Investigating aerosol–cloud interactions

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Abstract

Microphysical and dynamical interactions between clouds and aerosols are associated with some of the largest uncertainties in projections of future climate. Many possible aerosol effects on clouds have been suggested, but much more research is needed in order to estimate the size of these effects globally. In order to improve model projections of future climate, it is essential that we improve our quantitative understanding of these effects. Several studies investigating interactions between observed cloud and aerosol properties have been published in recent years. However, the observed correlations are not necessarily due to microphysical effects. They may be due to cloud flagging errors, retrieval errors, seasonal factors, humidity conditions or synoptic effects. Further work needs to be done in order to identify and quantify the physical reasons for these observations.

Using Moderate Resolution Imaging Spectroradiometer (MODIS) daily gridded satellite data, we present a preliminary analysis of aerosol–cloud interactions, providing an overview of some correlations between aerosol and cloud properties in different regions of the world. We find a globally persistent positive correlation between cloud fraction and aerosol optical depth and a negative correlation between cloud top pressure and aerosol optical depth. We also observe a strong negative correlation between cloud fraction and cloud top pressure. However, this preliminary analysis does not allow us to conclusively identify the reasons for these correlations. We outline a more detailed approach by which we intend to investigate the role of synoptic conditions.

Contents

Chapter 1

Introduction

1.1 Aerosols, clouds and climate

1.1.1 What are atmospheric aerosols?

The Intergovernmental Panel on Climate Change (IPCC) define *aerosols* as "airborne solid or liquid particles, with a typical size between 0.01 and 10 μm that reside in the atmosphere for at least several hours" [Forster et al., 2007]. The Earth's atmosphere contains many different kinds of aerosols, of both anthropogenic and natural origin. Some categories of aerosol which are considered to play an important role with regards climate are sulphate, organic carbon, black carbon, nitrate, mineral dust and sea-salt aerosols [e.g., Haywood and Boucher, 2000].

1.1.2 Direct and indirect aerosol effects on climate

The optical and microphysical properties of aerosols may give rise to several radiative effects in the Earth's atmosphere. Figure 1.1 contains a schematic illustrating some of these potential effects.

The *direct aerosol effect* refers to scattering and absorption of shortwave and longwave radiation by atmospheric aerosols. The direct effect radiative perturbation due to a given aerosol is dependent on the vertical distribution of the aerosol, the *albedo* (reflectivity) of the Earth's surface beneath and any clouds present [e.g. Haywood and Shine, 1995, Stier et al., 2007]. For example, for aerosols over a surface with a high albedo, such as snow or cloud, any absorption by the aerosol may dominate over scattering effects, leading to a net warming effect.

Heating due to the absorption of shortwave radiation by tropospheric aerosols can lead to increased temperatures in the aerosol layer, decreasing relative humidity and changing tropospheric stability. This can significantly affect cloud lifetime, and is known as the *semi-direct aerosol effect*. For example, a modelling study showed that absorption of shortwave radiation by black carbon aerosol can lead to an enhanced daytime clearing of trade cumulus clouds over the northern Indian Ocean [Ackerman et al., 2000].

Indirect aerosol effects are the radiative effects which aerosols can have through microphysical interactions with clouds. Many aerosols are effective cloud condensation nuclei. A strong correlation between cloud condensation nuclei concentrations and aerosol optical depth has been observed [Andreae, 2009]. Increasing the aerosol concentration in a cloud can lead to increased numbers of cloud condensation nuclei competing for water vapour which, in a cloud of constant liquid water content, in turn leads to a greater number of smaller droplets. This increases the albedo of the cloud [Twomey, 1977], resulting in more shortwave radiation being reflected to space, and is referred to as the *cloud albedo effect*. Precipitation can be suppressed and the lifetime of the cloud can be affected [Albrecht, 1989], an effect known as the *cloud lifetime effect*. The cloud top height can also be affected [Pincus and Baker, 1994]. Other indirect effects have been suggested, such as the *glaciation*

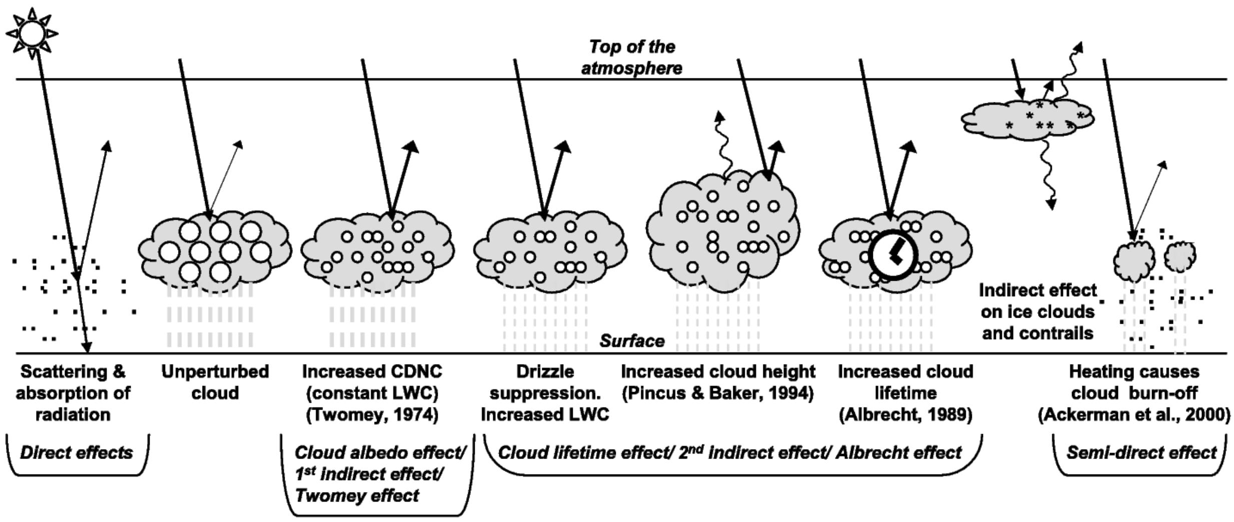


Figure 1.1: Schematic diagram showing the various radiative mechanisms associated with cloud effects that have been identified in relation to aerosols. The small black dots represent aerosol particles; the larger open circles cloud droplets. Straight lines represent the incident and reflected solar radiation, and wavy lines represent terrestrial radiation. The filled white circles indicate cloud droplet number concentration. The unperturbed cloud contains larger cloud drops as only natural aerosols are available as cloud condensation nuclei, while the perturbed cloud contains a greater number of smaller cloud drops as both natural and anthropogenic aerosols are available as cloud condensation nuclei. The vertical grey dashes represent rainfall, and LWC refers to liquid water content. [Figure taken from Forster et al., 2007.]

inirect effect, proposed by Lohmann [2002], whereby black carbon aerosols act as effective ice nuclei and can therefore enhance precipitation in the ice phase.

A complex interplay between different aerosol effects and feedbacks may exist in any given real-world situation. For example, Koren et al. [2008] propose that for clouds in smoky conditions over the Amazon, microphysical (indirect) aerosol effects dominate for low aerosol conditions and radiative (semi-direct) aerosol effects on the clouds dominate in high aerosol conditions.

It is worth noting that most aerosols have a much shorter lifetime than greenhouse gases, and so have a much stronger regional distribution and remain in the atmosphere for a much shorter period of time. Since many aerosol species can lead to health problems [see e.g. Bell et al., 2004, Kennedy, 2007], increasingly cleaner technologies are being employed in order to decrease aerosol emissions. Indeed, European emissions of sulphur have decreased by 70 % since 1980 [Grennfelt and Hov, 2005]. As a result, due to their short lifetime, the atmospheric concentration of aerosols is unlikely to increase significantly. Under these conditions, future warming due to the increasing greenhouse gas concentrations will increasingly dominate over aerosol radiative effects [Andreae et al., 2005, Kiehl, 2007]. Uncertainties in the size of aerosol radiative effects lead to uncertainties in the sensitivity of the climate's response to greenhouse gas forcing. In order to accurately forecast future warming trends, it is therefore important to quantify the significance of aerosol effects in the present, and to reduce the large uncertainty in the radiative forcing due to aerosols.

1.1.3 Radiative forcing, climate feedbacks and projection uncertainties

Radiative forcing is "the change in the net, downward minus upward, irradiance (expressed in Wm^{-2}) at the tropopause due to a change in an external driver of climate change" [Forster et al., 2007]. Positive radiative forcings lead to a warming of the climate system. Significant forcings regularly considered in attribution studies are those caused by changes in the solar constant, greenhouse gas

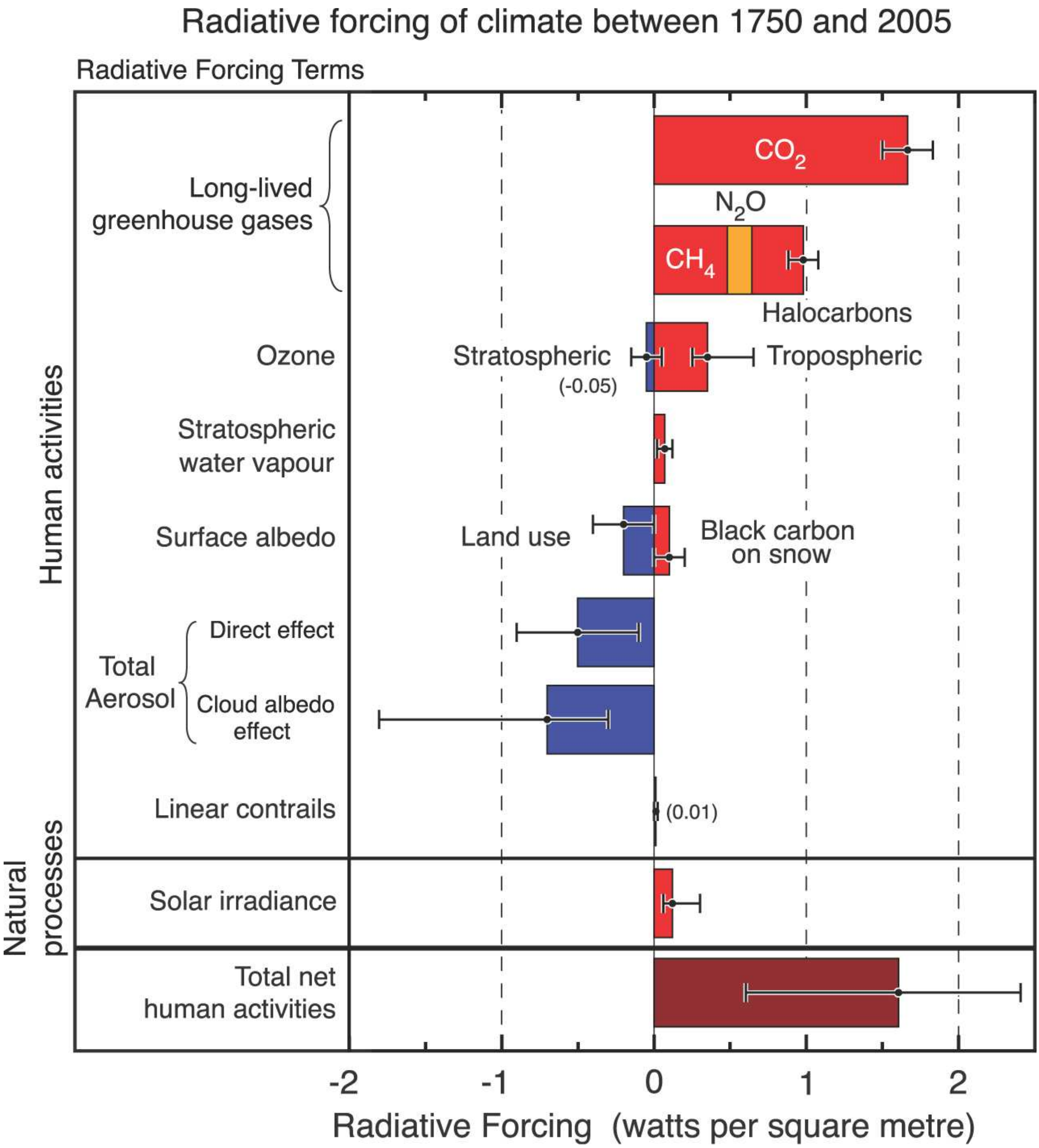


Figure 1.2: Summary of the principal components of the radiative forcing of climate change. The values represent the forcings in 2005 relative to the start of the industrial era (about 1750). Human activities cause significant changes in long-lived gases, ozone, water vapour, surface albedo, aerosols and contrails. The only increase in natural forcing of any significance between 1750 and 2005 occurred in solar irradiance. The thin black line attached to each coloured bar represents the range of uncertainty for the respective value. [Figure taken from Forster et al., 2007.]

concentrations, tropospheric sulphate aerosols and stratospheric volcanic aerosols [e.g. Stone et al., 2007]. Figure 1.2 summarises the climatic radiative forcing components identified by Forster et al. [2007].

Many mechanisms act either to intensify or to oppose changes in the climate system. These effects are known as *climate feedbacks*. For example, if a surface-air temperature increase in polar regions reduces snow cover, the surface albedo will be reduced and less solar radiation will be reflected, creating a positive feedback mechanism leading to a greater warming in these regions. Several major climate feedbacks are those associated with clouds, water vapour, the lapse rate and the cryosphere [Bony et al., 2006].

As shown in Figure 1.2, Forster et al. [2007] estimate an aerosol direct effect radiative forcing of -0.5 [-0.9 to -0.1] Wm^{-2} and a cloud albedo effect of -0.7 [-1.8 to -0.3] Wm^{-2}. Due to their interaction with the hydrological cycle, the other indirect effects and semi-direct effect are not considered as radiative forcings by Forster et al. [2007]. However, Lohmann and Feichter [2005] estimate an effective cloud lifetime effect radiative forcing of -1.2 [-1.9 to -0.5] Wm^{-2}, leading to a total net anthropogenic radiative forcing of 0.4 [-0.3 to 2.4] Wm^{-2}. As can be seen, there is a large uncertainty in the present day total anthropogenic radiative forcing, and most of this uncertainty is due to uncertainties in the size of aerosol effects.

1.2 Observational tools

Observations play an integral role in helping us to improve our understanding of the highly complex processes which occur in the atmosphere. They often highlight new avenues for investigation. They form the basis for scientific hypotheses. And they act as the plumbline by which theories must be assessed. As the physicist Richard Feynman famously said, “if it disagrees with experiment it is wrong” [Feynman, 2007]. This applies as much to atmospheric science as it does to fundamental physics. Observations help scientific research to remain rooted in reality.

Many observational datasets of the atmosphere exist. Some are produced using in situ surface, ship and aircraft measurments, such as those gathered during the Variability of the American Monsoon System (VAMOS) Ocean–Cloud–Atmosphere–Land Study Regional Experiment (VOCALS-REx) field campaign [Wood et al., 2007]. Other datasets contain data that has been remotely sensed from the surface, such as the aerosol data retrieved from the AErosol RObotic NETwork (AERONET) [Holben et al., 1998]. However, the observational datasets most widely used for aerosol–cloud interaction studies are those retrieved from satellite data.

1.2.1 Satellite platforms

Terra

Terra, also known as Earth Observing System (EOS) AM-1, was launched on 18th December 1999 and acts as a platform for a number of instruments: ASTER (Advanced Spaceborne Thermal Emission and Reflection Radiometer), CERES (Clouds and Earth’s Radiant Energy System), MISR (Multiangle Imaging SpectroRadiometer), MODIS (MODerate resolution Imaging Spectroradiometer) and MOPITT (Measurements Of Pollution In The Troposphere) [NASA, 2009b]. It is in a sun-synchronous orbit with a 10:30 A.M. equatorial crossing time [Kaufman et al., 1998].

Envisat

Envisat, launched in March 2002, is in a sun-synchronous orbit with a southwards equatorial crossing time of 10:00 A.M. [ESA, 2009]. It carries ASAR (Advanced Synthetic Aperture Radar), MERIS (Medium Resolution Imaging Spectrometer), AATSR (Advanced-Along Track Scanning Radiometer), RA-2 (Radar Altimeter 2), GOMOS (a medium resolution spectrometer), MIPAS (Michelson Interferometer for Passive Atmospheric Sounding), SCIAMACHY (an imaging spectrometer), DORIS (Doppler Orbitography and Radio-positioning Integrated by Satellite) and LRR (Laser Retro-Reflector).

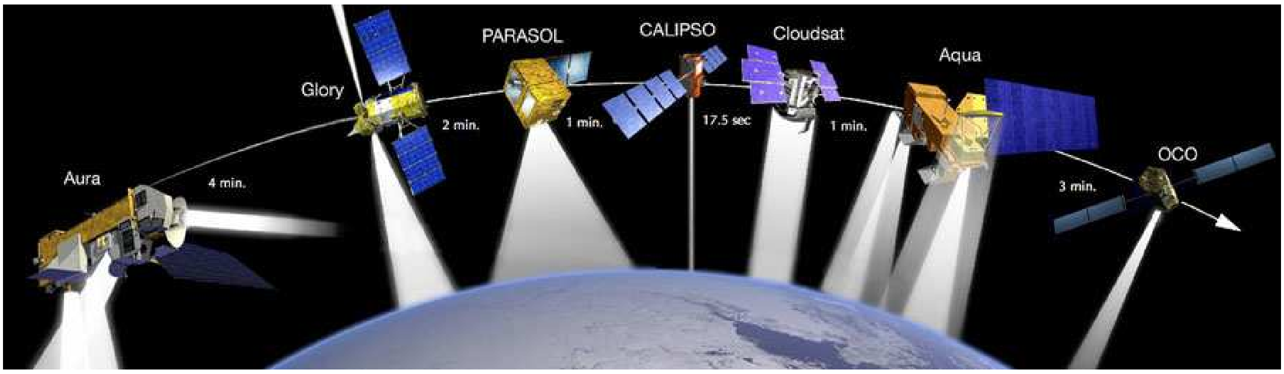


Figure 1.3: This graphic depicts the satellites that make up the Afternoon Constellation—"The A-train". Listed next to each satellite is the time lag, in terms of equator crossing time, between it and the preceeding satellite. The Oribiting Carbon Observatory (OCO) failed to reach orbit, so Aqua is the leading satellite in the A-train. [Figure taken from NASA, 2009c. Caption adapted from NASA, 2003.]

Aqua

Aqua, sometimes referred to as EOS PM-1, was launched on 4th May 2002 [Parkinson, 2003]. Like Terra, it is has CERES and MODIS instruments onboard, in addition to AIRS (Atmospheric Infrared Sounder), AMSR-E (Advanced Microwave Scanning Radiometer for the Earth Observing System), AMSU-A (Advanced Microwave Sounding Unit-A) and HSB (Humidity Sounder for Brazil). Aqua flies in a sun-synchronous orbit with a northwards equatorial crossing time of 1:30 P.M.. It is the first member, with respect to both launch date and orbital position, of the afternoon A-Train constellation of satellites (Figure 1.3). These satellites fly in close formation, so there is potential to combine data from different instruments in order to produce improved data products [e.g. Jeong and Hsu, 2008].

PARASOL

Another member of the A-train, Polarization and Anisotropy of Reflectances for Atmospheric Sciences coupled with Observations from a Lidar (PARASOL), was launched on 18th December 2004 [CNES, 2009b]. It caries a POLDER (POLarization and Directionality of the Earth's Reflectances) instrument.

CloudSat

CloudSat, also part of the A-Train, was launched on 28th April 2006, and carries the first satellite Cloud Profiling Radar (CPR) [Colorado State University, 2009].

CALIPSO

The Cloud Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) satellite, another part of the A-Train, was also launched on 28th April 2006 [CNES, 2009a]. The payload consists of CALIOP (Cloud-Aerosol LIdar with Orthogonal Polarization), IRR (Imaging Infrared Radiometer) and WFC (Wide Field Camera), all co-aligned [NASA, 2009a].

1.2.2 Satellite instrumentation

ATSR, ATSR-2 and AATSR

The Along-Track Scanning Radiometer (ATSR) was launched on ERS-1; ATSR-2 is on ERS-2; the Advanced Along-Track Scanning Radiometer (AATSR) is on ENVISAT [RAL, 2003]. The primary purpose of ATSR is to measure sea-surface temperatures [Mutlow, 1993]. However, the Global Retrieval of ATSR cloud Parameters and Evaluation (GRAPE) project [University of Oxford, 2009a] is developing a retrieval scheme capable of retrieving aerosol and cloud parameters (Oxford-RAL Aerosol and Cloud, ORAC [University of Oxford, 2009b]). A new aerosol algorithm has now been developed for AATSR [Sayer, 2008].

MODIS

The MODerate resolution Imaging Spectroradiometer (MODIS) has 36 bands in the visible and infrared, with nadir resolutions of 250–1000 m, and has a cross track swath width of 2330 km [Barnes et al., 1998]. There are MODIS instruments on both Terra and Aqua. They provide complete global coverage every 1–2 days [NASA, 2009g]. Terra MODIS data from 24th February 2000 – present and Aqua MODIS data from 3rd July 2002 – present are available [NASA, 2009d]. MODIS atmospheric data products include aerosol and cloud properties, as well as water vapour [NASA, 2009f]. MODIS also provides land data, such as vegetation indices [Huete et al., 2002], and ocean data, such as chlorophyll concentration [Esaias et al., 1998].

MISR

The Multi-angle Imaging SpectroRadiometer (MISR), on Terra, images the Earth in nine different view directions [Diner et al., 1998]. It can provide cloud height data, in addition to other cloud and aerosol properties [NASA, 2009e]. The swath width is 360 km.

POLDER-1, POLDER-2 and PARASOL

The POLarization and Directionality of the Earth's Reflectances (POLDER) instrument, designed to measure polarized and directional reflected solar radiation, has a swath width of approximately 2200 km and a resolution of 6×7 km^2 [Deschamps et al., 1994]. Measuring polarized and directional reflectances allows it to differentiate between radiation scattered in the atmosphere and that reflected by the Earth's surface, allowing improved observations of clouds, aerosols, the land surface and oceans [CNES, 2009c]. Data is available from three POLDER instruments: POLDER-1, on ADEOS-1, from November 1996 – June 1997; POLDER-2, on ADEOS-2, from April 2003 – October 2003; the third POLDER instrument, on PARASOL, from December 2004 onwards [Université Lille 1, 2009]. PARASOL is intended to observe clouds and aerosols [CNES, 2009b].

CloudSat CPR

The CloudSat Cloud Profiling Radar (CPR) operates at 94-GHz (3 mm wavlength), has a horizontal footprint of 1.4 km $\times$ 3.5 km and has a vertical resolution of 250 m [Posselt et al., 2008]. It measures several cloud parameters, and a cloud classification product is available.

CALIOP

The Cloud-Aerosol LIdar with Orthogonal Polarization (CALIOP), on CALIPSO, has been producing cloud and aerosol data since 2006, much of which has a horizontal resolution of 330 m and a vertical resoltion of 30–60 m [Winker et al., 2007].

1.2.3 Assessment of using satellite data for aerosol-cloud studies

In comparison to field campaigns, which provide datasets limited in both spatial and temporal extent, satellite sensors can provide global datasets extending over several years. Although AERONET releases aerosol data retrieved at many locations on Earth, the coverage is severely limited compared to that of many satellite datasets. The large datasets offered by satellite instruments are invaluable for statistical studies of the atmosphere and the processes which occur there.

However, satellites have a number of limitations. Cloud profiling radars, such as that onboard CloudSat, and lidars, such as CALIOP, can provide vertically resolved cloud and aerosol data, but they resolve only one horizontal dimension; radiometers, such as MODIS, often have large two-dimensional horizontal coverage but generally have no vertical resolution when retrieving aerosol and cloud properties. One limitation of particular interest here is that radiometers cannot retrieve both aerosol and cloud properties simultaneously: if a pixel is identified as cloudy, a cloud retrieval will be attempted; if a pixel is identified as cloud-free, an aerosol retrieval will be attempted. So retrieved aerosol and cloud properties are never completely co-located horizontally[1]. Also, radiometer products rarely allow vertical co-location to be tested: retrieved cloud properties are often that of the top of the highest layer of cloud and may not be representative of cloud below; retrieved aerosol properties are often column averages or totals and do not provide information on the location of the aerosol, which may be in defined layers.

1.2.4 Choice of satellite data

For large-scale statistical studies of aerosols and clouds, datasets providing extensive horizontal coverage are highly desirable. Several satellite radiometers fulfil this criterion. MODIS data is available at two overpass times (Terra 10:30 A.M.; Aqua 1:30 P.M.), the data covers several years, the spatial resolution is high and the swath is wide compared to many radiometers. MODIS aerosol and cloud products have undergone several validation and comparison studies [e.g Remer et al., 2005, Wu et al., 2009], and these products are commonly used in aerosol–cloud interaction studies (see Section 1.4). We have therefore chosen to use MODIS data at the beginning of this research, although data from other satellite sensors may also be used at a later stage. Since Aqua is in the A-Train, data from other A-Train instruments could readily be used in combination with Aqua MODIS.

1.2.5 Retrieved aerosol and cloud properties

Many aerosol and cloud properties are retrieved using satellite radiometer data. Some of the commonly retreived properties which are of particular interest when investigating aerosol–cloud interactions are as follows: *aerosol optical depth* (AOD) is the total extinction at a given wavelength due to aerosol in an atmospheric column; *aerosol index* (AI) is another measure of aerosol load in an atmospheric column; *liquid cloud droplet effective radius* (CER_{liquid}) is a retrieved estimate of the size of the droplets near the top of liquid water clouds; *liquid water path* (LWP) is the total mass of liquid water in an atmospheric column; *cloud optical depth* (COD) is the total extinction at a given wavelength due to cloud water (liquid, ice or both) in an atmospheric column; *cloud fraction* (CF) gives the total fractional cover of all clouds (liquid and ice) in a given region; *cloud top pressure* (CTP) is the average pressure at the top of all clouds (liquid and ice) in a given region.

[1]This is not completely true. By using PARASOL data together with other data from the A-train, Waquet et al. [2009] have demonstrated that it may sometimes be possible to retrieve properties of aerosols above clouds.

Although not a directly retrieved quantity, *cloud droplet number concentration* (CDNC), N_d, is sometimes estimated using the adiabatic approximation [Brenguier et al., 2000, Quaas et al., 2006]:

$$N_d = \gamma \tau_c{}^{\frac{1}{2}} r_e{}^{-\frac{2}{5}} \tag{1.1}$$

where τ_c is COD, r_e is $\text{CER}_{\text{liquid}}$ and $\gamma = 1.37 \times 10^{-5}$ $\text{m}^{-\frac{1}{2}}$.

1.3 Modelling tools

1.3.1 General circulation models and parameterization uncertainties

Modern climate prediction, like weather forecasting, relies heavily on complex numerical models used to simulate the Earth's atmosphere-ocean system [Meehl et al., 2007]. There are a number of state-of-the-art models which couple a dynamic ocean to a dynamic atmosphere, so-called *general circulation models* (GCMs), currently being used by different research facilities around the world [see e.g. Table 8.1 of Randall et al., 2007]. Each model simulates physical processes slightly differently.

Many physical processes, such as cloud formation, occur on scales too small to be resolved by GCMs, so these processes must be parameterized [e.g. Arakawa, 2004]. There is a wide range of potentially valid parameter values that can be employed for a given GCM [Randall et al., 2007]. One technique that is regularly used in probabilistic climate forecasting is to run an experiment using several different GCMs, generating a *multi-model ensemble* [e.g. Lambert and Boer, 2001]. An alternative approach is to vary the values of parameters in a single model, testing different parameter combinations, generating a *perturbed physics ensemble* [e.g. Stainforth et al., 2005, Piani et al., 2005]. Unfortunately, generating even modest-sized ensembles for a given experiment can be computationally expensive.

Climate*prediction*.net is a distributed computing project, using volunteers' computers distributed around the world to generate large perturbed physics ensembles for a number of climate experiments [Stainforth et al., 2002]. Using climate*prediction*.net data, Sanderson et al. [2008b] find that the response to greenhouse-gas forcing is dependent on two climate feedbacks largely regulated by two parameters: the *entrainment coefficient*, associated with the cloud convection parametrization, and the *ice fall speed*, which affects cloud cover and humidity. Much of the variation in *climate sensitivity*, the equilibrium temperature response for a doubling of atmospheric carbon dioxide, also appears to be due to parameters associated with clouds [Sanderson et al., 2008a].

Bony and Dufresne [2005] demonstrate that marine boundary layer clouds are a dominant source of uncertainty in tropical cloud feedbacks in GCMs. A review by Stephens [2005] highlights that much of the uncertainty in model projections of climate change is due to cloud feedbacks in models. Much work needs to be done in order to improve cloud parameterizations in GCMs, and hence improve predictions of future climate.

1.3.2 Using general circulation models for aerosol–cloud studies

Many GCMs have aerosol and cloud modules which attempt to simulate aerosol indirect effects [Penner et al., 2006]. Unlike the real atmosphere, it is often possible to switch GCM microphysics components on or off. This can allow the reasons for correlations between aerosol and cloud properties to be probed [e.g. Lohmann et al., 2006].

1.3.3 Cloud models and large-eddy simulations

Due to the computational expense of running simulations on a global scale, GCMs cannot normally be run at a high enough resolution to resolve cloud processes such as convection, so clouds must be parametrized. However, numerical models have been designed to explicitly simulate individual clouds. For example, Altaratz et al. [2008] use an axisymmetric model with high radial and vertical resolutions of 50 m, and a domain size of 4000 m in the radial direction and 5000 m in the vertical. Cloud-scale convection can be explicitly analysed and the effects of detailed microphysics shemes can be tested.

Large-eddy simulations are capable of resolving large-scale turbulence but require small-scale turbulence to be parametrized [Jacobson, 2005].

1.4 Published aerosol–cloud interaction studies

Aerosol effects on clouds and their potential significance to the climate system are currently poorly understood, as shown in Section 1.1. Much work needs to be done in order to improve our understanding of these effects. Several studies investigating interactions between observed cloud and aerosol properties have been published in recent years. We provide a brief survey of the findings of some of these studies, categorising them according to potential aerosol–cloud interactions and implications.

It is worth noting from the outset that an observed correlation between an aerosol property and a cloud property may not necessarily be due to aerosol effects on the cloud. We explore potential explanations for spurious correlations in Section 1.5.

1.4.1 Cloud albedo effect: CDNC and CER_{liquid}

Conceptually, the cloud albedo effect predicts that higher aerosol loads should lead to higher liquid cloud droplet number concentrations (CDNC) and smaller liquid cloud droplet effective radii (CER_{liquid}).

Quaas et al. [2008] find that higher MODIS AODs are generally associated with a higher CDNC (calculated using Equation 1.1). Similarly, a surface remote sensing and in situ study has shown that, for stratus clouds off the Californian coast, a positive correlation between AOD and CDNC exists [McComiskey et al., 2009].

Selecting North Atlantic stratiform clouds, Kaufman et al. [2005] find a negative correlation between CER_{liquid} and AOD in MODIS gridded daily data.

Using Along Track Scanning Radiometer ATSR-2 data for different regions and seasons, Bulgin et al. [2008] generally observed negative correlations between CER_{liquid} and AOD, although positive correlations were also often observed.

Ravi Kiran et al. [2009] claim that a decrease in CER_{liquid} observed during break spells in the indian monsoon is due to an increase in aerosol transportation to the continental tropical convergence zone during the break spells.

Using POLDER satellite data, Bréon et al. [2002] observe a negative correlation between CER_{liquid} and aerosol index (AI).

Suzuki et al. [2008] show that negative correlations exist between CER_{liquid} and AI for different liquid water clouds in data produced by NICAM-SPRINTARS (Nonhydrostatic Icosahedral Atmospheric Model - Spectral Radiation-Transport Model for Aerosol Species), a global cloud resolving model with a horizontal resolution of 7 km coupled with an aerosol transport model. They show that these findings are consistent with similar results observed in MODIS data.

Menon et al. [2008] compare aerosol-cloud relationships observed in MODIS and CERES satellite data with those produced by the GISS (Goddard Institute for Space Studies) GCM. They run three

GCM simulations: one with no aerosol microphysical effects on clouds; one with aerosol microphysical effects on low-level liquid clouds; one with winds nudged to reanalysis data, in addition to the microphysical effects. They observe that CER_{liquid} decreases with increasing AOD in the satellite data. A similar relationship, although weaker, is present in the data from GCM runs which include aerosol–cloud microphysical effects.

1.4.2 Cloud lifetime effect: LWP and CF

Conceptually, the cloud lifetime effect predicts that, due to the suppression of precipitation, a positive correlation should exist between liquid water path (LWP) and AOD. Increased cloud lifetime should also lead to a positive correlation between cloud fraction (CF) and AOD. Although the cloud lifetime effect is associated with suppression of precipitation and cloud height, we focus on these later.

Using MODIS gridded daily mean data, Koren et al. [2005] find a positive correlation between CF and AOD for convective clouds over the North Atlantic Ocean. However, using the ECHAM4 GCM and running a simulation with aerosol microphysical effects on clouds switched off, Lohmann et al. [2006] demonstrate that much of this increase in CF associated with high AOD conditions may be due to dynamical rather than microphysical effects.

The study by Menon et al. [2008] (see above) finds that for the MODIS and CERES data, high AOD conditions correspond to high CF. However, the nudged GISS GCM simulation results suggest that much of this correspondence may due to synoptic conditions.

Surprisingly, the study by Suzuki et al. [2008] (see Section 1.4.1) suggests that LWP may decrease slightly with increasing AI, a relationship observed in both the MODIS and NICAM-SPRINTARS data. This is inconsistent with the basic conceptual model of the cloud lifetime effect.

Using a single cloud model and LESs, Jiang et al. [2006] find that aerosols may increase the lifetime of shallow cumulus clouds. However their study suggests that, in addition to suppressing precipitation, smaller droplets can also lead to increased evaporation which would act to reduce the lifetime of the cloud. In some situations, it is therefore possible that an inverse cloud lifetime effect may sometimes occur.

Long-term trends in global cloud cover have been observed, although it may be unlikely that these trends are due to aerosol effects [Warren et al., 2007].

1.4.3 Aerosol effects on precipitation and CTP

Rosenfeld and Lensky [1998] observe that precipitation forming processes appear to be different for marine and continental convective clouds, and that marine clouds are modified as they move inland into more continental aerosol conditions. They also find that high aerosol conditions, due to biomass burning and urban air pollution, can significantly suppress precipitation. Air pollution has been observed to completely inhibit precipitation in some cases [Rosenfeld, 2000].

However, Khain et al. [2005], who use a non-hydrostatic two-dimensional cloud model to investigate the dynamical effects of aerosols on clouds, conclude that sometimes aerosols can enhance convection and lead to the formation of squall lines and intense precipitation.

An in situ aircraft study over the Amazon [Andreae et al., 2004] found that smoke aerosols decrease cloud droplet sizes and suppress precipitation. As a result, clouds could extend to greater heights, transporting water, aerosols and latent heat higher in the atmosphere, and leading to an increase in thunderstorms and hail. Rosenfeld et al. [2008] provide a similar conceptual model as to why increased cloud condensation nuclei concentrations can suppress precipitation in some clouds but enhance convection and precipation in other clouds (Figure 1.4).

The findings of Meskhidze et al. [2009], who use Terra and Aqua MODIS data to look at morning–afternoon differences, support the theory that aerosols may enhance convection over the Amazon.

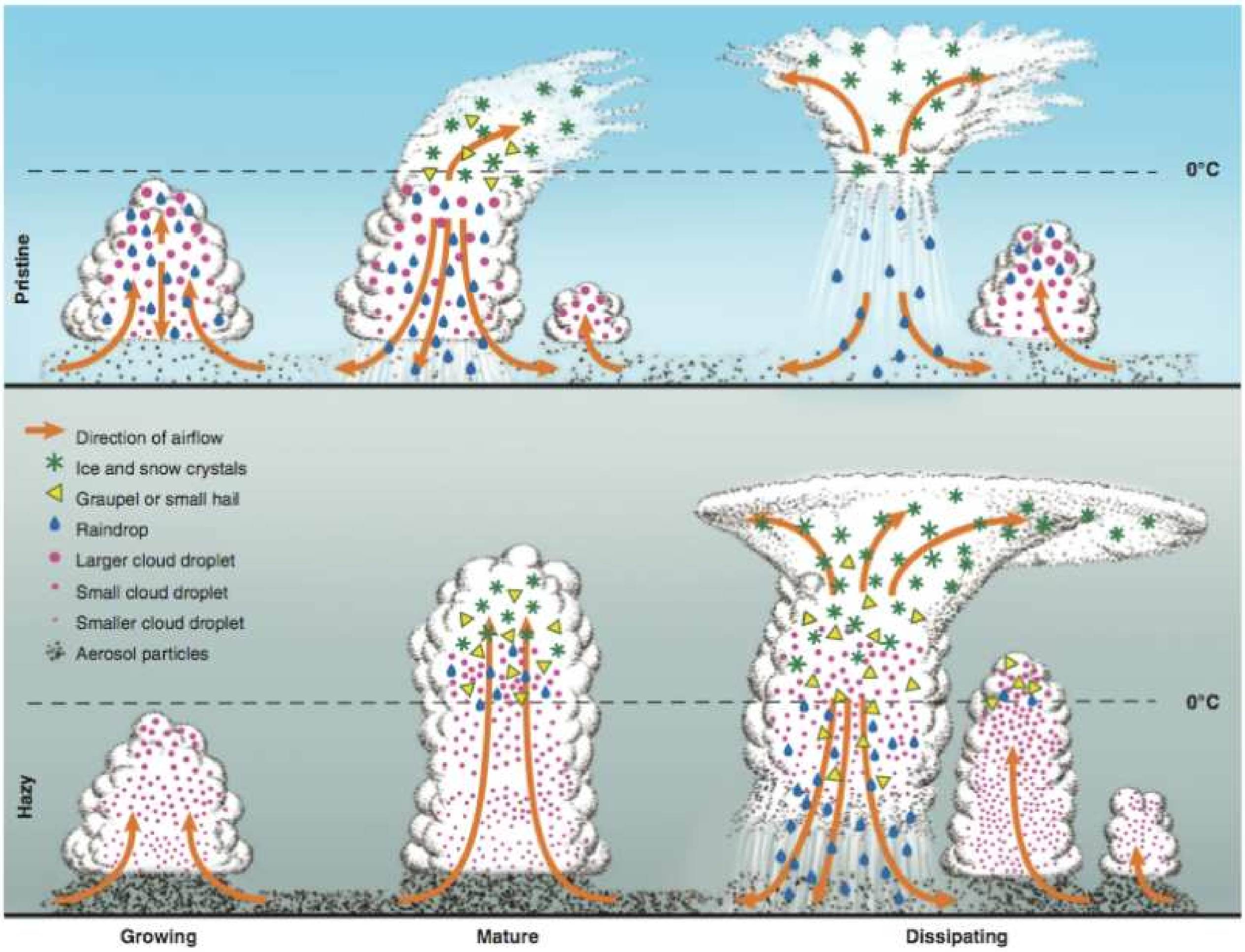


Figure 1.4: Conceptual hypothesis of the evolution of deep convective clouds developing in the pristine (top) and polluted (bottom) atmosphere. Cloud droplets coalesce into raindrops that rain out from the pristine clouds. The smaller drops in the polluted air do not precipitate before reaching the supercooled levels, where they freeze onto ice precipitation that falls and melts at lower levels. [Figure taken from Rosenfeld et al., 2008.]

Koren et al. [2005] (see Section 1.4.2) find a negative correlation between CTP and AOD, suggesting that suppression of precipitation may allow convective clouds to extend to greater heights in the North Atlantic.

1.4.4 Local aerosol effects with a global scope

Andreae et al. [2004] (see Section 1.4.3) suggest that the suppression of precipitation in the Amazon could affect the global circulation and water cycle. An ECHAM4 GCM study by Nober et al. [2003] supports the idea that the suppression of precipitation and enhancement of convection by aerosols can affect the global circulation.

Other mechanisms may exist by which aerosol effects could have a global scope. Evan et al. [2008] use satellite observations and a simple model to investigate how aerosol changes have affected cyclone development in the tropical Atlantic. They argue that a reduction in aerosol direct radiative forcing would lead to higher sea-surface temperatures which could result in an increased frequency of tropical cyclones. Since tropical cyclones probably play an important role in regulating stratospheric humidity [Romps and Kuang, 2009], they may in turn affect climate and stratospheric ozone [Shindell, 2001].

1.4.5 Relative humidity and the direct effect

Interactions between aerosols and clouds may also be indirectly contributing to underestimates of direct aerosol radiative forcing. AOD has been observed to increase near clouds in both photometer and MODIS data, although MODIS retrievals may not be accurate in such cases [Redemann et al., 2009]. This increase is most likely due to the swelling of aerosols in the high humidity environments near clouds. Koren et al. [2007] argue that satellite measurements of AOD are biased towards cloud-free environments, and these measurements are therefore unrepresentatively low because they do not include scenes where aerosols are hygroscopically large in the high humidity environments near clouds. Using aircraft observations collected during the Indian Ocean Experiment (INDOEX), Twohy et al. [2009] provide evidence that relative humidity effects on aerosols in the vicinity of clouds can lead to a 35–65 % enhancement in direct radiative forcing, and that this is not properly accounted for in global radiative forcing estimates.

1.4.6 Dependence on specific conditions

Predicting how a cloud will respond to aerosol is complicated.

Cui et al. [2006] conducted a study using an axisymmetric model of a mixed phase convective cloud in low wind-shear continental conditions and found that increasing aerosols led to stagnated cloud development, a lower cloud top, weaker updrafts and suppressed precipitation. They mention that cloud response to aerosols appears heavily dependent on the type of cloud and conditions.

Using a numerical model of convective clouds, Altaratz et al. [2008] show that aerosol effects are dependent on relative humidity conditions.

Jones et al. [2009], who use MODIS data together with reanalysis data, suggest that synoptic conditions, aerosol type and the vertical location of an aerosol layer may be much more significant factors than AOD.

Using MODIS satellite and SPRINTARS model data, L'Ecuyer et al. [2009] suggest that sulphate and sea-salt aerosols may have opposite effects on clouds: sulphate aerosol can decrease precipitation and enhance vertical development, whereas sea-salt can increase precipitation and supress vertical development. This is probably because hydrophilic sea-salt can act as a giant cloud condensation nucleus on which large droplets can grow.

1.5 Possible reasons for aerosol-cloud correlations

Many relationships between aerosols and cloud properties have been observed, and many potential implications have been suggested in recent years. However, the observed correlations are not necessarily due to microphysical effects. They may be due to erroneous satellite data, seasonal factors, humidity conditions or synoptic effects.

Satellite datasets are not completely reliable. One potential problem of interest here is that of cloud-flagging errors associated with false identification of cloud or aerosol. Thin cloud, which may well be in a broken cloud field, may be flagged as cloud-free and a high AOD may therefore be retrieved, resulting in a false correlation between high AOD conditions and cloud cover.

Another problem is that retrievals are not always accurate. As can be seen in Figure 1.5, MODIS and ATSR CER_{liquid} retrievals do not agree. (Note: the ATSR CER_{liquid} retrieval is still a work in progress.) Bréon and Doutriaux-Boucher [2005] find a poor correlation between MODIS and POLDER CER_{liquid} over land, with a better correlation over ocean, although MODIS CER_{liquid} is generally higher. Polder is limited to homogeneous cloud fields. Marshak et al. [2006] suggest that the MODIS CER_{liquid} retrieval may not be reliable for inhomogeneous cloud fields.

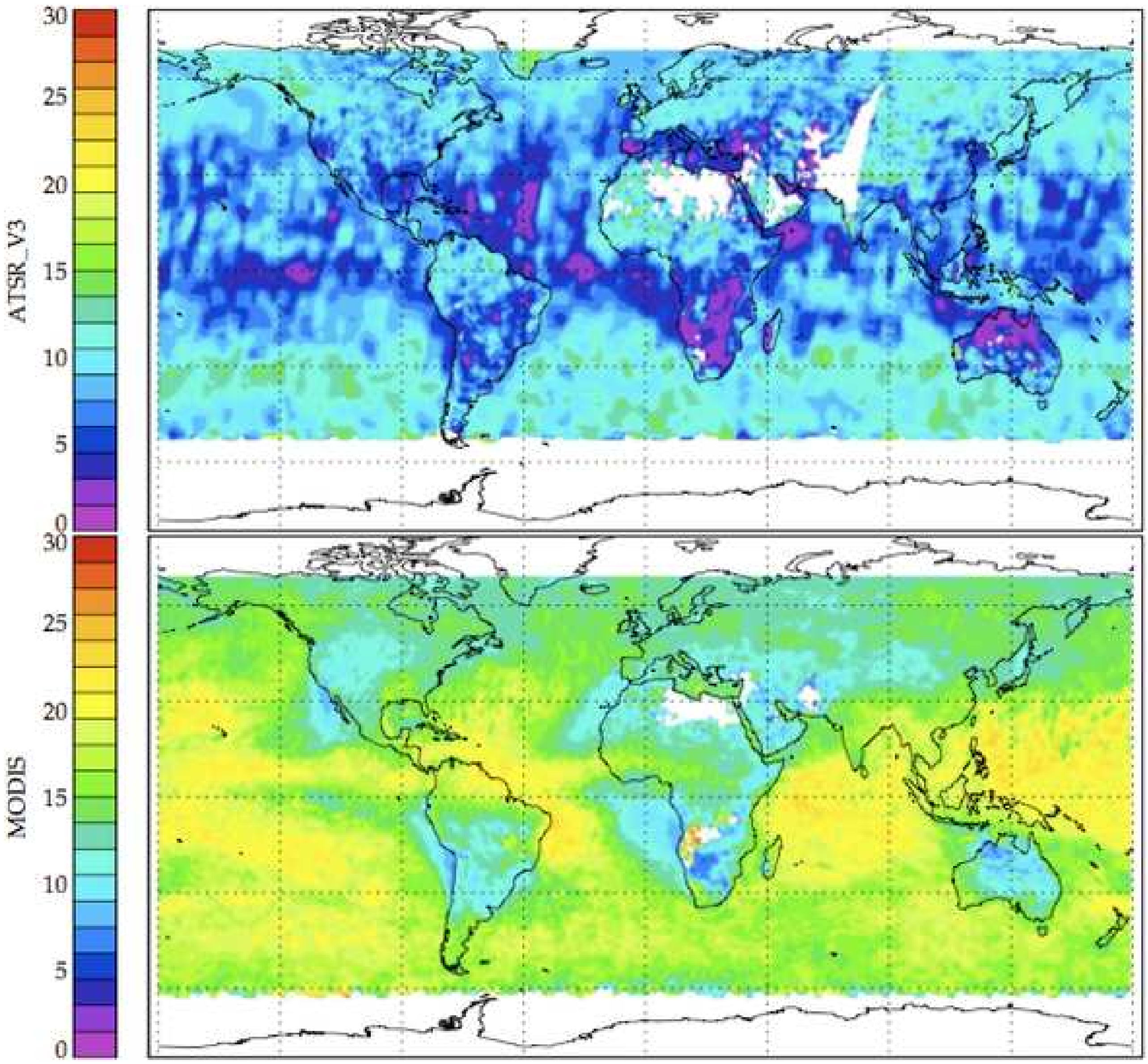


Figure 1.5: Comparison between monthly mean CER_{liquid} (μm) retrievals for ATSR (top; work in progress) and MODIS (bottom) for July 2000. [Figure courtesy of C. Poulson.]

Many regions of the world experience seasonal changes which may include pronounced seasonal cycles of aerosol and cloud properties. Many studies already aim to remove this factor by looking at individual seasons.

As mentioned in Section 1.4.5, aerosols often swell hygroscopically in high humidity conditions, significantly increasing AOD. Since clouds also occur in high relative humidity environments, this could lead to spurious correlations between AOD and cloud properties.

It is also possible that synoptic conditions may often lead to spurious correlations between aerosols and clouds.

Further work needs to be done in order to identify and quantify the physical reasons for the observed correlations between aerosol and cloud properties.

1.6 Aims of this project

The primary aim of this project is to improve our quantitative understanding of aerosol–cloud interactions, particularly those involving convective clouds, and their role in the Earth's climate system. By developing new methods of analysing available satellite data, we will explore some of the possible reasons for different correlations between aerosol and cloud properties. One of our main areas of focus will be to investigate the role of synoptic conditions. We will use GCM simulations to test our hypotheses.

Chapter 2

Aerosol-cloud interactions: A global panorama

We present our preliminary analysis of aerosol–cloud interactions, providing an overview of some relationships between aerosol and cloud properties in different regions of the world.

2.1 Method, data and properties used

We use six years (2003–2008) of the $1^\circ \times 1^\circ$ gridded daily mean Level 3 data product from the MODIS instrument onboard the Aqua satellite[1] (MYD08_D3, collection 5).

For each MODIS Level 2 pixel, the scene is identified as either cloudy or cloud-free, and a corresponding cloud or aerosol retrieval is attempted. Aerosol and cloud properties are not retrieved simultaneously. However, since several Level 2 pixels contribute to each Level 3 grid box, it is possible for both aerosol and cloud data to be available for a given $1^\circ \times 1^\circ$ grid box and day. We treat aerosol and cloud retrievals within a grid box as if they are co-located, based on the assumption that aerosol retrievals made in one part of a grid box are representative of the aerosol conditions throughout the grid box [Anderson et al., 2003].

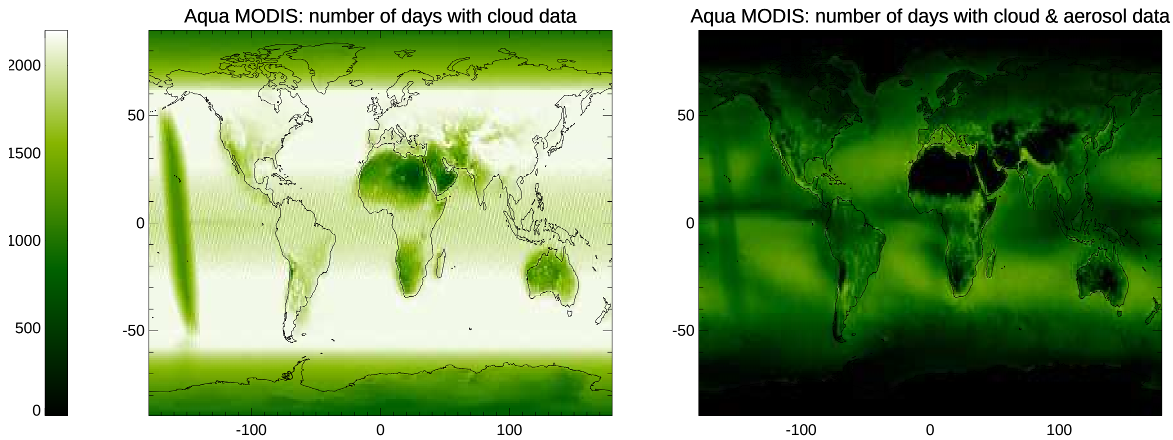
Aerosol data is available for the daytime only. We have chosen to limit our cloud data to daytime only retrievals, in order to focus on clouds which are present at the same time as the aerosol.

In this chapter, we focus on the following aerosol and cloud properties, available as Scientific Data Sets (SDSs) in the MYD08_D3 files:

- Aerosol optical depth (AOD). We use SDS Optical_Depth_Land_And_Ocean_Mean. We limit our data to grid boxes and days where at least six native resolution pixel aerosol retrievals contribute to the grid box mean, as suggested in Hubanks et al. [2008].
- Liquid cloud droplet effective radius ($\mathrm{CER}_{\mathrm{liquid}}$).
  We use SDS Cloud_Effective_Radius_Liquid_Mean.
- Cloud fraction (CF) for all clouds (liquid and ice). We use SDS Cloud_Fraction_Day_Mean.
- Cloud top pressure (CTP) of all clouds (liquid and ice).
  We use SDS Cloud_Top_Pressure_Day_Mean.

Figure 2.1(a) shows the number of days of available daytime cloud data for each $1^\circ \times 1^\circ$ grid box. As can be seen, there is greatest coverage in the midlatitudes, where the orbital swathes overlap. The slight reduction in coverage in the tropics is because the orbital swathes no longer overlap and

[1] If Terra MODIS is used, the results generally appear very similar to those presented here. At some point, we may identify and analyse differences between Terra MODIS and Aqua MODIS data.



(a) Cloud (b) Cloud and aerosol

Figure 2.1: Number of days of available Aqua MODIS daily gridded $1^\circ \times 1^\circ$ Level 3 data for 2003–2008.

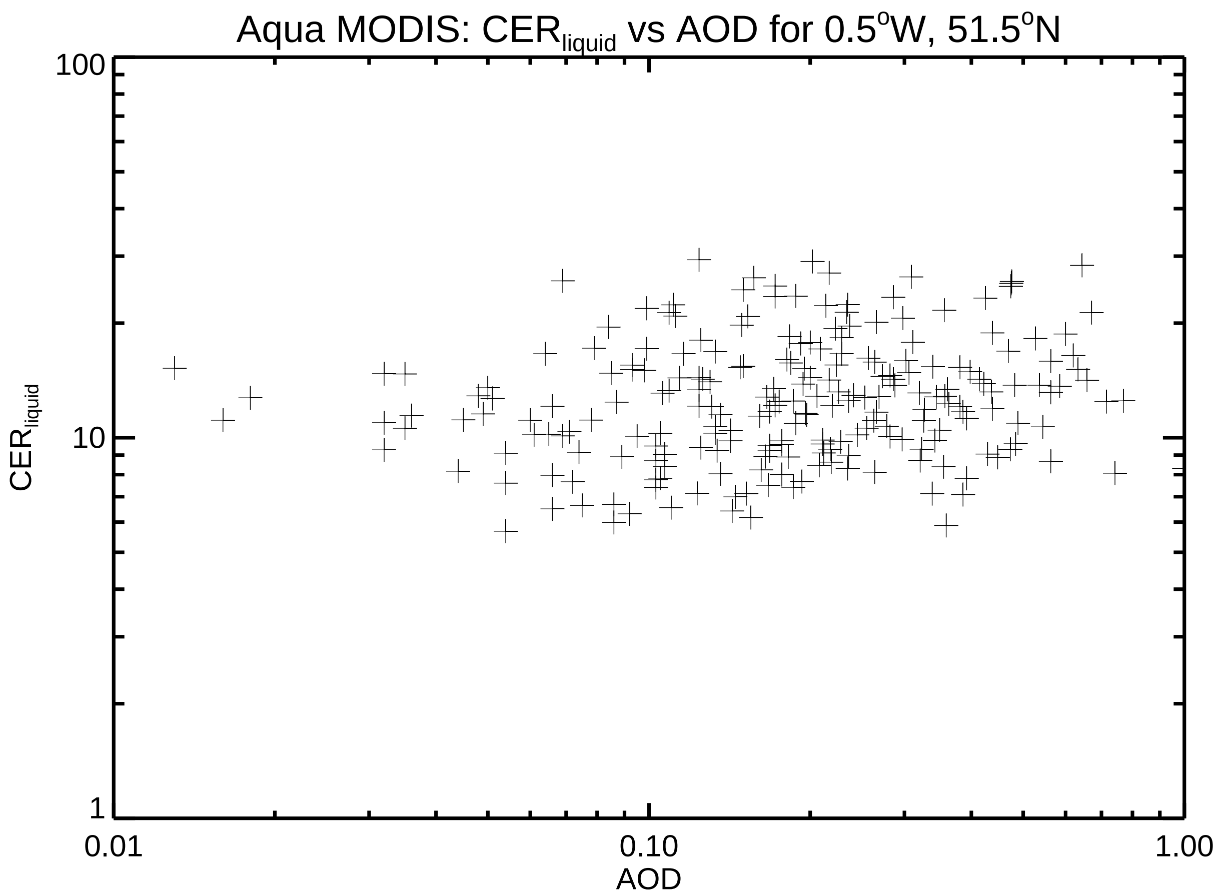


Figure 2.2: Scatter plot of CER$_{\text{liquid}}$ (μm) vs AOD (dimensionless) for the $1^\circ \times 1^\circ$ grid box centred on 0.5°W and 51.5°N (near Oxford). Each cross represents a day for which both aerosol and cloud data exist at this location. Here, the correlation (Linear Pearson correlation coefficient) $= 0.141$ and sensitivity ($\frac{\text{dln(CER}_{\text{liquid}})}{\text{dln(AOD)}}$) $= 0.056$.

have gaps between them due to the Earth's geometry. (1° longitude is wider at the equator than it is near the poles.) The reduction towards the poles is due to the reduction in day length during each hemispheric winter. We are currently unsure about the origin of the stripe of reduced data coverage over the Pacific Ocean, although it may be due to orbital gaps[2].

Figure 2.1(b) shows the numbers of days where both aerosol and daytime cloud retrievals are available. The very low coverage over snow (the poles, Greenland and high mountains) and desert (e.g. North Africa and the Arabian peninsula) is most likely due to the high reflectance of these surfaces leading to failed AOD retrievals [von Hoyningen-Huene et al., 2003]. Areas which often have complete cloud cover, such as the Intertropical Convergence Zone and the stratocumulus regions (for example, off the Californian coast [Albrecht et al., 1988]), have correspondingly fewer days where aerosol retrievals are available.

For each $1^\circ \times 1^\circ$ grid box, we calculate linear Pearson correlation coefficients and linear regression slopes between the natural logarithms of different cloud properties and AOD. Figure 2.2 shows $\mathrm{CER}_{\mathrm{liquid}}$ vs AOD for one grid box. For the remainder of this chapter, we refer to linear Pearson correlation coefficients simply as *correlations*. Following the method of Quaas et al. [2008], we define the *sensitivity*, b, of a cloud property, ϕ, to AOD, τ_a, as

$$b = \frac{\mathrm{d}\ln\phi}{\mathrm{d}\ln\tau_a}. \tag{2.1}$$

We make global maps of these correlations and sensitivities.

Figure 2.3 shows a map of the sensitivity of $\mathrm{CER}_{\mathrm{liquid}}$ to AOD. In the red and yellow regions, such as the Mediterranean, $\mathrm{CER}_{\mathrm{liquid}}$ increases as AOD increases; in blue and purple regions, $\mathrm{CER}_{\mathrm{liquid}}$ decreases as AOD increases.

[2]Personal correpsondence with Paul Hubanks (NASA) and Thorwald Stein (Reading).

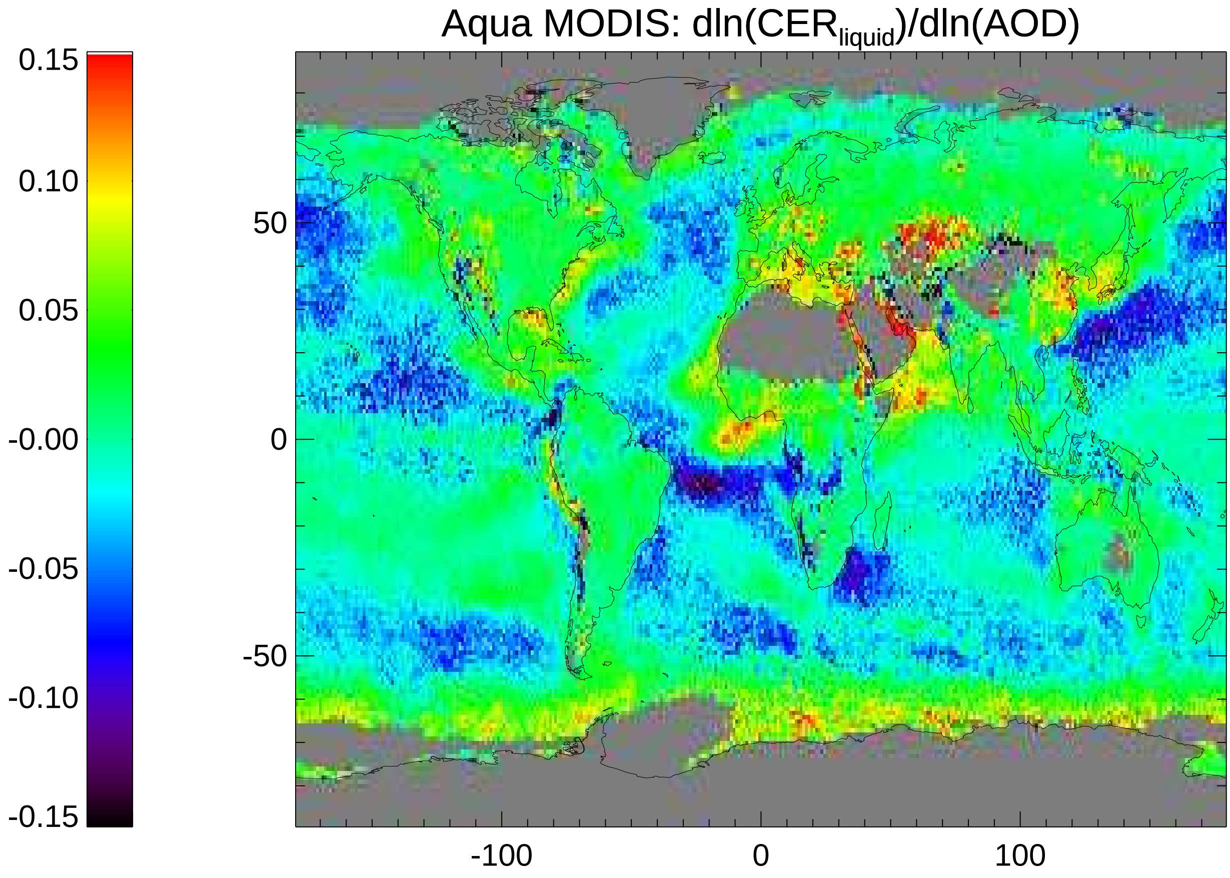
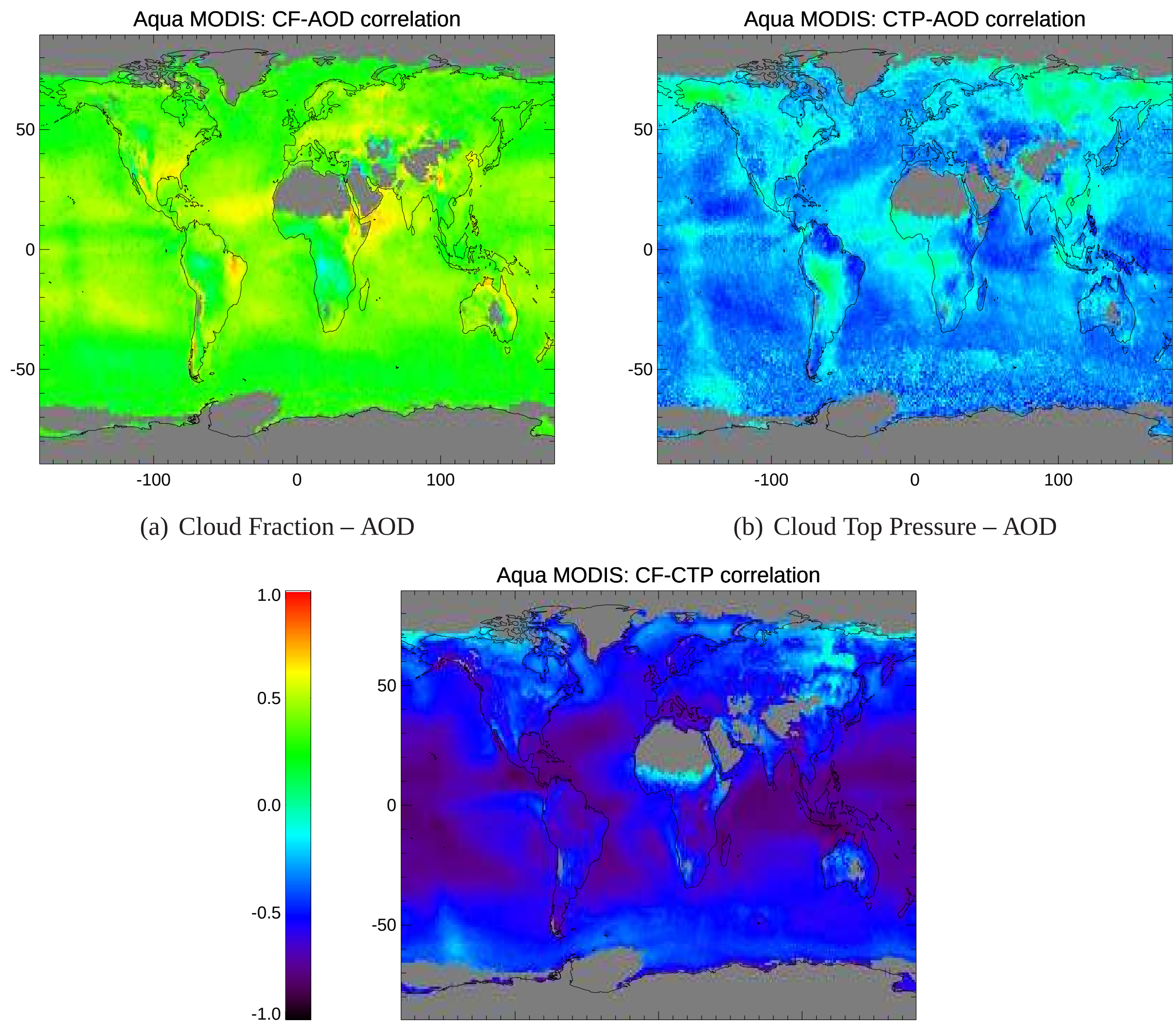


Figure 2.3: Sensitivity (dimensionless) of CER_{liquid} to AOD for each $1^\circ \times 1^\circ$ grid box using six years (2003–2008) of MODIS daily Level 3 data.



(a) Cloud Fraction – AOD

(b) Cloud Top Pressure – AOD

(c) Cloud Fraction – Cloud Top Pressure

Figure 2.4: Linear Pearson correlation coefficients (dimensionless) between different aerosol and cloud properties for each each $1^\circ \times 1^\circ$ grid box using six years (2003–2008) of MODIS daily Level 3 data.

2.2 Discussion

Considering only the cloud albedo effect, we would expect that higher AOD conditions should lead to smaller CER_{liquid} measurements. So we anticipate that the sensitivity of CER_{liquid} to AOD should be negative. However, as can be seen in Figure 2.3, positive sensitivities are observed in many regions. These findings appear to contradict those presented by Quaas et al. [2008] and Quaas et al. [2009], who use a similar (but not identical) MODIS dataset. The following points should be considered:

- We have not yet looked at individual seasons. It is possible that part of the signal in Figure 2.3 is due to a seasonal cycle.
- Quaas et al. conduct their analysis over large-scale regions, whereas as we use individual $1^\circ \times 1^\circ$ grid boxes.
- Quaas et al. use the Single Scanner Footprint Edition 2B data set, whereas we use MODIS MYD08_D3.
- The reliability of the MODIS CER_{liquid} data may be questionable (see Section 1.5).

However, our preliminary further investigations still suggest that our results are irreconcilable with those of Quaas et al. [2008] and Quaas et al. [2009]. One of our primary aims for the near future is to re-examine our method and identify the reasons for these differences (see Section 2.3).

Figure 2.4(a) shows that for most of the world, higher AOD conditions are associated with greater cloud cover. Figure 2.4(b) shows that, in general, higher AOD conditions are associated with higher clouds. These results are in agreement with those of Koren et al. [2005] who studied convective clouds in the North Atlantic. However, as discussed in Section 1.5, these observed correlations are not necessarily due to microphysical effects. They may be due to cloud flagging errors, retrieval errors, seasonal factors or synoptic effects. Figure 2.4(c) shows a strong and globally persistent negative correlation between CF and CTP, showing that higher cloud tops generally correspond to greater total cloud cover. Since synoptic conditions are likely to lead to correlations between cloud properties such as CF and CTP, this illustrates the potential for spurious correlations to be present.

The relationships between cloud and aerosol properties are complicated, and there is potential for oversimplification to occur if large regions are averaged over. Based on this preliminary analysis, it is very difficult to identify and distinguish microphysical, seasonal, synoptic, humidity and retrieval-related effects.

2.3 Future plans

In July 2009, the author submitted a research proposal to the Max Planck Institute for Meteorology, proposing to investigate the correlations between cloud and aerosol properties in more depth. The stated objectives of the project are as follows:

1. Our preliminary cloud–aerosol sensitivity results, calculated at the University of Oxford, appear inconsistent with those published in Quaas et al. [2008] and Quaas et al. [2009]. Our first objective is to identify the reasons for these discrepancies, by comparing specific methodologies and datasets.
2. Remove the seasonal component in observed sensitivities. We will initially attempt to do this by calculating sensitivities for different seasons separately, and exploring robustness with respect to season definitions. There is also potential to explore the possible application of statistical methods such as cyclo-stationary EOF analysis.

3. Explore the effect of regional domain choices on calculated sensitivities. We will do this by conducting our analysis for regional domains of differing sizes, ranging from $1^\circ \times 1^\circ$ grid boxes to entire continents and oceans.

4. Explore the effect of organising the data into different cloud regimes, either based on 500hPa vertical motion analysis data, lower tropospheric stability or the application of an ISCCP[3]-type classification to the satellite data.

5. Explore the effect of organising the data into different aerosol types, by looking at regions and seasons where specific aerosol types are known to dominate.

6. Test the robustness of sensitivities with respect to spatial and temporal resolution, by calculating sensitivities for a given region using different resolutions for the data.

7. Test the robustness of sensitivities with respect to choice of year, by investigating inter-annual variations.

8. Test the robustness of these results with respect to the choice of observing instrument and retrieval. We will investigate similarities and differences between MODIS and ORAC ATSR-2 results. Data from MISR may also be used.

9. Identify regions where more detailed study would be particularly beneficial.

The author has subsequently been awarded an International Max Planck Research School on Earth System Modelling (IMPRS-ESM) scholarship to visit the Max Planck Institute for Meteorology for a nine week period during October–December 2009, to work with Stefan Kinne, Johannes Quaas and their respective groups.

[3]International Satellite Cloud Climatology Project.

Chapter 3

A storm-centric approach

When correlations between aerosol and cloud properties are observed, it is difficult to conclusively identify reasons for the observations. One possible way to investigate the importance of synoptic conditions would be to organise the data based on vertical motion or lower tropospheric stability, as suggested in Section 2.3. Another would be to classify the data according to cloud type, although the ISCCP classification scheme is not effective at distinguishing between low cloud classes, and different cloud types often occur together [Hahn et al., 2001]. In this chapter we begin exploring another two possibilities: storms (extratropical and tropical cyclones) and fronts.

3.1 Aerosols, clouds, storms and fronts

It is possible that storms may lead to spurious correlations between aerosol and cloud properties. For example, the high windspeeds associated with storms may lead to high AODs, due to sea-salt over oceans and dust over continents. Since storms often produce high clouds with a high coverage, this may contribute to the negative CTP–AOD correlations and the positive CF–AOD correlations we see in Figure 2.4.

The frontal systems associated with extratropical cyclones often produce large bands of convective cloud. These clouds have the potential to both cycle and remove aerosol from the atmosphere. It would be interesting to investigate whether there are any significant differences between pre-frontal and post-frontal AOD. If is possible that such differences could contribute to spurious correlations between aerosol and cloud properties.

Although several studies have focused on the effect that climate change, including aerosol effects, may have on storms [e.g. Evan et al., 2008, Bengtsson et al., 2007b], comparatively little research has been done exploring the effect that storms may have on aerosols.

Over land, cold fronts have been observed to remove aerosols near the surface [Jia et al., 2008]. However, the question remains as to how they affect the total aerosol column burden.

In a case study in West Africa, Crumeyrolle et al. [2008] investigate how a mesoscale convective system effects the properties of aerosol layers, through gust generation of dust aerosol, washout, cloud processing and the mixing of layers. During a cyclone over India, an increase in ground-level PM_{10}[1] was observed although the total AOD decreased, possibly due to winds leading to increased ventilation; over the Bay of Bengal, the aerosol load and dust increased, as did atmospheric water-vapour [Badarinath et al., 2008]. High windspeeds can lead to significantly increased AODs over the ocean, due to both hygroscopic growth of aerosols and increased sea salt aerosol mass [Glantz et al., 2009]. Above a certain windspeed threshold, it is possible that sea salt concentrations near the surface may decrease due to scavenging by spray droplets [Pant et al., 2008], but it is unlikely that this decrease would be observed in column measurements of AOD.

[1]Particulate matter with a diameter of less than 10 μm.

Case studies are useful, but a statistical compositing approach should yield more robust and potentially more interesting results. As far as the author is aware, no previous research has considered this approach to investigating aerosol–cloud interactions.

3.2 Possible methods

Several methods of compositing cyclones have been developed and used for meteorological research. Lau and Crane [1995] use ISCCP data to build composites of marine tropical and extratropical cyclones based on peaks in timeseries of cloud optical depth. Lau and Crane [1997] use the same method to compare surface observations to ISCCP. Norris and Iacobellis [2005] use a similar method, but choose warm and cold advection as the compositing variable. Minima in surface pressure can also be used to detect cyclone centres for compositing [Wang and Rogers, 2001]. Chang and Song [2006] build monthly cyclone composites of precipitation using ECMWF (European Centre for Medium-Range Weather Forecasts) ERA-40 reanalysis data, and then use satellite and surface observations for comparison. Field and Wood [2007] use NCEP-NCAR (National Centers for Environmental Prediction - National Center for Atmospheric Research) reanalysis surface pressure to locate the centres of approximately 1500 cyclones, and then build composites based on strength and moisture categories. Field et al. [2008] use the same compositing technique to compare output from different versions of the CAM (Community Atmosphere Model) GCM with satellite data.

A general feature detection and tracking method has been developed for cyclones [Hodges, 1994]. It uses relative vorticity, so can detect both tropical and extratropical cyclones. This method has been used successfully in a number of studies [e.g. Bengtsson et al., 2007a,b]. It is this method that we intend to use for our research. Kevin Hodges has given us access to his tracking code, known as TRACK [Hodges, 2008], and has helped us to configure it to track extratropical cyclones in ECMWF operational analysis data. After removing tracks which persist for less than two days or move a distance of less than 1000 km, we have 1758 northern hemisphere and 1739 southern hemisphere probable extratropical cyclone tracks for the year 2007. As an example, Figure 3.1 shows one of the storm tracks detected when we apply our configuration of TRACK to some regridded ECMWF operational analysis relative vorticity data. This particular cyclone was detected in the north-eastern Atlantic Ocean and moved eastwards, as expected. We intend to develop our own compositing methodology using the storm track output.

Many attempts have been made to develop objective front identification methods [Hewson, 1998]. Objective front products are beginning to be used in scientific research [ECMWF, 2009]. It is possible that we may be able to obtain a license to use Met Office objective front code at some point in the future[2], and look at aerosols and aerosol-cloud interactions on a pre-frontal and post-frontal basis.

[2]Personal correspondence with Tim Hewson (ECMWF) and Richard Swinbank (Met Office).

Figure 3.1: Example of an extratropical cyclone track identified by our current TRACK configuration. The track starts at the most westward lozenge at 1200 UTC on 2nd January 2007 and finishes at the most eastward lozenge at 1200 UTC on 5th January 2007. Each lozenge is separated by a time of 3 hours.

Chapter 4

Skills and future plans

4.1 Transferable skills

The previous year has provided many opportunities to develop and consolidate many computing skills, including the use of Linux systems, Mac OSX, LaTeX, Microsoft PowerPoint, IDL, Python and shell scripts.

Attending weekly modelling group meetings, EODG group meetings, a National Centre for Earth Observation meeting (21$^{\text{st}}$–22$^{\text{nd}}$ May 2009) and the Royal Meteorological Society's (RMetS) 2009 Conference (30$^{\text{th}}$ June – 2$^{\text{nd}}$ July 2009) has provided opportunities to discuss the work presented in this report and the work of others. Presenting a poster at the RMetS Conference helped with the development of scientific communication skills.

Writing and submitting a research proposal to the Max Planck Institute for Meteorology was a good exercise in written communication, as was writing the current report.

In Trinity Term 2009, the author attended a weekend intensive German language course at the University of Oxford Language Centre, in preparation for the proposed visit to Germany in Michaelmas Term 2009.

4.2 Thesis outline

4.2.1 Chapter 1: Introduction

Chapter 1 will provide an introduction to aerosols, clouds and climate, including a literature review. The aims of the project will be outlined.

4.2.2 Chapter 2: Satellites and observational methods

Chapter 2 will introduce the satellites and observational datasets of interest to the research.

4.2.3 Chapter 3: Models

Chapter 3 will introduce any general circulation models (GCMs) of interest to the research. If relevant, a section on cloud-resolving models may also be included.

4.2.4 Chapter 4: Aerosol-cloud interactions

Chapter 4 will present analysis and discussion of the preliminary investigation into the correlations between aerosol and cloud properties. MODIS Level 3 data, and probably similar ATSR data, are expected to be the major contributing datasets.

4.2.5 Chapter 5: Storm-centric analysis of interactions

Chapter 5 will present results from investigating the aerosol and clouds using a storm centric approach. Fronts may also be investigated. It is anticipated that ATSR Level 2 and MODIS Level 2 datasets will be used extensively for this chapter, although data from other satellites may also be used.

4.2.6 Chapter 6: Comparison to a GCM

Depending on the findings of Chapter 5, some of the analyses may be extended to investigate output from a GCM. GCMs allow different processes, such as aerosol–cloud microphysics, to be turned on or off, thus allowing hypotheses to be tested. These results will be presented in Chapter 6.

4.2.7 Chapter 7: Conclusion

In Chapter 7, the main findings will be discussed and summarised.

4.3 Timetable

4.3.1 Summer 2009

- Submit first year report and have viva.
- Present poster at the European Space Agency Atmospheric Science Conference, Barcelona, Spain, 7$^{\text{th}}$–11$^{\text{th}}$ September 2009.
- Present talk at AOPP Annual Retreat, 1$^{\text{st}}$–2$^{\text{nd}}$ October 2009.

4.3.2 Michaelmas Term 2009

- Continue investigating aerosol–cloud correlations during visit to the Max Planck Institute for Meteorology (MPI-M), Hamburg, Germany. It is hoped that most of the research for Chapter 4 of the thesis will have been completed by the end of Michaelmas Term 2009.
- Present a talk at MPI-M.
- Submit a report to MPI-M following visit, and possibly write a paper for submission to a journal.

4.3.3 Hilary Term 2010

- Continue developing and testing a storm-centric methodology, based on previous storm tracking work done this year.
- Present Chapter 4 work, including that conducted at MPI-M, at a student seminar in AOPP during the Easter vacation 2010.

	2009		2010				2011		
	Summer	MT	HT	TT	Summer	MT	HT	TT	Summer
Chapter 1					W				
Chapter 2					W				
Chapter 3							W		
Chapter 4	R	R		W					
Chapter 5			R	R	R	W			
Chapter 6						R	R	R,W	
Chapter 7									W

Table 4.1: Timetable for research (R) and writing (W) of thesis chapters.

4.3.4 Trinity Term 2010

- Continue storm-centric analysis of satellite data, including starting to work on fronts.
- Complete draft of Chapter 4.

4.3.5 Summer 2010

- Finish storm-centric analysis of satellite data, completing main body of research for Chapter 5 of the thesis.
- Complete drafts of Chapter 1 and Chapter 2.
- Submit second year report.

4.3.6 Michaelmas Term 2010

- Begin storm-centric analysis of GCM data.
- Complete draft of Chapter 5.

4.3.7 Hilary Term 2011

- Continue storm-centric analysis of GCM data.
- Complete draft of Chapter 3.

4.3.8 Trinity Term 2011

- Finish storm-centric analysis of GCM data, completing research for Chapter 6.
- Finish draft of Chapter 6.

4.3.9 Summer 2011

- Finish draft of Chapter 7.
- Submit thesis at the end of September 2011.

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