and 0.17–0.28 (Figure 8 of Zelinka et al., 2011a). Differences between models have not reduced since AR4 [TBC] and dominate the spread in model climate sensitivities. However, new approaches to diagnosing cloud feedback in GCMs have clarified robust cloud responses, while continuing to implicate low cloud cover as the most important source of intermodel spread in simulated cloud feedbacks. Some new

approaches to observation and mechanistic understanding of cloud feedback have been taken, using a hierarchy of models and observations on local and global scales.

Several physical processes contribute to a positive net cloud feedback, as summarized in Figure 7.6. First, high clouds to rise in altitude and thereby exert a stronger greenhouse effect in warmer climates. This altitude feedback mechanism is well understood, has theoretical and observational support, occurs consistently in GCMs and other models, and explains about half of the multimodel-mean positive cloud feedback. Second, middle and high level cloud cover tends to decrease in warmer climates even within the storm tracks and ITCZ, consistent with the reduction of upward, cloudy air fluxes required to balance the hydrological cycle in an atmosphere with more water vapour. This decreases the albedo more than the greenhouse effect, adding positive feedback. Third, observations and most models suggest storm tracks shift poleward in a warmer climate, drying the subtropics and moistening the high latitudes, which causes further positive feedback via a net shift of cloud cover to latitudes that receive less sunshine. Most GCMs also predict low cloud amount decreases especially in the subtropics, another source of positive feedback, though there is wide spread in this feedback and it lacks a well-accepted theoretical basis. Over high latitudes, models suggest warming-induced transitions from ice to water clouds may cause clouds to become more reflective, but this optical depth feedback operates over too small a region to be globally significant. While other feedbacks remain possible and uncertainties remain –particularly for low cloud amount– there is no consistent observational evidence, consensus among models, nor robust physical argument that would favour these having a negative feedback contribution over a positive one.

Since all established globally-significant feedback mechanisms are positive, a positive overall cloud feedback is more likely than a negative one. How much more likely depends on the likely strength of any other mechanisms that may exist. Our chief guidance on this question is from GCMs. As of yet, the wide variety of formulations and resulting behaviour in these models has not produced a single example [TBC] where other mechanisms have been strong enough to outweigh unequivocally the known sources of positive feedback. In the absence of supporting evidence, we judge that cloud feedback outside the range of any current CMIP3 GCM is unlikely (less than 33% probability), and that such cloud feedback is no more likely to lie below the CMIP3 range than above it. The lowest cloud feedback for any CMIP3 model was slightly positive (Soden and Held, 2006), so this reasoning implies that the net cloud feedback is likely (83% chance) positive. Note that this conclusion is independent of constraints on climate sensitivity from observed trends or palaeoclimate information as discussed in Section 12.4.

[INSERT FIGURE 7.6 HERE]

Figure 7.6: Robust cloud responses to greenhouse warming simulated by the CMIP3 multimodel ensemble. Panel (a) is a schematic latitude-altitude section showing typical cloud types in a pre-industrial climate. Grey (white) indicates clouds composed predominantly of liquid water (ice). Raindrops and snowflakes indicate the typical precipitation type. Dotted line indicates the typical freezing level, and purple dashed line indicates the tropopause. Panel (b) shows the same cross section for a warmer climate, with arrows denoting the movement of different boundaries. Tropical deep convection regions narrow and intensify, the subsidence regions of the subtropics widen poleward, with most GCMs projecting low cloud decreases in this area, and storm track cloud and precipitation also shift poleward. Cirrus cloud tops rise in lockstep with the tropopause, helping induce positive longwave cloud feedbacks. The rising freezing level causes more cloud to become liquid, contributing to increased optical thickness of high latitude clouds in the CMIP3 multimodel mean. [PLACHOLDER FOR SECOND ORDER DRAFT: CMIP5]

7.2.5 Basis of Precipitation Changes in Cloud Physical Processes

This section reviews fundamental process-level knowledge relevant to changes of precipitation characteristics in warmer climates and their relation to clouds. Observed trends in mean and extreme precipitation are discussed in Chapters 2 and 10; changes in hurricanes and other dynamical phenomena, and precipitation projections for specific regions, are discussed in Chapters 11, 12 and 14.

7.2.5.1 Coupling of Large-Scale Trends in Clouds and the Hydrological Cycle

In a perturbed climate, regional changes in cloud cover and type will inevitably accompany regional changes in the hydrological cycle. The CMIP3 coupled climate models as a group predict that in 21st century global warming scenarios, the ITCZ will narrow and intensify, the subtropics will dry, and the storm tracks will move poleward (Held and Soden, 2006; Meehl et al., 2007); helping induce the cloud responses discussed in

Section 7.2.4 and shown in Figure 7.6. The strong geographical relation between projected precipitation and cloud changes can be seen in CMIP multimodel mean results (IPCC 2007) [PLACEHOLDER FOR SECOND ORDER DRAFT: CMIP5 updates].

Within the tropics, individual climate models simulate regional changes in the zonally asymmetric part of mean rainfall patterns and circulation as the climate warms. The detailed geographical pattern of these changes is model-dependent (Bony et al., 2004; Neelin et al., 2006), and they lead to annual-mean deep convective cloud changes in each model that are quite similar to the corresponding changes in mid-tropospheric vertical motion (Bony et al., 2004), such that up to 70% of the longwave and shortwave cloud radiative forcing changes at a typical location over the tropical oceans can be ascribed to changes in the vertical motion at that location (Wyant et al., 2006). This point motivates such analysis techniques as vertical-velocity binning (Bony et al., 2004) for separating local dynamically driven cloud changes from the residual 'thermodynamic' cloud changes. Over the scale of the tropics or the entire globe, regions of enhanced mean ascent balance those of more mean descent, so these dynamical effects on cloud response largely cancel while the thermodynamic cloud changes need not (Bony et al., 2004).

7.2.5.2 Changes in Precipitation Extremes

Studies since AR4 have focused new attention on the climate-dependence of rainfall extremes down to hourly time scales. The expectation has been that extreme precipitation is limited by the precipitable water content of the atmosphere in the storm environment, which increases at roughly 7% per °C of warming ("Clausius-Clapeyron" or CC scaling) if relative humidity is constant. However, not only can relative humidity vary significantly, but intensified latent heating could feed back on cloud-scale dynamics so as to alter the efficiency with which precipitable water is condensed (O'Gorman and Schneider, 2009); such changes are sensitive to factors other than surface temperature (Berg et al., 2009).

Lenderink and van Meijgaard (2008, 2010) report that extremes of daily rain accumulation at several sites in Europe show approximate CC scaling with respect to weather variations (not temperature trends), but summertime extremes at the hourly time scale showed a scaling twice this strong, probably due to stronger convective storms on hotter days. Lenderink and van Meijgaard (2010) link this to atmospheric humidity, finding an even stronger supersensitivity of 17% per °C of dewpoint. These results were roughly reproduced by regional model simulations. Similar supersensitivities appear in maritime tropical rainfall rates (Allan and Soden, 2008; Allan et al., 2010). However, they do not appear at most sites in the U.S. (Shaw et al., 2011) or Australia (Jones et al., 2010b), where hourly rain scaling is more erratic but close to CC on average. In some datasets, particularly the Australian data, scaling turns negative at the highest temperatures which are associated with dry conditions.

The interpretation of these results is not straightforward, especially since weather relationships cannot be assumed to apply directly to climate changes. Berg et al. (2009) point out that the temperature scaling inferred from the European data can weaken significantly if one controls for other factors such as storm type. Allan et al. (2010) note that modelled and observed trends in maritime convective extremes since the 1980's are both close to CC, but that weather-related fluctuations show scaling several times stronger; this reflects the relatively strong sensitivity of tropical convection to local maxima of temperature as opposed to overall means (Johnson and Xie, 2010). Idealised calculations using regional (Muller et al., 2011) or cloud-resolving (Romps, 2011) models show extremes scaling with or somewhat below CC in idealised global warming scenarios, possibly because humidity increases more slowly near the surface than aloft.

Regional models are inconsistent in predicting how rainfall extremes will change with global warming, ranging from CC scaling to nearly double this; predicted changes to extremes are also regionally diverse, exceeding CC by factors of several in regions that receive more rain but decreasing in those receiving less (Lenderink and van Meijgaard, 2010). This makes generalised statements difficult. However, all results point to greater rainfall extremes in warmer climates except in places of significant rainfall decline. The best guess as to the magnitude of the increases for typical conditions (e.g., where rain climatology matches the global mean) is probably close to 7% per °C, but this remains uncertain especially for extremes on sub-daily time scales.

7.2.6 Anthropogenic Sources of Cloudiness

Human activity can be a source of additional cloudiness through specific processes involving a source of water vapour in the atmosphere. We discuss here the impact of aviation and irrigation on water vapour and cloudiness. The impact of water vapour sources from fossil fuel combustion at the Earth's surface is thought to be negligible.

7.2.6.1 Contrails and Contrail-Induced Cirrus

Aviation jet engines emit hot moist air which can form line shaped persistent condensation trails (contrails) under certain atmospheric conditions. These have been observed to spread into large cirrus sheets which may persist for several hours, and observational studies confirm their overall positive net radiative forcing impact (Haywood et al., 2009). Aerosol emitted within the aircraft exhaust can also affect cloudiness. This last effect is classified as an aerosol-cloud interaction and is treated as part of Section 7.4.

Persistent contrails can form in the upper troposphere in air that is already supersaturated with respect to ice and colder than –40°C. They are composed of ice crystals that are typically smaller than background cirrus (Frömming et al., 2011; Heymsfield et al., 2010), they trap longwave radiation and reflect solar radiation leading to a pronounced diurnal cycle in radiative forcing (Burkhardt and Kärcher 2011; Rap et al., 2010b; Stuber and Forster, 2007).

Forster et al. (2007) estimated the 1750–2005 radiative forcing from persistent linear contrails as 0.01 W m^{-2} (0.003 to 0.03 W m^{-2} 90% uncertainty range). They also estimated a range of forcing from 1750 to 2000 for aviation induced cirrus of 0.03 W m^{-2} (with a 0.01 to 0.08 W m^{-2} 90% uncertainty range). Lee et al. (2009) scaled the IPCC AR4 estimates of the persistent contrail RF to account for revised fuel use estimates, propulsive efficiency and flight routes, which resulted in a 18% increase for 2005. Rap et al., (2010b) found that contrails preferentially formed in conditions where there was already considerable cloud present, which contributed to lessen their radiative forcing. Kärcher et al. (2010) corrected previous estimates for inconsistencies between contrail cover and averaged optical depth and came up with a range of 0.008 to 0.020 W m^{-2} for the reference year 2000. Overall we adopt a RF estimate of 0.02 ± 0.01 W m^{-2} (90% uncertainty range) for persistent linear contrails for the period 1750–2010.

Rap et al. (2010a) strengthened the assessment that aviation contrails are very unlikely at current levels of coverage to have an observable effect on surface temperature and diurnal temperature range. They also found a very small efficacy for the contrail radiative forcing of 31%, smaller than the 59% found in the ECHAM4 model (Ponater et al., 2005).

Contrail induced cirrus forcing estimates have been based on correlating observations (e.g., Boucher, 1999) and may have included cirrus changes that were not directly caused by aviation. Burkhardt and Kärcher (2011) have removed the need for such gross assumptions by modelling the global impact of both line-shaped and spreading contrails within a climate model (Burkhardt and Kärcher, 2009). They estimated a RF of 0.037 W m^{-2} for contrails and contrail cirrus but did not fully quantify the uncertainty. They also found the contrails to reduce the background cirrus cloudiness in the main traffic areas and suggest this gives a negative forcing of roughly –0.007 W m^{-2}. Compounding their assessed errors from spreading rate, optical depth, ice particle shape and radiative transfer would give an error estimate of 67% in agreement with Markowicz and Witek (2011). As this study is a major step forward from previous work we employ their estimate for this assessment, and assess a combined contrail and contrail-induced cirrus AF to be 0.03 W m^{-2} with a 90% uncertainty range of 0.01 to 0.06 W m^{-2}. The upper bound for this forcing mechanism is not well constrained however.

7.2.6.2 Irrigation-Induced Cloudiness

Boucher et al. (2004) estimated a RF due to water vapour from irrigation in the range of 0.03 to 0.10 W m^{-2} but the net climate effect was dominated by the evaporative cooling at the surface and by atmospheric thermal responses to low-level humidification. Regional surface cooling was confirmed by a number of more recent regional and global studies (Kueppers et al., 2007; Lobell et al., 2009). It was also found that the resulting increase in water vapour could induce a small enhancement in precipitation downwind of the major

irrigation areas (Puma and Cook, 2010), as well as some regional circulation patterns (Kueppers et al., 2007). Sacks et al. (2009) reported a 0.001 increase in cloud fraction over land (0.002 over irrigated land). This points to an AF no more negative than –0.1 W m^{-2}.

7.3 Aerosols

7.3.1 Introduction

This section assesses the role of aerosols in the current climate system, focusing on anthropogenic changes in aerosols and their effects. In particular it covers the direct radiative forcing of aerosols, their effects on atmospheric heating and snow/ice surfaces, as well as Earth system feedbacks involving natural and anthropogenic aerosols. Cloud microphysical effects of aerosols are discussed in Section 7.4 although some of the relevant properties of aerosols, such as cloud condensation nuclei (CCN) and ice nuclei (IN), are documented here. The time evolution of aerosols and their forcing are discussed in Chapters 2 and 8, with Chapter 8 also covering changes in natural volcanic aerosols.

The Section covers material previously assessed in the AR4 of the IPCC (IPCC, 2007). Chapter 2 (Forster et al., 2007) assessed the total direct RF from aerosol to be –0.5 ± 0.4 W m^{-2} and broke this down into components associated with several species. Land albedo changes associated with black carbon (BC) on snow were assessed to be +0.1 ± 0.1W m^{-2}. The semi-direct effect and its contribution to the AF were discussed but not explicitly quantified in Chapter 7 (Denman et al., 2007). Chapter 7 also gave a preliminary assessment of possible earth system feedbacks involving aerosols. The RF uncertainty estimate in AR4 was based on the range of model results and remote observations.

Since AR4 in-situ and remote observations of aerosol have improved and global aerosol models have become considerably more complex and evaluated against observations. Earth system models are continuing to be developed and some have been used in CMIP5, which allows the investigation of possible biogeochemical aerosol feedbacks on climate. To help improve the forward assessment of uncertainty in aerosol forcing this chapter discusses aerosol science in more detail than AR4, from both an observational and a modelling perspective.

Radiative forcing estimates rely on knowledge of aerosol emissions and aerosol properties. Aerosol sources and properties are discussed in Section 7.3.2 while Section 7.3.3 assesses the key climate-relevant aerosol properties and Section 7.3.4 discusses aerosol distributions. Section 7.3.5 covers the direct and semi-direct effects of aerosols, which provides the basis for evaluating the anthropogenic component to that effect, in terms of RF and AF; this section also assesses the forcing contribution that results from aerosol-induced changes to the surface properties of snow and ice. Finally Section 7.3.6 updates our understanding of potential aerosol-climate feedbacks.

7.3.2 Aerosol Sources and Processes

Atmospheric aerosol particles, whether natural or anthropogenic, originate from two distinctively different pathways: direct emissions of primary particles and secondary aerosol formation from gaseous precursors (Figure 7.7). Secondary aerosol formation is initiated by gas phase chemistry that produces a large number of organic and inorganic compounds of different level of oxidation and volatility. Some of these compounds form secondary particulate matter by condensing onto or reacting with pre-existing aerosol particles or cloud droplets. A small fraction of gaseous compounds is capable of producing new aerosol particles by nucleation. Both primary and secondary particles grow in size in the atmosphere by condensation, coagulation and cloud processing. These processes also affect the aerosol chemical composition, size, shape and mixing state. Aerosols are removed from the atmosphere through dry deposition at the surface and wet deposition (including in-cloud and below-cloud rainout).

[INSERT FIGURE 7.7 HERE]

Figure 7.7: Overview of atmospheric aerosol processes and meteorological variables influencing the aerosol semi-direct, direct and indirect aerosol effects. Red designates gas phase processes and variables; blue designates particulate (aerosol) phase processes and variables; processes and variables relevant to the aerosol direct and semi-direct effects appear in black, while those relevant to the aerosol indirect effects appear in green.

The main chemical constituents of the atmospheric aerosol are inorganic species (such as sulphate, nitrate, sea-salt), organic species (also termed organic carbon or organic aerosol), black carbon (BC), and mineral species (mostly desert dust). BC, sea salt and dust are introduced into the atmosphere as primary particles, whereas sulphate and nitrate are formed almost entirely in the atmosphere by secondary aerosol formation processes. Organic aerosol (OA) has both primary (POA) and secondary (SOA) sources. The majority of BC, sulphate and nitrate come from anthropogenic sources, whereas sea salt and most of dust is of natural origin. Atmospheric POA is likely to be dominated by anthropogenic sources, whereas SOA is to a larger extent of natural origin (Farina et al., 2010; Hallquist et al., 2009; Kroll and Seinfeld, 2008). Despite earlier recognition of their climatic importance (Adams et al., 2001), it is only recent that nitrate aerosols are being considered in a wider set of models. Emissions of aerosols and aerosol precursors are summarized in Table 7.1.

[INSERT TABLE 7.1 HERE]

Table 7.1: Global and regional anthropogenic emissions important for aerosol formation and tropospheric chemistry. The maximum and minimum values from available inventories are presented. Units for NO_x are Tg NO yr^{-1}, other units are Tg yr^{-1}. Adapted from Granier et al. (2011).

The climate effects of atmospheric aerosol particles depend on their atmospheric distribution, along with their optical properties and ability to act as CCN or IN. Key quantities for aerosol optical and cloud forming properties are the particle number size distribution, chemical composition, mixing state and shape. These properties are determined by a complex interplay between their sources, atmospheric transformation processes and their removal from the atmosphere (Figure 7.7).

Sea salt particles are produced at the sea surface by breaking waves (de Leeuw et al., 2011). Emissions of sea salt particles depend mainly on the surface wind speed, and to a lesser extent on the temperature and composition of the sea water; however the effective flux in the atmosphere also depends on atmospheric stability. Since AR4, substantial progress has been made in constraining the total mass, number size distribution and chemical composition of emitted sea salt particles (Evan et al., 2008; Fuentes et al., 2011; Keene et al., 2007; Ovadnevaite et al., 2011). A major new observation has been the frequent presence of organic material in submicron sea salt particles, especially at smallest particle sizes. Process-based estimates of sea salt emissions continue nevertheless to have a significant uncertainty (de Leeuw et al., 2011). These uncertainties in the chemical composition and size of emitted sea-salt particles translate into a large uncertainty on the natural level of CCN in the marine atmosphere, which unlike the aerosol optical depth cannot be constrained from space observations.

Dust particles are produced by disintegration of aggregates following creeping and saltation of larger soil particles over desert and other arid surfaces (Kok, 2011; Zhao et al., 2006). The magnitude of dust emissions to the atmosphere depends on the surface wind speed and many soil-related factors such as its texture, moisture and vegetation cover. Since AR4, new dust emission schemes have been implemented in large-scale models and compared against surface measurements and observation by remote sensing (Cheng et al., 2008; Darmenova et al., 2009; Huneeus et al., 2011b). The range of estimates for the global dust emission span a factor of about 5 (Huneeus et al., 2011b).

Dimethylsulfide (DMS) emitted from the oceans is a major natural contributor to the atmospheric sulphate aerosol burden. Emissions of DMS to the atmosphere depend on its surface water concentration and its sea-to-air transfer velocity. Since AR4, progress has been made in understanding how DMS surface water concentrations are affected by solar radiation and atmospheric CO_2 concentration (Derevianko et al., 2009; Kim et al., 2010). Prognostic models for predicting the spatial and temporal variation of the DMS surface water concentration have been developed further and inter-compared for the first time (Le Clainche et al., 2010). New DMS air-to-sea transfer models have been developed and evaluated (Huebert et al., 2010; Vlahos and Monahan, 2009), and an updated climatology for surface ocean DMS concentrations and emission fluxes has been developed (Lana et al., 2011). The range of the current estimates of global DMS emissions span a factor of about 2. The majority of DMS emitted into the atmosphere is transformed to particulate sulphate via gas-phase chemistry and cloud processing (Karl et al., 2007). Sulphate produced from DMS has a major influence on marine CCN concentrations, but the relation between the DMS

emissions and CCN concentrations appears to be more complicated than previously thought (Korhonen et al., 2008; Thomas et al., 2010; Woodhouse et al., 2010).

Biogenic volatile organic compounds (BVOC) are globally the most important precursors for SOA. Emissions of BVOC to the atmosphere depend on the amount and type of vegetation, temperature, radiation and several environmental factors such as the ambient CO_2 concentration (Grote and Niinemets, 2008; Pacifico et al., 2009). While speciated BVOC emission inventories have been derived for some continental regions, global emission inventories or schemes are available only for isoprene and some monoterpenes (Guenther et al., 2006; Muller et al., 2008). The total global BVOC emissions have large uncertainties, despite the apparent convergence in different model-based estimates (Arneth et al., 2008). In the atmosphere, BVOCs are transformed to SOA with variable yields via gas-phase chemistry followed by gas-particle partitioning, and via other ageing mechanisms. The global SOA production resulting from BVOCs has been constrained substantially since AR4, the recent estimates being mostly within a factor 3 (Farina et al., 2010; Hallquist et al., 2009; Heald et al., 2010; Spracklen et al., 2011).

Formation of new particles with 1–2 nm diameters by nucleation and their initial growth to detectable sizes is called new particle formation (NPF). Atmospheric NPF has been observed to take place in a large number of different environments (Kulmala and Kerminen 2008; Manninen et al., 2010; O'Dowd et al., 2010) and substantial progress in our understanding has been made since AR4. Multiple lines of evidence indicate that sulfuric acid plays a central role in atmospheric NPF (Kerminen et al., 2010; Sipila et al., 2010), with observed particle formation rates scaling mostly to the power 1 to 2 of the gaseous sulphuric acid concentration (Kuang et al., 2008; Paasonen et al., 2010). Other compounds capable of influencing NPF include low-volatile organic vapors (Metzger et al., 2010; Paasonen et al., 2010; Wang et al., 2010b), ammonia and amines (Benson et al., 2011; Berndt et al., 2010; Kirkby et al., 2011; Kurten et al., 2008; Smith et al., 2010). Ion-induced nucleation is very likely to contribute to NPF throughout the atmosphere (Kirkby et al., 2011; Yu 2010), but NPF seems to be dominated by neutral nucleation pathways in continental boundary layers (Hirsikko et al., 2011; Kazil et al., 2010). NPF is not thought to affect directly the mass burden or direct RF of atmospheric aerosols. Model studies suggest, however, that NPF is a dominant source of the particle number concentration in the global atmosphere (Spracklen et al., 2006; Spracklen et al., 2010; Yu et al., 2010a) and a potentially significant contributor to global CCN (Kazil et al., 2010; Merikanto et al., 2009; Pierce and Adams 2009b). The growth of newly-formed particles to CCN sizes depends crucially on organic vapors (Riipinen et al., 2011), and is therefore strongly tied with atmospheric SOA formation.

Aerosol sinks are generally better understood than aerosol sources, and there has been correspondingly less progress in understanding and modeling since AR4. Improved dry deposition models, which depend the particle size as well as the roughness properties of the surface, have been parameterized and are increasingly being used in global aerosol models (Feng 2008; Kerkweg et al., 2006; Petroff and Zhang 2010). For the largest particles in the coarse mode, it is important to consider sedimentation throughout the atmosphere and its role in dry deposition at the surface. Aerosol wet deposition (in-cloud and below-cloud) is affected by the size and chemical composition of particles. For insoluble primary aerosol species like BC and dust, wet deposition depends also strongly on their degree of mixing with soluble species. Wet deposition of aerosols remain a key source of uncertainty in aerosol models which affects the vertical distribution and long-range transport of aerosols (Lee et al., 2011; Vignati et al., 2010). In particular, many models treat convective transport and scavenging sequentially rather than as a coupled process.

7.3.3 Progresses and Gaps in Understanding Climate Relevant Aerosol Properties

Since AR4, understanding some of the key aerosols properties has been greatly improved by laboratory and field experiments using advanced measurement technologies and theoretical studies. These experimental studies have stimulated improvement in the model representations of the aerosol physical, chemical and optical properties, and their role as cloud and ice nuclei (Figure 7.7; Table 7.2). Some of the field experiments have also been used to evaluate aerosol modules used in regional and global scale models. We focus our assessment on some of the key issues where there has been progress since AR4, focusing on BC, organic aerosols, aerosol size distribution and mixing state.

[INSERT TABLE 7.2 HERE]

Table 7.2: Key aerosol properties of the main aerosol species in the troposphere. Brown carbon is a particular type of OA but is treated here as an additional component because it is light absorbing. The estimate of aerosol burdens and lifetimes in the troposphere are based on the AeroCom models.

7.3.3.1 Chemical Composition

The gas-aerosol partitioning of semi-volatile inorganic species (H_2SO_4, HNO_3, HCl, and NH_3) is governed by gas-particle mass transfer kinetics, equilibrium thermodynamics, and heterogeneous chemical reactions. These processes are determined by meteorological conditions, such as temperature and relative humidity, and chemical conditions, such as gas-phase concentrations, aerosol size distribution and phase, and pH of particles. Aerosol models are increasingly calculating the gas-aerosol partitioning by assuming thermodynamic equilibrium (e.g., Nenes et al., 1998; Wexler and Clegg, 2002) or by solving gas-particle mass transfer dynamically (e.g., Zaveri et al., 2008; Zhang et al., 2004). Although a number of climate and global models assume thermodynamic equilibrium for semi-volatile inorganic species (or do not include semi-volatile inorganic species), the equilibrium assumption is not always valid even under typical atmospheric conditions.

BC is a distinct type of carbonaceous material, formed only near flames during combustion of carbon-based fuels. It strongly absorbs visible light and it is refractory. These physical properties allow at least in principle for a strict definition of BC. Direct measurement of individual BC-containing particles is possible with a single-particle soot photometer (SP2) based on laser-induced incandescence (Gao et al., 2007; Moteki and Kondo, 2010; Moteki et al., 2009; Schwarz et al., 2008; Schwarz et al., 2006). This technique has enabled accurate measurements of the size of BC cores (mean number diameters of 50–80 nm for fossil-fuel sources and 120 nm for biomass burning). However spatial and temporal coverage of the measurements are still limited (Schwarz et al., 2010).

Condensation of gas-phase species on BC and coagulation with other particles alter the mixing state of BC (e.g., Adachi et al., 2010; Li et al., 2003; Pósfai et al., 2003). In polluted urban air, BC becomes internally mixed on a timescale of ~12 h (McMeeking et al., 2010; Moteki et al., 2007). The resulting BC-containing particles become more hydrophilic which can lead to a reduced lifetime and atmospheric loading (Stier et al., 2006b). In addition, internal mixing can enhance the BC mass absorption efficiency by up to a factor of 2 (Bond and Bergstrom, 2006; Cross et al., 2010), with typical values of about 8–20 $m^2 g^{-1}$ at a wavelength of 532 nm.

Long-range transport of BC has been observed in some conditions, e.g., from biomass burning in Russia to the Arctic in spring (Kondo et al., 2011; Warneke et al., 2009; Warneke et al., 2010), or in northeastern China in spring when humidity was relatively low (Oshima et al., 2011). However it is thought that rapid coating of BC by soluble material reduce the effectiveness of the vertical transport by convection.

Formation processes of OA remain highly uncertain, which is a major weakness in the present understanding of atmospheric aerosol (Hallquist et al., 2009; Kanakidou et al., 2005). Measurements by Aerosol Mass Spectrometers have provided some insights into sources of OA (Lanz et al., 2007; Ulbrich et al., 2009; Zhang et al., 2005b). Measurements at continental midlatitudes including urban and rural/remote air suggest that majority of SOA is likely to be oxygenated OA (Zhang et al., 2005a; Zhang et al., 2007). Measurements within and downstream of urban air indicate that under most circumstances SOA substantially contributes to the total OA mass (de Gouw et al., 2005; Volkamer et al., 2006; Zhang et al., 2007).

The contribution of secondary organic aerosol (SOA) to the total OA is larger than previously thought, but the split between primary organic aerosols (POA) and SOA has remained somewhat ambiguous (Jimenez et al., 2009; Robinson et al., 2007). POA had been assumed to be completely non-volatile. However, it has been found that a certain amount of POA may evaporate, followed by gas-phase photochemical oxidation of the vapors, to produce even less volatile SOA (Robinson et al., 2007), as shown in Figure 7.7.

Some of OA is light absorbing (brown carbon; BrC), which is produced mainly by lower temperature combustion, such as biomass burning (Andreae and Gelencser 2006). The absorption properties of BrC can be attributed to water soluble organic compounds and humic-like substances (Graber and Rudich 2006; Kirchstetter et al., 2004), although they are poorly quantified (Alexander et al., 2008; Flowers et al., 2010).

There is a large range in the complexity with which OA are represented in global aerosol models. Some complex, yet still parameterized, chemical schemes have been developed recently which account for multigenerational oxidation (Donahue et al., 2011; Jimenez et al., 2009; Robinson et al., 2007). Some regional and global models use semi-empirical schemes, where semi- or non-volatile organic compounds (SVOC) are produced from parent VOCs by oxidation processes and partitioned between the aerosol and gas phases (Fan et al., 2005; Heald et al., 2005; Russell and Allen, 2005; Tsigaridis and Kanakidou, 2003). The representation of secondary organic aerosols in many of the models used in CMIP5 is either non-existent or very crude in that the source terms are prescribed and/or the models ignore the complex chemical and aging processes.

7.3.3.2 Mixing State

There are multiple observations that show aerosols to be internally mixed (that is with multiple materials in individual particles) relatively soon after emission. In general, organic and black carbon, in biomass burning aerosols were found frequently internally mixed with ammonium, nitrate, and sulphate (Deboudt et al., 2010; Pratt and Prather 2010). Studies over urban locations revealed that as much as 90% of the particles are internally mixed with secondary inorganic species (Bi et al., 2011). Likewise there is evidence of internal mixing between dust and biomass burning aerosols when these aerosol types age together (Hand et al., 2010). Studies have shown that state of mixing can alter particle hygroscopicity and hence their ability to act as CCN (Wex et al., 2010).

A common form of mixing is coating of soluble material (secondary aerosol) over a primary aerosol such as a BC or dust core. The solid to liquid transition of such mixed aerosols as the RH increases influences their light scattering properties through changes in particle shape, size, and refractive index (Freney et al., 2010). Another form of coating involves fine BC contaminating large dust particles. The optical properties of an internal mixture are known to be significantly different from those of an external mixture (Garcia et al., 2011; Jacobson, 2001; Shiraiwa et al., 2008) with an increase in the mass absorption efficiency.

Global aerosol models are increasingly treating aerosols as an internal mixture, which enables a consistent treatment of aerosol hygroscopicity, scavenging and optical properties (i.e., Stier et al., 2006a). Commonly-used modal approaches do not allow to represent discrete variations of mixing state across an aerosol mode, though.

7.3.3.3 Aerosol Size Distribution

Aerosol size distribution is one of the key characteristics required for estimating the spectral optical and CCN properties of aerosols. Though ground-based measurements of aerosol size distribution are being carried out as part of various field campaigns (Haywood et al., 2011), such information is still sparse and there have been only few attempts to build a climatology of aerosol size distribution (e.g., Heintzenberg et al., 2000). There have been improvements in the algorithms to derive column-averaged volume size distribution from sun photometer measurements (Dubovik et al., 2006). Validation studies show agreement for the derived size distribution against in situ (surface as well as aircraft-based) measurements (Gerasopoulos et al., 2007; Haywood et al., 2011; Radhi et al., 2010; Smirnov et al., 2011), but these inversion products have not been comprehensively validated. There has also been some progress in measuring the aerosol size distribution down to a few nanometre size (e.g., Feldpausch et al., 2006) which revealed large Aitken mode concentrations in the tropical upper troposphere.

Satellite sensors such as MODIS and POLDER also provides the Angström coefficient and AOD fine mode fraction (i.e., contributed by those aerosols with diameter less than about 1 μm). Even though aerosol fine mode fraction has been validated over ocean (Remer et al., 2005), accurate determination of fine mode fraction over land still remains a challenge (Jethva et al., 2009; Levy et al., 2007).

Measurements have been used to evaluate the ability of the new generation of global aerosol models to simulate aerosol number concentration and the aerosol size distribution (Bauer et al., 2008; Liu et al., 2005b; Spracklen et al., 2005a, 2005b; Spracklen et al., 2010; Stier et al., 2005; Wang et al., 2009; Yu and Luo, 2009; Zhang et al., 2010b). Such evaluations place powerful constraints on the representation of new particle

formation in the models, which is a major source of uncertainty in estimates of the aerosol indirect effects (see Section 7.4). Simulated aerosol number concentrations at the surface are usually found to be too low unless empirical particle nucleation schemes are used. Models are also hampered by the lack of size-resolved inventories of aerosol emissions.

7.3.3.4 CCN Properties of Aerosols

A proportion of aerosol acts as CCN. The ability of an aerosol particle to take up water and subsequently activate is determined by its size and composition. Common CCN in the atmosphere are composed of sea salt, sulphates and sulphuric acid, nitrate and some organics (Table 7.2). The size of the CCN was found to be more important than their chemical composition at one continental location as larger particles are more readily activated than smaller particles because they require a lower critical supersaturation (Dusek et al., 2006). However, the chemical composition influences the aerosol size distribution and may be important in other locations. The CCN activity of aerosols can be characterised by a hygroscopicity parameter κ that can be derived from measurements (Petters and Kreidenweis, 2007).

CCN activity of inorganic aerosols is relatively well understood. Lately most attention has been paid to the CCN activity of mixed organic/inorganic aerosols (e.g., King et al., 2010; Prisle et al., 2010). Uncertainties in our current understanding of CCN properties are associated with SOA (Good et al., 2010), mainly because OA are still poorly characterized (Jimenez et al., 2009). For SOA it is not clear how important surface tension effects and bulk to surface partitioning are and if the water activity coefficient changes significantly as a function of the solute concentration (Good et al., 2010; Prisle et al., 2008). Thus, when the aerosol is dominated by organics, discrepancies between values of κ obtained directly from both CCN activity measurements and sub-saturated particle water uptake measurements have been observed in some instances (e.g., Irwin et al., 2010; Prenni et al., 2007; Roberts et al., 2010), whereas in other studies closure could be obtained (e.g., Duplissy et al., 2008; Rose et al., 2011).

Pringle et al. (2010) used surface and aircraft measurements to evaluate the values of the hygroscopicity parameter simulated by a global aerosol model, and found generally good agreement. Spracklen et al. (2008) used surface and aircraft measurements to evaluate simulated distributions of CCN concentration, and found good agreement between simulated and observed concentrations.

7.3.3.5 IN Properties of Aerosols

Aerosols that act as IN are solid substances at atmospheric temperatures and supersaturations. Mineral dust and primary biological particles are typically known as good IN. Conflicting evidence has been presented for the ability of BC, organic and biomass burning particles to act as IN. Four heterogeneous ice nucleation modes are distinguished in the literature. Immersion freezing refers to freezing that is initiated from within a cloud droplet, condensation freezing refers to freezing during droplet formation, and contact freezing is initiated when an IN collides with a supercooled cloud droplet. Deposition nucleation refers to the direct deposition of vapour onto IN. Lidar observations revealed that liquid cloud droplets are present before ice crystals form via heterogeneous freezing mechanisms (Ansmann et al., 2008; de Boer et al., 2011) indicating that deposition nucleation does not seem to be important for mixed-phase clouds.

A compilation of onset temperatures and relative humidities for deposition/condensation nucleation versus immersion freezing for bacteria, mineral dust and soot based on recent data from laboratory data is shown in Figure 7.8. It should be noted that the reported nucleation onset points for the different materials differ with respect to the fraction of ice nucleating particles (e.g., due to the detection thresholds of measurement methods). Bacteria initiate immersion freezing at the highest temperatures, followed by mineral dust and then soot. It is obvious that for dust and bacteria, heterogeneous ice nucleation occurs at considerably higher temperatures than homogeneous freezing. Remote sensing observations confirm that ice clouds in air containing dust can be found at significant warmer temperatures than in dust-free conditions (Choi et al., 2010; Sassen et al., 2003; Seifert et al., 2010). Laboratory results indicate that, in comparison with natural IN such as mineral dust and biological particles, soot initiates ice at the coldest temperatures. Deposition nucleation of ice on most types of soot particles is not important above –30ºC and below water saturation (Dymarska et al., 2006; Friedman et al., 2011). However, in-situ observations indicate an enrichment of BC

in atmospheric ice particle residuals in tropospheric mixed phase clouds (Cozic et al., 2008; Targino et al., 2009; Twohy et al., 2010), thus there must be some mechanism for BC to enter ice clouds.

Lately, a variety of other substances such as crystalline ammonium sulphate at cirrus temperatures (e.g., Abbatt et al., 2006; Baustian et al., 2010; Wise et al., 2009), oxalic acid (Wagner et al., 2010, 2011; Zobrist et al., 2006), marine diatoms (Knopf et al., 2011) as well as fulvic and humic acid (Wang and Knopf, 2011) have been identified as possible IN. Biological particles in significant concentrations have also been observed in residues from ice crystals (Pratt et al., 2009; Prenni et al., 2009).

IN can either be bare or mixed with other substances. As bare particles age in the atmosphere, they acquire liquid surface coatings by condensing soluble species and water vapor or by scavenging soluble particles, which may transform IN from deposition or contact nuclei into possible immersion nuclei. This transformation or the chemical processing with ozone or other oxidizing substances may dampen the ice-forming ability of some IN types (Chernoff and Bertram, 2010; Cziczo et al., 2009b; DeMott et al., 2009; Eastwood et al., 2009; Mohler et al., 2005; Sullivan et al., 2010b; Wang and Knopf, 2011). IN have been observed to be less efficient after chemical aging with nitric acid (Sullivan et al., 2010a) with ammonia gas (Salam et al., 2007) and of BC particles with aqueous sulphuric acid for nucleation in cirrus clouds (DeMott et al., 1999; Koehler et al., 2009).

[INSERT FIGURE 7.8 HERE]

Figure 7.8: The onset temperatures and relative humidities for deposition/condensation freezing and immersion freezing for bioaerosols (Ahern et al., 2007; Diehl et al., 2001; Iannone et al., 2011; Kanji et al., 2011; Mohler et al., 2008; Mortazavi et al., 2008; von Blohn et al., 2005; Yankofsky et al., 1981), mineral dusts (Archuleta et al., 2005; Bundke et al., 2008; Connolly et al., 2009; Cziczo et al., 2009a; Field et al., 2006; Kanji and Abbatt 2006; Kanji et al., 2011; Knopf and Koop 2006; Koehler et al., 2010; Kulkarni and Dobbie 2010; Lüönd et al., 2010; Mohler et al., 2006; Murray et al., 2011; Niedermeier et al., 2010; Niemand et al., 2011; Roberts and Hallett 1968; Salam et al., 2006; Schaller and Fukuta 1979; Welti et al., 2009; Zimmermann et al., 2008), organics (Baustian et al., 2010; Kanji et al., 2008; Petters et al., 2009; Prenni et al., 2007; Shilling et al., 2006; Wagner et al., 2010, 2011; Wang and Knopf 2011; Zobrist et al., 2007), solid ammonium sulphate (Abbatt et al., 2006; Baustian et al., 2010; Mangold et al., 2005; Shilling et al., 2006; Wise et al., 2009; 2010) and BC (soot) (Crawford et al., 2011; DeMott 1990; DeMott et al., 1999; Diehl and Mitra 1998; Dymarska et al., 2006; Fornea et al., 2009; Gorbunov et al., 2001; Kanji et al., 2011; Mohler et al., 2005), from a compilation of experimental data of sub- and super-micron aerosol particles in the literature (for references see supplementary material). The large range of observed ice nucleation onset conditions is due to different experimental setups, particle sizes, activated fractions and chemical composition. Only those IN species for which at least three papers exists are shown. The dashed line refers to the homogeneous freezing of solution droplets after (Koop et al., 2000).

7.3.4 Aerosol Distributions

This section assesses our current understanding of aerosol distributions from in-situ and remote-sensing measurements, and the current ability of global aerosol models to simulate such distributions. Aerosol predictions in the CMIP5 models are assessed in Chapter 12. Since the AR4, new and improved observational tools for global aerosol model evaluation have emerged.

The most commonly used in-situ measurements for global aerosol model evaluation are mass concentrations and deposition fluxes. Long-term aerosol mass concentrations have been measured at the surface by global and local networks, as well as individual efforts with improved coordination and quality assurance. Climatologies of the main aerosol types can be constructed from such measurements (e.g., Jimenez et al., 2009; Figure 7.9). They show a wide spatial variability in the dominant aerosol type and aerosol concentrations worldwide. Mineral aerosol is the largest aerosol component in most areas with higher concentrations in Urban S. Asia and China, accounting ~ 35% of PM10. Aerosol fraction in rural U.S. and South America are composed mainly (i.e., ~20%) of OC, and the OC fractions also rank second or third with a mean of ~16% in other areas of the world. Sulfate normally accounts for ~10–30%, except for the areas in rural Africa, high Asia, urban Oceania and South America with less than ~5%. The fractions of nitrate and ammonium are only around 6% and 4%, in average respectively. In most areas, BC fractions are less than 5%, although this percentage may be larger in South America, urban Africa, urban Europe, South-East and East Asia and urban Oceania. Sea-salt aerosols can be dominant at oceanic sites. Measurements of dry and wet deposition on the surface have also been made (e.g., Hjellbrekke, 2001; McConnell et al., 2007) and

used to evaluate global aerosol models (e.g., Boucher and Pham, 2002; Easter et al., 2004; Flanner et al., 2007; Huneeus et al., 2011b).

[INSERT FIGURE 7.9 HERE]

Figure 7.9: Bar chart plots summarizing the annual, seasonal or monthly mean mass concentration (μg m^{-3}) of six major types of aerosol particles in diameter smaller than 10 μm with at least an entire year data from various rural and urban sites in nine continental areas of the world. These include: 1) rural U. S. (Chow et al., 1993; Liu et al., 2005a; Malm and Schichtel 2004; Malm et al., 1994); urban U. S. (Chow et al., 1993; Ito et al., 2004; Kim et al., 2000; Liu et al., 2005a; Malm and Schichtel 2004; Sawant et al., 2004); 2) South America (Artaxo et al., 1998; Artaxo et al., 2002; Bourotte et al., 2007; Celis et al., 2004; Fuzzi et al., 2007; Gioda et al., 2011; Mariani and Mello 2007; Martin et al., 2010; Morales et al., 1998; Souza et al., 2010); 3) rural Europe (Gullu et al., 2000; Hueglin et al., 2005; Kocak et al., 2007; Putaud et al., 2004; Puxbaum et al., 2004; Querol et al., 2001; Querol et al., 2009; Querol et al., 2004; Rodriguez et al., 2002; Rodrıguez et al., 2004; Salvador et al., 2007; Theodosi et al., 2010; Viana et al., 2008; Yin and Harrison 2008; Yttri 2007); urban Europe (Hueglin et al., 2005; Lenschow et al., 2001; Lodhi et al., 2009; Lonati et al., 2005; Perez et al., 2008; Putaud et al., 2004; Querol et al., 2001; Querol et al., 2006; Querol et al., 2004; Querol et al., 2008; Rodriguez et al., 2002; Rodrıguez et al., 2004; Roosli et al., 2001; Viana et al., 2007; Viana et al., 2006; Yin and Harrison 2008); 4) rural Africa (Maenhaut et al., 1996; Mkoma 2008; Mkoma et al., 2009a; Mkoma et al., 2009b; Nyanganyura et al., 2007; Weinstein et al., 2010); urban Africa (Favez et al., 2008; Mkoma 2008; Mkoma et al., 2009a); 5) high Asia, with altitude larger than 1680 m. (Carrico et al., 2003; Decesari et al., 2010; Ming et al., 2007a; Qu et al., 2008; Ram et al., 2010; Rastogi and Sarin 2005; Rengarajan et al., 2007; Shresth et al., 2000; Zhang et al., 2001; Zhang et al., 2008; Zhang et al., 2011a); 6) rural China (Hagler et al., 2006; Hu et al., 2002; Zhang et al., 2011a); urban China (Cheng et al., 2000; Hagler et al., 2006; Oanh et al., 2006; Wang et al., 2003; Wang et al., 2005b; Wang et al., 2006; Xiao and Liu 2004; Yao et al., 2002; Ye et al., 2003; Zhang et al., 2002; Zhang et al., 2011a; Zhang et al., 2011b); 7) South-East and East Asia (Han et al., 2008; Khan et al., 2010; Kim et al., 2007; Lee and Kang 2001; Oanh et al., 2006); 8) urban South Asia (Chakraborty and Gupta 2010; Khare and Baruah 2010; Kumar et al., 2007; Lodhi et al., 2009; Raman et al., 2010; Rastogi and Sarin 2005; Safai et al., 2010; Stone et al., 2010); 9) urban Oceania (Chan et al., 1997; Maenhaut et al., 2000; Radhi et al., 2010; Wang and Shooter 2001; Wang et al., 2005a).

Measurements of aerosol optical depth (AOD) are retrieved at the surface by AERONET (e.g., Holben et al., 1998; Holben et al., 2001), other ground-based networks and an increasing number of satellite missions. Retrievals from aerosol-dedicated instruments such as MODIS (Kleidman et al., 2011; Levy et al., 2010; Remer et al., 2005), MISR (Kahn et al., 2005; Kahn et al., 2007) and POLDER/PARASOL (Tanré et al., 2011). Other instruments such as AVHRR (Geogdzhayev et al., 2002; Jeong and Li 2005), TOMS (Torres et al., 1998; Torres et al., 2002) and ATSR/AATSR have longer measurements records. Composite aerosol datasets have also been developed (see Chapter 2). While each aerosol retrieval shows some skill against AERONET measurements, there are still large differences in regional and seasonal patterns because of differences in sampling, cloud screening and treatment of the surface reflectivity (Kokhanovsky et al., 2010).

Field experiments involving research aircraft such as TRACE-P (Jacob et al., 2003), INTEX-A (Singh et al., 2006), ARCTAS (Jacob et al., 2010), ARCPAC (Warneke et al., 2010) and HIPPO1 (Schwarz et al., 2010; Schwarz et al., 2006), commercial aircraft (Brenninkmeijer et al., 2007) and spaceborne lidars (Omar et al., 2009; Winker et al., 2009) can provide measurements of the aerosol vertical profile. For example, Koch et al. (2009b) and Schwarz et al. (2010) used BC measurements by an SP2 instrument on aircraft to evaluate AeroCom model simulations of the vertical distribution of BC aerosol in many regions, and found that most models simulate too much BC in the upper troposphere (Schwarz et al., 2010, see Figure 7.10). Koch et al. (2009b) also used AERONET retrievals of aerosol absorption optical depth (AAOD) to show that most AeroCom models underestimate AAOD in many regions. Yu et al. (2010b) and Koffi et al. (2011) found that global aerosol models tend to have a positive bias on the aerosol extinction scale height in some (but not all) regions resulting in an overestimate of aerosol concentrations above 6 km.

[INSERT FIGURE 7.10 HERE]

Figure 7.10: Comparison of BC profiles as measured during the ARCTAS, HIPPO1 and FORCE-A campaigns and simulated by a range of global aerosol models. [PLACEHOLDER FOR SECOND ORDER DRAFT: will be updated from AeroCom and CMIP5 models]

Overall aerosol measurements have been widely used in the evaluation of aerosol models both within the AeroCom activity (e.g., Huneeus et al., 2011b; Kinne et al., 2006; Koch et al., 2009b) and elsewhere (e.g., Mann et al., 2010; Myhre et al., 2007; Wang et al., 2009). This has contributed to continuous model improvement since AR4 but models are still imperfect [PLACEHOLDER FOR SECOND ORDER DRAFT:

AeroCom reference]. In a few studies the measurements are also used directly to improve aerosol distributions through data assimilation (e.g., Benedetti et al., 2009; Huneeus et al., 2011a). The lack of an observational constraint on the pre-industrial aerosol distribution is a continuous source of uncertainty when estimating aerosol radiative forcing.

7.3.5 Aerosol Radiative Effects

7.3.5.1 Direct Radiative Effect

Direct radiative effect (DRE) is the change in radiative flux caused by the combined direct effect of all anthropogenic and natural aerosols. The physics behind the DRE is robust and the DRE is close to being an observable quantity, yet our knowledge of aerosol and environmental characteristics needed to quantify the DRE at the global scale remains incomplete (Anderson et al., 2005; Jaegle et al., 2011; Satheesh and Moorthy 2005). The DRE of aerosols depends on the distribution of their optical properties and the scattering and absorbing properties of the environment (i.e., surface, atmospheric molecules and clouds) as illustrated on Figure 7.7. More specifically the DRE requires the knowledge of the spectral variations in aerosol extinction coefficient, single scattering albedo and phase function, which can in principle be estimated from the aerosol size distribution, shape, chemical composition and the state of mixing. In the solar spectrum, the DRE is typically negative at the top-of-atmosphere but gets weaker and can become positive with increasing aerosol absorption, decreasing upscatter fraction, or increasing albedo of the underlying surface. DRE is largest in cloud-free conditions, but can be significant in the presence of a thin cloud layer. Top-of-atmosphere DRE can become positive when the aerosols are located above clouds (e.g., Chand et al., 2009). The DRE at the surface is almost always negative and its magnitude can be much larger than the DRE at top-of-atmosphere when there is aerosol absorption (Li et al., 2010). In the longwave spectrum, the DRE is generally positive at the top-of-atmosphere but is only significant for desert dust and sea-salt (Reddy et al., 2005).

There have been many measurement-based estimates of the DRE (Bauer et al., 2011; Bergamo et al., 2008; Di Biagio et al., 2010; Yu et al., 2006) although some studies involve some degree of modelling. There is generally a good agreement between observed and calculated shortwave radiative fluxes when aerosol properties are known (e.g., Osborne et al., 2011). Global observational estimates of the DRE rely on satellite remote sensing of aerosol properties and/or measurements of the Earth's radiative budget (Chen et al., 2011; Haywood et al., 2011). The shortwave clear-sky top-of-atmosphere DRE over the ocean is estimated to be in a range from –4 to –6 W m^{-2}, with sea-salt being the main contributor (Bellouin et al., 2005; Loeb and Manalo-Smith 2005; Myhre et al., 2007; Yu et al., 2006). Uncertainties are larger over land because satellite retrievals are more difficult and the surface is less well characterised (Chen et al., 2009; Jethva et al., 2009) despite recent progress in inversion algorithms (e.g., Dubovik et al., 2011). Attempts to estimate the DRE in cloudy-sky remain elusive (e.g., Peters et al., 2011) although remote sensing of aerosols over clouds is now possible from passive (Waquet et al., 2009) and active (Omar et al., 2009) methods. Notable areas of positive DRE include the Arctic, where absorbing fossil-fuel aerosols overlie ice surfaces (Stone et al., 2008), and off the shore of Namibia, where absorbing biomass-burning aerosols seasonally overlie stratocumulus clouds. While AOD and aerosols size are relatively well constrained, uncertainties in the aerosol single-scattering albedo (Loeb and Su, 2010) and vertical profile (e.g., Zarzycki and Bond, 2010) contribute significantly to the overall uncertainties in DRE, especially for all-sky estimates.

7.3.5.2 Aerosol Semi-Direct Effect and its Impact on Precipitation

BC, dust and other absorbing aerosols perturb the temperature structure of the atmosphere through radiative heating, modify the surface fluxes, and thus can influence cloud cover, precipitation and atmospheric dynamics. This effect, known as the semi-direct effect, can be considered as a rapid response associated to the direct effect and as such it can be accounted for through the concept of AF introduced in Chapters 1 and 8. Denman et al. (2007) assessed the semi-direct aerosol forcing to be small and of indeterminate sign and attached a very low understanding to it. This very low understanding resulted from the differences between cloud resolving and global model studies.

Since AR4 the observational evidence for the semi-direct effect has strengthened. There are additional observations of variations in cloud cover correlating with variations in the amount of absorbing aerosols

(e.g., Brioude et al., 2009; Wilcox, 2010) for which the semi-direct effect has been offered as an explanation. Such a causal effect between aerosol atmospheric heating (and surface cooling) and cloudiness has been confirmed by modelling studies on eddy-resolving, regional and global scale, which have revealed a more complicated picture than initially anticipated (e.g., Hill and Dobbie, 2008; Sakaeda et al., 2011; Zhuang et al., 2010).

Absorbing aerosol modifies atmospheric stability; the effect of this on cloud cover depends on the height of the aerosol relative to the clouds (Allen and Sherwood, 2010; Yoshimori and Broccoli, 2008). Both scattering and absorbing aerosols are very effective at reducing the downwelling solar radiation at the surface such that their changes contribute to solar dimming and brightening (see also Chapter 2). Balance in the net surface radiation therefore requires that either the upwelling surface sensible or latent heat fluxes decrease. Over land, where surface heating is a primary driver for convective clouds, this will affect cloud fraction and depth. Together the mechanisms of stabilisation and reduction in surface fluxes provide a means for aerosols to significantly modify the cloud fraction of surface-forced continental clouds (Feingold et al., 2005; Sakaeda et al., 2011). The microphysical properties of the cloud can modify entrainment rates and have also been shown to affect the semi direct effect when BC occurs within the cloud layer (Hill and Dobbie, 2008).

Cloud cover is expected to decrease if absorbing aerosol is embedded in the cloud layer. This has been observed (Koren et al., 2004) and simulated (e.g., Feingold et al., 2005) for clouds over the Amazon forest in the presence of smoke aerosols. In the stratocumulus regime, absorbing aerosol above cloud-top strengthens the temperature inversion, reduces entrainment and tends to enhance cloudiness. Satellite observations (Wilcox, 2010) and modelling (Johnson et al., 2004) of marine stratocumulus show a thickening of the cloud layer beneath layers of absorbing smoke aerosol, which gives a local negative semi-direct effect.

GCMs lack the ability of large eddy simulation models to represent smaller scale cloud processes and this gives low confidence in their ability to simulate the semi-direct effect (Johnson, 2005). Conversely, the GCM response allows for large scale circulation changes that can have significant additional effects which are not captured in large eddy models. Importantly, GCM results have shown that semi-direct effects are not confined to absorbing aerosol, CO_2 and potentially all mechanisms have an associated tropospheric adjustment (Section 7.2.4.3.6). Further, the semi-direct effect is not only associated with cloud changes that are co-located with the forcing, it also leads to circulation changes that effect clouds and land surface properties remotely.

There is a clear link between changes in atmospheric heating and global mean precipitation changes (Andrews et al., 2010; Lambert and Allen, 2009; Liepert and Previdi, 2009; Stephens and Ellis, 2008; Takahashi, 2009; Wild and Liepert, 2010). Andrews et al. (2010) showed that absorbing aerosols lead to an initial reduction in global precipitation which is not completely offset by the slow global precipitation change that responds linearly to the global mean surface temperature change.

7.3.5.3 Estimates of Aerosol Radiative and Adjusted Forcings

Building on our understanding of aerosol concentration distributions (Section 7.3.4) and their radiative effects (Sections 7.3.5.1 and 7.3.5.2), the section combines the direct RF with the adjustment effect to produce an AF. The subsection additionally assesses the forcings from absorbing aerosol (BC and dust) on both snow and ice.

In Forster et al. (2007) these adjustment effects were not included in the forcing term but were evaluated as part of the response. Instead, adjustment was accounted for by applying an efficacy term that modified the climate sensitivity. For consistency with Chapter 8, all quoted ranges represent a 2-σ uncertainty (i.e., 5% to 95% probability and we evaluate radiative forcings between 1750 and ~2010. Note that for several aerosol species (such as biomass burning) this does not quite equate to the anthropogenic effect as emissions started to be influenced by humans before the industrial revolution. Many models estimate the aerosol radiative forcings between 1850 and present-day and conversion to a radiative forcing between 1750 and present-day contributes to increase the uncertainty (Bellouin et al., 2008).

The estimate of the total aerosol direct radiative forcing in Forster et al. (2007) combined estimates from 9 AeroCom models, 8 other model studies and 3 observationally-based estimates. The model direct radiative forcing was estimated to be –0.4 W m^{-2} with a 0 to –0.8 W m^{-2} range. However, the observationally-based estimates gave a more negative radiative forcing, so overall a best estimate for the aerosol direct forcing was given as –0.5 ± 0.4 W m^{-2}. Due to the extra evidence from observations for the total forcing there was more confidence placed in the total aerosol direct forcing than in that from individual aerosol species.

Observationally-based estimates of the direct RF are not completely independent of global aerosol models (Loeb and Su, 2010; Myhre, 2009). They employ satellite data of aerosol optical depth in combination with either aerosol optical properties from AERONET (Bellouin et al., 2008) or observationally-derived aerosol radiative efficiency (Quaas et al., 2008). In the observationally-based method, satellite retrieval constraints over land necessitate the use of information from global aerosol models in order to derive the change in AOD that is due to anthropogenic activity. The aerosol optical properties (single scattering albedo and asymmetry factor) have been held constant between preindustrial and present time, and radiative forcing calculations have been performed with radiative transfer schemes (Bellouin et al., 2008), or the aerosol radiative efficiency has been assumed to be constant over the industrial era (Quaas et al., 2008).

Since AR4 further observationally-based estimates of the total direct radiative forcing have been made. These employ both improved observations and also a more robust methodology for combining with models to infer radiative forcing. Zhao et al. (2008) derived a clear-sky anthropogenic aerosol direct forcing over ocean from a combination of CERES/MODIS data and the GOCART aerosol model finding radiative effects that would broadly agree with equivalent estimates from Bellouin et al., (2008). Two studies (Bellouin et al., 2008; Myhre, 2009) identified a number of causes for the larger magnitude in the observationally-based radiative forcing estimate compared with modelling-based estimates within AR4. Compared to Bellouin et al., (2005), Bellouin et al., (2008) refined their calculation of the direct radiative forcing using updated satellite retrievals, finding an all sky radiative forcing of –0.65 W m^{-2}. They were able to attribute a more negative observationally-based forcing to i) a positive cloudy-sky direct radiative forcing in the model, ii) the exclusion of high surface albedo regions, and iii) the fact that the model forcing was compared to 1750 rather than present day natural aerosol for the observed estimate. Myhre (2009) extended this analysis and used model results to adjust their observed estimate to make it representative of a globally averaged all-sky present – preindustrial radiative forcing. Making these changes reduced the magnitude of the observationally-based radiative forcing estimate from –0.65 W m^{-2} to –0.3 ± 0.2 W m^{-2}. An additional comparison of the Oslo model with various observations suggested that the estimate of total direct RF was rather robust, giving increased confidence to their estimate.

Loeb and Su (2010) have challenged the uncertainty ranges presented in Forster et al. (2007) and Myhre (2009). They use an estimate of present-day AERONET uncertainties in aerosol optical depth, single-scattering albedo, asymmetry parameter, aerosol scale height, and anthropogenic fraction. The aerosol optical depth was perturbed by ±0.01, the asymmetry parameter was perturbed by ±0.02, and the single-scattering albedo was perturbed by ±0.06 over ocean and ±0.03 over land. The scale height was perturbed by 0.8 km and the anthropogenic fraction by 10%. Their analysis does not take into account additional uncertainties in the estimate of direct radiative forcing such as preindustrial emissions. Nevertheless, they find that uncertainties in aerosol optical depth alone can lead to errors of ±0.2 W m^{-2}. The single-scattering albedo error was found to be much more significant, contributing an error of greater than ±0.5 W m^{-2}. Our own assessment of uncertainty (see below) falls between those from past studies, indicating that the Loeb and Su (2010) approach is valid but likely over-estimated the single scattering albedo uncertainty.

Lohmann et al. (2010) found direct RFs in five GCMs ranged from –0.1 to –0.4 W m^{-2}. A second phase of AeroCom model results is being compiled, based on updated emissions and model codes. Simulations of the secondary components nitrate and SOA are now included in the analysis. Estimates of the radiative forcing for the total anthropogenic direct aerosol effect range from –0.06 to –0.49 W m^{-2}. Figure 7.11 shows the zonal mean total aerosol RF for all participating AeroCom models. Most of the models have a maximum negative radiative forcing around 20–50°N, in the region with highest aerosol concentrations. Several models show a positive radiative forcing at high latitudes due to the higher surface albedo there. We combine our uncertainty analysis with the best estimate of –0.3 W m^{-2} from the satellite derived estimates, reinforced by the RF from the Lohmann et al. (2010) and the new AeroCom modelling studies, to estimate an aerosol direct RF of –0.3 ± 0.3 W m^{-2}.

[INSERT FIGURE 7.11 HERE]

Figure 7.11: Zonal mean total aerosol direct radiative forcing from the different AeroCom models. No adjustment for missing species has been applied.

The semi-direct effect in GCMs has been estimated from both fixed sea surface temperature experiments and regression methods (e.g., Allen and Sherwood, 2010; Andrews and Forster, 2008; Lohmann et al., 2010). Most GCM studies indicate regional variations in the cloud response but generally increased cloud cover over oceans, especially in low and mid level clouds (Allen and Sherwood, 2010; Sakaeda et al., 2011). Over land the response is more dependent on the type of forcing (Allen and Sherwood, 2010; Koch and Del Genio, 2010; Sakaeda et al., 2011). Overall the result is thought to be a small and possibly net negative semi-direct effect feedback from the cloud response to absorbing aerosols (Koch and Del Genio, 2010). Five GCMs were analysed for RF and AF in Lohmann et al. (2010). One GCM (CSIRO) had a large difference between its RF and AF, giving a significant and negative semi-direct effect for the direct aerosol forcing of around –0.3 W m^{-2}, traced to a longwave cloud adjustment. The other four GCMs analysed exhibited both positive and negative semi-direct effects but none were significant. Based on current understanding, there is high confidence at the local scale that in-situ heating by absorbing aerosol can cause cloud to both increase and decrease, depending on specific conditions. However, there is low confidence in a globally-significant effect as GCMs differ in their responses and are not able to adequately represent some of the important cloud processes. Thus the semi-direct effect can be significant regionally and can be of either sign. Globally the semi-direct effect is likely to contribute an AF that is small (smaller than 0.1 W m^{-2}) and not significantly different than zero. In conclusion, adding the semi-direct effect to the RF increases the uncertainty range, but does not alter the best estimate, which gives an AF of –0.3 ± 0.4 W m^{-2}.

7.3.5.4 Aerosol Direct Radiative Forcing by Species

AeroCom studies have calculated the direct radiative forcing using preindustrial and present-day simulations with the same meteorology and no microphysical changes to isolate direct effects. For the RF of the individual species (SO_4, BC fossil-fuel, OC fossil-fuel, biomass burning or BB, SOA, NO_3) simulations are performed as a difference between the control and a new simulation with only the aerosol component in question set to emissions as in the preindustrial simulation (Figure 7.12).

For sulphate AeroCom models give a RF of –0.2 to –0.6 W m^{-2} 90% range with a –0.3 W m^{-2} median estimate and we adopt this our our best estimate.

Ramanathan and Carmichael (2008) suggest a radiative effect of +0.9 W m^{-2} for the total (anthropogenic plus natural) BC using an analysis of AERONET data. However, they likely overestimate the aerosol optical depth as their results may be contaminated by dust. The current AeroCom models have components of BC from fossil fuel and biomass burning. We take the estimate of RF from BC aerosol from fossil fuel from AeroCom as +0.0 to +0.4 W m^{-2} 90% range with a +0.2 W m^{-2} median estimate. Zarzycki and Bond (2010) assess uncertainty in the BC forcing-efficiency associated with the vertical profile of BC. They find that uncertanity in the vertical profile contributes an uncertainty of around 25% to the forcing but also a positive bias of around 15% in model-estimated radiative forcing, as models tend to have too much BC in the upper troposphere compared to observations (see Section 7.3.4). When the AeroCom BC fossil fuel and BC from biomass burning are combined and scaled by AERONET observations, we derive an anthropogenic BC radiative forcing of 0.4 ± 0.2 W m^{-2}, considering uncertainties associated with model diversity, AERONET-model bias, AERONET representativeness and clear sky bias (e.g., AERONET does not observe over ocean or on cloudy days), dust contamination of AERONET data, anthropogenic fraction, vertical distribution of BC, underlying surface albedo, radiative transfer and covariance of aerosol with clouds.

For organic carbon aerosol from fossil fuel AeroCom models give a –0.0 to –0.1 W m^{-2} (90% range) with a –0.05 W m^{-2} median estimate from fossil fuel emissions and we adopt this our our best estimate.

For biomass burning aerosol AeroCom models give a –0.15 to +0.1 W m^{-2} (90% range) with a –0.01 W m^{-2} median estimate and we adopt this our our best estimate.

For SOA AeroCom models give a –0.03 to –0.07 W m^{-2} 90% range with a –0.04 W m^{-2} median estimate and we adopt this our our best estimate.

We combine the AeroCom range with earlier estimates from Adams et al. (2001), Bauer et al. (2007) and Myhre et al. (2009) give a RF estimate of –0.1 ± 0.08 W m^{-2} for nitrate aerosols.

Anthropogenic sources of mineral aerosols can result from changes in land use and water use or climate change. Estimates of the anthropogenic radiative forcing of mineral aerosols is highly uncertain, because natural and anthropogenic sources of mineral aerosols are often located close to each other (Mahowald et al., 2009). Using a compilation of observations of dust records over the 20th century with model simulations, Mahowald et al. (2010) deduce an 1750–2000 change in mineral aerosol direct radiative forcing including both natural and anthropogenic changes of –0.14 ± 0.11 W m^{-2}. This is consistent within the AR4 estimate of –0.1 ± 0.2 W m^{-2} (Forster et al., 2007) which we retain here. Note that some of this forcing could be due to feedback processes (see Section 7.3.6).

[INSERT FIGURE 7.12 HERE]

Figure 7.12: Median, full range and 5%–95% range of AeroCom model direct radiative forcing by species and the total direct forcing. The total direct forcing has been adjusted to take account of missing species in some models by adding the median value of the species forcing from the remaining models.

7.3.5.5 Absorbing Aerosol on Snow and Sea-Ice

Forster et al. (2007) estimated the radiative forcing for surface albedo changes associated with BC on snow to be 0.10 ± 0.10 W m^{-2}, with a low level of understanding. This estimate was largely based on calculations with the GISS climate model (Hansen and Nazarenko, 2004) and a prognostic estimate by Jacobson (2004). Since AR4, understanding, observations and modelling have all improved such that a more robust assessment can be made. Further, additional effects of BC in snow and ice have been observed and estimated. The role of non-BC constituents has also been investigated.

Global model studies have either scaled albedo changes based on model-derived BC deposition rates (Hansen et al., 2005; Shindell and Faluvegi, 2009) or have prognostically determined the concentrations of BC in snow (Flanner et al., 2009; Jacobson, 2004; Koch et al., 2009b; Rypdal et al., 2009) then calculated the change in snow albedo, radiative forcing, and climate response. All of these indicate that BC in snow produces warming both in the Arctic and across the northern hemisphere and that the climate efficacy (change in temperature per unit forcing) of BC in snow is 2–4 times greater than that of CO_2 or other climate forcers. This high efficacy occurs primarily because all of the forcing energy is deposited directly into the cryosphere, whose evolution drives a positive albedo feedback on climate.

Radiative forcing by BC in the cryosphere was calculated by Jacobson (2004) and Flanner et al. (2009) using a baseline snowpack which included light-absorbing soil dust, which reduces the impact of other light-absorbing particles by approximately 20%. Large uncertainties persist in the distribution of dust mass and absorptivity. Some model studies calculated radiative forcing due to fossil fuel and biofuel emissions only (Jacobson, 2004; Rypdal et al., 2009), but others calculated the effect of all sources: fossil fuel, biofuel and biomass burning (Flanner et al., 2009; Hansen and Nazarenko, 2004; Hansen et al., 2005; Koch et al., 2009a). Finally, while most studies have calculated the total forcing for a given (near present-day) year, Hansen et al. (2005) and Koch et al. (2009a) quantified the change in forcing from pre-industrial to present, defined as 1880–2000 and 1890–1995, respectively. Key uncertainties are BC concentrations in snow and ice, BC mixing state and optical properties, snow and ice area coverage and patchiness the background particles already present in the snow pack, snow effective grain size and its influence on albedo reduction from impurities, the masking of snow surfaces by clouds and vegetation, and the accumulation of BC at the top of snowpack induced by melting and sublimation. Biases in forcing estimates arise from the model assumptions cited above. We use a field survey of Arctic snow samples collected during 2005–2009 (Doherty et al., 2010) to adjust the model studies based on biases in Arctic BC snow concentrations. This leads to a radiative effect of of +0.04 W m^{-2} (with a 0.01–0.09 W m^{-2} 90% uncertainty range) caused by present-day sources of fossil fuel, biofuel, and biomass burning BC, where the range represents the combination (in quadrature) of all individual uncertainties listed above except for snow patchiness and masking by vegetation and clouds. This estimate includes forcing from snow on land and lying on sea-ice.

About 80% of this forcing is exerted on land-based snow (Flanner et al., 2007; Flanner et al., 2009). We also make a crude estimate of the effect of BC within snow-free sea ice, estimating a present-day effect of +0.010 (0.003–0.032) W m^{-2}, leading to a combined (snow and sea-ice) present-day cryosphere radiative effect from BC of +0.05 W m^{-2}. Finally, by scaling the forcing contributions from biomass burning, biofuel, and fossil fuel BC emissions according to their 1750–2010 changes, we arrive at an industrial-era RF estimate of +0.04 (0.01–0.10) W m^{-2} for the combined influence of BC in sea-ice and in snow overlying land and sea-ice. Note that not all of this RF is necessarily directly anthropogenic.

Filter measurements indicate that a large fraction of the aerosol light absorption (~30–50%) in Arctic snow is due to non-BC constituents (Flanner et al., 2009). Other studies show that dust is the dominant source of light absorption in some continental snowpacks (Painter et al., 2010). Moreover, Hegg et al. (2010) attribute most of the non-BC light absorption in Arctic snow samples to brown carbon (BrC), with crop and biomass burning identified as the primary sources. These agents also contribute a positive snow radiative forcing, although some of this absorption may be implicitly accounted for in the BC emission inventories applied in modelling studies. As it is not clear to which extent changes in dust emissions are anthropogenically driven we refrain from providing a RF best estimate for the effect of anthropogenic dust on snow and sea-ice, but it is considerably smaller than the effect of BC (Flanner et al., 2009).

In summary we assess that anthropogenic absorbing aerosols (BC/BrownC) on snow/ice are responsible for a positive RF of +0.04 W m^{-2}, with a 0.01–0.10 W m^{-2} 5%–95% uncertainty range. It is important to note that this forcing is 2–4 time more effective at causing global mean temperature changes than an equivalent forcing from CO_2.

7.3.6 Aerosol-Climate Feedbacks

7.3.6.1 Introduction

Changes in climate parameters can modify the sources of natural aerosols and the atmospheric lifetime of anthropogenic and natural aerosols, which may in turn feedback on the climate system through their interactions with radiation and cloudiness and their contribution to nutrient cycling (Carslaw et al., 2010; Jickells et al., 2005, see also Chapter 6). The climate drivers of changes in aerosols can be split into physical changes (temperature, humidity, precipitation, soil wetness, solar radiation, wind speed, sea ice extent, etc…), chemical changes (availability of oxidants) and biological changes (vegetation cover and properties, plankton abundance and speciation, etc…). The response of aerosols to climate change may constitute a feedback loop whereby climate processes amplify or dampen the initial perturbation. We will assess here the relevance and strength of aerosol-climate feedbacks in the context of future climate change scenarios.

7.3.6.2 Changes in Aerosol Concentrations with Climate

7.3.6.2.1 Sea salt and mineral dust

Climate change influence atmospheric burden of sea salt by altering emissions, transport, and deposition. There is no agreement among climate models about the strength of such an effect, with estimates ranging from an overall 19% reduction in global sea salt burden from the present-day to year 2100 (Liao et al., 2006) because of a reduction in surface wind speed and an increase in scavenging, to little sensitivity (Mahowald et al., 2006a), or a sizeable increase in Bellouin et al. (2011) because of a decrease in sea ice cover despite a decrease in wind speed over most of the tropical and mid-latitude oceans (Jones et al., 2007). Given that sea salt particles comprise a significant fraction of CCN concentrations over the oceans, such large changes are likely to feedback on climate through changes in cloud drop number (Korhonen et al., 2010b). There is little understanding on how surface wind speed may change over the ocean in a warmer climate with some observations suggesting an increase in wind speed over the last two decades (Young et al., 2011) and some models predicting a widespread decrease in ice-free oceanic regions.

Studies of the effects of climate change on dust loadings give a wide range of results from large increases (e.g., Woodward et al. (2005) find a factor of 3 increase in 2100) to moderate (–10 to –20%) increase or decrease (e.g., Liao et al., 2006; Liao et al., 2009; Tegen et al., 2004) and to large decreases (e.g., Mahowald and Luo (2003) and Mahowald et al. (2006b) find a 60% decrease under double CO_2 concentration). The large range reflects different responses of the atmosphere and vegetation cover to climate change forcings.

For example, Mahowald (2007) found that the consideration of the CO_2 fertilization effect is important for predicting desert response to future climate change.

7.3.6.2.2 Sulfate, ammonium and nitrate aerosols

The magnitude and sign of the DMS-sulfate-cloud-climate feedback loop remain uncertain despite two decades of research (see Ayers and Cainey, 2007 and Carslaw et al., 2010 for a review). It is now realized that the feedback could operate in numerous ways through changes in temperature, solar radiation dose, mixed layer depth and nutrient recycling, sea-ice extent, wind speed, shift in marine ecosystems due to ocean acidification and climate change, atmospheric processing of DMS into CCN, and no study has included all the relevant effects. There is however some consistency among Earth System models to simulate a weak feedback due to i) a weak sensitivity of DMS production to climate change, and ii) a weak sensitivity of CCN population to changes in DMS emissions (Carslaw et al., 2010; Woodhouse et al., 2008). However regional effects could be larger (Bopp et al., 2004).

In the atmosphere chemical production of sulfate increases with temperature (Aw and Kleeman, 2003; Dawson et al., 2007; Kleeman, 2008), due to faster SO_2 oxidation (higher rate constants and higher oxidant concentrations). Changes in O_3 and H_2O_2 with climate also influence sulfate through in-cloud sulfate formation. Most studies to date predicted a reduction of 0–9% in global sulfate burden, mainly because of the future increases in precipitation (Liao et al., 2006; Pye et al., 2009; Racherla and Adams, 2006; Unger et al., 2006). However Rae et al. (2007) found a small increase in global sulfate burden from 2000–2100 because the simulated future precipitation was reduced in regions of high sulphate abundance.

Changes in temperature have a large impact on nitrate aerosol formation by shifting of aerosol thermodynamic equilibrium, because more HNO_3 remains in the gas-phase in a warmer climate. There is some agreement among global aerosol models that climate change will contribute to decrease nitrate concentrations (Bellouin et al., 2011; Liao et al., 2006; Pye et al., 2009; Racherla and Adams, 2006) with the exception of Bauer et al. (2007) who found little change in nitrate for year 2030. It should be noted however that changes in precursor emissions are likely to increase nitrate concentrations in the future (Bellouin et al., 2011).

Changes in sulfate and nitrate influence ammonium aerosol formation. The burden of ammonium was predicted to decrease by about 5% in Pye et al. (2009) from present day to 2050 under the A1B scenario. Changes in ammonium in Pye et al. (2009) are milder than those predicted by Racherla and Adams (2006) and likely reflect the fact that the sulfate burden is relatively insensitive to climate change under the A1B scenario.

7.3.6.2.3 Carbonaceous aerosols

Natural emissions of carbonaceous aerosols, such as wild fires and biogenic emissions are climate sensitive. There is evidence that future climate change could lead to increases in the occurrence of wildfires because of changes in fuel availability, readiness of the fuel to burn and ignition sources (Kloster et al., 2010; Marlon et al., 2008; Mouillot et al., 2006; Pechony and Shindell, 2010). However vegetation dynamics may also play a role which is not well understood. Increased fire occurrence would increase aerosol emissions, but decrease BVOC emissions. This could lead to a small positive or negative radiative effect (and feedback) depending on the sign of the net impact by biomass burning aerosols (Carslaw et al., 2010).

A large fraction of secondary organic carbon aerosol form from the oxidation of isoprene, sesquiterpenes and monoterpenes from biogenic sources. Emissions from vegetation can increase in a warmer atmosphere, everything else being constant (Guenther et al., 2006). Global aerosol models simulate an increase in isoprene emissions of 22–55% by 2100 in response to T change (Heald et al., 2008; Liao et al., 2006; Sanderson et al., 2003) and a change in global SOA burden of –6% to +11% through the climate-induced changes in aerosol processes and removal rates (Heald et al., 2008; Liao et al., 2006; Tsigaridis and Kanakidou, 2007). Increasing CO_2 concentrations are believed to inhibit BVOC emissions (Arneth et al., 2007) which could offset the T effect and adds significant uncertainty to future emissions. Future changes in vegetation cover, whether they are natural and anthropogenic, also introduce large uncertainty, with either a decrease (forest to cropland) or increase (some biofuels plantation) in emissions (Lathière et al., 2010). There

is little understanding on how the marine source of organic aerosol may change with climate, notwithstanding the large range of emission estimates for the present day (Carslaw et al., 2010).

7.3.6.3 Synthesis

There is no robust evidence to suggest that future changes in emissions of natural aerosols (or their precursors) could represent a significant climate feedback during the 21st century; the feedback factor is mostly bracketed within ±0.1 W m^{-2} K^{-1} (Carslaw et al., 2010). While some models simulate large changes (such as for dust emissions), these simulations are also associated with large uncertainties and there is little to no agreement among models. There is conflicting evidence of the strength of the feedback associated with future changes in precipitation and aerosol scavenging with one study (Liao et al., 2009) showing a significant positive feedback (feedback parameter of +0.04 to +0.15 W m^{-2} K^{-1} on a global mean basis) and other studies simulating smaller feedback of –0.02 to –0.08 W m^{-2} K^{-1} (Bellouin et al., 2011). However such feedbacks may be important for climate at the regional scale.

7.4 Aerosol-Cloud Interactions

7.4.1 Introduction

This section assesses our understanding of aerosol-cloud-precipitation interactions, emphasizing the ways in which anthropogenic aerosols may be affecting the distribution of clouds and precipitation. The idea that anthropogenic aerosols are changing cloudiness, and hence the planetary albedo, and thus contribute a substantial radiative forcing to the climate system is a longstanding one. For these reasons aerosol-cloud interactions have been addressed to varying degrees in all of the previous assessment reports.

Since the AR4, research has continued to articulate new pathways through which the aerosol may affect the radiative properties of clouds, but also (and increasingly so) patterns of precipitation (Rosenfeld et al., 2008), Global-scale modelling has advanced in its ability to represent a greater diversity of aerosol-cloud interactions, and with greater consistency. Observational studies continue to document strong local correlations between aerosol proxies and clouds or precipitation, but have become more quantitative and are increasingly identifying the methodological challenges associated with such correlations. Fine-scale modelling studies have begun to be used in a systematic manner, and among other things have shown how turbulent mixing, cloud and regional-scale circulations may buffer the effects of aerosol perturbations (Stevens and Feingold, 2009).

Aerosol-cloud interactions are highly heterogeneous and transient, which further complicates their quantification. While the top-of-atmosphere radiative effect is a well-established measure of the net impact of aerosol-cloud interactions on global temperatures (Lohmann and Feichter, 2005), no similar framework has yet established for addressing the hydrological and/or local effects of climatic perturbations. Global-mean precipitation is constrained approximately by the surface energy budget, hence an analogous framework developed around the adjusted surface forcing has been suggested (Andrews et al., 2010). But interest in precipitation tends to be local, where changes in horizontal energy transport within the atmosphere dominate over those of the surface energy budget (Muller et al., 2011). The idea of precipitation susceptibility, how the surface precipitation changes locally as a function of the aerosol burden, has been developed with these interests in mind (Feingold and Siebert, 2009), and as an extension of the idea of albedo susceptibility which has been used to explore the regional diversity of radiative effects stemming from aerosol-cloud interactions.

The remainder of this introductory section presents an overview of the framework through which the climatic effect of aerosol cloud interactions will be assessed, and highlights some of the general advances, and remaining challenges that face this topic of inquiry. In the subsequent sections advancements in our more specific understanding of aerosol-cloud, or aerosol-precipitation, interactions are assessed, from the perspective of their overall impact on the climate system.

7.4.1.1 Overview and Classification of Hypothesized Aerosol-Cloud Interactions

Denman et al. (2007) catalogued several possible contributions to aerosol indirect effects. Given the number of possible aerosol-cloud interactions, and the difficulty of isolating them individually, we see little value in attempting to assess each effect in isolation especially since modelling studies suggest that the effects may interact and compensate (Morrison and Grabowski, 2011; Stevens and Feingold, 2009). Instead, we group all radiative consequences of aerosol-cloud interactions into two broad categories: the immediate impact on radiative forcing in the absence of macrophysical changes to clouds, denoted "indirect radiative forcing" (iRF), and the final result including follow-on impacts of macrophysical responses to the initial change, denoted "indirect adjusted forcing" (iAF). The iRF represents the classical "Twomey" or cloud albedo effect whereby greater CCN numbers increase the droplet surface area, but is extended to include ice clouds and changes in the breadth of the size distribution. It is discussed further in Section 7.4.2. The iAF additionally accounts for any secondary effects that result as clouds adjust to the rapid changes in their environment accompanying an aerosol perturbation (Figure 7.1). Although iAF subsumes iRF, we retain an estimate of iRF for continuity with prior assessments and because it is better understood than the model-dependent effects determining iAF (Section 8.1.1). The iAF includes for example lifetime effects, wherein cloud macrostructure adjusts to changes in cloud microstructure (Albrecht, 1989; Liou and Ou, 1989; Pincus and Baker, 1994). Possible contributions to the iAF from liquid clouds are discussed in Section 7.4.3, separately from those associated with adjustments by ice or mixed phase clouds, which are presented in Section 7.4.4.

7.4.1.2 Advances and Challenges in Observing Aerosol-Cloud Interactions

Since the AR4 the characterization of aerosol sources, sinks, and composition has continued to advance. The on-going development of understanding of climate-relevant aerosol properties is reviewed in Section 7.3.3.

Progress has also been made in understanding how measurement artefacts affect retrievals of both aerosol (Kahn et al., 2005; Tanré et al., 1996; Tanré et al., 1997) and cloud properties (Platnick et al., 2003) in broken fields of clouds. Two key issues are that measurements classified as cloud-free often are not, and that aerosol measured in the vicinity of clouds is significantly different than it would be were the cloud field, and its proximate cause (high humidities), not present. The latter results from hydration effects on aerosol optical properties (Charlson et al., 2007; Twohy et al., 2009), contamination by undetectable cloud fragments (Koren et al., 2007) and the remote (non-columnar) effects of clouds on radiation through scattering (Wen et al., 2007). Photons scattered by cloud edges can interact with clear-sky aerosol layers as far as 15 km away, thereby biasing aerosol retrievals in ways that depend on the prevalence of clouds (Várnai and Marshak, 2009).

The use of active space-based remote sensing has also begun to address the coincidence problem, wherein satellite retrievals using passive sensors are unable to distinguish aerosol layers above or below fields of clouds from those intermingling with the cloud field (Anderson et al., 2005; Huffman et al., 2007; Stephens et al., 2002). Spectral polarization and multi-angular measurements provide much needed information on the phase function of particulate matter in the atmosphere, which can discriminate between clouds and aerosol and thus improve estimates of aerosol loading and absorption (Deuzé et al., 2001; Mishchenko et al., 2007). Field studies (Rauber et al., 2007; Wood et al., 2011) and laboratory investigations (e.g., Stratmann et al., 2009) of cloud aerosol interactions also continue to make important contributions to our understanding of how aerosols impact cloud processes. As a result our understanding of the distribution and properties of the aerosol in the vicinity of clouds, continues to improve apace with an appreciation of the limits of this understanding (Anderson et al., 2009).

The observational challenge of inferring causality from correlation remains a large, and limiting, one. Because the aerosol is a strong function of air-mass history and origin, and is strongly influenced by cloud and precipitation processes (Anderson et al., 2009; Clarke et al., 1999; Petters et al., 2006), and both are affected by meteorology (Engström and Ekman, 2010), correlations between the aerosol and cloud, or precipitation, cannot be taken as generally indicating a meteorological response to the aerosol. Furthermore, attempts to control for other important factors (air-mass history or cloud dynamical processes) are limited by a lack of understanding of cloud controlling factors in the first place (Anderson et al., 2009; Siebesma et al., 2009; Stevens and Brenguier, 2009). These problems greatly undermine confidence in observationally based inferences of aerosol effects on clouds and precipitation and must be considered with caution or interpreted with suitable models.

7.4.1.3 Advances and Challenges in Modelling Aerosol-Cloud Interactions

Fine-scale models, capable of resolving cloud-scale circulations have greatly advanced as a tool for testing the physical mechanisms proposed to govern aerosol-cloud-precipitation interactions (Ackerman et al., 2009; vanZanten et al., 2011). Aerosol-cloud interactions in climate models have largely been introduced based on simple calculations or highly idealized models (e.g., Albrecht, 1989; Pincus and Baker, 1994; Twomey, 1977). A general finding from explicit numerical simulations of clouds is that various aerosol impact mechanisms tend to be mediated (and often buffered) by interactions across scales not included in the idealized models that gave rise to the original idea (Stevens and Feingold, 2009). Specific examples involve the interplay between the drop-size distribution and mixing processes that determine cloud macrostructure (Ackerman et al., 2004; Bretherton et al., 2007; Small et al., 2009; Stevens et al., 1998; Wood, 2007), or the dependence of precipitation development in stratiform clouds on details of the vertical structure of the cloud (Wood, 2007). As a result it is more likely than not the physical system is less sensitive to aerosol perturbations than are large-scale models.

The representation of aerosol effects in large-scale models has also advanced. Most global models now represent an increasing number of hypothesized aerosol-cloud interactions, and through comparisons to data and to other models their evaluation has greatly advanced (Quaas et al., 2009b). In particular, global models are beginning to represent effects in convective, ice and mixed-phase clouds (e.g., Lohmann, 2008) and as described below in more detail. In addition "superparameterisation" approaches (Section 7.2.3.5.2) hold promise for treating aerosol-cloud interactions more comprehensively, but are computationally very expensive and have yet to be systematically applied to this problem.

Although advances have been considerable, the challenges remain formidable. The representation of clouds in large-scale models remains primitive (Section 7.2) and even if large-scale models were able to represent clouds with greater fidelity, fine-scale modelling suggests that the outcome of an aerosol perturbation depends on the details of the interaction of clouds, turbulence, radiation and precipitation processes on a range of scales not represented by large-scale models (vanZanten et al., 2011). For this reason it is not surprising that large-scale models exhibit a range of manifestations of aerosol-cloud interactions, which limits quantitative inference (Quaas et al., 2009b). However, by examining the interactions between cloud and large-scales in the global models, and between aerosol and turbulent processes in explicit cloud simulations, some progress has been made and is likely to continue even though no single model can simulate everything.

7.4.1.4 Combined Modelling and Observational Approaches

Combined approaches, which attempt to maximize the respective advantage of models and data, are beginning to add to understanding of aerosol-cloud interactions. These include inversions of the observed historical record using large-scale modelling studies, but also the use of reanalysis and chemical transport models to help interpret satellite records (Chameides et al., 2002; Koren et al., 2010; Mauger and Norris, 2010), field study data to help constrain fine-scale modelling studies (Ackerman et al., 2009; vanZanten et al., 2011), or satellite climatologies to constrain large-scale modelling (Quaas et al., 2009b).

7.4.2 Aerosol Effects on Liquid Cloud Albedo (Indirect Radiative Forcing – iRF)

7.4.2.1 The Physical Basis for the Indirect Forcing

The cloud albedo effect (Twomey, 1977), which here is simply called the indirect RF or iRF, is the mechanism by which an increase in aerosol number concentration leads to an increase in liquid cloud albedo (reflectance of incoming shortwave solar radiation) by increasing the cloud droplet number concentration and hence increasing total droplet surface area, with the liquid water content and cloud geometrical thickness held fixed. Although only the change in the cloud droplet concentration is considered in the original concept of cloud albedo effect, a change in the shape of droplet size distribution (such as breadth), which is directly induced by aerosols, may also be important (e.g., Feingold et al., 1997; Liu and Daum, 2002). Cloud albedo effects may also be manifest in ice, or mixed phase clouds, but these are discussed in Section 7.4.4.

The physical basis of the indirect forcing is generally well understood with research since the AR4 generally supporting the picture that had developed at that time. Detailed in-situ aircraft observations continue to show

that cloud droplet concentrations observed just above the cloud base generally agree with what would be predicted given the aerosol observed below the cloud base (cloud droplet closure), which is the fundamental link in the cloud albedo effect (e.g., Fountoukis et al., 2007). Vertical profiles of cloud effective radius also agree with those predicted by models which take into account the effect of entrainment (Lu et al., 2008), although uncertainties still remain in estimating the shape of the droplet size distribution (Hsieh et al., 2009), and the degree of non-adiabaticity within clouds. Multi-dimensional radiative transfer calculations have also been applied to estimate cloud albedo instead of using the traditional two-stream approximation to find that the latter could overestimate the albedo effect under certain conditions (Duda et al., 1996; Zuidema et al., 2008).

7.4.2.2 Observational Evidence for the Indirect Forcing

There is ample observational evidence for increases in aerosol resulting in an increase in drop concentration and decrease in drop size (for constant liquid water) and the main question is one of magnitude and spatial extent. Based on simple metrics, there is a large range of physically plausible responses with aircraft measurements (e.g., Lu et al., 2007; Lu et al., 2008; Twohy et al., 2005) tending to show stronger responses than satellite-derived responses (McComiskey and Feingold, 2008; Nakajima and Michael, 2009).

Radiative forcing associated with the Twomey effect is impossible to observe because of the rapid adjustments and further discussion is deferred to Section 7.4.3.2.

7.4.2.3 Advances in Process Level Understanding

At the heart of the albedo effect lie two fundamental issues. The first is the problem of droplet activation and its sensitivity to aerosol and meteorological parameters. The primary controls on drop number concentration are the aerosol number concentration (particularly at diameters > than ~ 80 nm) and updraft velocity. Aerosol size distribution can play an important role under high aerosol loadings, whereas aerosol composition is unimportant, except perhaps under very polluted conditions and low updraft velocities (e.g., Ervens et al., 2005). The second issue is the presence of condensed water that strongly determines how much energy can be reflected; there is no indirect effect unless clouds are present. Simple arguments show that the amount of reflected energy is approximately two-and-a-half times more sensitive to changes in the liquid water path than to changes in drop concentration. The magnitude of the indirect effect therefore rests mostly on dynamical forcing such as convective strength and entrainment that controls cloud amount, and a few key aerosol parameters such as aerosol number concentration and perhaps size distribution.

While observationally-based assessments of aerosol-cloud interactions have a long history, a more recent development is assessment of the ability of detailed models to reproduce the radiative fingerprints of aerosol-cloud interactions (Schmidt et al., 2009). This involves comparison between measurements of fields such as irradiance and comparison with the same fields calculated in finescale models that represent aerosol-cloud interactions. Such approaches identify key forcing parameters (e.g., cloud-field properties, aerosol hygroscopicity and absorption) and strengthen our understanding of the radiative forcing associated with aerosol effects on cloud microphysics.

7.4.2.4 Advances in and Insights Gained from Large-Scale Modelling Studies

Despite the consolidation in our understanding of the physical basis of the indirect forcing there still remain large uncertainties in quantification, because of the aforementioned difficulties in representing clouds and aerosol-cloud interactions in climate models. Recent estimates of the indirect forcing using satellite observations are systematically smaller than the AR4 estimates, which were derived from GCM calculations based on parameterization of physical processes or in-situ observations (Lohmann and Lesins, 2002; Quaas et al., 2008; Quaas et al., 2009b). The satellite data show lower susceptibility of cloud effective radius or droplet number concentration to aerosol optical depth or number concentration as compared with in-situ observations or detailed cloud parcel model calculations (McComiskey and Feingold, 2008), leading to the differences in the estimate of the iRF. This result is at least partly due to scale-related averaging biases in satellite retrievals (McComiskey and Feingold, 2011). It is also generally difficult to separate the iRF from the cloud fast feedbacks and meteorological effects in both observations and fully coupled numerical model

calculations (e.g., George and Wood, 2010; Lohmann et al., 2010). Estimates of the iRF are given in Section 7.4.6.

7.4.3 Adjustments in Liquid Clouds

7.4.3.1 The Physical Basis for Adjustments in Liquid Clouds

The effect of the aerosol on cloud amount is more multi-faceted than its effect on albedo alone, leading investigators to discuss such an effect in the plural. Such effects are often associated with changes in cloud lifetime and in the past have been referred to as 'lifetime' effects. However this nomenclature is misleading because it assumes a relationship between cloud lifetime and cloud amount. Moreover, the effect of the aerosol on cloud amount may have nothing to do with cloud lifetime per se (e.g., Pincus and Baker, 1994).

The traditional view has been that indirect adjustment effects of CCN will add to the initial albedo increase (iRF) by increasing cloud amounts. The chain of reasoning involves three steps. First that cloud droplet concentrations depend on the number of available cloud condensation nuclei; second that precipitation development is regulated by the cloud droplet concentration; and third that the development of precipitation reduces cloud amount (Stevens and Feingold, 2009).

Of the three steps the first has ample support in both observations and theory, as reviewed in the previous section. More problematic are the second two links in the chain of reasoning. The physical basis for an unambiguous, and positive, dependence of cloud amount on the available cloud condensation nuclei is weak. Although the idea that increased cloud droplet concentrations inhibit the initial development of precipitation in single clouds is longstanding, it is not clear that such effects reduce the precipitation overall. Some modelling studies suggest the opposite, wherein increased aerosol concentrations actually promote the development of deeper clouds, thereby invigorating precipitation (Rosenfeld et al., 2008; Stevens and Seifert, 2008, see also a more extensive discussion of this point in Section 7.4.5). The idea that cloud amount is a decreasing function of precipitation efficiency is even less clear. Although the original studies that hypothesized cloud amount effects (Albrecht, 1989; Liou and Ou, 1989) are often taken as demonstrative of this point, there is limited unambiguous observational evidence (exceptions to be given below). Many climate models assume such an effect *a priori,* which likely influences their forcing estimates.

A general statement of how precipitation affects cloud amount remains elusive. It appears increasingly likely that cloud amount effects, which almost certainly exist locally, vary from one cloud regime to the next. As such their quantification globally is a more ambitious task than originally anticipated, as it requires models to correctly represent the distribution of cloud regimes evincing such effects, and knowledge of how such effects manifest themselves across these regimes. Then again, because diverse effects offer the possibility of compensating one another, at least globally, it seems possible that lifetime effects may be considerably less important than previously thought (Stevens and Feingold, 2009).

7.4.3.2 Observational Evidence of Adjustments in Liquid Clouds

We discuss a sample of observations for which there is clear evidence of the aerosol interacting with clouds, in the broader sense of iRF and iAF since the latter subsumes the former. The cloud albedo effect is best manifested in so-called ship tracks, which are bright lines of clouds behind ships. As shown during the Monterey Area Ship Track (MAST) experiment, many ship tracks are characterized by an increase in the cloud droplet number concentration resulting from the increase in aerosol number concentration and an absence of drizzle size drops, which leads to a decrease in the droplet radius and an increase in the cloud albedo (Durkee et al., 2000). The global radiative forcing of visible ship tracks has been estimated from satellite and found to be insignificant at 0.5 m W m^{-2} (Schreier et al., 2007), although there is some concern that this analysis may not have identified all shiptracks. The new A-Train satellites offer the possibility to study the signal of long-term degassing of low-lying volcanic aerosol on stratocumulus (Gasso, 2008) and trade wind cumuli (Yuan et al., 2011). The stratocumulus respond with smaller drop sizes but ambiguous changes in cloud fraction and cloud water. The trade cumuli respond with smaller droplet size, decreased precipitation efficiency, increased cloud amount and higher cloud tops.

The development of precipitation in stratocumulus appears to be associated with more heterogeneous cloud features, including breaks in the cloud coverage (Comstock et al., 2005; Sharon et al., 2006; vanZanten et al., 2005). In some cases pronounced reversals in the cellular polarity become evident, where in compact regions (pockets), open cellular convection is surrounded by regions of closed cellular convection. Closed cellular convection in stratocumulus is characterized by a high albedo as broad regions of mesoscale ascent are covered by relatively optically thick clouds and narrow regions of mesoscale descent define the cell boundaries, and are characterized by optically thin clouds or even clear air. In contrast, open cellular convection in stratocumulus typically has a much lower albedo; broad regions of mesoscale descent are largely cloud free, and clouds are confined principally to cell boundaries in regions of mesoscale ascent. The development of pockets of open cells in broad regions of closed cellular convection appears closely linked to the development of precipitation (Savic-Jovcic and Stevens, 2008; Stevens et al., 2005; Wang and Feingold, 2009). The lack of any apparent difference in the large-scale environment of the open cells, versus the surrounding closed cellular convection, implies the existence of multiple equilibria. This raises the possibility that the onset of precipitation can lead to a chain of events that leads to a large-scale reduction of cloudiness, in agreement with original work from Liou and Ou (1989), Albrecht (1989) and Baker and Charlson (1990).

A number of hypotheses have been advanced, and explored, to explain why precipitation initially develops in regions of open cells (Rosenfeld et al., 2006; Wang et al., 2010a). If broad regions of open cells were much more characteristic of the pre-industrial atmosphere it would suggest that the anthropogenic aerosol has been responsible for large changes in planetary albedo. Cloud resolving modelling studies suggest however that even in relatively clean air in the South East Pacific, CCN concentrations are not low enough to support broad regions of open cells, hence this scenario appears unlikely. Indeed, one hypothesis for the formation of pockets of open cells is that they carry the imprint of airmasses in which sustained precipitation was dynamically triggered thereby leading to sufficient depletion of the aerosol to sustain the observed pockets of open cells (Wang et al., 2010a).

Precipitation from shallow convection, and clouds, prove difficult to observe in the trades, as the clouds tend to be small, and the small footprint of many precipitating trade-wind clouds does not lend itself well to space-based remote sensing techniques. [PLACEHOLDER FOR SECOND ORDER DRAFT: CloudSat reference]. However unlike stratocumulus, where more precipitation favours fewer clouds, in the much broader trade wind regions cloudiness tends to increase with precipitation amount, most likely because processes which favour precipitation development also favour clouds (Nuijens et al., 2009) and because precipitating trade cumulus tend to regenerate through colliding outflows.

7.4.3.3 Advances in Process Level Understanding

Since the AR4, there has been progress understanding of some basic processes relevant to cloud amount effects. One basic question is how susceptible precipitation is to cloud-droplet number concentrations, and by inference the available aerosol. Simple models for the conversion of cloud-water into rain-water predict that the rate of this process autoconversion process scales with the square of the inverse of the droplet number concentration. However theoretical work, that incorporates a fuller description of rain formation processes suggests that this strongly over-estimates the sensitivity of rain formation in shallow clouds (Stevens and Seifert, 2008), and that rain formation scales with a critical liquid water content that increases with approximately the inverse square-root of the droplet number concentration (Kostinski, 2008; Seifert and Stevens, 2010). Note that thicker, liquid clouds generate rain via accretion of cloud drops by raindrops, a process that is relatively insensitive to droplet concentration, and therefore to aerosol perturbations.

A number of observational studies have likewise found that the rain-rate from stratiform clouds scales with the liquid-water path of the cloud layer to the 3/2 power, and with the inverse of the droplet concentration (Comstock et al., 2005; Pawlowska and Brenguier, 2003; vanZanten et al., 2005). But because the observations also encapsulate the tendency of liquid water and cloud condensation nuclei to be removed by rain they likely over-estimate the sensitivity of rain formation to the droplet concentration, and under-estimate its sensitivity to liquid water. Some of the effects that reduce and even eliminate the sensitivity of rain formation to the auto-conversion process have begun to be incorporated in parameterizations used by large-scale models (Posselt and Lohmann, 2009).

Recent small-scale studies tend to confirm two responses of the cloud liquid water to increasing aerosol. Under clean conditions when clouds are prone to precipitation, an increase in the aerosol tends to increase cloud amount. Under non-precipitating conditions, clouds tend to thin in response to increasing aerosol (Ackerman et al., 2004; Small et al., 2009; Xue et al., 2008). Treatment of the subtlety of these responses and associated detail in small-scale cloud processes is not feasible in GCMs.

Since AR4, cloud resolving model simulation has begun to stress the importance of scale interactions when addressing aerosol-cloud interactions. Large model domains (order 100 km) allow mesoscale circulations to develop in response to changes in the aerosol. These dynamical responses may have a significant impact on cloud morphology and radiative forcing. Examples include the dramatic changes in cloud morphology associated with changes in cellular structure discussed above and the cloud-free shadows that appear alongside ship tracks (Wang and Feingold, 2009). These underscore the folly of applying simplistic rules for aerosol-cloud interactions.

7.4.3.4 Advances in and Insights Gained from Large-Scale Modelling Studies

Attempts to quantify cloud-mediated aerosol effects using global models suggest that lifetime effects contribute between –0.3 and –1.3 W m^{-2} to the adjusted forcing from the anthropogenic aerosol (Lohmann and Feichter, 2005). Because such studies usually neglect processes that may generate positive forcing usually involving ice (e.g., Storelvmo et al., 2008a) and because models are tuned to preclude estimates of the cloud-mediated aerosol radiative forcing that are too high (Hoose et al., 2009), it is difficult to use the model-based estimates as an independent constraint. By using satellite data to rescale relations that emerge from an ensemble of models several studies have argued for weaker cloud-mediated aerosol effects, and by extension weaker cloud lifetime effects (Quaas et al., 2009b).

Regional and global models systematically misrepresent the distribution of clouds, and cloud processes, especially those for shallow maritime clouds. One persistent shortcoming of global models is the tendency to only treat aerosol-cloud interactions in terms of large-scale, but not convective clouds. Recent efforts to consistently address both types of cloud representations represent a significant advance in large scale modelling (Lohmann, 2008). Nonetheless our understanding of aerosol-cloud interactions is incomplete, and what is well-understood is incompletely represented in large scale models. For these reasons, and because lifetime effects depend critically on the interplay of uncertainly parameterized physical processes, global-model based estimates of lifetime effects remain controversial.

7.4.4 Adjustments in Cold Clouds

7.4.4.1 The Physical Basis for Adjustments in Cold Clouds

In a water-saturated environment both liquid water and ice can co-exist, at least on timescales relevant for cloud processes, at temperatures between 0°C and –38°C. Clouds in which both liquid water and ice particles are present are referred to as mixed-phase clouds. At warmer temperatures ice rapidly melts, whereas at colder temperatures liquid water will freeze homogeneously. The formation of ice in the range of temperatures between 0°C and –38°C depends on heterogeneous freezing whereby a foreign medium, usually insoluble aerosol particles collectively referred to as ice nuclei (IN), initiate the freezing processes. Soluble matter or physiochemical transformations can hinder glaciation by depressing the freezing temperature of super-cooled drops (e.g., Baker and Peter, 2008; Girard et al., 2004). The same process can occur in cirrus clouds (Crawford et al., 2011) but there the lack of natural IN in the atmosphere makes homogeneous freezing the preferred pathway in cirrus clouds (Kärcher and Strom, 2003). Hence anthropogenic perturbations to the aerosol have the potential to affect when and where clouds become glaciated. For cirrus clouds this could inhibit homogeneous nucleation. For both mixed-phase and ice clouds, anthropogenic perturbations affect cloud optical properties, and can contribute to an albedo effect.

Because the enthalpy of liquid water is larger than that of ice, freezing is associated with a transfer of enthalpy from the particles to their environment. Hence glaciation affects cloud dynamics. Moreover, in a cloud consisting of supercooled liquid water the equilibrium vapour pressure will correspond to water saturation, which is significantly oversaturated with respect to ice. Thus the initiation of ice in a supercooled liquid cloud will cause vapour to diffuse rapidly toward ice particles at the expense of the liquid water in the

cloud, a process known as the Bergeron-Wegener-Findeisen process. This favours the depositional growth of large ice crystals, which may sediment away from the saturated region of the atmosphere, influencing the subsequent evolution of the cloud. Hence the ease with which ice forms may be influenced by anthropogenic perturbations to the aerosol, and in turn may regulate cloud amount (Lohmann, 2002a; Storelvmo et al., 2011; see also Section 7.2.2.4), or upper tropospheric humidity.

The types of aerosol particles that contribute to the distribution of IN apart from mineral dust are poorly understood. For example, the role of biological particles acting as IN remains controversial. While such particles have been found to be negligible for realistic concentrations of bacteria (Diehl and Wurzler, 2010; Hoose et al., 2010a; 2010b; Phillips et al., 2009; Sesartic et al., 2011) some investigators (Ariya et al., 2009; Sun et al., 2010) argue that biological particles even in low concentrations may still be important because they can trigger ice multiplication. The presence of more soluble aerosol particles would make it harder to form atmospheric ice homogeneously, but the primary production of insoluble aerosol particles could help initiate ice through poorly known heterogeneous freezing mechanisms. If IN become coated with soluble material they may also become less effective as IN (see Section 7.2.2.4, Hoose et al., 2008; Lohmann and Hoose, 2009; Storelvmo et al., 2008a). Anthropogenic changes to the biosphere could conceivably also make biological IN less prevalent. Our poor understanding of the climatology and life-cycle of aerosol particles that can serve as IN complicates attempts to generally assess what constitutes an anthropogenic perturbation to the IN population, let alone the effect of such a perturbation.

7.4.4.2 Observations of Aerosol Effects on Ice and Mixed-Phase Stratiform Clouds

Arctic mixed-phase clouds have received a great deal of attention since AR4, with major field programs conducted in 2004 (Verlinde et al., 2007) and 2009 (Brock et al., 2011; Jacob et al., 2010; McFarquhar et al., 2011) in addition to longterm monitoring at Barrow, Alaska (Shupe et al., 2008). Mixed-phase Arctic clouds persist for extended periods of time (days and even weeks), in spite of the inherent instability of the ice-water mix. We focus here on the role of the aerosol and refer to Section 7.2.1.3.3 for a discussion of meteorological aspects. The subset of the aerosol that act as IN exists in concentrations of 10^{-5} to 10^{-1} cm^{-3}, i.e., only about 1 in a million particles acts as an IN. Even at such low concentrations, they have an important influence on cloud persistence, with clouds tending to glaciate and disappear rapidly when IN concentrations are relatively high and/or updraft velocities too small to sustain a liquid water layer. The details of the heterogeneous ice-nucleation mechanism remain controversial but there is increasing evidence that ice forms in Arctic stratus via the liquid phase (immersion freezing) so that the CCN population also plays an important role (de Boer et al., 2011; Lance et al., 2011). If ice indeed forms via the liquid phase this represents a self-regulating feedback that helps sustain the clouds: as ice forms, water is depleted, which restricts further ice formation and competition for water vapour via the Bergeron-Wegener-Findeisen process.

7.4.4.3 Advances in Process Level Understanding

Since the AR4 research on ice-microphysical processes has been active, to a large degree with an eye toward a better representation of such processes in models. Korolev (2007) developed a theoretically based parameterization of the Bergeron-Wegener-Findeisen process that has lately been employed in different GCMs (Lohmann and Hoose, 2009; Storelvmo et al., 2008b). As some mixed-phase clouds have been found to be long-lived, Korolev and Field (2008) derived a theoretical framework to explain this phenomenon. A recent review (Morrison et al., 2012) discusses the myriad processes that create a resilient mixed-phase cloud system.

An example of the level of detail associated with mixed–phase clouds that may be required for adequate representation in models is that the dependence of ice particle growth by vapour diffusion depends strongly on particle habit (Harrington et al., 2009) and may even have equal weight to ice nucleation mechanism vis-à-vis glaciation times (Ervens et al., 2011).

For the mixed phased processes where the Bergeron-Wegener-Findeisen process makes ice-initiation interesting, heterogeneous freezing parameterizations employed in cloud or larger-scale models remain mostly empirical (e.g., DeMott et al., 2010; Gettelman et al., 2010; Hoose et al., 2008; Lohmann and Diehl, 2006; Phillips et al., 2008; Salzmann et al., 2010; Storelvmo et al., 2008a), although some recent work

attempts to ground the development of parameterisations in concepts derived from classical nucleation theory (Hoose et al., 2010b).

Ice nucleation in cirrus clouds (at temperatures less than –35°C) depends crucially on the cloud updraft velocity and hence the supersaturation with respect to ice. For homogeneous nucleation, the threshold relative humidities have been parameterized using results of parcel model simulations (e.g., Barahona and Nenes, 2009; Sassen and Dodd, 1988), airborne measurements in cirrus or wave clouds (Heymsfield et al., 1998; Heymsfield and Miloshevich, 1995), extensions of classical homogeneous ice nucleation theory (Khvorostyanov and Sassen, 1998; Khvorostyanov and Curry, 2009), and data from laboratory measurements (e.g., Bertram et al., 2000; Friedman et al., 2011; Koop et al., 2000; Magee et al., 2006; Mohler et al., 2003). If ice nuclei are present, then heterogeneous nucleation is the preferred freezing pathway because it requires lower threshold relative humidities (or higher threshold temperatures) than homogeneous nucleation. The threshold relative humidities (or temperatures) for heterogeneous nucleation depend on the type and size of the involved ice nuclei (Figure 7.8 and Section 7.3.3.5).

7.4.4.4 Advances in and Insights Gained from Large-Scale Modelling Studies

Studies of the iAF that depend on ice microphysical pathways are considerably less advanced than those involving only liquid clouds, but have come into increasing focus since the AR4. Penner et al. (2009) obtained a rather large iAF of anthropogenic ice-forming aerosol on upper tropospheric clouds. However, they ignore potential compensating effects on lower lying clouds and therefore should be regarded with caution. The climate impact of anthropogenic lead-containing mineral dust particles, among the most efficient ice-forming substances, has been investigated. In the extreme scenario in which 100% of ice-forming mineral dust particles in cirrus clouds contained lead, up to 0.8 W m^{-2} more long-wave radiation was emitted to space as compared to pure mineral dust particles (Cziczo et al., 2009b).

Since the AR4 a number of studies have contributed to our ability to quantify aerosol effects on cirrus. Because such clouds usually only involve ice-phase microphysical processes, the physical pathways are somewhat simpler. BC can impact background cirrus by affecting ice nucleation properties but its effect remains uncertain (Kärcher et al., 2007). There is some evidence of a statistically significant impact on cirrus coverage (Hendricks et al., 2005). However, Liu et al. (2009) examined their role in radiative forcing of cirrus clouds and found it to be very small.

7.4.5 Aerosol-Cloud Microphysical Effects on Precipitating Systems

7.4.5.1 The Physical Basis for Aerosol Effects on Precipitating Convection

Deep convective clouds arise in response to the differential heating of the surface versus the atmosphere, largely as a result of radiative processes. This differential heating leads to the development of convective instabilities. The enthalpy of vaporization, and the much smaller enthalpy of fusion, which is released through the development of precipitation, is a primary mechanism for transporting enthalpy from the surface to the atmosphere and hence consuming the column instability. Through their effect on the microphysical development of clouds, for instance whether or not precipitation sized particles are readily formed, or if ice is initiated, the aerosol may modify the vertical distribution of condensate, and thus modify cloud and precipitation development.

Microphysically, the suppression in drop size in response to an aerosol perturbation is expected to suppress drop growth and precipitation formation and favour enhanced water loading in the upper portions of the cloud, which may limit cloud development, but favour the formation of convective downdrafts through more ready mixing with the environment. However, the lack of precipitation may allow clouds to grow deeper and help generate more precipitation (Nuijens et al., 2009). Then again, the cold pools associated with precipitation may be crucial for secondary precipitation (Lee et al., 2010; Matheou et al., 2011). As a result, whether more efficient precipitation production fosters or limits cloud development remains very uncertain.

Similar arguments can be extended to the ice phase. Once the cloud passes the freezing level, the effect of the aerosol on ice initiation becomes important, with the ensuing complications discussed in the previous section. If, however, changes in the anthropogenic aerosol make the initiation of ice less effective the

development of precipitation may be further retarded (Rosenfeld and Woodley, 2001), and vice versa. By delaying the initiation of ice the additional enthalpy of fusion will be introduced higher in the convective updraft, thereby suppressing the development of clouds initially, but promoting their development when they reach the level where ice forms. Although this is a rather small effect, relative to the tenfold larger enthalpy release from ongoing condensation, it has nonetheless been hypothesized to have a critical effect on cloud development (Andreae et al., 2004; Rosenfeld et al., 2008).

Finally, by determining whether liquid, mixed, or ice-phase processes are dominant, perturbations to the aerosol may also affect the distribution of precipitation. Studies in the southwestern Amazon region during the transition from dry to wet seasons (Martins et al., 2009) invoke such mechanisms to explain why higher CCN concentrations are associated with less frequent low-to-moderate rainfall rates and more frequent high rainfall rates.

To date, the issue of aerosol effects on precipitation remains an open question (Levin and Cotton, 2009). There are many uncertainties related to the physical mechanisms involved and to the observational and numerical tools (e.g., Khain, 2009).

7.4.5.2 Observations of Aerosol Effects on Precipitating Systems

Numerous observational studies have examined the links between aerosol particles and deep convective cloud properties, seeking to find robust associations between the aerosol and cloud properties that can be interpreted to be a result of aerosol effects on clouds. The availability of satellite data with global coverage, ever finer footprints, and a richer palette of spectral, angular, and polarization information has improved the retrieval of aerosol and cloud properties, and continues to advance the field, with the major strides since the AR4 coming from the use of active remote sensing, angular, and vertically, resolved data.

The analysis of satellite data, and *in situ* measurements, shows that an increase in aerosol loading is associated on average with smaller particles, taller invigorated convective clouds with larger cloud fraction and more extensive ice portions. These associations are found over the tropical Atlantic (Jenkins et al., 2008; Koren et al., 2005), Europe (Devasthale et al., 2005), North and South America (Andreae et al., 2004; Bell et al., 2008; Koren et al., 2008; Lin et al., 2006; Lindsey and Fromm 2008) and appear for all types of aerosol particles: biomass burning smoke, urban/industrial aerosol and desert dust. These more invigorated clouds would be expected to create more precipitation (Rosenfeld et al., 2008). For example an analysis of satellite data over the entire Brazilian Amazon during the dry, biomass burning, season found that elevated aerosol loading was associated with increased rainfall amounts and a shift to higher rainfall rates (Lin et al., 2006). Likewise changes in rainfall during the week over the south-eastern portion of the United States found an increase in rainfall for afternoon storms during the midweek compared to the weekend (Bell et al., 2008).

Observational studies examining the aerosol effect on precipitation of mixed phase clouds, often however report rain suppression in polluted atmospheres. Satellite data measured at various geographical locations suggest suppressed precipitation associated with polluted clouds (Jiang et al., 2008). Investigation of orographic clouds showed a reduction in the annual precipitation over topographical barriers downwind of major urban areas (Givati and Rosenfeld, 2004; Jirak and Cotton, 2006). Other studies showed no response of orographic rain to pollution (Halfon et al., 2009).

However, whether these observations should be interpreted as an effect of the aerosol on clouds, vice versa, or due to a third factor remains controversial. Causal pathways have been proposed that would be consistent with all of the hypotheses (Rosenfeld et al., 2008; Stevens and Feingold, 2009). What is clear is that a consistent picture has yet to emerge, perhaps because different pathways are active in different cloud regimes, or at different stages in the life-cycle of cloud systems (Koren et al., 2008; Stevens and Seifert, 2008).

7.4.5.3 Advances in Process Level Understanding

Modelling studies suggest that the thermodynamic environment in which the clouds grow is an important factor in the determination of the aerosol effect on the ground precipitation (Khain et al., 2005; Lynn et al., 2005; Tao et al., 2007). For clouds developing in dry unstable air, there is a decrease in the accumulated

precipitation with an increase in aerosol loading. For deep maritime clouds modelling studies suggest that aerosol perturbations can lead to an increase in precipitation. The important environmental factors are humidity, that influences the evaporation of cloud liquid (and sublimation of ice), and wind shear that can modulate the entrainment of dry air into clouds and the transport of cloud liquid into unsaturated areas (Fan et al., 2009).

Weekly cycles in aerosol properties and precipitation have emerged to tackle this problem. While all studies support a weekly cycle in aerosol properties, the results for weekly cycles in precipitation are conflicting (Barmet et al., 2009; Bäumer et al., 2008; Hendricks Franssen et al., 2009; Sanchez-Lorenzo et al., 2008; Stjern 2011).

Recently the tools of numerical weather prediction have been applied to this problem. Looking at summer season forecasts using a cloud-resolving regional model the question has been posed as to whether changes to the aerosol systematically affect precipitation over large regions. Little to no systematic effect of the aerosol could be documented (Seifert et al., 2011).

Cloud-resolving modelling by Lynn et al. (2007) suggests that aerosol perturbations will result in precipitation being displaced to the leeward side of the mountain, with the impact of the aerosol being strongest under drier conditions and weaker horizontal winds. Regional studies confirm a reduction of precipitation on the windward side of a mountain barrier and a tendency for the precipitation to shift downstream to the leeward side but the magnitude of this depends on the importance of the ice phase in these orographic clouds (Muhlbauer and Lohmann, 2009; Zubler et al., 2011).

As computational resources have increased it has become increasingly clear that some aerosol effects appear as transients in short duration simulations and that long (multi-day) simulations are necessary to get a more complete picture. Moreover the need to consider cloud systems, as opposed to individual clouds, when attempting to establish aerosol influences must be emphasised (Morrison and Grabowski, 2011). Parallels can be drawn with shallow systems in which mesoscale organization triggered by aerosol perturbations amplifies the response beyond what one might have expected from microphysical considerations alone (see Section 7.4.3.3). The large-scale constraint on surface precipitation makes it much more likely that one might get changes in the frequency and spatial distribution of rain, than changes in the total amounts.

7.4.5.4 Advances in and Insights Gained from Large-Scale Modelling Studies

Fast feedbacks associated with the aerosol indirect effects do not cause much change in precipitation if an average over a big enough domain is considered. Slow feedbacks through aerosol-induced changes in surface temperature/surface energy budget and changes in circulation can cause regional and global changes in precipitation. These latter effects have been estimated in AR4. Here the decrease in the global annual mean shortwave radiation at the surface since pre-industrial times due to scattering and absorbing aerosols amounted to –2.3 W m^{-2} with a range between –1.3 to –3.3 W m^{-2} (Denman et al., 2007). The associated change in the global mean precipitation amounts to between 0 and –0.13 mm day^{-1} (Denman et al., 2007).

Studies with a climate model coupled to an advanced representation of the aerosol, wherein convection is parameterized, also find little evidence of a weekly cycle in precipitation (Quaas et al., 2009a).

7.4.5.5 Large-Scale Convective Systems

Several studies since the AR4 highlight the possibility of aerosol effects on Atlantic hurricane activity, either by altering radiative heating or through microphysical effects on clouds. Year-to-year variations in hurricane activity and prevalence of dusty Saharan air layers (SAL) are significantly anticorrelated (Dunion and Velden, 2004; Evan et al., 2006). Periods of prominent SAL activity correlate with reduced sea surface temperatures, which would tend to suppress strong storm activity (Lau and Kim, 2007); dust also absorbs sunlight, thus warming the troposphere and further decreasing instability (Jury and Santiago, 2010; Wong et al., 2009). Evan et al., (2008) estimated that one third of the increase in the hurricane power dissipation index during the preceding 25 years was statistically attributable to the absence of the SAL. Since the SAL is warm and dry, it would be expected to suppress convection even in the absence of any dust (Sun et al., 2008; Wong and Dessler, 2005), which complicates any attribution of observed behaviour to dust. Moreover, Folz

and McPhaden (2008) showed that the correlation between Saharan dust and a cooler sea surface was mainly because both were caused by stronger winds, rather than dust causing the cooling. Thus while it is likely that dust variations have to some extent affected tropical Atlantic sea surface temperature trends and therefore Atlantic hurricane activity, it remains unclear how important this is.

7.4.6 Synthesis of Aerosol Effects

There are different ways to report the iRF and the iAF at the top-of-the-atmosphere since pre-industrial times. In Figure 7.13a we show estimates of the change in the net radiation if possible. If that estimate is not available the next options, taken in this order, are the change in the net shortwave radiation, the change in the net cloud forcing and last, the change in shortwave cloud forcing. Given that there is practically no longwave signal associated with the iRF and there are no changes in the clear-sky, all estimates of the iRF are comparable. However, for the iAF when fast feedbacks are included, there can be changes in the clear-sky and the longwave radiation, especially if aerosol effects on mixed-phase and ice clouds are considered or if local circulations respond rapidly to changes in the shortwave forcing.

Ensemble-averaged global-mean model estimates of the iRF have remained rather constant over time (Figure 7.13a) and amount to roughly –1 W m^{-2}. This estimate is obtained from the average over all published estimates, treating each of them as equal (one vote per model per paper). The –1 W m^{-2} estimate is slightly stronger than the estimate of the indirect forcing in AR4 where a different weighting procedure was used (Forster et al., 2007). If the iRF studies from GCMs are divided into those published prior to TAR (1993–2000), between TAR and AR4 (2001–2006) and since 2007, the median indirect forcing remains within 0.15 W m^{-2}. It is most negative and exhibits the largest variability between 2001 and 2006 (Figure 7.13b). This increase in variability reflects the increase in complexity with which aerosol-cloud interactions are simulated. Whereas early models used offline three-dimensional sulphate fields, state-of-the art GCMs have their own aerosol schemes and consider sea salt, mineral dust and carbonaceous aerosols in addition to sulphate. There does not seem to be a systematic tendency for models that use a parameterization based on cloud parcel models instead of empirical relationships between the aerosol mass/number concentration with the cloud droplet number concentration to have a larger or smaller indirect aerosol effect. Sensitivity studies did show that the iRF is larger if the background aerosol concentration is low (Chen and Penner, 2005) as this increases the cloud susceptibility. As shown by Storelvmo et al. (2009) different empirical relationships that are used to bypass cloud activation can cause a difference of 1.3 W m^{-2} in the iRF. The iRF also depends strongly on the assumed minimum cloud droplet concentration because that determines the susceptibility of the cloud (Hoose et al., 2009).The iRF is smallest (–0.4 W m^{-2}) if model data are rescaled to conform with observational constraints.

In response to the aerosol, there are multiple possible adjustments, such as changes to the cloud lifetime (cloud lifetime or second indirect aerosol effect), reduction in cloud cover due to absorption of solar radiation by BC or other absorbing aerosols (semi-direct effect) and aerosol effects on mixed-phase, ice and convective clouds (Denman et al., 2007) that are included in the iAF. The iAF amounts to –1.5 W m^{-2} if either changes in cloud lifetime alone or changes in cloud lifetime together with the direct and semi-direct aerosol effects are included in GCMs. Note that GCMs that use autoconversion rates of cloud droplets to form rain drops which depend inversely on the cloud droplet number concentration build in a cloud lifetime effect. In small-scale studies this does not lead to an increase in lifetime because small droplets also evaporate more readily (Jiang et al., 2006) but rapid timescale processes of this kind are not represented in GCMs (Lohmann and Feichter, 2005). The iAF tends to be smaller if changes to the cloud droplet size distribution (dispersion) are considered (e.g., Rotstayn and Liu, 2005) or if a prognostic equation for precipitation is introduced (Posselt and Lohmann, 2009) because that shifts the emphasis from the autoconversion rate to the accretion between cloud droplet and rain drops in better agreement with observations (Wood, 2005).

iAF is considerably smaller if aerosol effects on mixed-phase clouds are included in addition to the above mentioned adjustments in liquid clouds. The GCM average amounts to –1.1 W m^{-2}, only slightly larger than the iRF implying that the rapid adjustments (and inclusion of the direct aerosol effect) almost cancel each other. The spread between the different studies in the liquid+mixed category depends on the frequency of glaciation of supercooled clouds. If more IN are available in a polluted climate, supercooled clouds glaciate more readily and precipitate (see Section 7.4.4.1). In these cases an additional cooling stems from more

longwave radiation being emitted to space. If on the contrary IN become coated with soluble material and become less efficient, supercooled clouds remain longer in the atmosphere, which enhances the iAF of liquid clouds but leads to a small positive longwave effect that slightly reduces the shortwave cooling.

The iAF increases substantially in magnitude if aerosol particles are also allowed to change convective clouds. However, the uncertainty associated with these estimates is the largest and also it is unclear if convective clouds are characterized in sufficient detail in GCMs to warrant such estimates. As in the case of the iRF, rescaling model based estimates of the iAF, so as to match constraints from satellite retrievals, also reduces their magnitude (from –1.1 to –0.7 W m^{-2}, Figure 7.13a).

A complementary approach to estimate the iAF is to infer it as a residual using the observed temperature record over land, and estimates of the ocean heat uptake and the evolution of greenhouse gas and solar radiative forcing (Anderson et al., 2003; Hegerl et al., 2007). These approaches are called inverse estimates. They normally involve models of intermediate complexity. The only inverse study that obtained the iAF (Knutti et al., 2002) bracketed the iAF to be between 0 and –1.2 W m^{-2}. All estimates of iAF that involve satellite data fit into this range as do the mean values of the different iAF groups. An inverse estimate that is obtained purely from an energy balance perspective limits the iAF (including the direct effect plus other unknown residuals that are assumed to be small) since 1950 to be between –0.7 to –1.5 W m^{-2} (Murphy et al., 2009). The 5–95% confidence interval of all inverse estimates of the iAF is –0.4 to –1.3 W m^{-2} (Figure 7.13b). Again, all estimates of iRF that involve satellite data fit into this range as does the GCM average of the iAF that includes mixed-phase clouds. However, all other iAF averages exceed the 95% confidence interval of the inverse estimates. This is likely reflects limitations in our ability to parameterize clouds, aerosols, and aerosol-cloud interactions in GCMs.

Because GCMs tend to include negative forcings but not positive ones they tend to produce larger forcings than small-scale studies that include compensating processes and than inferred from observations (e.g., ship track studies or pure GCM estimates of the iAF and iRF vs. those that include satellite data). Therefore we use the GCM estimates of the iRF (average of iRF-TAR, iRF-AR4 and iRF-AR5 in Figure 7.13) as providing the lower bounds of –1 and –1.5 W m^{-2}. The lower bound of the iAF is obtained from the average of iAF-liquid-AR4 and iAF-liquid-AR5 as most of the other studies also include the direct effect. It amounts to –1.5 W m^{-2}. The upper bounds have been put at the smallest GCM estimates of –0.1 W m^{-2} and 0 W m^{-2} for the iRF and iAF, respectively, because of the indications that GCMs overestimate the forcings and that very small forcing values cannot be ruled out. The studies that take satellite data into account therefore arrive at much smaller forcings with a median of –0.4 W m^{-2} and –0.7 W m^{-2} for the iRF and iAF, respectively and an upper bound for the iRF of –0.2 W m^{-2}. We regard the median values of the studies including satellite data as a plausible but yet more uncertain estimate of the lower bound. Based on these arguments, we assess the iRF and iAF as follows: iRF is very likely between –1 and –0.1 W m^{-2}, the lower bound being based on GCMs, and likely between –0.4 and –0.2 W m^{-2}, the lower bound being based on studies that take satellite data into account. Following the same line of argumentation, iAF is very likely between –1.5 and 0 W m^{-2} and likely between –0.7 and –0.2 W m^{-2}.

[INSERT FIGURE 7.13a HERE]

Figure 7.13a: Model, satellite and inverse estimates of the iRF and the iAF since 1993. For each paper the best estimate per model is shown as a plus-sign. If multiple estimates or uncertainties are given in a paper, the ranges are shown as vertical lines bounded by diamonds. The thin horizontal lines denote the average of the respective group and the width of the coloured box denotes its standard deviation. The iRF studies from GCMs are divided into those published prior to TAR: iRF-TAR (Boucher and Lohmann 1995; Chuang et al., 1997; Feichter et al., 1997; Jones et al., 1994; Kaufman and Chou 1993; Kiehl et al., 2000; Lohmann and Feichter 1997; Lohmann et al., 2000; Rotstayn 1999), between TAR and AR4: iRF-AR4 (Chen and Penner 2005; Chuang et al., 2002; Ghan et al., 2001; Hansen et al., 2005; Jones et al., 2001; Kristjansson 2002; Ming et al., 2005; Penner et al., 2006; Quaas and Boucher 2005; Quaas et al., 2004; Rotstayn and Penner 2001; Rotstayn and Liu 2003; Suzuki et al., 2004; Takemura et al., 2005; Williams et al., 2001) and since 2007: iRF-AR5 (Barahona et al., 2011; Bellouin et al., 2011; Haerter et al., 2009; Kvalevag and Myhre 2007; Lohmann et al., 2007; Lohmann et al., 2010; Penner et al., 2011; Rotstayn and Liu 2009; Storelvmo 2011; Storelvmo et al., 2009; Wang and Penner 2009). iAF studies on liquid clouds that include the cloud albedo and cloud lifetime effect are also divided into those published until 2006: iAF-liquid-AR4 (Easter et al., 2004; Ghan et al., 2001; Johns et al., 2006; Jones et al., 2001; Kristjansson 2002; Kristjansson et al., 2005; Lohmann 2002b; Lohmann and Feichter 1997; Lohmann et al., 2000; Menon et al., 2002; Ming et al., 2005; Peng and Lohmann 2003; Penner et al., 2003; Penner et al., 2006; Quaas et al., 2006; Rotstayn 1999; Rotstayn and Penner 2001; Rotstayn and Liu 2005; Storelvmo et al., 2006; Takemura et al., 2005; Williams et al., 2001) and since 2007: iAF-liquid-AR5 (Chen et al., 2010; Ghan et al., 2011b;

Hoose et al., 2009; Kirkevag et al., 2008; Makkonen et al., 2011; Menon and DelGenio 2007; Ming et al., 2007b; Penner et al., 2011; Quaas et al., 2009b; Rotstayn and Liu 2009; Storelvmo et al., 2008a); iRF and iAF estimates that involve satellite data are shown in pink: iRF-satellites (Dufresne et al., 2005; Lebsock et al., 2008; Quaas and Boucher 2005; Quaas et al., 2008; Quaas et al., 2009b; Storelvmo et al., 2009) and iAF-satellites (Lohmann and Lesins 2002; Quaas et al., 2006; Quaas et al., 2009b; Sekiguchi et al., 2003), inverse estimates for the iRF and iAF are shown in turquoise: iRF-inverse (Knutti et al., 2002) and iAF-inverse (Anderson et al., 2003; Andronova and Schlesinger 2001; Church et al., 2011; Forest et al., 2006; Forest et al., 2002; Gregory et al., 2002; Hansen et al., 2011; Harvey and Kaufmann 2002; Huber and Knutti 2011; Libardoni and Forest 2011; Murphy et al., 2009; Shindell and Faluvegi 2009; Stott et al., 2006) in turquoise. iAF studies that include the direct and semi-direct effect in lilac: iAF-liquid+dir+SD (Ghan et al., 2011a; Lohmann and Feichter 2001; Lohmann et al., 2007; Posselt and Lohmann 2008; Posselt and Lohmann 2009; Quaas et al., 2004; Quaas et al., 2006; Quaas et al., 2009b; Rotstayn et al., 2007; Salzmann et al., 2010), those that additionally consider aerosol effects on mixed-phase clouds in purple: iAF-liquid+mixed (Hoose et al., 2008; Hoose et al., 2010b; Jacobson 2006; Lohmann 2004; Lohmann and Diehl 2006; Lohmann and Hoose 2009; Lohmann and Ferrachat 2010; Salzmann et al., 2010; Storelvmo et al., 2008a; Storelvmo et al., 2008b), and those that treat aerosol effects in stratiform and convective clouds in green: iAF-liquid+conv (Koch et al., 2009a; Lohmann 2008; Menon and Rotstayn 2006; Menon and DelGenio 2007; Unger et al., 2009; Wang et al., 2011b). For the inverse estimates no best estimate is shown and the turquoise colour box denotes the average of the lower and upper bounds of these studies, respectively.

[INSERT FIGURE 7.13b HERE]

Figure 7.13b: Box plots of model, satellite and inverse estimates of the IF and the AIF since 1993 for the same groups of estimates as in Figure 7.13a provided at least 6 estimates are available. Displayed are the averages (red stars), median values (blue lines), 33% and 67% percentiles (box boundaries) and 5% and 95% percentiles (ends of vertical lines) except for the inverse estimates, which is an expert assessment of the combined estimate of multiple inverse estimates.

Finally Table 7.3 provides values of the aerosol AF as diagnosed in simulations of the CMIP5 models which provide climate projections in this report. [PLACEHOLDER FOR SECOND ORDER DRAFT: to discuss differences between our best estimates and CMIP5]

Table 7.3: Estimates of aerosol AF (in W m^{-2}) in the CMIP5 models. The AF are estimated from fixed-SST experiments using the atmosphere-only version of the models listed. Different models include different aerosol effects. [PLACEHOLDER FOR SECOND ORDER DRAFT: table will be updated as data become available on the CMIP5 archive.]

Modelling Group Model	CCCma CanESM2	CSIRO-QCCCE CSIRO-Mk3-6-0	IPSL IPSL-CM5A-LR	MOHC HadGEM2-A	NCC NorESM1-M	MPI-M
Anthropogenic sulfate aerosol		–1.10	–0.71	–1.16		
All anthropogenic aerosol	–0.87	–1.41		–1.22	–0.99	–0.35

7.4.7 Impact of Cosmic Rays on Aerosols and Clouds

A high solar activity leads to a more complex magnetic configuration of the heliosphere, which reduces the flux of galactic cosmic rays (GCR) in the Earth's atmosphere. It has been hypothesised that a lower flux of GCR would modify cloudiness in a way that would amplify the warming effect expected from an increase in solar activity. There have been many studies aiming to test this hypothesis since AR4, which fall in two categories: i) studies that seek to establish a causal relationship between cosmic rays and aerosols/clouds by looking at correlations between the two quantities on timescales of days to decades, and ii) studies that test through observations or modelling one of the physical mechanisms that have been put forward. We assess these two categories of studies in the next two sections.

7.4.7.1 Correlations Between Cosmic Rays and Properties of Aerosols and Clouds

Many empirical relationships or correlations have been reported between GCR or cosmogenic isotope archives and some aspects of the climate system, such as SSTs in the Pacific Ocean (Meehl et al., 2009), some reconstruction of past climate (Kirkby, 2007) or tree rings (Dengel et al., 2009). We focus here on observed relationships between GCR and aerosol- and cloud-properties. Such relationships have focused on

decadal variations in GCR induced by the 11-year solar cycle, shorter variations associated with the quasi-periodic oscillation in solar activity centred on 1.68 years or sporadic variations associated with so-called Forbush decrease events that happen on timescales of days. It should be noted however that such correlations could arise for reasons related to changes in atmospheric heating and circulation, rather than an impact of cosmic rays.

Some studies have shown co-variation between GCR and low-level cloud cover using global satellite data over periods of typically 5–10 years (e.g., Marsh and Svensmark, 2000; Svensmark and FriisChristensen, 1997). Such correlations have not proved to be very robust when extending the time period under consideration (Agee et al., 2011), restricting the analysis to particular cloud types (Kernthaler et al., 1999) or locations (Udelhofen and Cess, 2001; Usoskin and Kovaltsov, 2008). Some authors have attributed the purported correlations to other factors such as El Nino (Farrar, 2000) and artefacts of the satellite data due to the solar cycle cannot be ruled out (Pallé, 2005). However statistically significant correlations have been found in some locations. Harrison and Stephenson (2006) examined the relationship between diffuse radiation and cosmic rays at the surface for some UK sites between 1951 and 2000. They found that the 87% of days with the highest cosmic ray flux had a 19±4% higher chance of being overcast and the diffuse fraction in incoming surface solar radiation increases by 2±0.3%. Moreover Forbush reductions in cosmic rays corresponded to a simultaneous decrease in diffuse fraction. Similarly Harrison (2008) found a unique 1.68 year cosmic ray periodicity in surface radiation for two different UK sites between 1978 and 1990, which they believe is associated with variations in cloudiness. These cloud responses were found to occur within a day (Harrison and Ambaum, 2010). However, large reduction events in cosmic rays did not occur frequently enough in the record to generate robust statistics. Svensmark et al. (2009) found large global reductions in the aerosol Angström exponent from AERONET, liquid water path from SSM/I, and cloud cover from MODIS and ISCCP after large Forbush decreases with a lag of 5 to 9 days. This finding has been questioned by Laken et al. (2009). The study by Kristjansson et al. (2008) suggests a weaker impact of Forbush decrease event on clouds over the Southern Ocean. Moreover the 5–9 day lag found by Svensmark et al. (2009) does not match the rapid response observed in Harrison and Stephenson (2006) and Harrison and Ambaum (2010). Further the studies of Calogovic et al. (2010) and Kristjansson et al. (2008) did not find a global cloud effect. Laken et al. (2010) and Rohs et al. (2010) found a very weak but significant positive correlation between GCR and high- and mid-altitude clouds. A problem with all these studies is that very few large Forbush decrease events have occurred during the satellite era (~6) and the statistics are sensitive to how the Forbush events are selected (Laken et al., 2009). Finally it should be noted that Kulmala et al. (2010) found no connection between GCR and new particle formation over a solar cycle (1996–2008); however the measurements of new particle formation were restricted to one surface station in Finland and may not be representative of new particle formation in the free troposphere.

7.4.7.2 Physical Mechanisms Linking Cosmic Rays to Cloudiness

Several physical mechanisms have been put forward to explain the possible link between GCR and cloudiness. The most widely studied of these is the “ion-aerosol clear air” mechanism, in which atmospheric ions produced by GCR facilitate aerosol nucleation and growth with a further impact on CCN concentrations and cloud properties (Carslaw et al., 2002; Usoskin and Kovaltsov, 2008). The variability of atmospheric ionization rates due to GCR changes can be considered relatively well quantified (Bazilevskaya et al., 2008), whereas resulting changes in aerosol nucleation rates are very poorly known (Enghoff and Svensmark, 2008; Kazil et al., 2008). The Cosmics Leaving OUtdoor Droplets (CLOUD) experiment at CERN has been designed to simulate conditions close to the real atmosphere and is currently the only laboratory experiment on ion-induced nucleation in which it has been possible to isolate the role of ions produced by GCRs (Kirkby et al., 2011). The CLOUD experiment indicates that GCR-induced ionization enhances water–sulphuric acid nucleation in the middle and upper troposphere, but is very unlikely to give a significant contribution to nucleation taking place in the continental boundary layer. Field measurements support qualitatively this view but cannot provide any firm conclusion on the role of ions because of the scarcity and other limitations of free-troposphere measurements (Arnold, 2006; Mirme et al., 2010) and difficulties in separating GCR-induced nucleation from other nucleation pathways in continental boundary layers (Hirsikko et al., 2011; Manninen et al., 2010). Regardless of the exact aerosol formation mechanism, a significant fraction of CCN in the boundary layer may originate from aerosol particles nucleated in the free troposphere (Merikanto et al., 2009).

Observational evidence for particle nucleation in the cloudy marine boundary layer is relatively rare with some notable exceptions (e.g., Hegg et al., 1990; Petters et al., 2006). Nucleation (neutral and charged) appears to require special conditions, namely low ambient aerosol, high DMS, photochemically produced OH (to oxidise DMS to SO_2), relatively high RH and relatively low temperature. The previously discussed open-cellular cloud structure appears to provide adequate conditions for nucleation. Updrafts in open cell walls supply DMS produced at the ocean surface to an ultra-clean layer near the top of the boundary layer that has been scavenged of aerosol by precipitation. High actinic flux because of the presence of adjacent clouds produces OH, which oxidises the DMS to SO_2 and H_2SO_4. Petters et al. (2006) observed new particle formation within open cells and more recently Kazil et al. (2011) modeled the formation of particles via neutral and charged H_2SO_4/H_2O nucleation in a mesoscale cloud-resolving model representing the aerosol lifecycle. The relative importance of surface aerosol production, entrainment from the free troposphere, and nucleation are shown depend on factors such as surface wind-speed, entrainment rates, and just how favourable ambient conditions are for nucleation. To date, no process-level models have looked at the role of the phase of the solar cycle, via its influence on nucleation rates, on this balance of aerosol sources and their importance for maintaining boundary layer clouds.

Our understanding on the "ion-aerosol clear air" as a whole relies on a few model investigations that simulate GCR changes over a solar cycle (Kazil et al., 2006; Pierce and Adams, 2009a; Snow-Kropla et al., 2011) or during strong Forbush decreases (Bondo et al., 2010; Snow-Kropla et al., 2011). Although all model studies found a detectable connection between GCR variations and either CCN changes or column aerosol properties, the response appears to be too weak to cause a significant radiative effect because of the low sensitivity of CCN concentrations to the nucleation caused by GCR.

A second mechanism has been proposed by which ionization in the atmosphere may have an impact on clouds. GCR ionization modulates the fair-weather current in the global electrical circuit and it has been hypothesised that droplet charging could modify supersaturation and temperature at the cloud base. Harrison and Ambaum (2010) found some observational evidence of this with a small reduction in downward longwave radiation associated with variations in surface current density, but the evidence remains very low.

7.4.7.3 Synthesis

In summary, there is evidence from laboratory, field and modelling studies that ionization from cosmic rays may enhance aerosol nucleation in the free troposphere. However there is medium evidence and high agreement that the cosmic ray-ionization mechanism is too weak to influence global concentrations of CCN or their change over the last century or during a solar cycle in a climatically-significant way. Moreover it should be noted that one study infers no trend in cosmic ray intensity over the last 50 years (McCracken and Beer 2007).

7.5 Solar Radiation Management and Related Techniques

7.5.1 Introduction

Geoengineering is a term often used to describe the deliberate large scale intervention in the Earth system to counter undesirable impacts of climate change on the planet (e.g., Keith, 2000). One class of geoengineering methods is based upon manipulating the energy budget of the planet. Solar Radiation Management (SRM) methods aim to achieve a planetary cooling by reducing the amount of solar energy absorbed by the climate system. They are discussed here because some of these methods involve clouds and/or aerosols. A related technique is also discussed that seeks to deliberately decrease the greenhouse effect in the climate system by altering high-level cloudiness. The other class of geoengineering methods known as Carbon Dioxide Reduction (CDR) is discussed in Chapter 6.

This section restricts its assessment to a "physical climate" perspective of methods published in the scientific peer-reviewed literature that appear to influence components of the energy budget by at least a few tenths of a W m^{-2} in the global mean and for which a plausible technology exists. Cost, implementation and governance issues are beyond the scope of this section. [PLACEHOLDER FOR SECOND ORDER DRAFT: reference to WGIII AR5 (tbc)]. Geo-engineering techniques were not discussed in previous WGI assessments but were mentioned in WGII (Klein et al., 2007) and WGIII (Barker et al., 2007) AR4.

Virtually all research on SRM has followed one of two paths: i) theoretical and modelling studies and ii) a search for analogues to SRM, and the impact of those phenomena to the planet (e.g., large volcanic eruptions in the stratosphere such as Pinatubo, ship tracks, inadvertent climate modification from anthropogenic aerosols, etc). The scientific tools required to explore SRM methods and their impact on the planet are essentially the same as those needed to understand and predict the impacts of other forcing agents. The predicted climate changes in response to SRM are subject to the same limitations and uncertainties as predictions of future climate change. We assess geoengineering from two points of view: i) issues surrounding the production of the radiative forcing and ii) the climate response and other impacts generated. As with other types of climate forcing, there are serious limitations in our capability to predict regional responses of the climate system to SRM.

7.5.2 Idealised Experiments

Some aspects of SRM methods can be explored in an idealised way by artificially reducing the solar constant in a climate model. This results in a global cooling, albeit with a climate efficacy generally smaller than 1, and a reduction in the global-mean precipitation. Earlier (Bala et al., 2008) and more recent (GeoMIP, Kravitz et al., 2011) model experiments where the RF by CO_2 is exactly balanced by a reduction in the solar constant show some residual surface temperature changes, especially at high latitudes, and a reduction in the global-mean precipitation which can be explained by arguments on the energy budget of the atmosphere (Allen and Ingram, 2002; Andrews et al., 2010; Held and Soden, 2006). Figure 7.14 shows the annual-mean temperature and precipitation changes produced by experiment G1 of the Geoengineering Model Intercomparison Project for a subset of 4 models (GeoMIP, Kravitz et al., 2011).

[PLACEHOLDER FOR SECOND ORDER DRAFT: a more complete discussion of impacts on T, P, sea-ice and soil moisture in the GeoMIP models.]

[INSERT FIGURE 7.14 HERE]

Figure 7.14: Multi-model mean of the residual surface temperature and precipitation changes from GeoMIP simulations with a simultaneous fourfold increase in CO_2 and a reduction in solar forcing which has been adjusted in each model to maintain the top of atmosphere net flux imbalance within ±0.1 W m^{-2} (Kravitz et al., 2011).

Idealized experiments have also been conducted where solar radiation is reduced only over the ocean (e.g., to mimic the effects of marine cloud seeding or increased sea foam). Bala et al. (2010) suggested that although global-mean precipitation might decrease, precipitation over land might actually increase, because of increased gradients in RF between land and ocean. Other studies have explored albedo changes over land (e.g., to represent a surface albedo increase through plant albedo and desert regions) and over particular latitudinal bands (Caldeira and Wood, 2008) or regions (Irvine et al., 2010; Irvine et al., 2011).

7.5.3 Stratospheric Aerosols

Observations in the aftermath of major volcanic eruptions like Pinatubo demonstrate that increasing stratospheric aerosols will cool the planet and it has been suggested that global warming might be deliberately countered with a continuous release of sulphur species (Budyko, 1974; Crutzen, 2006) to mimic this. Most of the research on stratospheric aerosol SRM to date has explored the possibility of forming sulphuric acid aerosols by injecting sulphur containing gases in the stratosphere, although using BC (Crutzen, 2006; Keith, 2010) or metal oxides (Keith, 2010) has also been suggested.

There are many subtleties to SRM by stratospheric aerosols (Rasch et al., 2008b). The evolving size of the particles in the stratosphere has profound effects on the viability of the strategy, with impacts on radiative forcing by unit of injected mass, stratospheric ozone chemistry, and climate response. Initial modelling studies recognized this fact but prescribed the aerosol size, assuming it would range between small sizes characteristic of background conditions and the larger sizes observed soon after a major volcanic eruption (Rasch et al., 2008a). Heckendorn et al. (2009) found that particle size from a continuous injection of gases oxidizing to sulphuric acid particles could be very inefficient, because much of the sulphuric acid would condense on particles already present from earlier SRM emissions. Particles would be likely to grow larger, become less efficient per unit mass at scattering energy back to space, and sediment quicker out of the

stratosphere as the injection flux increases. They found that at least a four times higher injection rate would be required to double the forcing for the scenario they explored. Pierce et al. (2010) have suggested a way around this by introducing a source of sulphuric acid gas that immediately condenses to particles, bypassing some coagulation and deposition processes that lead to particle growth from gaseous sulphur precursors.

Several modelling studies suggest that it would be possible to stabilize global average surface temperature at least through a doubling of CO_2 concentrations using stratospheric sulphate aerosol with many signatures and planetary consequences that are similar to those found in the idealized studies, i.e., a residual warming at high latitudes (assuming a more or less homogeneous distribution of stratospheric aerosols) and a reduction in the global-mean precipitation.

Dispersion and lifetime of the aerosol dispersed in the stratosphere is a strong function of height and latitude of the injection, with high latitude and lower injection altitudes being less effective. Local injections of aerosol precursors at high latitudes will produce aerosol extending over sizable fraction of a hemisphere. Early studies (Jones et al., 2010a; Rasch et al., 2008b; Robock et al., 2008) used somewhat different experimental protocols and found significant disagreement in regional responses to stratospheric aerosols. It is unclear whether differences in the regional responses are due to the experimental protocol, or to model differences. [PLACEHOLDER FOR SECOND ORDER DRAFT: GeoMIP update]

Observations also show other impacts from volcanic eruptions like Pinatubo. There are measurable effects on the hydrologic cycle (Trenberth and Dai, 2007) similar to those found in idealised experiments, impacts on stratospheric ozone, and the ratio of direct to diffuse sunlight reaching the Earth's surface. These effects are also expected to occur from geoengineering with sulphate aerosols (see Rasch et al., 2008b and references therein). Tilmes et al., (2009) used a model with a well resolved middle atmosphere and stratospheric chemistry to explore responses to stratospheric aerosol SRM and found discernable shifts in tropopause altitude (lifting by 1 km), and changes in ozone abundance (depletion at high latitudes and increases in the tropics). This change is ozone might have discernable impacts on UV light reaching the surface, although some earlier calculations suggest that there is some degree of compensation between increases in UV associated with ozone depletion and decreases associated with attenuation by the aerosols themselves (Vogelmann et al., 1992). A decrease in direct radiation and increase in diffuse radiation reaching the Earth's surface may increase photosynthesis in terrestrial ecosystems (Mercado et al., 2009; see Chapter 6) and impact some systems that exploit renewable solar energy [WGII Chapter xx].

7.5.4 Cloud Brightening

Boundary layer clouds act to cool the planet, and relatively small changes in cloud albedo, lifetime, or areal extent can have profound effects on the energy budget of the planet (e.g., Slingo, 1990). Latham (1990) suggested that it might be possible to deliberately increase cloud albedo as a mechanism for countering global warming by introducing additional sea salt particles into the marine boundary layer, to act as CCN, "brightening" clouds through the aerosol-cloud indirect mechanisms described in Section 7.4. The idea has been examined using models at various scales (cloud parcel models, large eddy simulations, and climate models (e.g., Latham et al., 2008). Examples of cloud brightening include shiptracks produced in marine stratocumulus clouds by emissions of particles from freighters and changes in trade cumulus cloud properties produced by a relatively weak but continuous volcanic eruption of SO_2 (Yuan et al., 2011).

Changing cloud morphology (e.g., from open to closed cells) or changing low-liquid water clouds to high liquid water clouds have the potential to create large radiative forcings. Our current understanding suggests that marine stratocumulus clouds are an optimal cloud type for brightening because of their relatively low values of CDNC and the longer lifetime of sea-salt particles in non-precipitating environments. However these clouds occupy a relatively small fraction of the planet and large RF (30–100 W m^{-2}) would be required locally to produce globally-averaged changes of the order of 1–5 W m^{-2}. Studies cited in Section 7.4 highlight the importance of the details of aerosol-cloud interactions in influencing cloud albedo and lifetime producing very large uncertainties about the viability of cloud brightening for SRM. Wang et al. (2011a) explored the sensitivity of marine stratocumulus in various meteorological regimes and levels of background aerosol amounts to aerosol injection strategies using a cloud system resolving model. That work demonstrated very strong interactions between aerosol distribution, precipitation and cloud morphology and differing responses in each scenario. Korhonen et al. (2010a) showed that "competition effects" between the

geoengineering aerosol and ambient aerosol could be important, with the ambient and SRM aerosol populations competing for liquid water, sometimes reducing, rather than enhancing albedo in some circumstances.

[PLACEHOLDER FOR SECOND ORDER DRAFT: contribution of geo-engineered sea-salt particles to the direct effect to be discussed].

Climate model studies (Jones et al., 2009; Latham et al., 2008; Rasch et al., 2009) that assumed geoengineering would directly influence cloud drop number changed cloud albedo and produced global average RF as negative as –5 W m^{-2}. These studies also indicated some changes in regional precipitation patterns, although the sign and amplitude of the changes differed between studies.

7.5.5 Surface Albedo Changes

It has also been suggested that planetary albedo can be increased by local changes to the albedo of urban areas, croplands, grasslands, deserts and the ocean surfaces.

Rosenfeld et al. (1998) proposed to increase the albedo of urban areas as a way to improve air quality and make cooling-energy savings in buildings especially during summertime. Hamwey (2007) estimated the potential RF from whitening roofs and pavements at –0.17 W m^{-2} but more recent estimates accounting for more realistic artificial surface area per capita and appropriate atmospheric radiative transfer suggest significantly less negative values (Akbari et al., 2009; Lenton and Vaughan 2009; Oleson et al., 2010).

Hamwey (2007) further suggested that increasing the albedo of the world's grassland (meaning open shrubland, grassland and savannah) by replacing native species by other natural or bioengineered species. Their RF estimate of –0.59 W m^{-2} (assuming the grassland albedo can be increased by 25% from an average value of 0.17) has been revised to –0.5 W m^{-2} by Lenton and Vaughan (2009) when accounting for the atmospheric absorption of the radiation reflected by the surface. Ridgwell et al. (2009) and Doughty et al. (2011) extended the concept of increased surface albedo to croplands and found the maximum effect over summertime mid-latitudes (e.g., 0.25 K per 0.01 increase in surface albedo in regions north of 30°N). Ridgwell et al. (2009) estimated a global-mean surface cooling of 0.11 K for a +0.04 increase in cropland albedo. Both studies pointed out to potential feedbacks in low-latitude regions with a reduction in soil moisture, cloud cover and precipitation. The potential for increasing crop and grassland albedo across a wide variety of species remains unproven.

Irvine et al. (2011) tested the impact of increasing desert albedo up to 0.80 in the HadCM3 model. This cooled surface temperature by –1.1 K (versus –0.22 and –0.11 K for their largest crop and urban albedo change). They also simulate significant land precipitation changes, with large reduction in rainfall over the Indian and Sahel regions.

The low albedo of ocean surfaces and large areal extent mean only a small increase in albedo could be sufficient to offset several W m^{-2} of RF by greenhouse gases. Engineering techniques (Evans et al., 2010; Seitz, 2011) have been proposed to increase the fraction of the oceans covered with foam because of its large albedo (Whitlock et al., 1982). However ocean foam is short-lived and artificial foam would somehow have to be engineered to last longer. Neither the extent of foam generation and persistence required for a significant climate impact nor the impact of artificial foam on the world's ocean (including ocean biology, air-sea fluxes of latent heat, sensible heat and trace gases, and surface emissivity) have been assessed in the peer-reviewed literature.

7.5.6 Cirrus Thinning

Cirrus clouds affect both outgoing longwave radiation (OLR) and absorbed solar radiation. Thin high cirrus above 300 hPa affects OLR more than incoming solar energy and thereby contribute to warming the climate (see Section 7.2). Reducing the coverage or longwave opacity of these clouds would therefore contribute a negative RF. Cirrus cloud coverage is sensitive to the ice fall speed which depends on ice crystal size. By increasing ice crystal size in the coldest cirrus ice crystals could fall out and reduce the overall coverage (Mitchell and Finnegan, 2009). Although an aerosol cloud seeding mechanism has been proposed to increase

crystal size, cirrus nucleation processes are not yet well enough understood to provide a firm basis to this method (Section 7.4).

[START FAQ 7.1 HERE]

FAQ 7.1: How do Aerosols Affect Climate and Climate Change?

[INSERT FAQ 7.1, FIGURE 1 HERE]
FAQ 7.1, Figure 1: Overview of aerosol direct and indirect effects on climate.

It is believed that man-made variations in atmospheric aerosols are responsible for a cooling which have partially masked the warming from man-made greenhouse gases.

Atmospheric aerosols are small particles suspended in the atmosphere with a typical lifetime of 1–2 weeks in the troposphere and 1–2 years in the stratosphere. There are many types of aerosols, which can be of natural (e.g., dust, sea-salt, some biogenic compounds) or anthropogenic (e.g., sulphates, soot, biomass burning aerosols) origin. Atmospheric aerosols exhibit large variations in size, shape and chemical composition. Changes in the climate can be caused either by emissions of anthropogenic aerosols and their precursors, or by changes in natural aerosols which themselves respond to other changes in the climate system (e.g., increase in dust due to a regional drying).

Aerosols affect climate in multiple ways. First they scatter and absorb sunlight which modifies the planet radiative balance, an effect known as the *aerosol direct effect* (see FAQ.7.1, Figure 1). Aerosol scattering generally results in a more reflective planet and a cooler global climate, while absorption results in a less reflective planet and a warmer climate. Sulphate aerosols from fossil fuel burning are especially important in scattering, while soot from some combustion sources is an important absorber. The balance between cooling and warming depends on the aerosol properties and the environmental conditions. Many observational and modelling studies have been done to quantify the global direct effect from anthropogenic and natural aerosols. While these remain uncertain, studies have consistently indicated that the direct effect from anthropogenic aerosols has been to cool the planet relative to what would otherwise have occurred. One of the remaining uncertainties comes from aerosol absorption, which is more difficult to measure than scattering and induces a specific cloud response (known as the *semi-direct effect*).

Since aerosols are distributed unevenly in the atmosphere, they can heat and cool the climate system in patterns that can drive subtle changes to the weather, affecting cloud or rainfall amounts. These effects are complex and hard to predict with current models, but several studies suggest significant effects on precipitation in certain regions.

Aerosols also serve as condensation and freezing sites for cloud droplet and ice particle formation (see FAQ.7.1, Figure 1). While it might seem that more condensation nuclei would increase the amount of low clouds, cloud formation is largely limited by dynamical processes so that the net effect on clouds of more aerosols is quite subtle and remains uncertain. A robust result is that more aerosols tend to produce liquid clouds with more numerous and smaller particles, everything else being equal. This and other impacts on clouds alter their reflectivity, producing what are called *aerosol indirect effects* on climate. Indirect effects can arise through many pathways, particularly in ice or mixed liquid and ice clouds where phase changes are sensitive to aerosols. Quantifying the overall impact is understandably more difficult than for the direct effect, but again available studies generally indicate that the net indirect effect of anthropogenic aerosols has been to further cool the climate system over the industrial period, enhancing their direct effect.

Because of their short lifetime, the abundance of aerosols and their climate effects have varied over time in rough concert with anthropogenic emissions of aerosols and their gaseous precursors, and variations in natural sources. Since anthropogenic emissions have increased substantially over the industrial period, this has very likely counteracted some of the warming that would otherwise have occurred from increased concentrations of long-lived greenhouse gases. Aerosols from volcanic eruptions such as those of the El Chichón and Pinatubo have also caused sporadic cooling periods. Trends in anthropogenic aerosol emissions over the last couple of decades have varied regionally (e.g., decreased emissions in industrialised countries,

increased emissions in developing countries) and it is difficult to assess whether the global impact has been to cool or warm the planet over the recent period. It is very likely, however, that emissions of anthropogenic aerosols will ultimately decrease. When this happens, decreasing aerosol emissions will begin to augment greenhouse-gas induced warming.

Some studies have hypothesised that climate change could feed back on the lifecycle of atmospheric aerosols, including natural aerosols such as sulphate, sea salt or biogenic aerosols. However there is contradicting evidence so far whether this could be a large effect at the global scale over the coming century. Moreover the sign of such a feedback is not known and could vary regionally.

[END FAQ 7.1 HERE]

[START FAQ 7.2 HERE]

FAQ 7.2: How do Clouds Affect Climate and Climate Change?

The importance of potential changes in cloudiness for the problem of climate change has been recognized as a key factor since the 1970s. Clouds affect the climate system in a variety of ways. They produce precipitation (rain and snow) that is necessary for life. They strongly affect the flows of both solar and infrared radiation through the atmosphere. Finally, they are intimately associated with powerful vertical motions that can carry air from near the surface to the upper troposphere in less than an hour. The strong vertical currents carry energy, moisture, momentum, and various chemical constituents, including aerosols. Each of the various cloud processes has the potential to change as the climate state evolves. Any change in a cloud process that is caused by a climate change and in turn influences climate represents a cloud-climate feedback.

Cloud feedbacks are of intense interest in the context of anthropogenic climate change. Many types of possible cloud-climate feedbacks have been identified. Broadly speaking, they would occur through changes in cloud amount, cloud top-height, and/or cloud optical properties. We still are not sure what types of cloud feedbacks will actually occur and how significant they will be for climate change. Nevertheless, all of the models used for the fourth IPCC Assessment produced either a positive or near-neutral cloud feedback. The differences in cloud feedbacks among the models strongly influenced their differences in climate sensitivity.

Low clouds reflect a lot of solar radiation back to space, but have only a weak effect on the infrared radiation emitted by the Earth. As a result, they tend to cool the Earth, in the present climate. In a future climate warmed by increasing greenhouse gases, an increase in low cloud amount would increase the cooling, and so could reduce the warming. On the other hand, a decrease in low-cloud amount would increase the warming.

Conversely, high cold clouds such as cirrus clouds are often somewhat transparent, so they do not reflect as much solar radiation, but they can still absorb the infrared radiation coming from the Earth's surface, leading to a warming near the cloud-base level and reducing the energy Earth loses to space. They therefore tend to warm the Earth as a whole. An increase in high cloud amount would tend to enhance greenhouse warming, while a decrease would tend to reduce it. Even if the high-cloud amount remained the same, high clouds could produce a positive feedback as the surface warms up, because they would prevent the extra infrared energy emitted by the warmer surface from leaving the climate system.

The amount of sunlight a typical cloud would reflect in a warmer climate could be different for many reasons. As an example, a warmer climate may see more clouds made of liquid drops, and fewer made of ice crystals. That could lead to a change in the overall amount of light reflected. Clouds are also affected in many ways by aerosols (see FAQ 7.1), which may have caused significant past changes (or may cause future changes) in cloud reflectivity independent of any caused by climate change. Subtle changes in wind patterns associated with transient or longer-term climate changes would also likely affect clouds.

In a climate change, there can be many different changes in the geographical patterns and seasonal distributions of both high and low clouds. The net cloud feedback results from the combined effect of these various changes.

For decades, climate scientists have been using observations to study how clouds change with the daily weather, with the seasonal cycle, and with year-to-year changes such as those associated with El Niño. We have also been working to improve the simulation of clouds in climate models. Many current models predict a moderately positive net cloud feedback, in which both low and high clouds feed back positively. Work continues to further evaluate and refine these results.

[PLACEHOLDER FOR SECOND ORDER DRAFT: addition of comment about confidence in current models is being considered]

The net feedback from clouds on global climate, if any, will almost surely result from the net effect of many diverse regional changes. This makes predicting the cloud feedback very difficult. While it would be desirable to infer this long-term cloud feedback somehow from observations, there is no way to do this that is broadly accepted as valid. To predict cloud phenomena comprehensively requires a global climate model; these models produce cloud fields that roughly resemble those observed, but are far from perfect. Models vary in how they predict clouds will change in a warmer climate, but so far no model has predicted changes in clouds so large that they significantly limit global warming, and nearly all models predict that cloud changes will actually amplify global warming.

[END FAQ 7.2 HERE]

[START FAQ 7.3 HERE]

FAQ 7.3: Could Geoengineering Counteract Climate Change and What Side-Effects Might Occur?

There are two different categories of geoengineering methods which are usually referred to as Solar Radiation Management (SRM, assessed in Chapter 7) and Carbon Dioxide Removal (CDR, assessed in Chapter 6). A less technical name for SRM is Sunlight Reflection Management. We discuss these in turn.

Carbon dioxide removal methods

By definition, CDR methods seek to accelerate the removal of CO_2 from the atmosphere and store it in land, ocean or geological reservoirs. Afforestation/reforestation, carbon sequestration in soils, bioenergy associated to carbon capture and storage, ocean fertilization, accelerated weathering of silicate and carbonate rocks and CO_2 air capture using chemical methods are some of the proposed CDR methods (see FAQ.7.3, Figure 1).

[INSERT FAQ 7.3, FIGURE 1 HERE]
FAQ 7.3, Figure 1: Overview of carbon dioxide removal methods.

CDR methods rely primarily on natural carbon cycle processes, either biological or chemical: enhanced biological production by photosynthesis on (1) land and (2) oceans, (3) accelerated chemical weathering reactions over (3) land and oceans and (4) enhanced solubility pump in the oceans. Direct air capture is an exception which relies on artificial chemical methods to remove CO_2 directly from air. Once captured, CO_2 would be stored within land and ocean reservoirs or geological formations. Land storage occurs in organic form but storage in oceans and geological formations is in inorganic forms.

To have a noticeable climate effect, CDR schemes should be able to remove several PgC per year from the atmosphere over several decades in this century. Important scientific considerations include the storage capacity and permanence of the reservoirs, and potential adverse side effects. CDR methods cause a "rebound effect": when carbon is stored in one reservoir, the concentration gradient between the atmosphere and carbon reservoirs is reduced and thereby the subsequent inherent rate of removal of CO_2 from the atmosphere by natural reservoirs is reduced or could even be reversed.

In general, CDR methods are believed to be relatively low risk in terms of unintended climatic side effects because they counter the root cause by reducing atmospheric carbon dioxide concentrations. CDR schemes

also reduce direct consequences of high CO_2 levels including surface ocean acidification. However, proposed CDR methods have limited potential to rapidly decrease the atmospheric concentration of CO_2. The large thermal inertia of climate system need to be also considered: many components of the earth system may continue to respond for decades or centuries to the original increases in CO_2 even after CDR is applied. Therefore, decreases in surface temperature would lag CDR-induced decreases in atmospheric CO_2 concentrations.

There are some potential climate or environmental effects from CDR methods. Some examples of the side effects are: 1) removal of atmospheric CO_2 would lead to a temporary acceleration in the global water cycle. 2) Implementation of CDR methods could lead to reduced plant productivity when compared to the elevated level expected with high CO_2 concentration. 3) Large scale biological production over land could have climate consequences by altering the surface characteristics such as surface reflectivity and evapotranspiration. For instance, many modelling studies have shown that afforestation in seasonally snow covered regions could accelerate global warming. 4) In the case of ocean fertilization, the utilization of macronutrients such as nitrogen and phosphate in the fertilized region can lead to a decrease in production "downstream" from the fertilized region. 5) Ocean iron fertilization could acidify the deep ocean by storing more dissolved inorganic carbon there. 6) Enhanced biological production over oceans could potentially lead to expanded regions with low oxygen concentration, increased production of N_2O and CH_4, and possible disruptions to marine ecosystems.

While the rate of removal of CO_2 through accelerated weathering and direct air capture methods are limited primarily by cost, energy and environmental constraints, those methods involving biological processes operate much more slowly with estimates of maximum physical potential for atmospheric CO_2 removal for each of the more effective biological CDR strategies to be on the order of 100 Gt C (~ 50 ppmv of CO_2) over a century, with similar limitations from cost, energy and environmental issues.

Solar radiation management methods

The average temperature of the planet is controlled by the amount of sunlight absorbed by the Earth's atmosphere and surface, and the degree to which gases and clouds in the atmosphere hinder the escape of the energy back to space. If less incoming sunlight reaches the surface (because of an increase in the reflectivity of the planet) or if it becomes easier for energy to escape (because of a decrease in heat trapping gases or some types of clouds), the average surface temperature will decrease.

Geoengineering methods relying on managing the Earth's radiative budget are based on this fundamental physical principle. These methods seek to increase the reflectivity of the planet by making the atmosphere, clouds or the surface more reflective or by suppressing cirrus clouds that hinder the escape of energy from the Earth system (see FAQ.7.3, Figure 2).

[INSERT FAQ 7.3, FIGURE 2 HERE]
FAQ 7.3, Figure 2: Overview of solar radiation management methods.

Basic physics tells us that if either of these changes is successful, the planet will cool. The picture is complicated, however, because of the multiple complex physical processes that govern the interactions between the flow of energy, the atmospheric circulation, weather and the resulting climate. While the average surface temperature of the planet responds to the energy budget in a rather straightforward way, the temperature at any particular location and time is influenced by many other factors. It is expected that any particular geoengineering technique will cool some regions more than others: the locations where radiation management cools the planet need not correspond to the locations and times where greenhouse gases produce a warming. For example, a change in the amount of sunlight via radiation management will operate only during daytime, but changes in greenhouse gases affect heating rates during both day and night. This inexact compensation will have some influence on the diurnal cycle of surface temperature at any given location, even if the average surface temperature is unchanged. This is a simple example of inexact compensation but other subtle changes may also occur.

Climate is much more than surface temperature, however; it is also characterized by precipitation patterns, the distribution of snowpack and sea-ice, and the frequency of occurrence of extreme events, just to name a

few. Both models and theory show that compensating an increased greenhouse effect with an increased planetary reflectivity will not maintain both the average surface temperature and the average precipitation rate. Regional changes in heating/cooling are expected to affect local precipitation rates, and other aspects of climate. The imprecise compensation in regional and global climate patterns make it unlikely that solar radiation management will produce a future climate that is "just like" the one we experience today, or have experienced in the past. The residual climate changes from inexact compensation may increase as the geoengineering effort is scaled up.

In addition, solar radiation management techniques may also have other side effects. For example, radiation management by stratospheric sulphate aerosols can produce stratospheric ozone depletion, and changes in the ratio of direct to diffuse sunlight reaching the surface that can affect terrestrial ecosystems. Moreover, radiation management will not have any impact on ocean acidification, which is driven by the atmospheric CO_2 concentration. A key unanswered question is whether or not the benefits of radiation management outweigh the associated risks, in light of the expected residual impacts.

[END FAQ 7.3 HERE]

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Tables

Table 7.1: Global and regional anthropogenic emissions important for aerosol formation and tropospheric chemistry. The maximum and minimum values from available inventories are presented. Units for NO_x are Tg NO yr^{-1}, other units are Tg yr^{-1}. Adapted from Granier et al. (2011).

Year 2000 emissions	CO		NOx		CH_4		NMVOCs		BC		OC		SO_2		NH_3	
Tg/year	*MIN*	*MAX*	*MIN*	*MAX*	*MIN*	*MAX*	*MIN*	*MAX*	*MIN*	*MAX*	*MIN*	*MAX*	*MIN*	*MAX*	*MIN*	*MAX*
Total	467.50	610.80	58.70	68.60	275.20	310.30	121.00	139.50	4.60	5.60	6.40	12.70	102.00	145.00	37.50	38.90
Western Europe	21.40	35.40	5.90	9.00	16.30	22.00	9.20	14.30	0.32	0.38	0.32	0.40	6.10	14.10	3.40	4.50
Central Europe	7.80	12.30	1.60	1.90	6.10	7.70	2.30	3.50	0.11	0.21	0.25	0.39	4.60	10.00	1.10	1.20
USA	55.90	94.40	11.50	14.10	26.20	40.70	13.00	17.50	0.27	0.40	0.36	0.51	13.50	17.80	3.30	4.40
Canada	4.20	11.20	1.20	1.70	3.90	5.00	1.50	3.40	0.04	0.04	0.03	0.06	2.20	2.90	0.51	0.60
Central America	10.00	15.10	1.50	2.10	8.40	9.10	2.90	4.10	0.11	0.11	0.17	0.35	3.70	4.10	1.10	1.10
South America	22.30	26.50	2.80	3.80	26.40	30.00	8.40	12.90	0.20	0.33	0.32	0.83	3.80	8.80	3.40	3.50
Africa	49.40	83.20	2.70	5.90	25.00	29.40	10.80	14.50	0.46	0.62	1.05	1.91	5.30	8.80	2.30	2.40
China	95.50	137.30	6.90	9.80	33.10	49.40	11.50	24.50	0.71	1.41	1.10	3.80	19.20	21.10	8.90	13.60
India	40.30	79.40	2.70	4.90	25.70	33.80	7.30	10.80	0.45	0.84	1.00	3.27	4.00	7.90	3.70	8.50
Oceania	2.60	5.70	1.10	1.90	6.40	6.80	0.00	1.50	0.03	0.04	0.04	0.08	2.40	2.70	0.72	0.72

Table 7.2: Key aerosol properties of the main aerosol species in the troposphere. Brown carbon is a particular type of OA but is treated here as an additional component because it is light absorbing. The estimate of aerosol burdens and lifetimes in the troposphere are based on the AeroCom models.

Aerosol Species	Global Burden	Mass Size Distribution	Sources	Sinks	Lifetime	Key Climate Relevant Properties
Black carbon		Freshly emitted: 0-80 nm Aged: accumulation mode	Combustion of fossil fuels, biofuels and biomass	Wet deposition Dry deposition	7–10 days	Large mass absorption efficiency in the visible
Brown carbon		Freshly emitted: 100-400 nm Aged: accumulation mode	Combustion of biofuels and biomass	Wet deposition Dry deposition	1 week	Medium mass absorption efficiency in the visible. Light scattering.
Organic aerosol		POA: Aitken mode SOA: nuclei mode Aged OA : accumulation mode Biogenic POA : coarse mode	Combustion of fossil fuel, biofuel and biomass. Continental and marine ecosystems. Some anthropogenic non-combustion activities.	Wet deposition Dry deposition	1 week	Light scattering. Lens effect when deposited on black or brown carbon. CCN active (depending on aging time and mechanism). IN active (biogenic POA)
Sulphate		Secondary: Nuclei, Aitken, and accumulation mode Primary: coarse mode	Primary: marine and volcanic emissions. Secondary: oxidation of SO_2 from natural and anthropogenic sources	Wet deposition Dry deposition	1 week	Light scattering. Very hygroscopic. Lens effect when deposited on black or brown carbon. CCN active.
Nitrate		Accumulation and coarse modes	Oxidation of NOx	Wet deposition Dry deposition	1 week	Light scattering. CCN active.
Dust	(sensitive to size cutoff)	Coarse and super-coarse modes, with a small accumulation mode	Wind erosion, soil resuspension. Some agricultural practices and industrial activities (cement)	Sedimentation Dry deposition Wet deposition	1 day to 1 week depending on size	IN active, light scattering and absorption, greenhouse effect.
Sea-salt	(sensitive to size cutoff)	Coarse mode and accumulation mode	Wave breaking. Wind erosion.	Sedimentation Wet deposition Dry deposition	1 day to 1 week depending on size	Light scattering. Very hygroscopic. CCN active.

Chapter 7: Clouds and Aerosols

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Figures

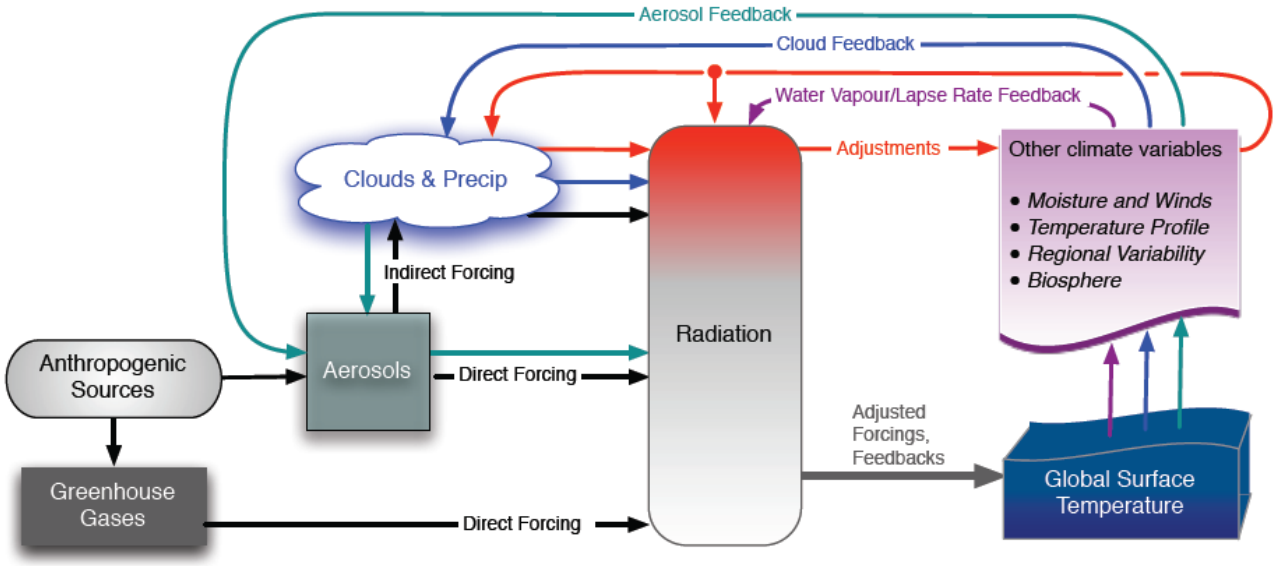


Figure 7.1: Overview of feedback and forcing pathways involving clouds and aerosols. Forcings are represented by dark arrows; forcing agents are boxes with grey shadows, rapid forcing adjustments (or rapid response) are red arrows and feedbacks are other-colored arrows. See text for further discussion.

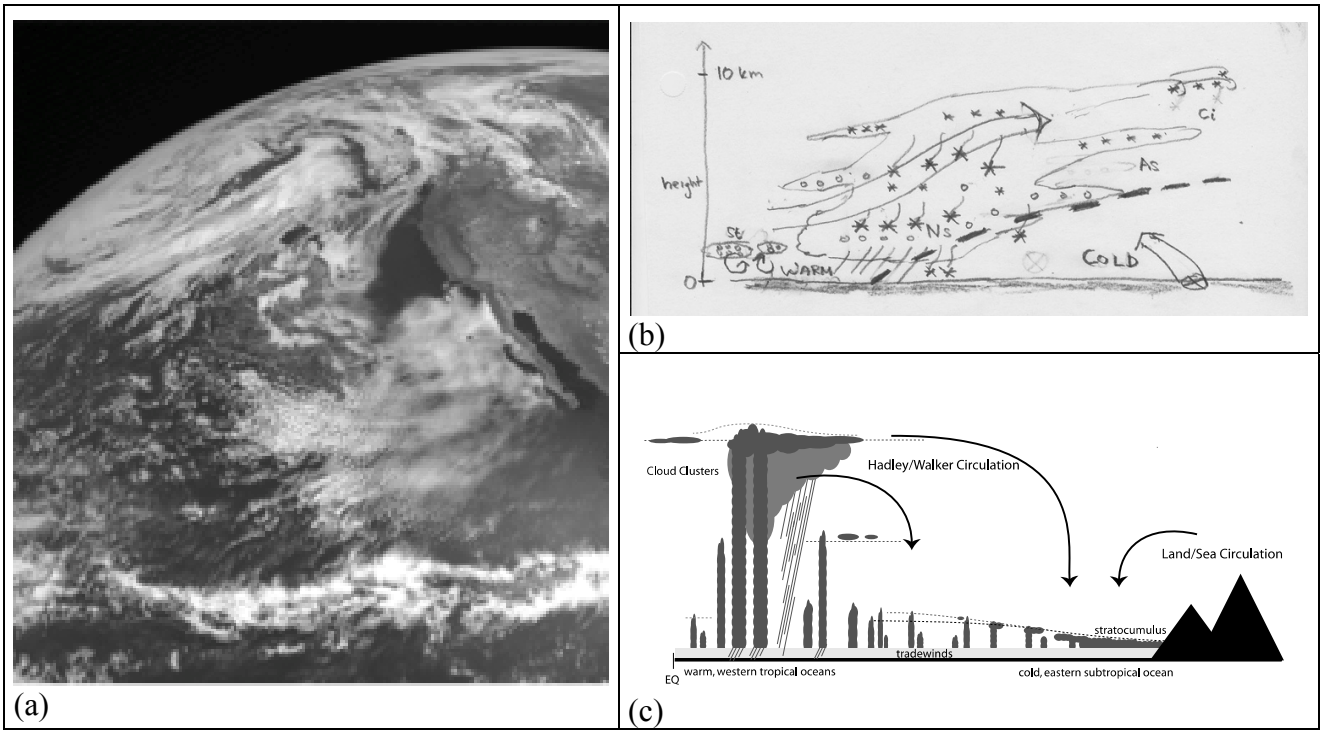


Figure 7.2: Diverse cloud regimes reflect diverse meteorology. (a) A visible-wavelength geostationary satellite image shows (from top to bottom) expanses and long arcs of cloud associated with extratropical cyclones, subtropical coastal stratocumulus near Baja California breaking up into shallow cumulus clouds in the central Pacific, and mesoscale convective systems outlining the Pacific ITCZ. (b) A schematic vertical section through a typical warm front of an extratropical cyclone shows multiple layers of upper-tropospheric ice (cirrus) and mid-tropospheric water (altostratus) cloud upwind of the frontal zone, an extensive region of nimbostratus associated with frontal uplift and turbulence-driven boundary layer cloud in the warm sector. (c) A schematic cross section along the low-level trade wind flow from a subtropical west coast of a continent to the ITCZ shows typical low-latitude cloud types, shallow stratocumulus in the cool waters of the oceanic upwelling zone near the coast, trapped under a strong subsidence inversion, shallow cumulus of warmer waters further offshore and a transition into precipitating cumulonimbus cloud systems with extensive cirrus anvils associated with rising air motions in the ITCZ.

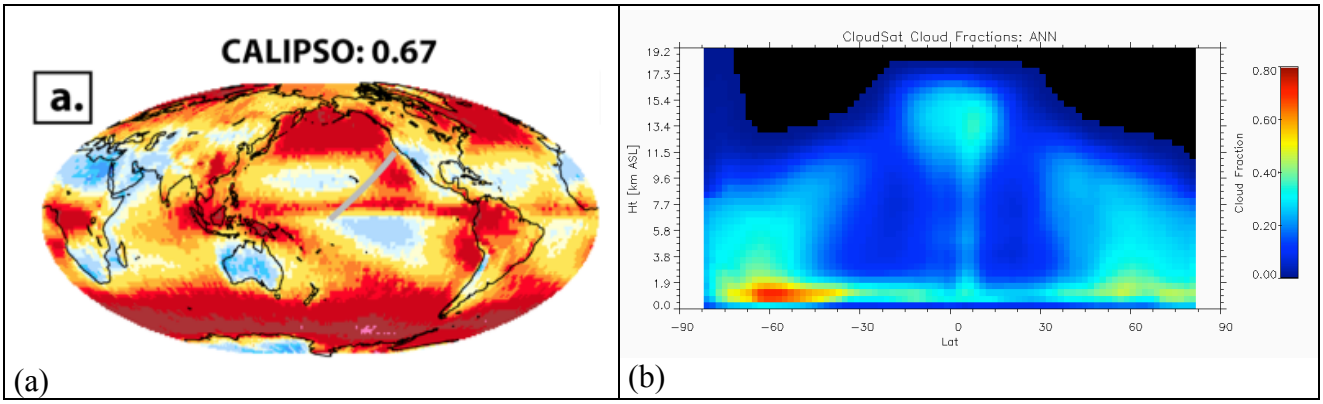


Figure 7.3: Annual-mean cloud fractional occurrence based on four years of satellite observations (June 2006-February 2011) from CloudSat and Calipso (Kay et al., 2011; COSP simulator). (a) Geographical mean, with thin cloud (SR < 5) removed; (b) latitude-height section of zonal mean cloud cover. [PLACEHOLDER FOR SECOND ORDER DRAFT: further graphical refinement.]

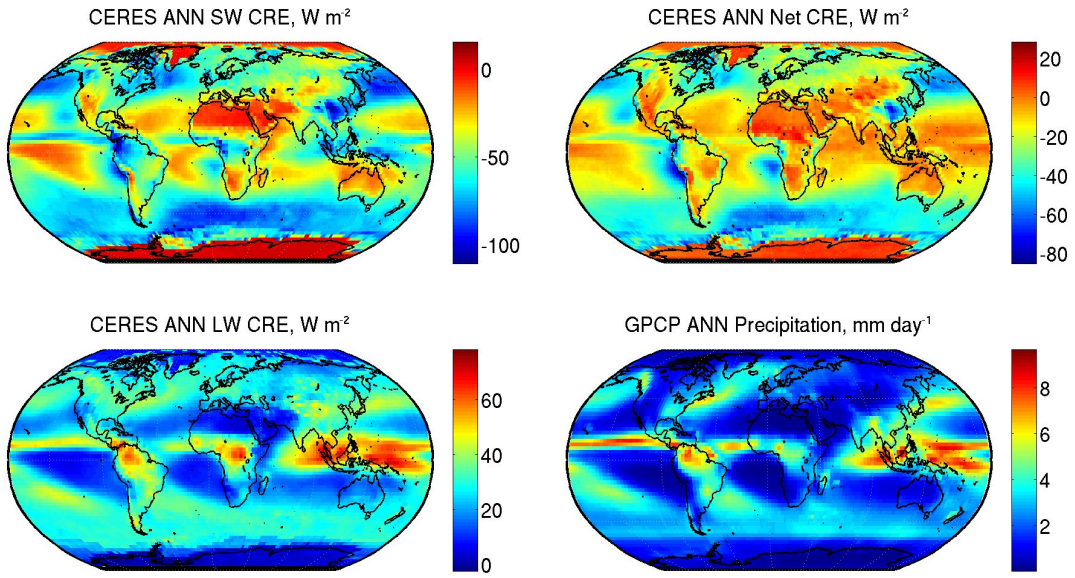


Figure 7.4: Distribution of annual-mean SWCRE, LWCRE, net CRE (from CERES-EBAF) and precipitation (from CMAP).

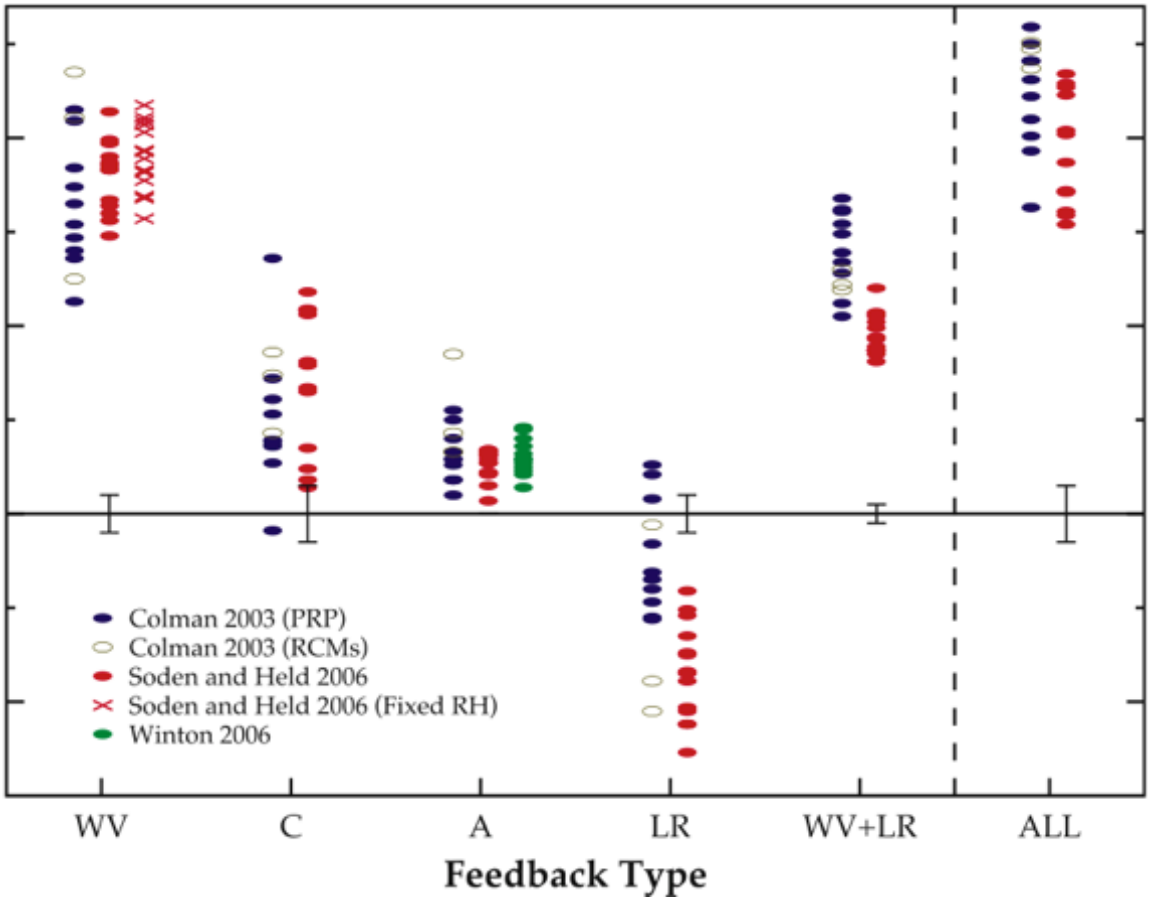


Figure 7.5: CFMIP figure on cloud feedbacks in CMIP5 models. [PLACEHOLDER FOR SECOND ORDER DRAFT: CMIP3 version used as placeholder.]

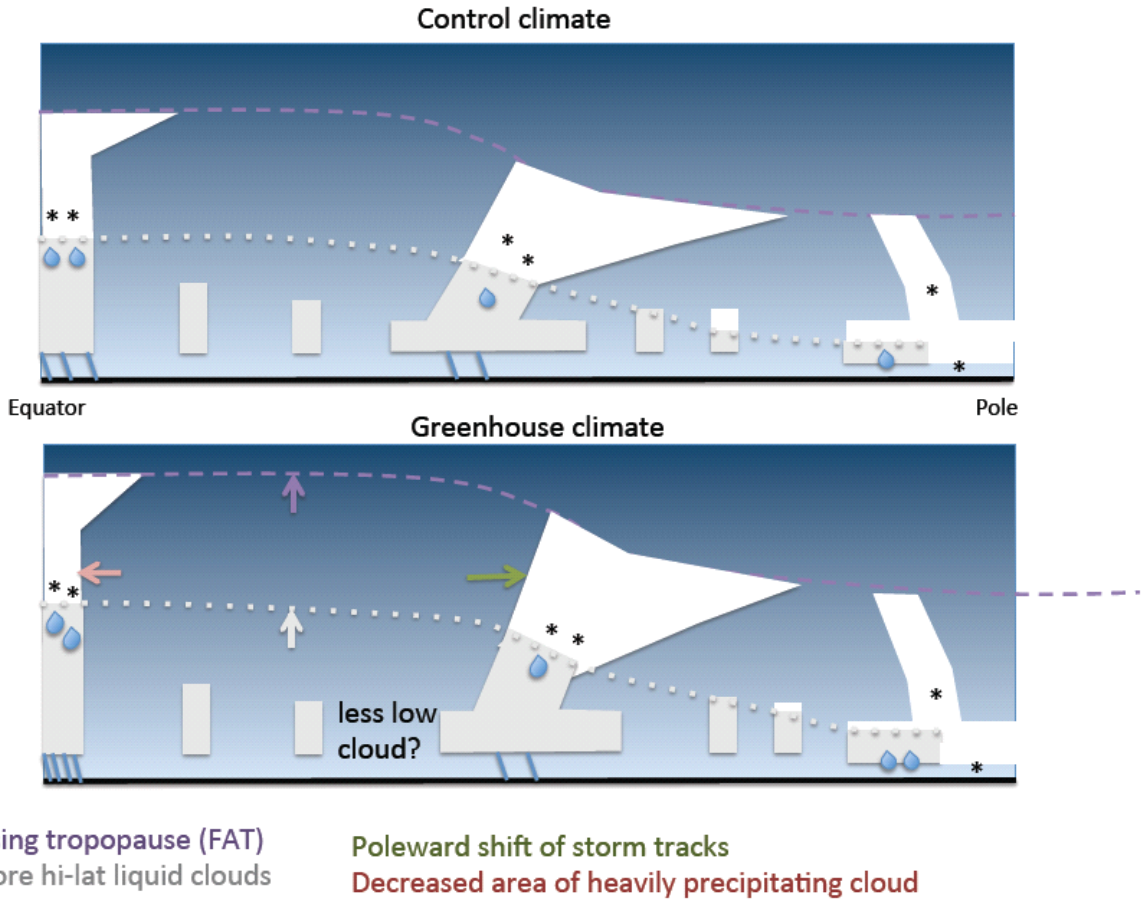


Figure 7.6: Robust cloud responses to greenhouse warming simulated by the CMIP3 multimodel ensemble. Panel (a) is a schematic latitude-altitude section showing typical cloud types in a pre-industrial climate. Grey (white) indicates clouds composed predominantly of liquid water (ice). Raindrops and snowflakes indicate the typical precipitation type. Dotted line indicates the typical freezing level, and purple dashed line indicates the tropopause. Panel (b) shows the same cross section for a warmer climate, with arrows denoting the movement of different boundaries. Tropical deep convection regions narrow and intensify, the subsidence regions of the subtropics widen poleward, with most GCMs projecting low cloud decreases in this area, and storm track cloud and precipitation also shift poleward. Cirrus cloud tops rise in lockstep with the tropopause, helping induce positive longwave cloud feedbacks. The rising freezing level causes more cloud to become liquid, contributing to increased optical thickness of high latitude clouds in the CMIP3 multimodel mean. [PLACHOLDER FOR SECOND ORDER DRAFT: CMIP5]

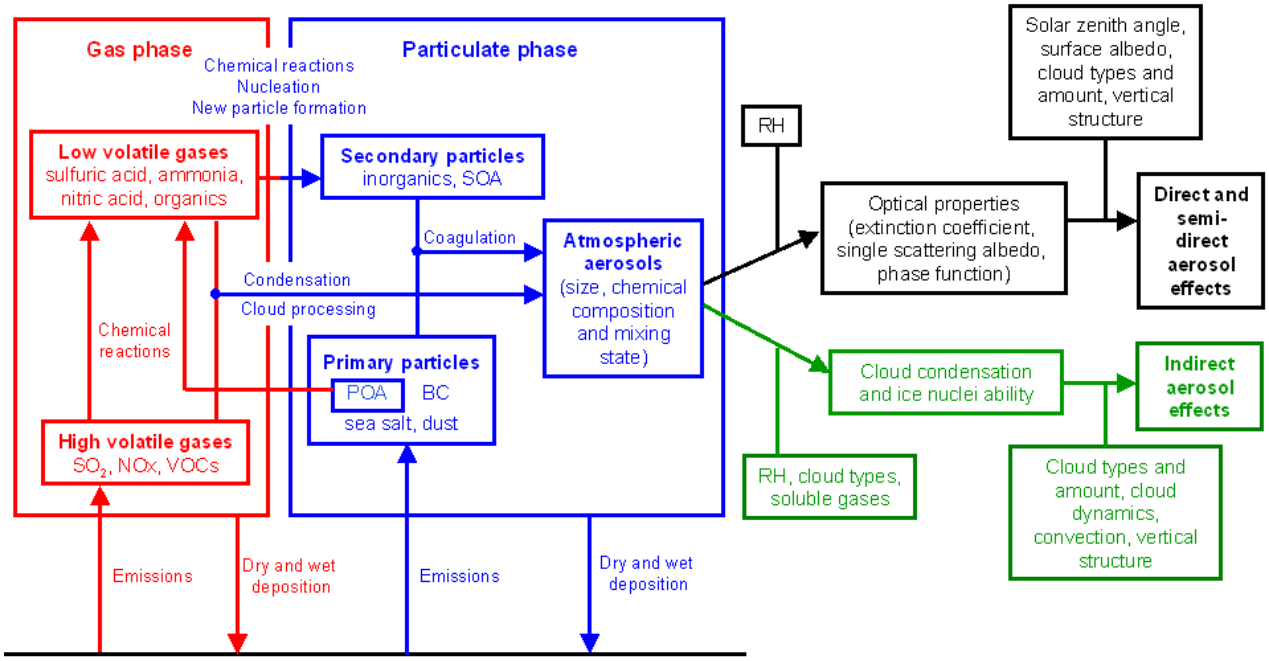


Figure 7.7: Overview of atmospheric aerosol processes and meteorological variables influencing the aerosol semi-direct, direct and indirect aerosol effects. Red designates gas phase processes and variables; blue designates particulate (aerosol) phase processes and variables; processes and variables relevant to the aerosol direct and semi-direct effects appear in black, while those relevant to the aerosol indirect effects appear in green.

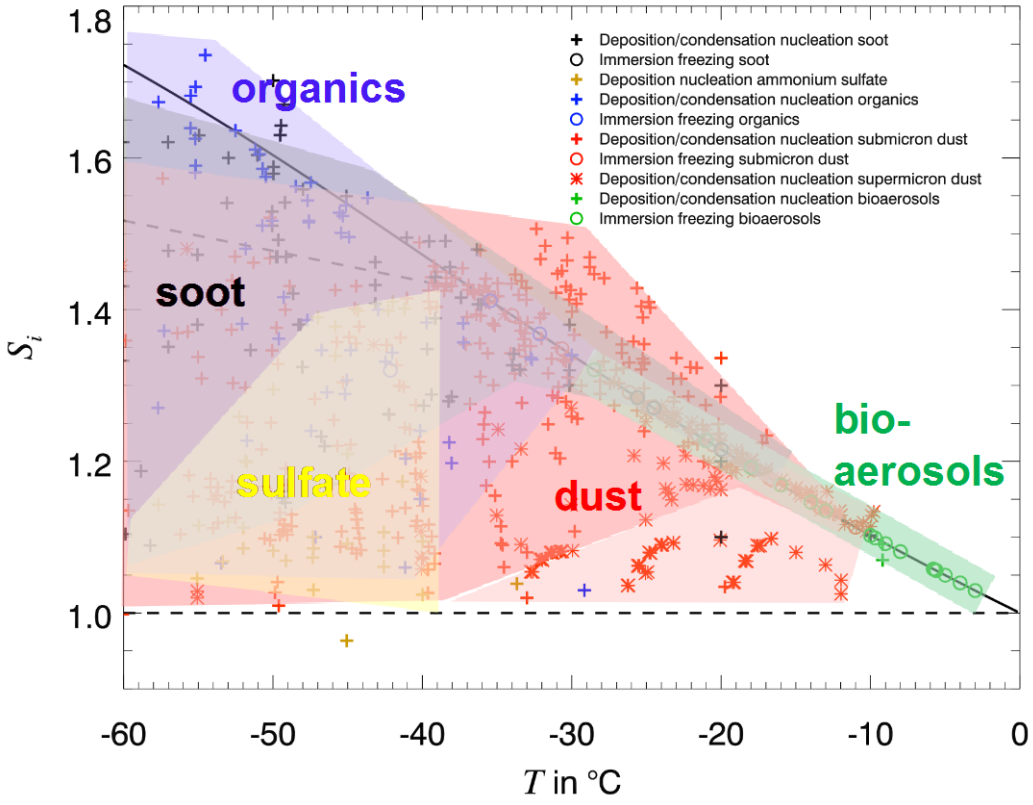


Figure 7.8: The onset temperatures and relative humidities for deposition/condensation freezing and immersion freezing for bioaerosols (Ahern et al., 2007; Diehl et al., 2001; Iannone et al., 2011; Kanji et al., 2011; Mohler et al., 2008; Mortazavi et al., 2008; von Blohn et al., 2005; Yankofsky et al., 1981), mineral dusts (Archuleta et al., 2005; Bundke et al., 2008; Connolly et al., 2009; Cziczo et al., 2009a; Field et al., 2006; Kanji and Abbatt 2006; Kanji et al., 2011; Knopf and Koop 2006; Koehler et al., 2010; Kulkarni and Dobbie 2010; Lüönd et al., 2010; Mohler et al., 2006; Murray et al., 2011; Niedermeier et al., 2010; Niemand et al., 2011; Roberts and Hallett 1968; Salam et al., 2006; Schaller and Fukuta 1979; Welti et al., 2009; Zimmermann et al., 2008), organics (Baustian et al., 2010; Kanji et al., 2008; Petters et al., 2009; Prenni et al., 2007; Shilling et al., 2006; Wagner et al., 2010, 2011; Wang and Knopf 2011; Zobrist et al., 2007), solid ammonium sulphate (Abbatt et al., 2006; Baustian et al., 2010; Mangold et al., 2005; Shilling et al., 2006; Wise et al., 2009; 2010) and BC (soot) (Crawford et al., 2011; DeMott 1990; DeMott et al., 1999; Diehl and Mitra 1998; Dymarska et al., 2006; Fornea et al., 2009; Gorbunov et al., 2001; Kanji et al., 2011; Mohler et al., 2005), from a compilation of experimental data of sub- and super-micron aerosol particles in the literature (for references see supplementary material). The large range of observed ice nucleation onset conditions is due to different experimental setups, particle sizes, activated fractions and chemical composition. Only those IN species for which at least three papers exists are shown. The dashed line refers to the homogeneous freezing of solution droplets after (Koop et al., 2000).

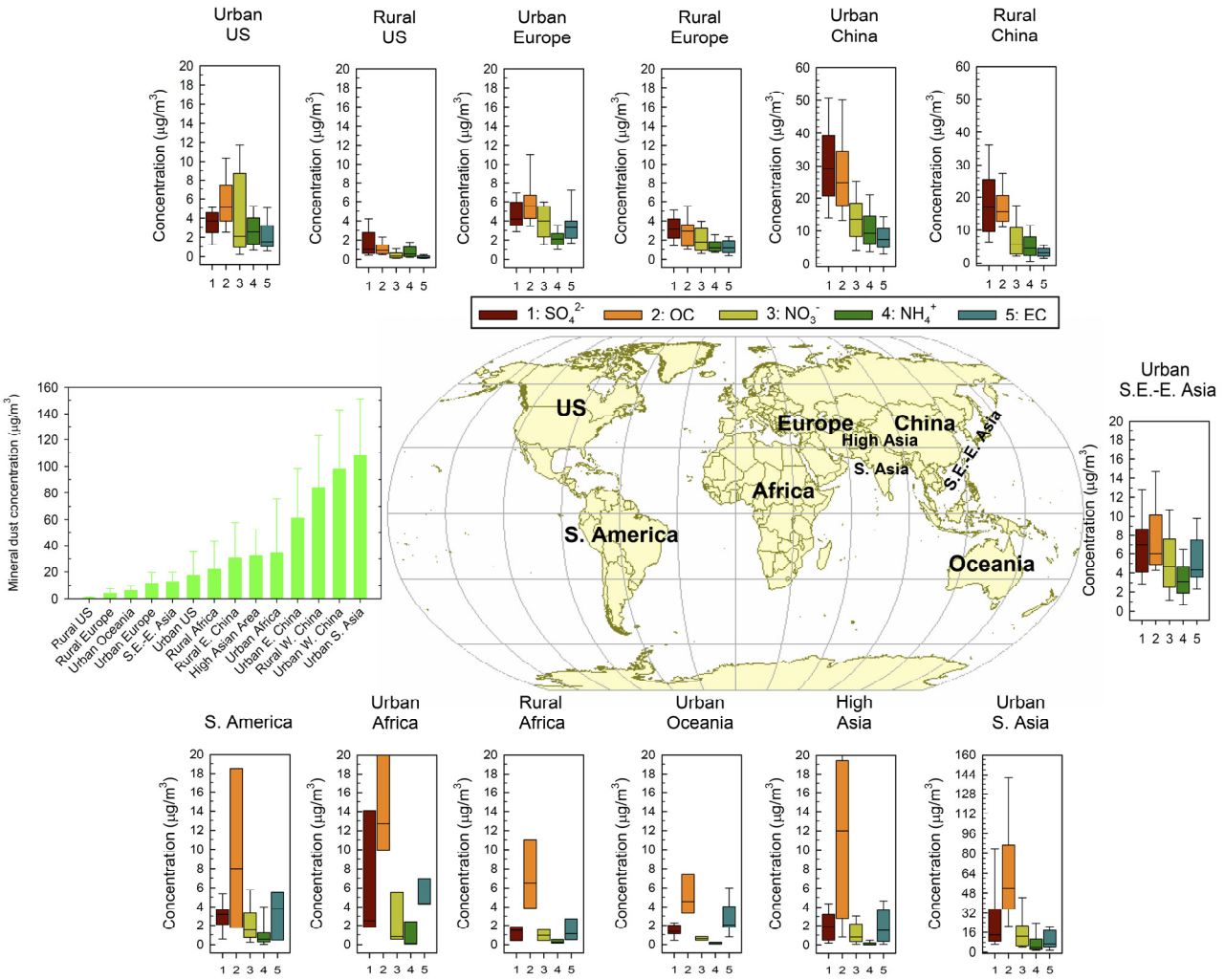


Figure 7.9: Bar chart plots summarizing the annual, seasonal or monthly mean mass concentration (μg m^{-3}) of six major types of aerosol particles in diameter smaller than 10 μm with at least an entire year data from various rural and urban sites in nine continental areas of the world. These include: 1) rural U. S. (Chow et al., 1993; Liu et al., 2005a; Malm and Schichtel 2004; Malm et al., 1994); urban U. S. (Chow et al., 1993; Ito et al., 2004; Kim et al., 2000; Liu et al., 2005a; Malm and Schichtel 2004; Sawant et al., 2004); 2) South America (Artaxo et al., 1998; Artaxo et al., 2002; Bourotte et al., 2007; Celis et al., 2004; Fuzzi et al., 2007; Gioda et al., 2011; Mariani and Mello 2007; Martin et al., 2010; Morales et al., 1998; Souza et al., 2010); 3) rural Europe (Gullu et al., 2000; Hueglin et al., 2005; Kocak et al., 2007; Putaud et al., 2004; Puxbaum et al., 2004; Querol et al., 2001; Querol et al., 2009; Querol et al., 2004; Rodriguez et al., 2002; Rodrıguez et al., 2004; Salvador et al., 2007; Theodosi et al., 2010; Viana et al., 2008; Yin and Harrison 2008; Yttri 2007); urban Europe (Hueglin et al., 2005; Lenschow et al., 2001; Lodhi et al., 2009; Lonati et al., 2005; Perez et al., 2008; Putaud et al., 2004; Querol et al., 2001; Querol et al., 2006; Querol et al., 2004; Querol et al., 2008; Rodriguez et al., 2002; Rodrıguez et al., 2004; Roosli et al., 2001; Viana et al., 2007; Viana et al., 2006; Yin and Harrison 2008); 4) rural Africa (Maenhaut et al., 1996; Mkoma 2008; Mkoma et al., 2009a; Mkoma et al., 2009b; Nyanganyura et al., 2007; Weinstein et al., 2010); urban Africa (Favez et al., 2008; Mkoma 2008; Mkoma et al., 2009a); 5) high Asia, with altitude larger than 1680 m. (Carrico et al., 2003; Decesari et al., 2010; Ming et al., 2007a; Qu et al., 2008; Ram et al., 2010; Rastogi and Sarin 2005; Rengarajan et al., 2007; Shresth et al., 2000; Zhang et al., 2001; Zhang et al., 2008; Zhang et al., 2011a); 6) rural China (Hagler et al., 2006; Hu et al., 2002; Zhang et al., 2011a); urban China (Cheng et al., 2000; Hagler et al., 2006; Oanh et al., 2006; Wang et al., 2003; Wang et al., 2005b; Wang et al., 2006; Xiao and Liu 2004; Yao et al., 2002; Ye et al., 2003; Zhang et al., 2002; Zhang et al., 2011a; Zhang et al., 2011b); 7) South-East and East Asia (Han et al., 2008; Khan et al., 2010; Kim et al., 2007; Lee and Kang 2001; Oanh et al., 2006); 8) urban South Asia (Chakraborty and Gupta 2010; Khare and Baruah 2010; Kumar et al., 2007; Lodhi et al., 2009; Raman et al., 2010; Rastogi and Sarin 2005; Safai et al., 2010; Stone et al., 2010); 9) urban Oceania (Chan et al., 1997; Maenhaut et al., 2000; Radhi et al., 2010; Wang and Shooter 2001; Wang et al., 2005a).

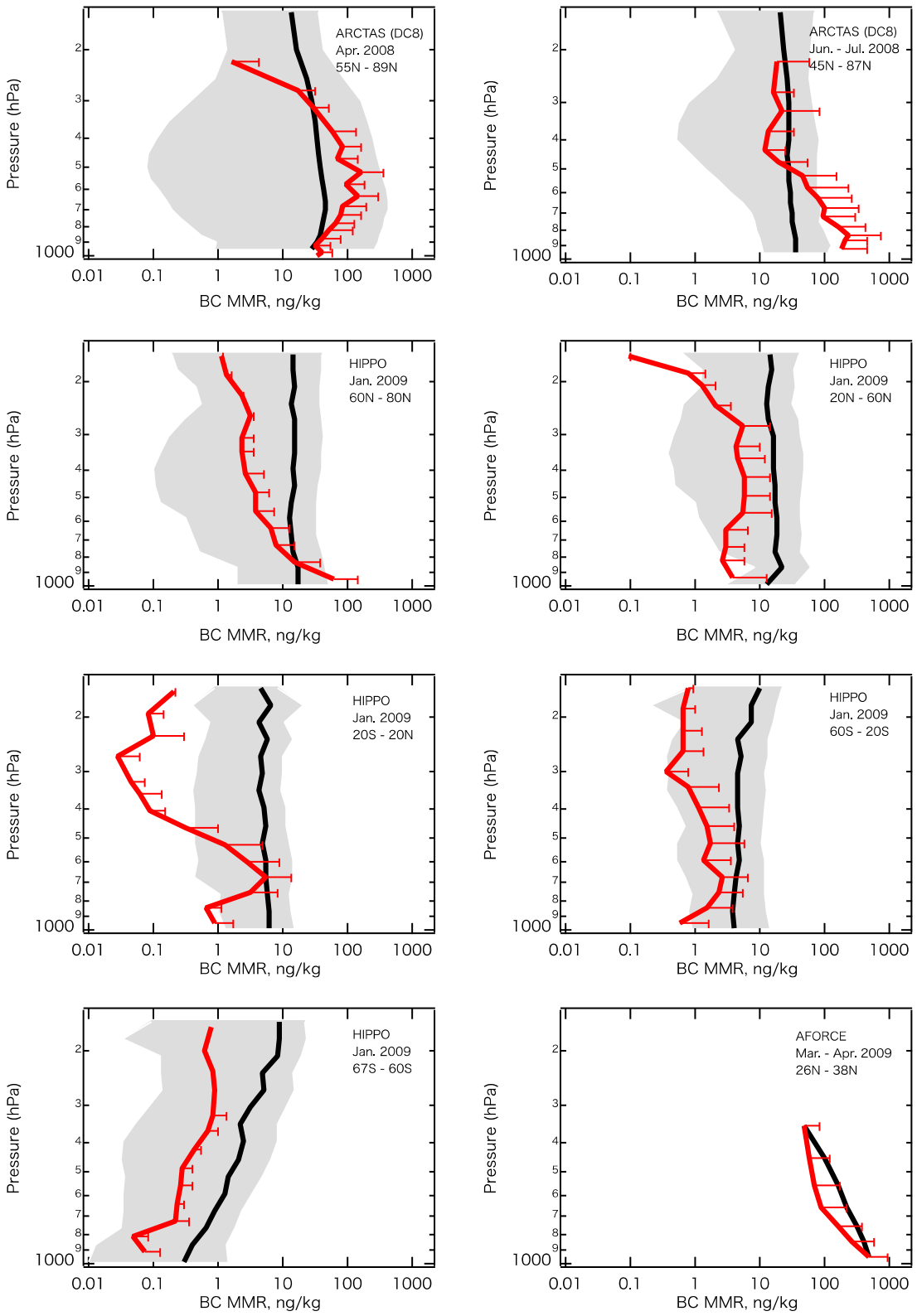


Figure 7.10: Comparison of BC profiles as measured during the ARCTAS, HIPPO1 and FORCE-A campaigns and simulated by a range of global aerosol models. [PLACEHOLDER FOR SECOND ORDER DRAFT: will be updated from AeroCom and CMIP5 models]

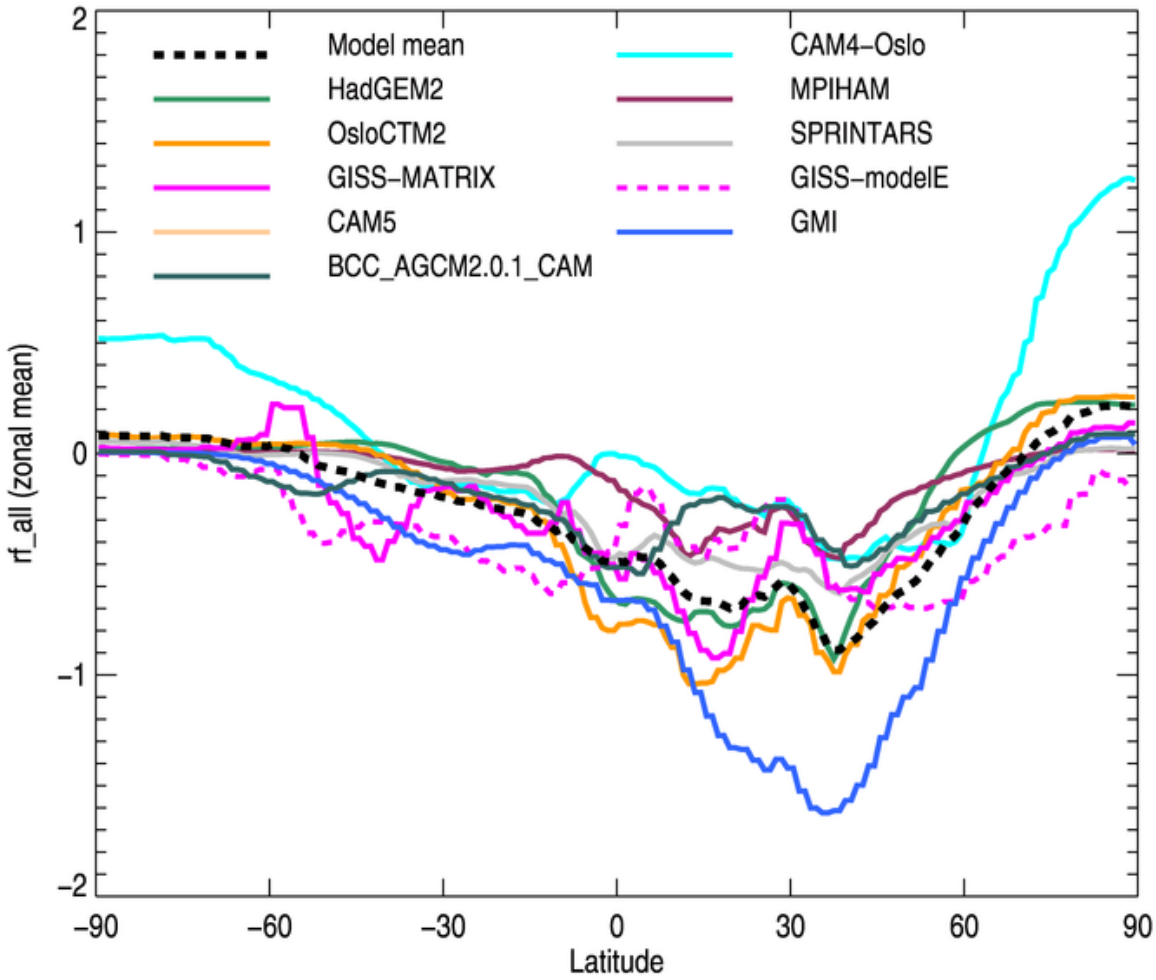


Figure 7.11: Zonal mean total aerosol direct radiative forcing from the different AeroCom models. No adjustment for missing species has been applied.

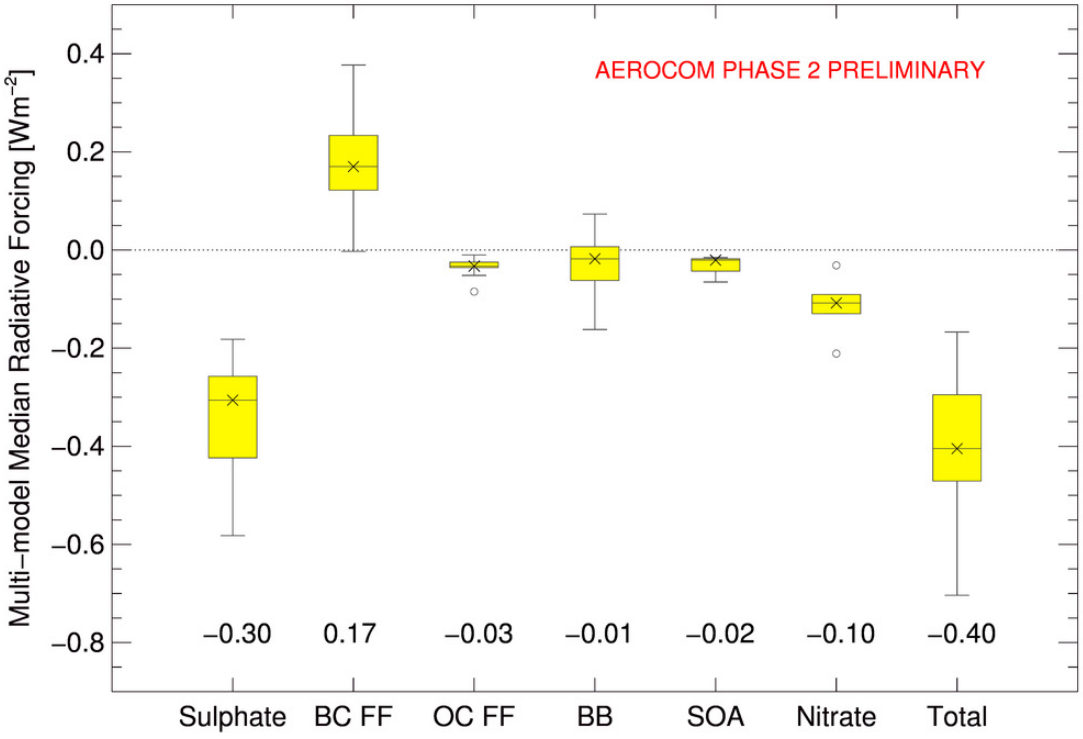


Figure 7.12: Median, full range and 5%–95% range of AeroCom model direct radiative forcing by species and the total direct forcing. The total direct forcing has been adjusted to take account of missing species in some models by adding the median value of the species forcing from the remaining models.

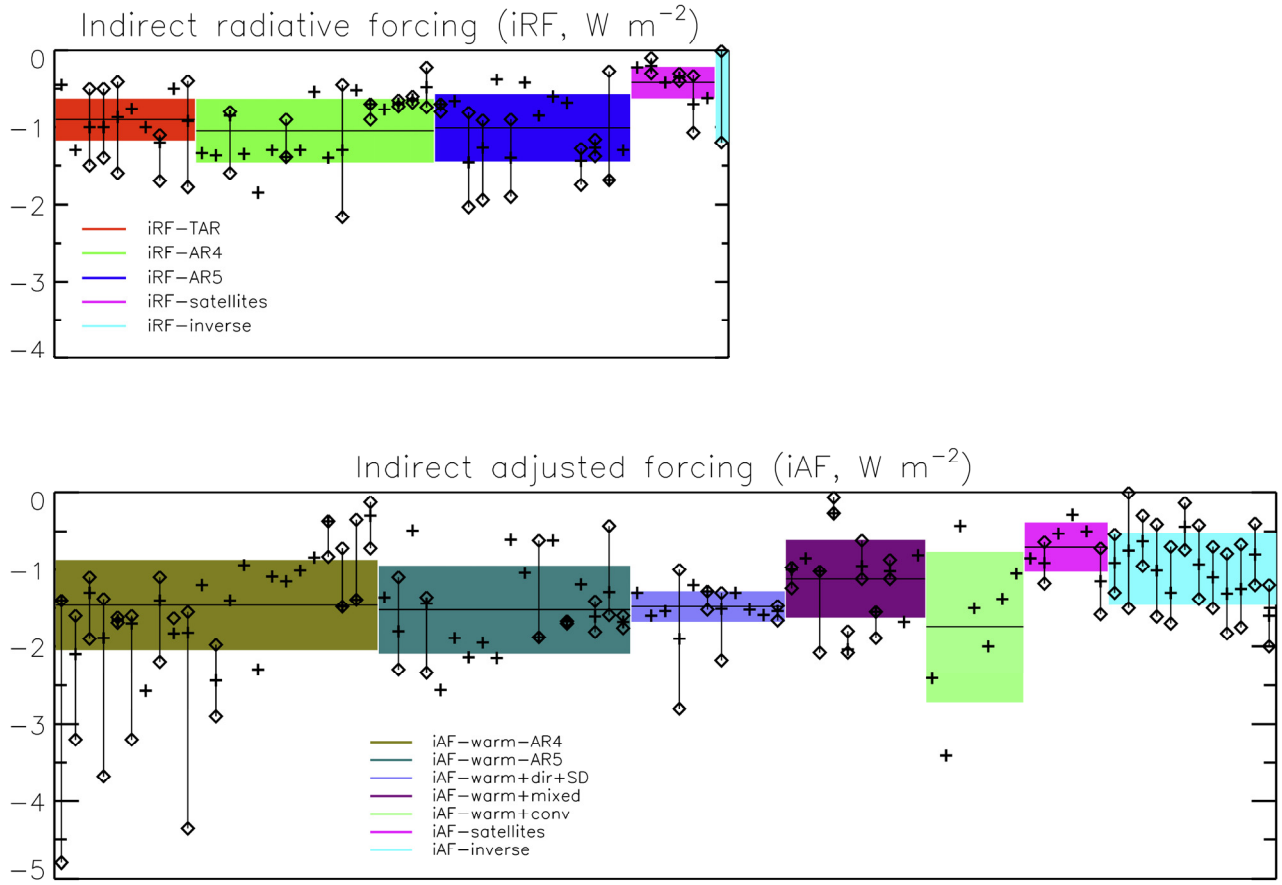


Figure 7.13a: Model, satellite and inverse estimates of the iRF and the iAF since 1993. For each paper the best estimate per model is shown as a plus-sign. If multiple estimates or uncertainties are given in a paper, the ranges are shown as vertical lines bounded by diamonds. The thin horizontal lines denote the average of the respective group and the width of the coloured box denotes its standard deviation. The iRF studies from GCMs are divided into those published prior to TAR: iRF-TAR (Boucher and Lohmann 1995; Chuang et al., 1997; Feichter et al., 1997; Jones et al., 1994; Kaufman and Chou 1993; Kiehl et al., 2000; Lohmann and Feichter 1997; Lohmann et al., 2000; Rotstayn 1999), between TAR and AR4: iRF-AR4 (Chen and Penner 2005; Chuang et al., 2002; Ghan et al., 2001; Hansen et al., 2005; Jones et al., 2001; Kristjansson 2002; Ming et al., 2005; Penner et al., 2006; Quaas and Boucher 2005; Quaas et al., 2004; Rotstayn and Penner 2001; Rotstayn and Liu 2003; Suzuki et al., 2004; Takemura et al., 2005; Williams et al., 2001) and since 2007: iRF-AR5 (Barahona et al., 2011; Bellouin et al., 2011; Haerter et al., 2009; Kvalevag and Myhre 2007; Lohmann et al., 2007; Lohmann et al., 2010; Penner et al., 2011; Rotstayn and Liu 2009; Storelvmo 2011; Storelvmo et al., 2009; Wang and Penner 2009). iAF studies on liquid clouds that include the cloud albedo and cloud lifetime effect are also divided into those published until 2006: iAF-liquid-AR4 (Easter et al., 2004; Ghan et al., 2001; Johns et al., 2006; Jones et al., 2001; Kristjansson 2002; Kristjansson et al., 2005; Lohmann 2002b; Lohmann and Feichter 1997; Lohmann et al., 2000; Menon et al., 2002; Ming et al., 2005; Peng and Lohmann 2003; Penner et al., 2003; Penner et al., 2006; Quaas et al., 2006; Rotstayn 1999; Rotstayn and Penner 2001; Rotstayn and Liu 2005; Storelvmo et al., 2006; Takemura et al., 2005; Williams et al., 2001) and since 2007: iAF-liquid-AR5 (Chen et al., 2010; Ghan et al., 2011b; Hoose et al., 2009; Kirkevag et al., 2008; Makkonen et al., 2011; Menon and DelGenio 2007; Ming et al., 2007b; Penner et al., 2011; Quaas et al., 2009b; Rotstayn and Liu 2009; Storelvmo et al., 2008a); iRF and iAF estimates that involve satellite data are shown in pink: iRF-satellites (Dufresne et al., 2005; Lebsock et al., 2008; Quaas and Boucher 2005; Quaas et al., 2008; Quaas et al., 2009b; Storelvmo et al., 2009) and iAF-satellites (Lohmann and Lesins 2002; Quaas et al., 2006; Quaas et al., 2009b; Sekiguchi et al., 2003), inverse estimates for the iRF and iAF are shown in turquoise: iRF-inverse (Knutti et al., 2002) and iAF-inverse (Anderson et al., 2003; Andronova and Schlesinger 2001; Church et al., 2011; Forest et al., 2006; Forest et al., 2002; Gregory et al., 2002; Hansen et al., 2011; Harvey and Kaufmann 2002; Huber and Knutti 2011; Libardoni and Forest 2011; Murphy et al., 2009; Shindell and Faluvegi 2009; Stott et al., 2006) in turquoise. iAF studies that include the direct and semi-direct effect in lilac: iAF-liquid+dir+SD (Ghan et al., 2011a; Lohmann and Feichter 2001; Lohmann et al., 2007; Posselt and Lohmann 2008; Posselt and Lohmann 2009; Quaas et al., 2004; Quaas et al., 2006; Quaas et al., 2009b; Rotstayn et al., 2007; Salzmann et al., 2010), those that additionally consider aerosol effects on mixed-phase clouds in purple: iAF-liquid+mixed (Hoose et al., 2008; Hoose et al., 2010b; Jacobson 2006; Lohmann 2004; Lohmann and Diehl 2006; Lohmann and Hoose 2009; Lohmann and Ferrachat 2010; Salzmann et al., 2010; Storelvmo et al., 2008a; Storelvmo et al., 2008b), and those that

treat aerosol effects in stratiform and convective clouds in green: iAF-liquid+conv (Koch et al., 2009a; Lohmann 2008; Menon and Rotstayn 2006; Menon and DelGenio 2007; Unger et al., 2009; Wang et al., 2011b). For the inverse estimates no best estimate is shown and the turquoise colour box denotes the average of the lower and upper bounds of these studies, respectively.

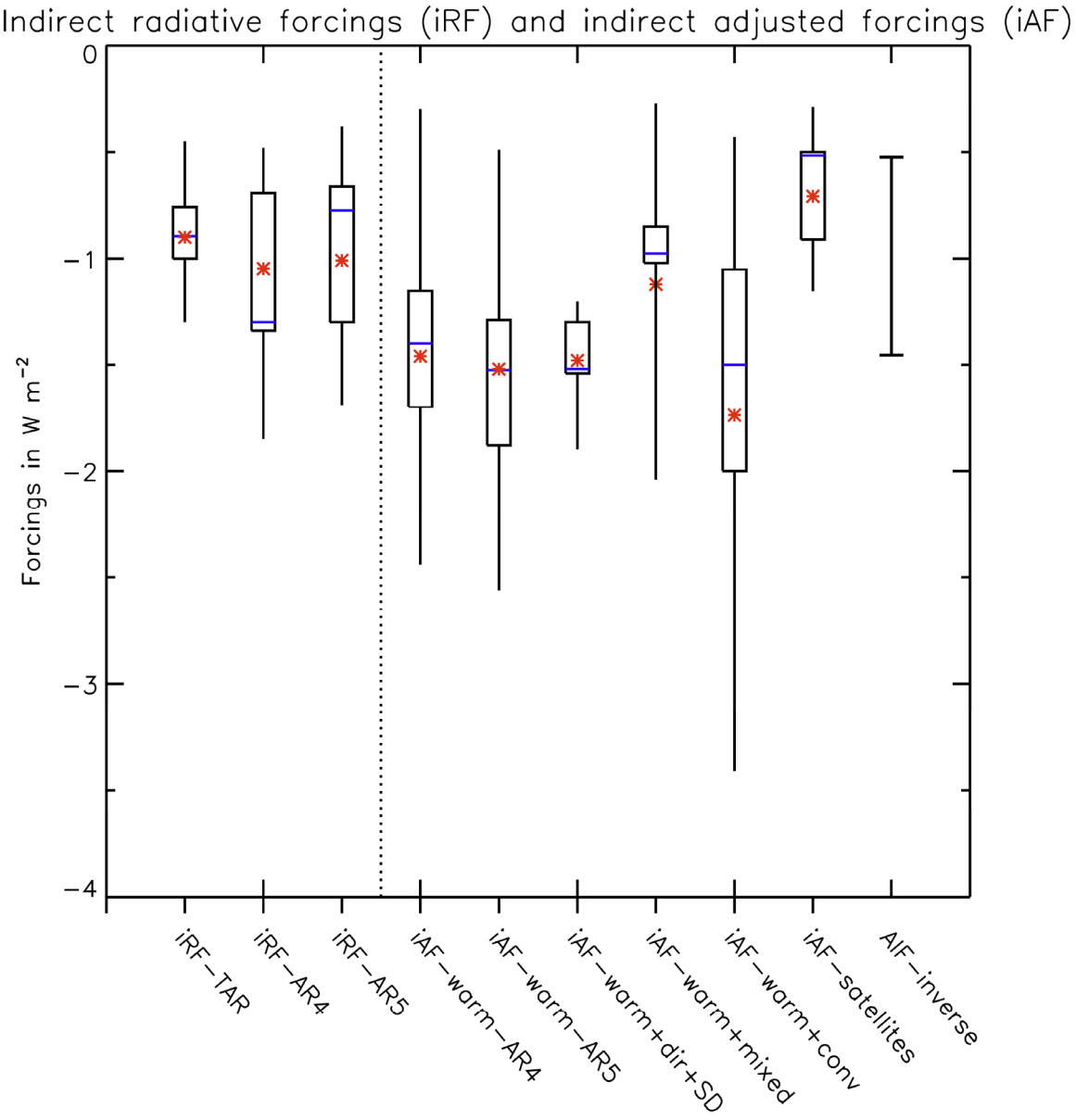


Figure 7.13b: Box plots of model, satellite and inverse estimates of the IF and the AIF since 1993 for the same groups of estimates as in Figure 7.13a provided at least 6 estimates are available. Displayed are the averages (red stars), median values (blue lines), 33% and 67% percentiles (box boundaries) and 5% and 95% percentiles (ends of vertical lines) except for the inverse estimates, which is an expert assessment of the combined estimate of multiple inverse estimates.

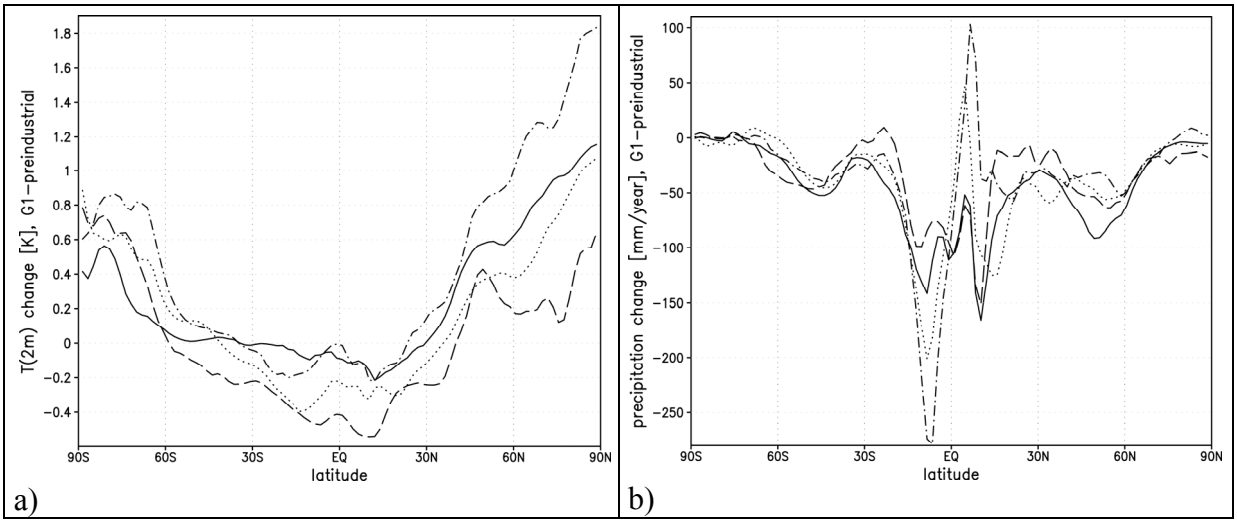
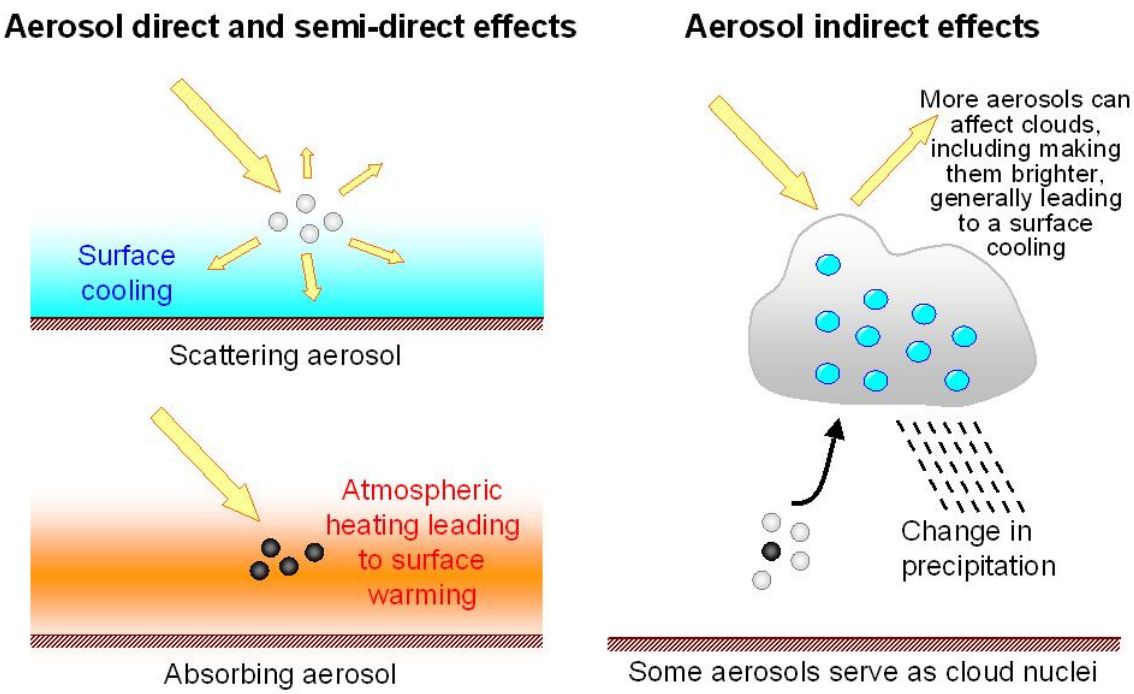
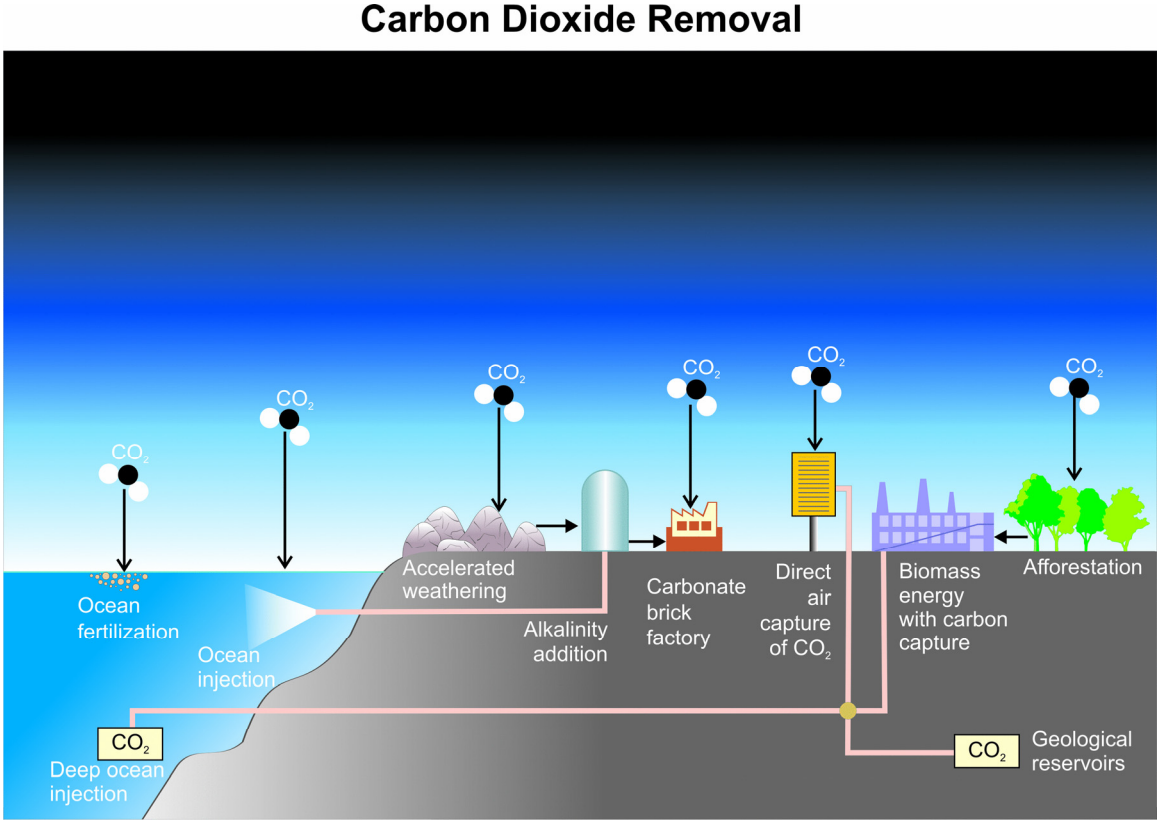


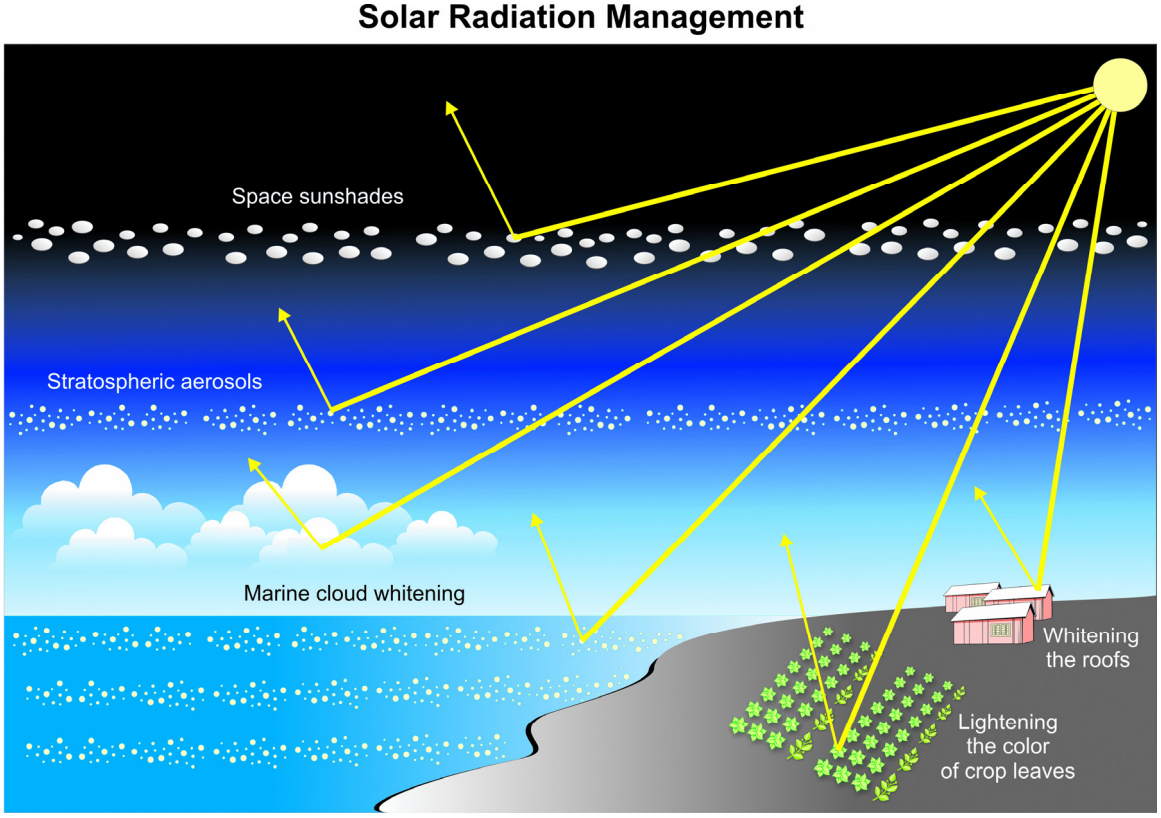
Figure 7.14: Multi-model mean of the residual surface temperature and precipitation changes from GeoMIP simulations with a simultaneous fourfold increase in CO_2 and a reduction in solar forcing which has been adjusted in each model to maintain the top of atmosphere net flux imbalance within ±0.1 W m^{-2} (Kravitz et al., 2011).



FAQ 7.1, Figure 1: Overview of aerosol direct and indirect effects on climate.



FAQ 7.3, Figure 1: Overview of carbon dioxide removal methods.



FAQ 7.3, Figure 2: Overview of solar radiation management methods.