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1. Introduction

The European Space Agency's (ESA's) Living Planet Programme includes two types of complementary user driven missions: the research-oriented Earth Explorer missions and the operational service oriented Earth Watch missions. These missions are implemented through the Earth Observation Envelope Programme (EOEP) and the Earth Watch Programme, where the Earth Explorer missions are completely covered by the EOEP.

Earth Explorer missions are divided into two classes, with Core missions being larger missions addressing complex issues of wide scientific interest, and Opportunity missions which are smaller in terms of cost to ESA and address more limited issues. Both types of missions address the research objectives set out in the Living Planet Programme document (ESA, 1998), which describes the plans for the Agency's strategy for Earth Observation in the new millennium time frame. All Earth Explorer missions are proposed, defined, evaluated and recommended by the scientific community.

The Earth Clouds, Aerosol and Radiation Explorer (EarthCARE) mission has been jointly proposed by European and Japanese scientists, following ESA's Call for Core mission ideas in 2000. EarthCARE is based upon several years of scientific exchanges and meetings between Japan and Europe, on the work previously carried on ATMOS/B1 (NASDA) and the Earth Radiation Mission (ESA) and on the joint preparation of the Report for Assessment for the User Consultation Meeting held in Granada in 2001 (ESA, 2001). EarthCARE was selected in 2000 for pre-feasibility study and, subsequently, in November 2001 for Phase A study. The Phase A studies have been completed in early 2004, forming the basis for the Reports for Mission Selection for all candidate missions (ESA, 2004). The candidate missions have been presented at the User Consultation Meeting at ESRIN, Frascati, Italy, in April 2004, and Swarm and EarthCARE were subsequently selected for implementation.

The purpose of this document is to summarize the mission objectives and scientific requirements of EarthCARE and to provide guidelines for the implementation of the mission during Phases B, C/D, and E. The document has been divided into the following chapters. Chapter 2 describes the scientific background and justification of the mission. Chapter 3 describes the mission research objectives. Chapter 4 quantifies the mission observational requirements and describes the measurement principles. Chapter 5 describes the data processing requirements and Chapter 6 summarises the mission requirements.

2. Scientific Background

The Earth radiation budget is determined by a large number of factors, including solar radiation, greenhouse gas concentration in the atmosphere and surface reflectivity. The most important atmospheric constituents are clouds and aerosols. Clouds and aerosols change significantly the planetary albedo by reflecting solar radiation back to space, but also trap outgoing long-wave radiation in a similar way as greenhouse gases. Clouds are crucially important for latent heat transport from the oceans to the atmosphere, and aerosols play a critical role in condensation and cloud formation. Cloud structures can exhibit very high variability in time and space. Vertical structures of clouds vary from (sub-visible) thin layers up to massive Cumulonimbus towers that could reach throughout the troposphere and penetrate the tropical tropopause. Multiple scattering of radiation within the three-dimensional structure of clouds makes radiative transfer modelling highly complex. The large variability of clouds, and to a lesser extent of aerosols, combined with their dominant role in atmospheric radiative processes, makes them both critically important for atmospheric modelling, and which are at present poorly quantified on a global scale.

Approximations in the representation of cloud formation, aerosol-cloud interaction and cloud radiative properties are the largest source of uncertainty (IPCC, 2001) in today’s climate models. Global Circulation Models (GCMs) with grid sizes in the order of 100km (horizontal) by 1km (vertical) cannot possibly reproduce realistic cloud structures, in particular not down to the scale of convective processes. Therefore, GCMs depend on parameterisations for cloud prediction, which introduces significant model uncertainties.

It is therefore of high scientific interest to model clouds and their radiative impacts properly. This requires comprehensive and reliable sets of measurements – of cloud and aerosol parameters together with the corresponding outgoing radiation at the top of the atmosphere – and improvements in cloud modelling in atmospheric models.

2.1 Radiation and the Need for Reliable Models

Whether or not recent global warming is entirely attributable to human activities, there is no reasonable doubt that continuation and acceleration of the already strong anthropogenic alteration of the atmosphere’s composition must lead to stronger climate change than was observed in the 20th century (IPCC, 2001). Such change will necessarily involve changes in the distributions of precipitation and runoff, and will modify distributions of events such as violent winter storms, tropical cyclones, extended heat and drought, and flood-producing extreme rainfall.

(a) Radiation

Radiation is the only relevant process by which the Earth exchanges energy with space. Radiative effects play a governing role in climate, constituting both the source (solar radiation) and the ultimate sink (thermal infrared radiation to space) of energy in the Earth-atmosphere system. Both *aerosols* and *clouds* are major actors in the climate-radiation connection.

(b) Radiative Forcing

The Third Assessment Report (TAR) of the Intergovernmental Panel on Climate Change defines radiative forcing as "an externally imposed perturbation in the radiative energy budget of the Earth's climate system" (IPCC 2001). Figure 2.1 shows that the largest component of the forcing, from anthropogenic increases in greenhouse gases, is the best understood and quantified. As one moves from left to right in this figure, the level of scientific understanding decreases, but some of the magnitudes remain large. Using a new approach, Bellouin et al. (2005) have shown that the direct forcing due to aerosols for the year 2002 could be as high as –1.9 Wm^{-2}. The error estimation in the analysis is for the first time less than 20% (0.3 $W.m^{-2}$) using a Monte Carlo approach. The corresponding change in the radiative forcing can be computed with reasonable confidence, e.g. for the year 2000 compared to 1750 (Figure 2.1).

Estimates of the *indirect effect* of aerosol on clouds are all negative and very uncertain, although some estimates are as large in magnitude as the forcing from the greenhouse gases. Coupled with the fact that the most uncertain component of the climate feedback comes from clouds, this result reinforces the conclusion that aerosols and clouds provide the largest source of uncertainty in both the radiative forcing and in the climate response. The EarthCARE mission is motivated in part by the need to reduce these uncertainties and thereby to provide a more secure foundation for predictions of future climate change.

Since the TAR, a great deal of work has been done to improve the estimates of the direct effect (e.g. Takemura et al. 2002) and indirect effects, but the range of estimates of the latter remain high (Nakajima et al. 2001, Nozawa et al. 2001, Sekiguchi et al. 2003, Takemura et al. 2005). A recent review of estimates of the indirect effect has been provided by Lohmann and Feichter (2005). A full assessment of this material will be provided in the IPCC Fourth Assessment Report, which is currently in preparation.

(c) Climate Sensitivity and Feedbacks

For a given emissions scenario (translated into time-dependent radiative forcing), different model simulations exhibit a wide range of global warming. A major factor in these differences is certainly the model's 'water vapour feedback', whose strength depends to a large extent on the parameterisation of the condensation/freezing processes in clouds. These also strongly influence the 'cloud-radiation feedback', for which even the sign is unknown. It is well known that different climate models, providing reasonable simulations of the present climate, give widely different changes (both in magnitude and sign) in their shortwave, longwave, and net cloud radiative forcing for a given scenario. Most recent results reported by Bony et al. (2006) show that the cloud feedback sensitivity obtained from GCMs, although smaller than for water vapour, is about 0.69 +/- 0.38 W m^{-2} K^{-1}on a global average, as related to a change in SST. It is comparable to previous estimates but very different from one model to another. This spread is at the source of the uncertainty in climate projections, but parametrisations may also lead to some uncertainties. As part of this analysis, it has been further shown that the feedback sensitivity related to low clouds could be a important issue. Indeed, analysis from ERBE satellite observations has shown that models strongly underestimate the net cloud feedbacks related to low clouds, which could be as high as 6 W m^{-2} K^{-1} (Bony and Dufresne, 2005).

The potential importance of aerosols and clouds in climate change has been highlighted by recent research on trends in downward solar radiation at the surface. Observations from the 1950s to around 1990 show a downward trend, often referred to as “global dimming”, which has been attributed to increases in the concentrations of aerosols from anthropogenic sources. Since 1990, both direct observations of solar radiation at the surface and retrievals from satellite data show that this trend has been reversed, implying reduced aerosol concentrations, possibly due to the effect of air quality regulations and the decline in the economies of the members states of the former USSR (Wild et al. 2005, Pinker et al. 2005). Recent analyses (Bellouin et al., 2005; Andreae et al, 2005) emphasize that the observed direct aerosol radiative forcing would be larger than estimated from the models. As a result of the continuing reduction in aerosol load, the observed temperature increase in the forthcoming decades would be larger than predicted.

At the same time, analysis of recent CERES observations suggests the existence of a downward trend in planetary albedo, probably related to a slight downward trend in cloud cover. If such a trend continues it is an aggravating factor in global warming and needs to be understood in terms of cloud and aerosol processes (Wielicki et al. 2005). Further observations are needed to provide much stricter constraints on the cloud process parameterisations that to a large extent determine the strength of the water vapour and cloud-radiation feedbacks.

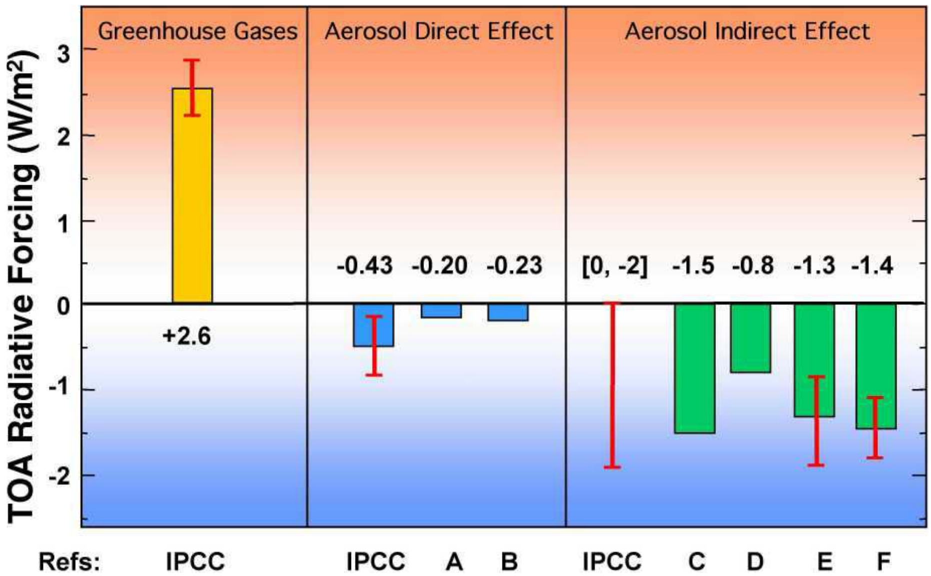

Figure 2.1: *Global annual mean radiative forcing at the top of atmosphere corresponding to changes from 1750 to 2000 following IPCC (2001) together with some more recent assessments of aerosol direct and indirect effects (modelling and satellite studies). The IPCC values show smaller direct forcings than recently derived by* Bellouin et al. *(up to -1.9 Wm^{-2}, see text). Note that although direct radiative forcing is fairly well known, indirect forcing of aerosols is extremely uncertain. (A: Takemura et al., 2002, B: Kaufman et al., 2003, C: Takemura, 2003a, D: Takemura, 2003b, E: Nakajima et al., 2001, F: Sekiguchi et al., 2003) Estimates of the SW, LWS and NET cloud feedbacks are much larger than the other values in the graph..*

(d) Model Needs

Although part of the huge range of global warming projections (viz. from 1.4 to 5.8 K for 2100) arises from the different emission scenarios, a significant range of climate model uncertainty remains. For an individual scenario, in addition to warming ranging from 2 to 4 K, different simulations give inconsistent projections of the changes in precipitation and runoff in important regions (e.g. eastern North America). This may depend in part on parameterisations of surface water processes, but certainly depends also on precipitation and so on clouds. The shortcomings in the treatment of cloud and aerosol processes in climate models arise from lack of observations to validate cloud and aerosol parameterisation schemes. The same difficulties bedevil numerical weather prediction models used for short and medium range and seasonal forecasting.

2.2 Numerical Weather Prediction and Climate Models

All such models divide the atmosphere into a series of grid boxes, typically around 50 km in the horizontal and 500 m in the vertical. By the time of launch of EarthCARE, global numerical models for weather forecasting are expected to have horizontal resolutions of 10 to 20 km and at least 90 levels in the vertical. The Earth Simulator in JAMSTEC is already able to run a non-hydrostatic global cloud system resolving climate model with a resolution of only 3.5km (Figure 2.2, Tomita et al., 2005). For each box, clouds are represented by prognostic variables such as fractional cloud cover, ice and liquid water content. The radiative transfer in atmospheric models also requires assumptions on cloud overlap for each vertical stack of grid boxes and particle sizes. The overlap assumptions affect both the radiative transfer and the precipitation efficiency of the clouds.

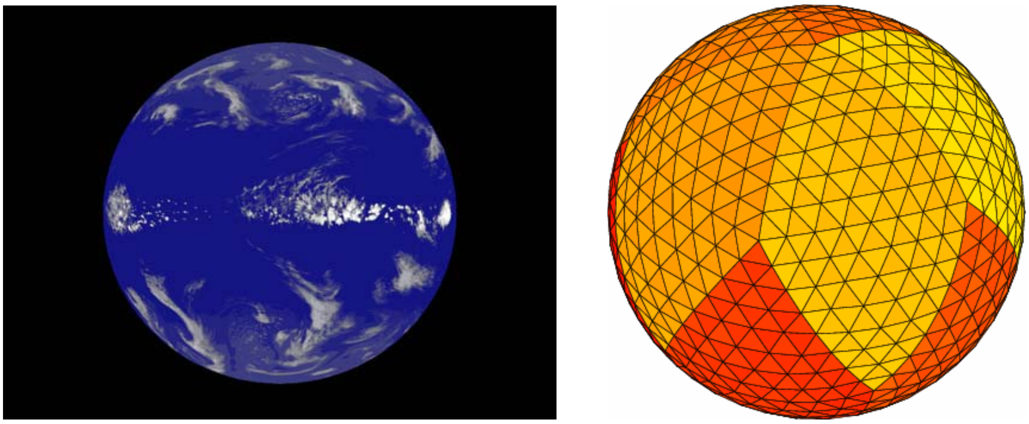

Figure 2.2: *Left: Cloud image (outgoing longwave radiation) produced with the 3.5km-mesh global cloud system resolving model "NICAM" under an aqua planet condition. Right: Structure of the icosahedral grid adopted in the global cloud-system resolving model "NICAM".*

A first step is to evaluate whether the *current* weather and climate is being correctly represented in numerical models to give us more confidence in the climate predictions. Recently Potter and Cess (2004) compared cloud-radiative forcing (CRF) predicted by a group of 19 atmospheric general circulation models. It was found that CRF at the top-of-the atmosphere predicted by many GCMs shows a significant negative bias where compared to ERBE measurements. This means that GCMs significantly overestimate cloud radiative cooling. Out of the group of 19 model analyses, those that did predict accurate top of the atmosphere (TOA) fluxes were found to agree with observations only due to compensating errors in either cloud vertical structure, cloud optical depth or cloud fraction. This clearly highlights the need for better measurements of cloud property vertical profiles on a global basis. Using ISCCP (International Satellite Cloud Climatology Project) cloud data in the AMIP (Atmospheric Model Intercomparison Project) comparisons, it has been shown that 'realistic' models correctly reproducing radiation fluxes at the top of the atmosphere strongly disagree in their assessments of cloud water (Figure 2.3). Clearly, stronger empirical constraints are needed to narrow this range, in particular measurements of total cloud water over land surfaces, where passive microwave radiometry is inadequate. Vertical cloud water profiles are needed to evaluate the model parameterisations of in-cloud processes and cloud-aerosol-radiation interactions. EarthCARE data products can be compared with GCMs and climate models to identify errors and biases in these models. This should lead to improved parameterisation schemes in the models and consequent improvements in weather forecasts, and more reliable predictions of temperature, precipitation and extreme events in the future climate.

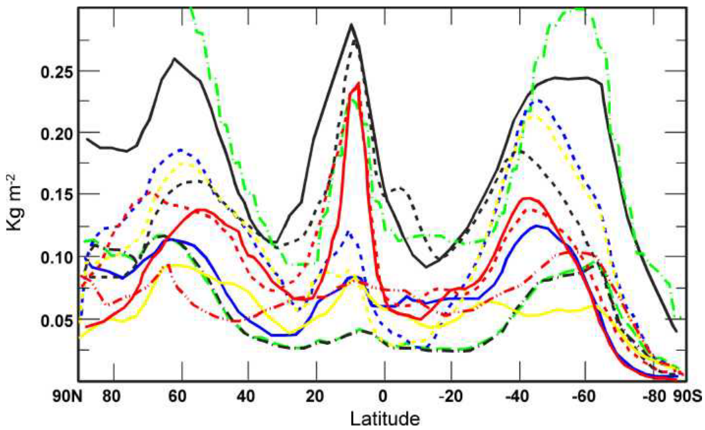

Figure 2.3: *Zonally and vertically integrated cloud water for 14 different AMIP2 climatologies for northern summer.*

2.3 Aerosols

Changes in aerosols directly modify solar radiation reaching the ground, and also affect microphysical, biochemical and photochemical processes in the atmosphere including processes affecting cloud properties. In particular, increases in aerosols resulting from human activities have 'indirect' radiative effects by (i) increasing cloud albedo by decreasing droplet size, and (ii) changing cloud lifetime. Indirect aerosol effects could be very important and must be taken into account in global and regional climate change forecasts. An example of the direct and indirect impact of anthropogenic aerosols (biomass burning and ship tracks) can be seen in the MERIS satellite image shown in Figure 2.4. Unlike greenhouse gases, aerosols have short atmospheric residence times, and thus are non-uniformly distributed in space and time, which complicates efforts to account for their effects.

Moreover, the presence of thin aerosol layers, difficult to detect by passive measurements, can induce strong errors in underlying cloud property retrievals. Anthropogenic aerosol and aerosol precursor emissions are likely to vary in time and space over coming decades, and it is essential to develop tools for evaluating to what extent passive measurements can monitor the associated forcing. New active measurements of aerosol profiles are essential, e.g. for evaluation of retrievals from passive measurements.

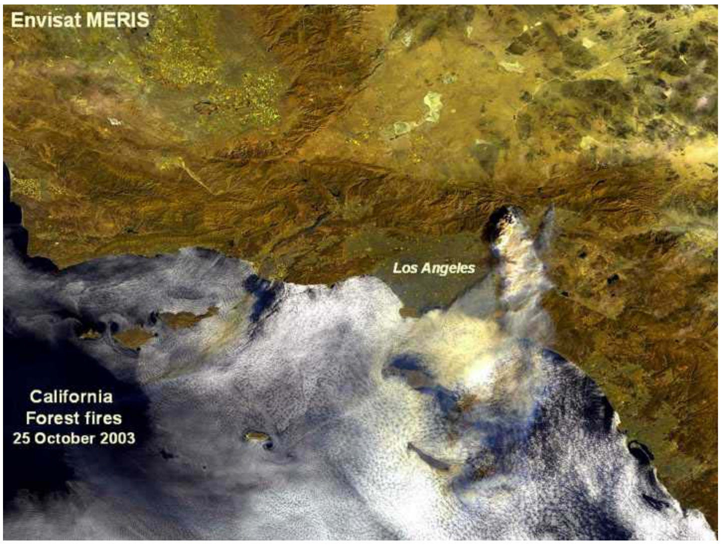

Figure 2.4: *Wild fires in California (25 October, 2003, ESA, MERIS). The aerosols are transported SSW and interact with low level cloud off the Californian coast. Aerosols above the stratocumulus cloud layer reduce their reflectivity (positive forcing), while aerosols within the cloud lead to brighter clouds (negative forcing). The same is true for the ship tracks visible in the left part of the image.*

2.4 Clouds

In general, low-level clouds cool the Earth by reflecting more sunlight back to space, but high level cold clouds tend to warm the Earth by losing less infrared radiation to space. The concept is explained schematically in Figure 2.5. Low altitude polar clouds are also important and currently not well observed.

Hence, if cloud properties change in response to any future climate change, then the 'cloud radiative feedback' can either amplify the original direct radiative forcing or partially counteract it. Changes in the vertical profile of clouds lead to different heating rates and consequent important changes in the atmospheric dynamics that then feed back to changes in the cloud profiles. Currently, satellite instrument measurements constrain the total incoming and outgoing radiation at the top of the atmosphere, but cannot provide sufficiently accurate determinations of cloud profiles and consequent energy heating profiles.

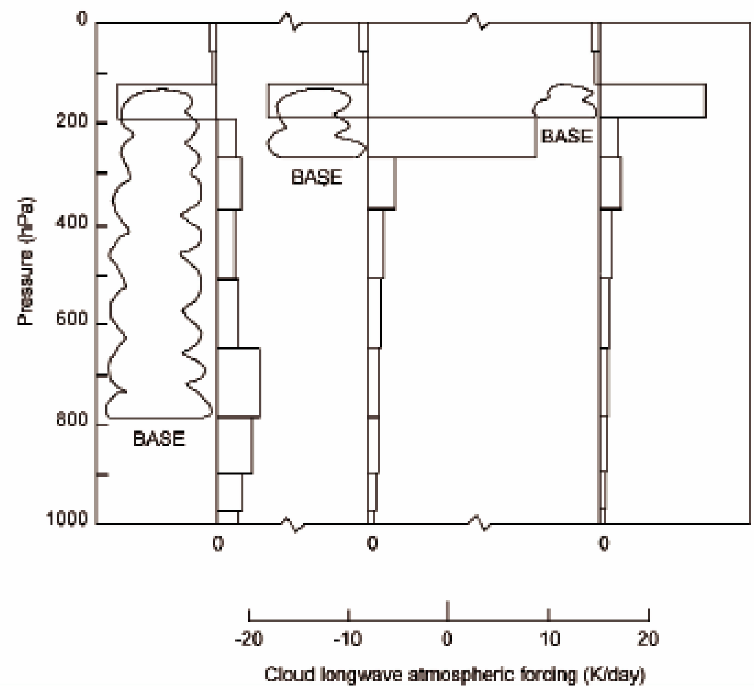

Figure 2.5: *Infrared radiative heating/cooling profiles, calculated for three different cloud base levels (after Slingo and Slingo, 1988). These profiles demonstrate the need for an accurate knowledge of upper and, in particular, lower cloud boundaries.*

Figure 2.6 displays vertical profiles of observed radar reflectivity, Doppler velocity and lidar backscatter and the derived values of ice particle effective radius (r_e) and ice water content (IWC) obtained from a nadir pointing radar and lidar on board an aircraft over-flying an extensive layer of cirrus during the EarthCARE Algorithm Validation (ECAV) campaign in Japan in 2003. The same retrieval algorithms will be used with the lidar and radar embarked on the EarthCARE satellite to provide global profiles of these parameters. The IWC values will provide data to evaluate the performance of the models shown in Figure 2.3. Global observations of IWC can help evaluate the current parameterisation of r_e as a function of temperature and condensate, and suggest improved parameterisation schemes. Ice sedimentation velocities are crucial in NWP models in fixing ice cloud cover and lifetime; global observations of these velocities using Doppler radar will improve this crucial aspect.

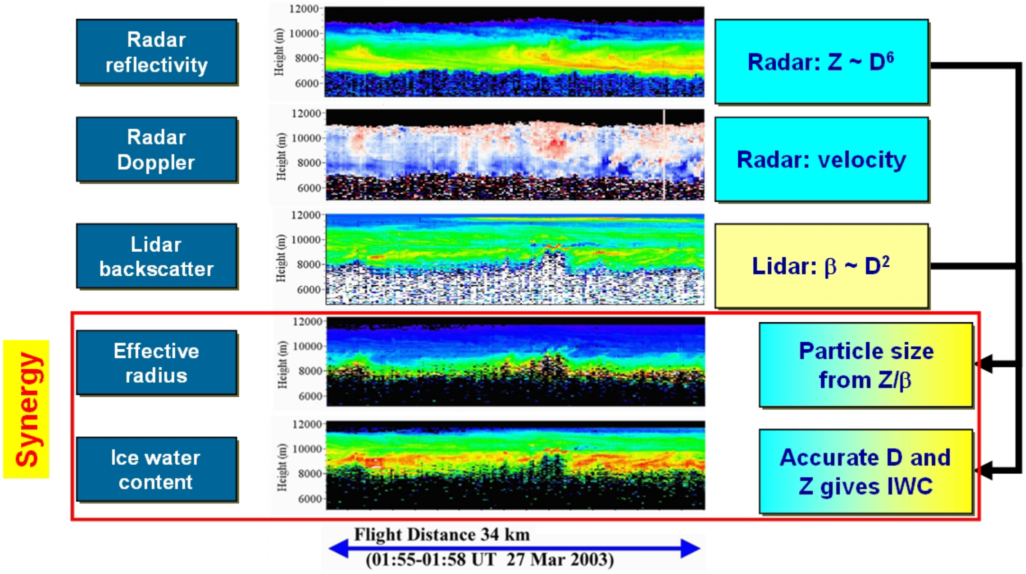

Figure 2.6: *Vertical profiles of observed radar reflectivity, Doppler velocity and lidar backscatter and derived values of ice particle effective radius (r_e) and ice water content (IWC) obtained from a nadir pointing radar and lidar on board an aircraft. (APEX/CAV Campaign, 2003, Japan.)*

2.5 Convection and Precipitation Processes

The majority of the Earth's rainfall results from convection – and especially so for the heavier precipitation such as displayed in Figure 2.7, which results in flash floods and loss of life. Yet convective precipitation is extraordinarily hard to represent within numerical models because it occurs on a scale which is much smaller than the grid box used in virtually all models, and is essentially a statistical and noisy phenomenon. The current almost universal approach is based on a mass flux scheme whereby the stability of the vertical profile for each stack of grid boxes is examined at each time step, and if it is unstable then the scheme removes a fraction of the instability through a vertical mass flux. If this flux produces supersaturation and cloud, then precipitation may result. The fraction of the box involved and hence the updraught and fractional cloud cover is not explicitly derived. The performance of such schemes is not always satisfactory, with such basic processes as the diurnal cycle of convection poorly represented. New more physical schemes being developed favour a statistical approach in which, rather than a deterministic mass flux, the convection process itself is treated as a probability distribution function (PDF). Convection is indeed a statistical phenomenon and these PDF approaches capture this aspect, but different schemes lead to different PDF's of mass flux and vertical velocities, both in magnitude, cross-sectional area and vertical profiles, and sensitivities to initialisation schemes. Global observations of mass flux and vertical velocity profiles, which can be derived from the observations in Figure 2.7, are needed to evaluate these schemes.

Apart from the occasional case study, there are currently no direct observations of velocities and fluxes.

The convection in Figure 2.7 overshoots the tropopause. Global observations are needed to quantify the amount of moisture introduced into the stratosphere by this process.

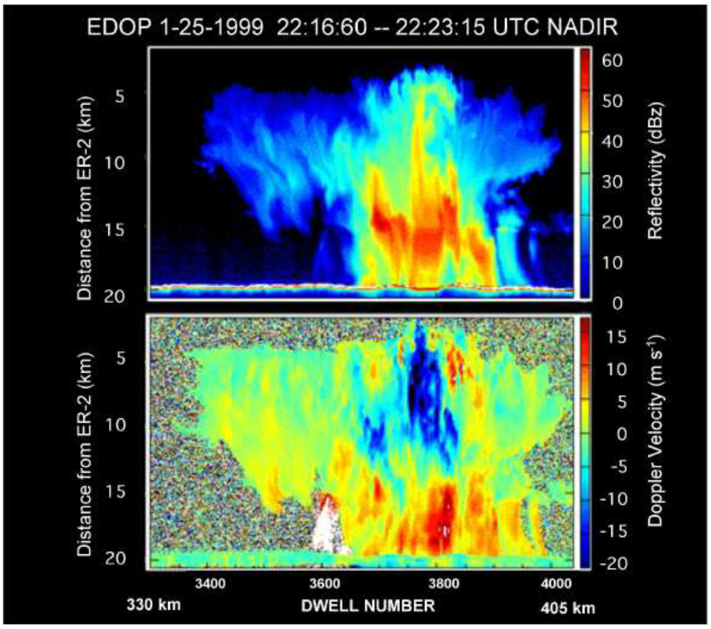

Figure 2.7: *Vertical cross section of radar reflectivity and vertical Doppler velocities observed from an X-band radar on the ER-2 aircraft overflying a severe tropical convective storm in Brazil (courtesy G. Heymsfield).*

2.6 Summary

Difficulties in representing aerosols, clouds, and convection in numerical models of the atmosphere seriously limit the ability to provide accurate weather forecasts on all timescales, and reliable predictions of future climate. These factors govern the radiation balance and hence the temperature of the Earth and are directly responsible for the production of precipitation and thus control the hydrological cycle. In summary:

(a) Aerosols
Aerosols have a direct radiative impact by reflecting solar radiation back to space, which leads to cooling. Absorbing aerosols, e.g. carbon from anthropogenic sources, can lead to local heating (direct effect) reducing lapse rate. Aerosols also control the radiative properties of clouds and their ability to produce precipitation (indirect

effect). The low concentration of aerosol particles in marine air leads to water clouds with a small number of relatively large droplets. In contrast, the high concentration of aerosols in continental and polluted air results in water clouds with a much higher concentration of smaller droplets. Continental clouds therefore not only have a higher albedo and reflect more sunlight back to space, but also are much more stable and long-lived and less likely to produce precipitation. Aerosols also control the glaciation process, yet their effect on the properties of ice clouds is essentially unknown. There is a need to quantify the degree to which aerosols are responsible for the observed rapid reduction in the albedo of freshly fallen snow. Present observations of global aerosol properties are limited to optical depth and a crude estimate of particle size. This is very unsatisfactory, since we need to know their chemical composition, whether they scatter or absorb, and their vertical and geographical distribution.

(b) Clouds

Clouds are the principal modulators of the Earth's radiation balance. Currently, there are global estimates of cloud cover, but little information on their vertical extent or the condensed mass of ice or liquid cloud water. Low clouds cool by reflecting short wave solar radiation back to space, whereas high clouds warm because they are cold and emit less infrared radiation to space. In the present climate, these two effects are large and have opposite signs. Any changes in the vertical distribution of clouds in a future warmer climate could lead to large changes in the net radiative forcing. Climate models disagree as to whether this attenuates or amplifies the effect of the original direct greenhouse gas warming. Uncertainties in the vertical profiles of clouds are largely responsible for the current unacceptable spread in predictions of future global warming, and also limit the accuracy of numerical weather prediction. Models are able to produce the present observed top-of-the-atmosphere radiation but with very different vertical profiles of clouds and water content. To evaluate models, observations of cloud profiles are urgently required, so that the ability of models to provide reliable weather forecasts and predictions of future global warming can be improved.

(c) Convection and Precipitation Processes

Clouds are the source of all precipitation, but details of this process are poorly understood. At present, models convert cloud condensate in a grid box to precipitation either by slow widespread ascent, or, for the more intense rainfall, by convection. Major difficulties include the large spread in model predictions of cloud condensate, the efficiency with which this is converted to precipitation, and the representation of sub-grid-scale convective motions. The extent to which vigorous tropical convection introduces moisture into the stratosphere is uncertain. It is also known that the modelled diurnal cycle of convection is incorrect. Understanding these processes is crucial for quantitative precipitation forecasting. Current models have major failings in their representation of organized tropical convection even on the largest scales (eg. the MJO); and these deficiencies reflect in systematic errors in the models basic climate simulations. The societal benefits in providing warning of flash flooding would be immense. Global observations of the probability distribution function of vertical motions within the grid box to constrain convective parameterisation schemes are needed.

3. Mission Objectives

The EarthCARE mission has been specifically defined with the basic objective of improving the understanding of cloud-aerosol-radiation interactions so as to include them correctly and reliably in climate and numerical weather prediction models. Specifically, the scientific objectives are (not in order of priority):

- The observation of the vertical profiles of natural and anthropogenic aerosols on a global scale, their radiative properties and interaction with clouds.
- The observation of the vertical distributions of atmospheric liquid water and ice on a global scale, their transport by clouds and their radiative impact.
- The observation of cloud distribution ('cloud overlap'), cloud-precipitation interactions and the characteristics of vertical motions within clouds, and
- The retrieval of profiles of atmospheric radiative heating and cooling through the combination of the retrieved aerosol and cloud properties.

The key parameters determining the radiative properties of clouds and aerosols are:

- The extinction and absorption properties of aerosols.
- Large-scale cloud structure, including cloud fraction and overlap.
- Cloud condensate content, particle size, shape and small scale cloud structure.

Note that macroscopic and microscopic cloud parameters depend in part on physical and chemical properties of aerosols acting as cloud condensation and/or freezing nuclei.

EarthCARE will meet these objectives by measuring simultaneously the vertical structure and horizontal distribution of cloud and aerosol fields together with outgoing radiation over all climate zones. More specifically, EarthCARE will measure:

1. Properties of aerosol layers:
   - (a) The occurrence of aerosol layers, their profile of extinction coefficient and boundary layer height, and
   - (b) The presence of absorbing and non-absorbing aerosols from anthropogenic or natural sources.
2. Properties of cloud fields:
   - (a) Cloud boundaries (top and base height) including multi-layer clouds.
   - (b) Height resolved fractional cloud cover and cloud overlap.

   (c) The occurrence of ice and liquid and of super-cooled cloud layers.
   (d) Vertical profiles of ice water content and effective ice particle size and shape.
   (e) Vertical profiles of liquid water content and effective droplet size.
   (f) Small scale (1 km or less) fluctuations in these cloud properties.

3. Vertical velocities to characterise cloud convective motions and ice sedimentation.
4. Drizzle rain rates and estimates of heavier rainfall rates.
5. Narrow-band and broad-band reflected solar and emitted thermal radiances at the top of the atmosphere.

4. Observational Requirements

4.1 Introduction

The EarthCARE mission focus on aerosols and clouds addresses the two largest sources of uncertainty in current climate predictions. Aerosols provide the largest source of error in representing the direct *radiative forcing* of the climate system, while clouds provide the largest uncertainty in representing the *radiative feedbacks* that can either enhance or reduce the sensitivity of climate to that forcing (IPCC 2001). A component of the cloud feedback is also believed to arise from the *indirect effect* of aerosols on cloud radiative properties, so the two uncertainties are inextricably linked. The need for reliable information on the vertical structure of clouds and aerosol layers demands a unified observational approach that is uniquely provided by EarthCARE.

The basic structure and the variables represented in Numerical Weather Prediction (NWP) models are very similar to those in climate models, so EarthCARE data will also be utilised in this important, operational activity. The data can be used to initialise the model forecasts and also in the off-line evaluation of the representation of clouds. Aerosols are just beginning to be represented explicitly in NWP models and by the time EarthCARE is operational their inclusion in order to model atmospheric visibility and air quality is expected to be routine. Other expected applications for EarthCARE data include process studies in conjunction with in-situ airborne measurements and/or surface-based remote sensing, and the provision of vertically resolved cloud and aerosol information for adding value to the data from other satellite missions.

The specific observational requirements may be derived with reference to the cloud and aerosol parameters used in climate and NWP models and the horizontal and vertical resolutions with which these parameters are represented. The accuracy requirements are derived from estimates of the impact of changes in these geophysical parameters on the magnitudes not only of the local radiative heating, but also of the radiative fluxes at the top of the atmosphere. The latter provides a particularly useful constraint, as it may be quantified both by the accuracy with which broad-band radiances need to be modelled and by the accuracy with which they can be measured by a broad band radiometer. In the following, we therefore firstly identify the required parameters and resolution and then proceed to estimate the required accuracy for each parameter.

4.2 Geophysical Parameters and Resolution

Climate and NWP models represent clouds in order to calculate the distribution of precipitation and the vertical profiles of solar and thermal radiative heating. Since the models cannot resolve individual clouds explicitly, the effects of such unresolved processes on the grid-scale are *parameterised*, using statistical relationships, which are derived from theory or observations. Similar considerations apply to aerosols.

For clouds, the variables that are represented explicitly in each grid box and at each vertical level are the fractional cloud cover, the liquid or ice water content and the size of water drops or ice crystals. The choice of ice water content terminal velocity has a profound effect on ice cloud lifetime. The parameterisations either assume or calculate the sub-grid-scale probability distribution function or PDF of water substance and of cloud-scale vertical velocities and mass fluxes within convective clouds. Assumptions are then made about the vertical overlap between cloud layers in order to calculate the precipitation and radiative heating rates. Passive remote sensing provides some information on the geographical distribution of these variables as inferred from space, but very limited information on the vertical structure. The simulated radiative fluxes at the top of the atmosphere can usually be adjusted to agree with satellite observations, but different models achieve this result through radically different vertical distributions of cloud. Vertically resolved observations of clouds are urgently needed, to constrain the models and to test the ability of the parameterisations to represent the profiles of all of the cloudy variables mentioned above.

Similar considerations apply to aerosols, which are included in models both to represent their *direct effect* via radiative fluxes and surface forcing, and their *indirect effect* on cloud particle size, precipitation efficiency and radiative properties. The aerosol mass and size must be represented for several different aerosol types (e.g. sea salt, sulphate, dust, volcanic and carbonaceous aerosols from both natural and anthropogenic sources). Passive observations provide estimates of the aerosol optical thickness but, as with clouds, vertical information is required to distinguish different layers. A great deal of aerosol remains in the planetary boundary layer and so provides a tracer of the boundary layer depth. However, aerosols can also be lofted by convection above the scavenging effects of clouds and form persistent layers, the radiative effects of which depend strongly on height, which again emphasizes the need for observations of the vertical structure and estimates of convective motions. Estimates of aerosol absorption are needed because they affect cloud lifetime and would provide valuable additional information on the aerosol type. High resolution in the horizontal would enable the direct effects to be distinguished from those of clouds and the indirect effects on clouds to be investigated.

Typical values of the horizontal and vertical resolutions of current climate and NWP models are shown in Table 4.1 together with values expected for the year 2010. Regional models with higher resolution are already used to provide local detail, or to resolve particular physical processes such as convection or severe storms. They will have resolutions close to 1 km by 2010. The use of such models provides two important links to EarthCARE. Firstly, by resolving explicitly more of the dynamical structures within which clouds form, the emphasis on model improvement moves to the representation of the small-scale physical processes within clouds and aerosol layers, which are the primary focus of the mission. Secondly, the regional model resolutions are approaching the basic footprint sizes of the elements of the EarthCARE mission, enabling direct model-to-satellite comparisons, which have hitherto been hindered by inadequate global model resolutions.

Table 4.1: *Typical resolutions of climate, NWP and regional models.*

Parameter	Climate Models		NWP Models		Regional	
	Now	2010	Now	2010	Now	2010
Horizontal resolution (km)	250	100	50	25	10	2
Number of vertical levels	40	80	50	100	50	100
Vertical resolution (upper troposphere)	1.5km	750m	1km	500m	1km	500m
Vertical resolution (boundary layer)	200m	100m	100m	50m	100m	50m

The values in Table 4.1 give the typical resolutions required for comparison of the EarthCARE products with models. However, the basic observational resolution needs to be as high as possible, not only to minimise non-linear effects on retrievals, but also so as to build up statistics on the sub-grid-scale distributions of cloud and aerosol quantities in 3 dimensions (e.g. the PDF of ice water content and cloud-scale vertical velocities).

Figure 4.1 summarises the scope of the EarthCARE mission.

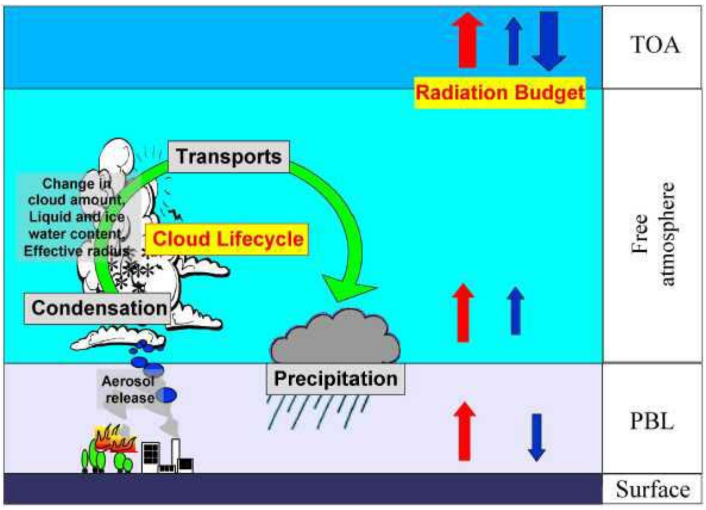

Figure 4.1: *The scope of the EarthCARE mission. The objective is to retrieve vertical profiles of cloud and aerosol, and characteristics of the radiative and micro-physical properties so as to determine flux gradients within the atmosphere and fluxes at the Earth's surface, as well as to measure*

directly the fluxes at the top of the atmosphere and also to clarify the processes involved in aerosol-cloud and cloud-precipitation-convection interactions.

4.3 Accuracy Requirements

The measurement requirements for EarthCARE are primarily based upon the need to measure height-resolved cloud and aerosol properties relevant to atmospheric radiative transfer. The accuracy placed upon these measurements is that which is consistent with a Top-of-Atmosphere (TOA) combined short and long-wave flux accuracy of about 10 Wm^{-2} for an instantaneous view with a footprint of 10 km^2. In other words, the accuracy of the measurements must be such that the atmospheric vertical flux profile (for a "snapshot" view on the scale of 10km x10km) may be reconstructed subject to a TOA accuracy of 10Wm^{-2}.

Specifying concise accuracy requirements for EarthCARE parameters is complicated by the fact that the impact of a given cloud or aerosol optical parameter at a given position in the atmospheric column (i.e. optical extinction) has on the radiative flux profile depends not only on the value of that parameter but also on the 3-D distribution of the fields of other relevant fields (as well as external factors such as solar elevation angle or surface albedo). For example, for an isolated homogeneous single cloud layer, the SW-TOA flux may change from 80 Wm^{-2} to 95 Wm^{-2} (above 10 Wm^{-2}) going from cloud optical depth 0.1 to 0.5, while going from optical depth 40 to 60 may change the TOA SW fluxes from 195 to 200 Wm^{-2}. In a similar fashion, in otherwise clear skies, aerosols may have an impact of 10 Wm^{-2} on the TOA fluxes. Furthermore, if varying aerosol amounts under a cloud layer will generally have a negligible effect on the TOA fluxes even under high loading conditions, their impact may be of importance above high albedo surfaces such as clouds, as the sign of the TOA forcing may be changed (Kiel et al., 2003).

Keeping the points made above in mind together with the need to be concise and traceable, it is necessary to specify two types of requirements. First, for each geophysical parameter, where appropriate, a minimum absolute "Detectability Threshold" is specified. The value of this threshold is consistent with producing a 10 Wm^{-2} change in the TOA compared with clear sky conditions. Secondly, a general "Accuracy" is specified which is appropriate for typical values of the cloud properties, which can be at least an order of magnitude larger than the threshold value.

The requirement for cloud related measurements are listed in Table 4.2. The numbers are based on part on the work of Slingo (1990) for liquid water clouds, Kristjánsson et al. (2000) for ice clouds, and calculations performed in earlier EarthCARE studies (i.e. Illingworth et al., 2000, Tinel et al. 2003, Donovan et al, 2004). The exception to this is the requirement for in-cloud vertical velocity (which is not directly driven by the 10 Wm^{-2} TOA requirement).

Table 4.2: *Accuracy requirements for cloud properties for a reference sample horizontally integrated over 10km.*

Property	Detectability Threshold[1]	Accuracy
Ice cloud top/base[1)]	N/A	300 m
Ice cloud extinction coefficient	0.05 km^{-1}	15%
Ice water content (IWC)	0.001gm^{-3}	± 30%
Ice crystal effective size	N/A	± 30%
Water cloud top/base	N/A	300m
Water cloud extinction coefficient	0.05 km^{-1}	15%
Liquid water content (LWC)	0.1gm^{-3}	±15-20%
Water droplet effective radius	N/A	± 1-2μm
Fractional cloud cover	5%	5%
Vertical velocity within clouds	N/A	± 0.2 to 1ms^{-1}

Profiles are required of all the quantities in Table 4.2 with a vertical resolution of 300m or finer. These accuracy requirements that are expressed as percentages are for typical values of the cloud properties, which can be at least an order of magnitude larger than the threshold value. The requirements for profile information are first expressed in terms of the radiative extinction coefficient, followed by the corresponding values for the geophysical properties (water content and particle size), separately for ice and water clouds.

The cloud products are required on a horizontal grid of 10km, but the individual measurements and retrievals should be made with a sampling of 1km or better, in order to capture adequately the internal spatial variability of clouds.

The overlap of cloud layers in the vertical is an important parameter, which is usually assumed in model parameterisations. The information on the cloud profiles in Table 4.2 will enable the overlap to be determined.

There is a need to infer vertical motion within clouds, in particular for characterising convection, estimating sedimentation velocity of ice particles in cirrus, quantifying drizzle fluxes in stratocumulus and estimates of heavier rain rates. The mission will therefore be equipped with a Doppler radar capability. The accuracy requirements are discussed in section 4.7.2.

The aerosol requirements, summarised in Table 4.3, have been calculated in earlier EarthCARE studies and may also be inferred from Cusack et al. (1998). The basic parameter required is the aerosol optical thickness profile. This needs to be converted into aerosol mass and size through additional information, for example on the aerosol extinction coefficient. Information on the aerosol type, even if only an indication of absorbing versus non-absorbing aerosol, would be extremely valuable. A great deal of

[1] Detectability is defined as the value measurable with not more than 100% RMS error.

aerosol resides in the planetary boundary layer, so a well resolved measure of the aerosol height would also provide information on the depth of the boundary layer.

Table 4.3: *Accuracy requirements for aerosol properties for a reference sample horizontally integrated over 10km.*

Property	Detectability threshold[2]	Accuracy
Boundary layer optical depth	0.05	10-15%
Top/base and profile	N/A	500m

This accuracy requirement is expressed as percentages for typical values of the aerosol properties, which can be at least an order of magnitude larger than the threshold value.

Other quantities include the solar and thermal broad-band radiances at the top of the atmosphere. The accuracy requirement given is that for an instantaneous measurement, consistent with an accuracy of 10 Wm^{-2} in the broad-band fluxes at the top of the atmosphere. Note that averaging in space and time provides fluxes with increased accuracy; for example, the values quoted for regional, monthly means are 5 Wm^{-2} from ERBE (Ramanathan et al. 1989) and 2 Wm^{-2} from CERES (Wielicki et al. 1996). However, the annual cycle of the global mean TOA net radiation has a reasonably well-determined peak-to-peak amplitude of approximately 15 Wm^{-2}, with maximum positive values occurring in late Southern Hemisphere summer. The nonzero values (a few watts per square meter) of the annual average global mean TOA net radiation found from ERBE and ScaRaB data are an indication of observational uncertainty rather than of global radiative imbalance (Kandel et al., 1998, Kandel and Viollier, 2005). The radiance requirement is based on the need to provide closure, by using the EarthCARE cloud and aerosol properties, together with additional information, to perform forward simulations of the observed radiances, as a final test of the retrieved properties. The specific value of 3 $Wm^{-2}sr^{-1}$ multiply by π is thus consistent with the general mission accuracy requirement of 10 Wm^{-2}. The accuracy requirements for the additional information needed to perform these calculations (surface temperatures and atmospheric temperatures and humidities) are not challenging and these quantities can be obtained from other sources, such as global NWP analyses.

The RMS error of the ECMWF operational temperature analysis is estimated to be smaller than 1.5 K (Simmons, 2003). Evaluation of the accuracy of the specific humidity analysis is more difficult, but an assessment as part of a new humidity analysis indicates that the RMS error is smaller than 10% (Holm, ECMWF personal communication).

[2] Detectability is defined as the value measurable with not more than 100% RMS error.

Table 4.4: *Accuracy requirements for other quantities.*

Quantity	Detectability threshold	Accuracy
TOA solar and thermal radiances	N/A	3 $Wm^{-2}sr^{-1}$
Surface and atmospheric temperatures	N/A	1.5K
Atmospheric humidity	N/A	10%

4.4 Space/Time Sampling Requirements

The EarthCARE concept is motivated by the same scientific requirement underpinning the US Atmospheric Radiation Measurement (ARM) programme: the need for measurements of all the radiatively important constituents of the atmosphere. This will enable forward simulations to be made of the observed radiances at the top of the atmosphere and of the profiles of solar and thermal radiative heating rates in the atmosphere. Although there are fewer instruments, EarthCARE essentially provides a space-borne ARM site that visits every climate region on the planet. To do this, it is essential that all instruments are embarked on a single platform, because this enables the fields of view of the different instruments to be co-located in space and time. Simultaneity between the various measurements is important because clouds vary rapidly in both space and time. The altitude of the platform will be as low as possible, so as to optimise the performance of the active instruments.

A near-polar orbit is required to provide global coverage. EarthCARE is a process study rather than a climatology mission. The aim of the mission is to provide numerous and comprehensive sets of samples of vertical profiles of clouds and aerosol properties constrained by TOA radiance measurements. In principle, the orbit could be slowly drifting in local time to enable sampling through the diurnal cycle. However, this would alias together the diurnal and seasonal variations so that the two are difficult to disentangle from the data. Therefore, the preferred option is a Sun-synchronous, near-polar orbit. The orbit equator crossing time should be 13:30 hours (see section 4.7.5 Common Measurement Requirements).

The orbit repeat cycle is not very strongly constrained by scientific requirements. EarthCARE objectives focus on atmospheric processes, therefore priority must be given to processes validation and quality control. It is important that sampling is unbiased in space and data be representative of all regimes found on Earth.

Overflights over ground-based observatories do not need to be considered for the orbit selection. It might seem that regular revisits of ground sites would be beneficial for periodical verification. However, considering that satellite sampling over a given site is very short (due to the high satellite speed) and generally poorly collocated (considering the high variability of clouds), ground-site overflights are generally not statistically significant and therefore only of limited value for validation.

Consequently, ground-site overflights are of limited value for validation and should not drive the selection of orbital parameters.

A mission lifetime of two (minimum) to three (optimum) years is required.

4.5 Data Delivery Requirements

Real-time access to the Level 1 EarthCARE data is necessary if the data are to be routinely monitored or assimilated into NWP models. Near-real-time access is needed for the evaluation of NWP models, to provide data for observational campaigns and for studies of current weather events. For climatological studies and the evaluation of climate and operational forecasting models, delayed mode access (days to one month) might be adequate, providing that the data products are of high quality and stability. However, for scientific data quality analysis and feedback it would be very important to access level 1 and level 2 data products within the shortest possible time, see section 5.1.

4.6 Measurement Principles

The observational requirements discussed above indicate the need for measurements from a single satellite platform of the vertical structure of aerosols and clouds, plus complementary information on cloud-scale vertical velocities and precipitation, and of the corresponding broad-band and narrow-band radiances at the top of the atmosphere. The profile information can only be provided through the analysis of back-scattered signals from an active technique. Furthermore, information is needed to discriminate absorbing from non-absorbing aerosols, on cloud-scale vertical velocities and precipitation (mainly drizzle) rates. Measurements are required to provide additional geographical coverage of aerosol and cloud optical properties. Broad-band measurements are required to measure radiances and to derive fluxes. These measurements need to be made for the whole globe and for a period long enough to ensure that all of the important climatic regimes are represented with the necessary statistical significance. This leads to the requirement for a single satellite mission with a lifetime of several years. Supplementary atmospheric information, namely temperature and humidity must be obtained from other sources.

The EarthCARE mission must detect clouds and aerosols that have an extinction coefficient above $0.05 km^{-1}$ and are therefore considered to be 'radiatively significant', because they produce changes in radiative flux of at least $10 Wm^{-2}$.

4.6.1 Broadband Radiometric Measurements

The TOA radiance shall be measured to derive fluxes of reflected short-wave (SW) and emitted long-wave (LW) radiation emerging from the same atmospheric region where the mission also observes clouds and aerosols. With this, the broadband measurements provide the boundary conditions for the fluxes calculations under consideration of the cloud/aerosol coverage and profiles measured by the other elements of the EarthCARE mission.

All energy exchange of significance between the Earth–atmosphere system and its cosmic environment is radiative. The system absorbs part of the incoming solar SW radiation flux and reemits this energy flux to space in degraded LW form with a spectrum characteristic of temperatures in the 200–300-K range. Reflected solar SW radiation flux, with a spectrum extending roughly from 0.2 to 5 μm, ranges from zero to 1000 Wm^{-2} locally and instantaneously. The Earth–atmosphere system emits LW thermal radiation over wavelengths mostly greater than 3.3 μm; outgoing LW flux ranges from 120 to 450 Wm^{-2} with a global annual mean value of 238 Wm^{-2}. Reflected SW and emitted LW spectra are fairly distinct, although there is overlap during daytime at wavelengths between 3.3 and 5 μm, where radiative flux density is relatively low for both spectra (Kandel et al., 1998).

The cloud and aerosol profiles shall be measured within the broadband measurement pixel so that their properties can be related to the measured TOA radiances. In order to derive physical properties and profiles of clouds and aerosols, the analysis of the back-scattered signal of appropriate active techniques is required.

4.6.2 Backscatter Measurements for Cloud and Aerosol Profiling

The EarthCARE mission shall measure the vertically resolved distribution and physical properties of clouds and aerosols, with the accuracy requirements given in Tables 4.2 and 4.3.

The cloud/aerosol profile measurements shall provide information on profiles of the liquid, supercooled and ice water content in clouds, cloud overlap, particle size within and extinction of clouds as well as vertical profiles of aerosols properties. Furthermore, convective updraught and ice fall speed shall be measured. This set of required parameters can only be satisfied using a combination of active techniques, namely, microwave Radio Wave Detection and Ranging (radar), for cloud measurements, and optical Light Detection and Ranging (lidar), for both cloud and aerosol measurements.

Radar

Vertical profiling of most cloud types can be achieved with an active technique of backscattered millimetre-wave radiation. Typically, 94GHz radars are used, which provides high sensitivity. The magnitude of the backscattered signal when expressed as the radar reflectivity factor, *Z*, is given by ΣND^6, where *N* is the concentration of particles of size *D*. *Z* is usually expressed in dBZ (=10 log Z) relative to the backscatter from a single raindrop of size 1 mm in a cubic metre assuming Rayleigh scattering. Liquid cloud droplets are typically 10 μm in size and are close to the threshold of detectability unless occasional drizzle drops are present. Ice particles with generally larger size have a much larger value of *Z*.

The net vertical motion of the cloud particles, which is the vector sum of their terminal velocity and the air motion, can be derived by analysing the difference in phase of backscattered radar signals between successively transmitted pulses. The EarthCARE radar would be the first to be flown in space with this 'Doppler capability'.

Microwave radar is a powerful technique for cloud profiling, however the link between optical characteristics is not direct (i.e. there is no simple relationship between cloud extinction and optical depth). Also, aerosol layers and certain clouds are comprised of particles too small when compared to radar wavelengths to make them visible to radar. Therefore, in addition to using microwave radar, aerosol and cloud measurements have to be done with shorter wavelengths than microwave, which implies an optical technique.

Lidar

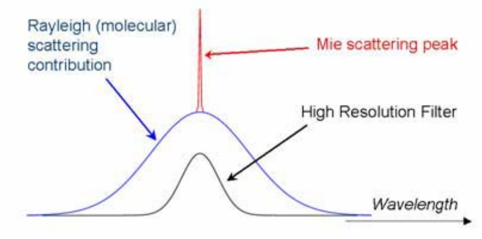

Figure 4.2: *Schematic of Rayleigh and Mie scattering contribution, which can be distinguished by a high spectral resolution technique.*

Lidar is the optical equivalent of microwave Radar. In general, an optical lidar signal will be backscattered by molecules (Rayleigh scattering) and particles (Mie scattering). An appropriate optical wavelength has to be selected in order to provide a sufficiently strong Rayleigh signal. Furthermore, the optical receiver has to be designed with high spectral resolution so that the separation of Rayleigh and Mie return signals is possible. The backscatter sensitivity has to be selected so that radiatively significant clouds or aerosols can be detected with sufficient accuracy, i.e. clouds/aerosols layers that change the flux by more than 10 Wm^{-2}, corresponding to an extinction coefficient of 0.05km^{-1}.

The detection of cloud/aerosol layers with an extinction 0.05 km^{-1} requires a sensitivity of the backscatter signal, $\beta = 8x10^{-7}(m\ sr)^{-1}$ assuming a ratio of the extinction (α) to backscatter, α/β, of the lidar signal, *S*, of 60.

By itself, the quantitative interpretation of the elastic backscatter only signal is difficult because of (i) attenuation in clouds, (ii) a variable and unknown value of *S*, and (iii) multiple scattering. For ice clouds, such as those shown in Figure 2.6, the microwave radar signal together with the lidar signal from cloud-free areas can be used to help correct for the optical attenuation and derive the profile of the extinction coefficient. The ratio of the microwave return ($Z \propto D^6$) to the corrected optical backscatter can then provide a measure of the mean ice particle size, and when combined with the microwave backscatter, the IWC and particle size can be estimated.

In the cases of aerosol and low radar reflectivity cirrus, no useful microwave radar return is available and thus the optical information alone must be relied upon. Conventional active optical techniques can only measure the total aerosol and molecular return signals, but because the extinction-to-backscatter ratio of aerosol (*S*) is quite variable, it is difficult to estimate aerosol extinction. The optical techniques must be able to separate the narrow band return from the slowly moving aerosol/cloud particles from the thermally Doppler broadened molecular signal as depicted in Figure 4.2. This 'High-Spectral Resolution Lidar' (HSRL) technique uses the derivative of the molecular return to directly determine the extinction, α, of aerosols and thin clouds. β is determined from the ratio of the Mie to the molecular return signal. The ratio of the retrieved extinction and backscatter values than directly lead to the profile of the lidar ratio S. The HSRL technique is similar in principle to the Raman lidar approach, in which the total Rayleigh and Mie return is separated from the wavelength shifted Raman scattered light from molecules. (See example in ESA, 2004, p. 22).

Information on the shape of the ice and aerosol particles can be obtained by measuring the backscattered optical signal in two polarisations. It also provides information on aerosol type and is complemented to this respect by the ratio of backscattering and extinction. The combination of measuring Rayleigh, Mie scattering and polarisation means that it will be possible to derive accurate measurements of aerosol and cloud optical depth and also provide an indication of aerosol composition and ice crystal habit.

The cloud and aerosol measurements need to be performed within each broadband measurement pixel as a two-dimensional (horizontal versus vertical) cut through the atmosphere along the flight direction. Therefore, the spatial requirements for the cloud/aerosol profile measurements can be expressed as along-track horizontal resolution and coverage requirements.

In practice, the footprint of the optical technique may turn out to be much smaller than the microwave footprint. Thus, it would be required that several successive optical measurements lie along-track within the footprint of every radar measurement.

4.6.3 Multi-Spectral Imaging

The broadband radiometric measurements and the active techniques' observations need to be collocated with measurements identifying the characteristics of the observed scene. This requires multi-spectral imaging over a swath significantly broader than the active instrument and broadband radiometer footprints, with spatial resolution significantly finer than the broadband measurements. This requirement responds to the need for a broader view of the atmospheric scenes observed by the narrow swath applying active techniques, and the need for information on the cloud variability within one broadband measurement and the conditions in its neighbourhood. With the help of multi-spectral imaging, it can be quantified how (in-)homogeneous an observed scene is and what possible impact on the scene observed by a broadband technique might result from the neighbourhood, e.g. shadow-casting of cumulonimbus towers. Therefore, the multi-spectral technique should view cross-track with a width significantly larger than the field of view of the broadband technique, and a resolution around 500m, below the resolution of the active microwave technique.

The multi-spectral imaging shall provide information on the horizontal variability of the atmospheric conditions and help to identify atmospheric components. Quantitative analysis of the measured reflected sunlight delivers information on the optical properties of the clouds and aerosols, while thermal infrared measurements yield information on temperature and infrared emissivity. Therefore, measurements in narrow spectral bands in the visible, near infrared, short wave infrared and thermal infrared are required.

Retrievals of multi-spectral imaging use the spectral information of radiances or apparent reflectance, which depend on the optical characteristics of atmospheric suspended particles, aerosol and cloud particles. For optically thin layers the radiances are mostly proportional to the sums of the differential cross sections and attenuated radiances reflected by surfaces. For the ocean surface, since the albedo is low, the reflected component is exactly estimated by the assumed particle parameters, whereas for a land surface the surface albedo has to be estimated in order to account for the generally non-negligible surface reflection. It is therefore difficult to retrieve the aerosol parameters over high reflectance surfaces such as desert and snow. Over dark vegetation surfaces, clear-day statistics of radiances and correlation between radiances at a specified wavelength and a longer near infrared wavelength, such as 1.6 μm and 2.2 μm, is used to retrieve the surface albedo. After removing the surface reflection component at each wavelength, the ratio of aerosol path radiances gives a good index of Ångström exponent or effective particle radius of aerosols. After obtaining the size index, reflectance is used to retrieve the optical thickness (Higurashi and Nakajima 1999).

On the other hand, the apparent reflectance for an optically thick cloud layer is mostly a function of optical thickness in a non-absorbing channel and a function of cloud particle size in the near-infrared channel. Therefore, use of a pair of wavelengths in the visible and near-infrared retrieves optical thickness and cloud particle size (e.g. Kawamoto et al. 2001). Cloud top temperature is also estimated by infrared channel accounting emissivity of cloud that depends of optical thickness and effective radius. The radiances of three thermal infrared channels are used to determine cirrus optical thickness, effective radius and temperature in daytime and night time (Katagiri and Nakajima 2004).

4.7 Scientific data observational requirements

The requirements discussed above can be achieved with a suite of instruments embarked on a single satellite platform, namely,

1. A broadband radiometer (BBR),
2. A cloud profiling radar (CPR),
3. An atmospheric backscatter Lidar (ATLID),
4. A multi-spectral imager (MSI)

The four instruments in concert provide the required geophysical parameters, as depicted in Figure 4.3.

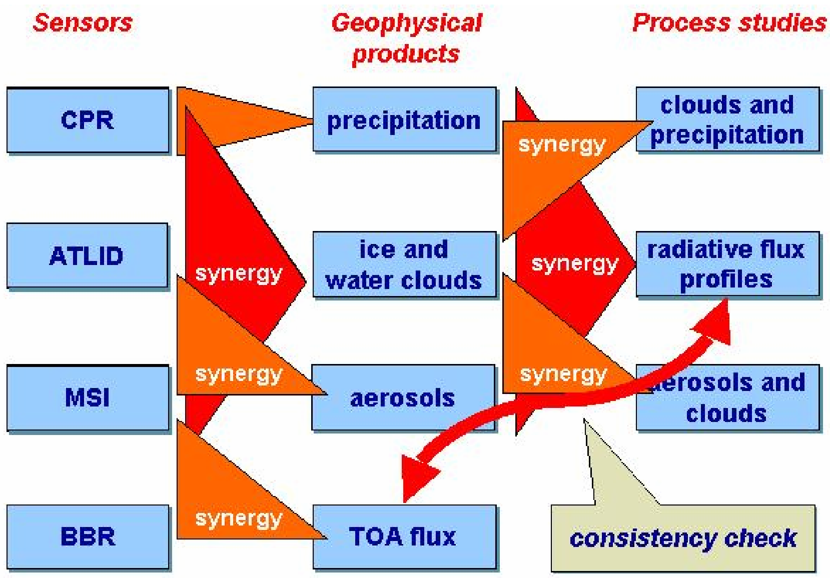

Figure 4.3: *The synergistic exploitation of the four instruments' observations will enable the retrieval of the required geophysical data products.*

While the detailed space segment system and instrument specification are not within the scope of this document, a number of high-level requirements on these instruments are necessary as far as they have been directly derived from scientific requirements analyses under consideration of the fact that the scientific mission goal can only be achieved through appropriate synergistic use of the measurements made by this instrument suite. The requirements have evolved from scientific analysis documented by ESA (2004) and other sources as respectively listed.

The following sub-sections summarise the measurement requirements taking into account synergistic exploitation of their respective observations.

4.7.1 Requirements for BBR measurements

The objective is to derive instantaneous broadband TOA fluxes with an accuracy better than $10 Wm^{-2}$.

The effectiveness of the principles and techniques of the broadband measurements using broadband observations have been demonstrated in the Nimbus/ERB, ERBE, ScaRaB, and CERES missions (e.g. Wielicki et al., 1996, Kandel et al., 1998, Duvel et al., 2001). In CERES, ScaRaB, and (in practice) ERBE, broadband LW radiance are usually obtained by subtraction of broadband SW radiance from total channel radiance (TW). The spectral response of the instrument shall cover the range 0.2-50 μm, with a possible extension up to 100 μm. The response shall be as flat as possible in order to be able to calculate the radiance of the scene from the radiance measured by the instrument and filtered by the instrument's spectral response. In the case of ScaRaB (Kandel et al., 1988), there was no abrupt longwave cutoff in spectral response of the TW channel although a gradual fall in response was expected as wavelengths become comparable to paint thickness. Even if response *were* zero at all wavelengths beyond 50 µm, assuming it to remain constant instead would lead to an error of 1-2 $Wm^{-2}sr^{-1}$ for Earth-emitted 'unfiltered' (i.e., spectrally corrected) LW radiances between 40 and 150 $Wm^{-2}sr^{-1}$. In the absence of reliable spectral response measurements beyond 50 µm, they assumed a continuation of the falloff observed between 20 and 50 µm, and estimated the maximum resulting error in unfiltered LW radiance to be less than 1%.

Spectral correction for residual spectral filtering and sensors is performed following procedures developed and applied in ERBE, ScaRaB, and CERES, involving knowledge of the instrument spectral properties together with scene identification to estimate scene spectral signature. Specific simulation estimations for the SW and for the LW have been developed by correlating 'unfiltered' BBR radiances from 'filtered' BBR and MSI radiances showing that the best correlation is a linear combination of BBR and MSI radiances. It is clear that the MSI radiances improve the *unfiltering* process by decreasing the noise error by a factor of two. Simulations of the *unfiltering* process for different input error levels have also been carried out for overcast conditions without distinction of the underlying surface (the most unfavourable cases) giving a good estimate of the maximum error thinking of the contribution to the error budget. Due to the nearly flat spectral responses over the SW and LW domains, these spectral corrections are relatively small and the error of the filtered BBR radiances propagates linearly through the correlations. (Lopez-Baeza et al., 2001).

Accurate *instantaneous* radiance-to-flux conversion shall be ensured thanks to the pointing capability of the BBR that observes along the scanning track with three views, namely a nadir view and two symmetric off-nadir views near 55º, and by viewing at three different angles along-track, building on experience from CERES, POLDER, and MISR (Lopez-Baeza et al., 2001, Bodas et al., .2003). The Agency has commissioned a study to improve the definition of the *Angular Dependence Models* (ADMs) which are needed for the specific characteristics of the BBR and taking into account 3-D cloud fields and surface BRDF (*Bidirectional Reflectance Distribution Functions*) effects into consideration (Lopez-Baeza et al., 2006).

One might consider that the pixel size of broadband measurements should be as small as possible in order to respond to the resolution needs of the models and in order to be as consistent as possible with the performance of the active instruments. However, even for a point at the top of the atmosphere, the outgoing flux is an integration of outgoing radiance over π steradians, and so depends on atmospheric conditions in a volume with horizontal extent that can be several tens of km. For example, in the SW, the relative amount of light scattered within the atmosphere into and out of the air column captured by the BBR is bigger for a smaller pixel. In the long-wave region, except for the case of optically thick high cloud, the flux at a point depends on the temperature/humidity profiles and cloud layers well outside the broadband pixel at TOA. Detailed scientific (Lopez-Baeza et al., 2001) and technical (R. Mercier Ythier et al., 2001) studies commissioned by the Agency in preparation of this mission have shown a good compromise could be achieved with a pixel size of 10km x 10km. This is a key issue because pixel size impacts flux retrieval significantly. The detailed justification for the BBR requirements can be found in the Technical Note E. Lopez-Baeza (2005).

Ideally, the whole pixel of the broadband measurements should be filled up with a full three-dimensional measurement of cloud and aerosol properties, however, as this may become prohibitively demanding on the engineering side, it would be sufficient to measure only a vertical slice through each broadband measurement pixel.

Both the misidentification of the scene for the different along-track views and the limited sampling of the active instruments over the BBR pixel make it necessary to define a certain degree of overlapping between the three views. This obviously helps to ensure co-registration conditions between the three views as well (parallax compensation). To improve co-registration with the other instruments, over-sampling will be used to provide 10 km square footprints (in nadir) every 1 km for each of the three views.

On-ground calibration is necessary to determine spectral response and calibrate in-flight calibration sources

In-flight calibration is needed to subtract the various instruments offset, to calibrate the instrument gain, to cross-calibrate the two spectral channels and to monitor possible spectral drifts of the instrument response. As a matter of fact, the spectral response can vary, mainly due to contaminants deposited on the optics, which can degrade the transmission when exposed to UV radiation. In-flight calibration sources: a black-body, the deep cold space and the sun via a diffuser.

Table 4.5: *Observational requirements for BBR measurements*

Parameter	Requirement	Reference/Comments
Spectral channels	SW: 0.2 – 4.0 μm LW: 4.0 – 50 μm	BBR shall estimate the Earth spectral radiance in the Long-Wave (LW) channel from the measurement in the Short-Wave (SW) (0.2 – 4.0 μm) and Total-Wave (TW) (0.2 – 50 μm) channels. The intermediate limit of around 4.0 μm is optimum to minimise terrestrial thermal radiation in the SW channel.
Dynamic range	SW: 0 – 450 $Wm^{-2}sr^{-1}$ LW: 0 – 130 $Wm^{-2}sr^{-1}$	To cover the full respective dyanmical ranges of wavelength-integrated reflected solar and emitted Earth radiation.

Absolute Accuracy	SW (threshold): 2.5 $Wm^{-2}sr^{-1}$ SW (goal): 2.0 $Wm^{-2}sr^{-1}$ LW: 1.5 $Wm^{-2}sr^{-1}$	Bozec et al., 2001, obtained a radiometric performance of 2.3 $Wm^{-2}sr^{-1}$ for the SW and 0.7 $Wm^{-2}sr^{-1}$ for the LW. This error is a 1 σ error estimated on the radiance measured by the BBR and filtered by the spectral response of the instrument. The absolute radiometric accuracy only includes all systematic errors after calibration and the instrument noise error (random error). Unfiltering process and radiance to flux conversion errors are not included here.
Angular sampling	3 pixels along-track: one at nadir, two at an angle of ±55 deg wrt local zenith	The selection of the optimum VZA is described in Lopez-Baeza et al., 2001 and in Bodas et al., 2003. Validation of this selection with CERES data is described in Bodas et al., 2003b
Horizontal along-track sampling	< 1km	To achieve co-registration of the 3 views and with other instruments, over-sampling will be used to provide a measurement every 1 km for each of the three views.
Geolocation knowledge (for each of the 3 BBR footprints)	better than 1km (goal 500m)	
Footprint size (at sea level)	10 km x 10 km in the three views.	Lopez-Baeza et al., 2001
Pixel co-registration	Registration error between the BBR nadir pixel and off-nadir pixel centres lower than 1 km (across-track)	See López-Baeza (2006). Note: the Earth rotation has to be taken into account and corrected for, in order to achieve the required overlap of fore-, nadir and aft-viewing pixels.
Instrument spectral response	The normalised spectral response of the SW and TW channels shall be "as flat as possible"	The goal of the BBR is to measure the integrated atmospheric radiation over the specified wavelengths ranges. This is done by the convolution of the atmospheric signal with the instrument's spectral response function in the respective channels. Considering that the Earth outgoing radiation shows very strong spectral and variable structures, the instrument response function must therefore be as flat as possible in order to approximate the real integral atmospheric radiances as closely as possible.

Across track sampling	No across-track sampling	

Spectral Correction (‘unfiltering’), which is part of the BBR retrieval, takes the optic component response into account to provide the physical (unfiltered) radiance that reaches the satellite.

4.7.2 Requirements for CPR measurements

The objective of the cloud radar is to probe and quantify clouds in synergy with the measurements of the backscatter Lidar. Furthermore, the cloud radar would be equipped with a Doppler shift measurements capability to measure vertical motions.

For cm wavelengths (weather radar) the value of the normalising dielectric factor, $|K|^2$, is 0.93. At 94GHz, $|K|^2$ is a function of temperature. However, we assume a constant value of 0.75 (value at 10°C). Using this value the power returned for a given Z is 1dB lower than for an assumed $|K|^2$=0.93. (See technical note by Lin, 2006)[3]

Parameter	Requirement	Reference/Comments
Radar reflectivity, dBZ	-35 dBZ at TOA over 10 km along-track horizontal integration.	To detect 98% of radiatively significant ice clouds and 40% of all stratocumulus (Stephens et al, 2002)
	-30 dBZ at TOA with 1km horizontal integration	Will detect 91% of significant ice clouds and 30% of all Sc.
Along-track sampling	1 measurement per 500m horizontal distance along-track	To determine subgrid scale structure. Hogan et al (2002) and to identify and correct for Doppler velocity biases.
Vertical range	The nominal vertical range shall be latitude dependent and extend from the surface (-0.5km) to 20 km in the tropics 16 km in the mid-latitudes 12 km in polar region. (The exact latitude bands shall be adjustable in flight.)	Vertical extent of tropical and mid latitude clouds from Okamoto et al (2005a, 2005b) and from the ARM and CloudNet web sites.
Vertical resolution	500m with sampling every 100m	Simple radiation calculations and aliasing studies by Schutgens and Donovan (2004)
Accuracy on altitude determination	100 m or better	See respective requirement in Table 4.7

[3] The requirements in Table 4.6 are based on definition of *Z*, assuming $|K|^2$ = 0.75 and SNR=3.35dB (speckle 1.65dB).

Doppler accuracy	Goal requirement: 1 m/s at 1km horizontal resolution (at -14dBZ for any PRF) and 0.2m/s at 10km horizontal resolution. (To be achieved at -14dBZ reflectivity and highest PRF of 7200Hz) Threshold requirement: 1m/s at 10km horizontal resolution. (To be achieved at -19dBZ for any PRF).	For studies of convective motion. (Igau et al, 1999). The 0.2m/s accuracy is required to determine ice crystal and drizzle sedimentation velocities. (Heymsfield and Iaquinta, 2000; O'Connor et al 2005). (See notes on beam direction reliability and reflectivity and velocity gradients below.)

Table 4.6: *Observational requirements for CPR measurements. Unless otherwise stated the vertical resolution requirement is 500m.*

Reflectivity

The requirement is to sense clouds occurring at all altitudes. The ARM and CloudNET programs maintain ground based cloud-profiling radars and lidars in tropical, mid-latitude and polar sites. Perusal of the monthly quicklook data[4] reveals that clouds extend to a height of 20km in the tropics but to less than 12km poleward of 60deg latitude. The upper limit of the vertical range as a function of latitude shall be adjustable in-flight. Present estimates suggest that the necessary vertical range for cloud measurements should have an upper boundary of 20km in the tropical and sub-tropical regions and 12km in regions poleward of about 60deg latitude. However, it has to be considered that on the one hand all significant clouds will have to be included in the vertical sampling range (calling for a large vertical range) and, on the other hand, Doppler measurements will benefit from a lower upper boundary (i.e. a denser vertical sampling within the vertical range). The exact latitudes where the altitude top is switched between 12km/16km or 16/20km should be based on analyses of in-flight measurements. The mission design and observation plan needs to take into account a flexible vertical range. The numbers given in the table are the default values.

The vertical resolution of 500m arises because a change of cloud height by this amount leads to a change in fluxes of $10W/m^2$. Studies by Schutgens and Donovan (2004) show that if the basic resolution from the radar pulse is 500m then oversampling improves the resolution to 100-200m provided the signal to noise is high enough.

[4] http://met.reading.ac.uk/radar/cloudnet/quicklooks

The radar sensitivity requirement for ice clouds is provided by Brown et al (1995) who analysed ice crystal images in cirrus clouds, computed the reflectivity, ice water content, and carried out radiative transfer calculations. They found that a change of $10W/m^2$ resulted from cloud and aerosol layers with an optical depth above 0.05, and that 98% of ice clouds with this optical depth would have a reflectivity above –35dBZ. These findings are confirmed by Hogan et al (2006) The situation with water clouds is more difficult, because the smaller size of the water droplets leads to lower radar reflectivities. A 12 month study with vertically pointing ground based radar and lidar (the lidar will, of course, detect all optically significant water clouds,) reported in Stephens et al, (2002) found that a detectability threshold of –35dBZ would sense about 40% of all stratocumulus clouds, this figure of 40% includes all the clouds which contain significant drizzle droplets.

The ice and water clouds in the above paragraph are quite extensive so a horizontal resolution of 10km is acceptable. However, this does mean that 1km horizontal resolution will be available for echoes above –30dBZ which encompasses 91% of the radiatively significant cirrus clouds and 30% of the low level liquid water stratocumulus clouds. 500m horizontal resolution is required to compute cloud fraction in a model grid box, and also to derive statistics on sub grid box scale inhomogeneitites and overlap in the vertical of such inhomogeneities (Hogan and Illingworth, 2003).

Doppler

Doppler information would be particularly useful in three areas: characterising convection, estimating the sedimentation velocity of ice particles in cirrus, and quantifying drizzle fluxes in stratocumulus. An extensive survey of convective motions over the tropical ocean by Igau et al (1999) who used the widespread definition of a convective core as a velocity of over 1m/s extending over a vertical distance of 500m, found that over 10% of updraught cores in the tropical pacific had a mean vertical velocity of 5m/s and a diameter over 2km.

Cirrus cloud particle terminal velocities are up to 2m/s, and because such clouds are extensive, useful information would be obtained for Doppler with an accuracy of 0.2m/s with a spatial resolution of 10km (Heymsfield and Iaquinta, 2000). O'Connor et al (2005) find that these specifications of an accuracy of 0.2m/s over a horizontal distance of 10km will also enable the radar to detect the presence of drizzling regions within stratocumulus, and also suggest how the Doppler and reflectivity measurements can be combined to provide an estimate of drizzle rain rate and liquid water flux.

From the satellite the ability to measure Doppler velocity is governed by the ratio of the Doppler width (depends upon the beamwidth) to the folding velocity (which depends upon the pulse repetition frequency of the radar and is of course affected by the signal to noise ratio (Kobayashi et al, 2002, 2003, Kumagai et al, 2002). Those studies suggest that it needs reflectivity of more than -30dBZ in order to get 1 m/s accuracy with 10km integration at the PRF of 7200Hz.

Another issue on vertical Doppler speed accuracy is *beam direction knowledge*. Even if the beam direction has 0.01 degrees inclination from the nadir, it will cause 1.3m/s component because of the satellite velocity. It should be possible to correct this using the satellite navigation data and ground return signal.

Even if the centre of the beam points exactly to the nadir, recent work has also suggested that *reflectivity and velocity gradients* across the beam on a scale of 1km can bias the retrieved Doppler velocity (Schutgens and Kumagai, 2005). If reflectivity or velocity of the forward side of the beam is much different from that on rearward side of the beam then the retrieved Doppler velocity will be biased. Sampling of the radar returns at a resolution of 500m may be necessary to identify when such biases may exist and correct the data accordingly. This aspect needs further investigation and evaluation.

Sato and Okamoto, 2005, suggest that if velocities with an accuracy of 0.1m/s are combined with the observed radar reflectivity, then an accuracy of 20% in r_e and 20% in IWC for the ice particles can be retrieved if r_e=100µm where the particle velocity is about -2m/s, -1.7m/s and -1.4m/s at the altitude of 15km, 12km and 8km, respectively. For ice particles with r_e=50µm (velocity is -1m/s, -0.8m/s and -0.7m/s at Z=15km, 12km and 8km), an accuracy of 20% and 50% in r_e and IWC, respectively, can be expected.

4.7.3 Requirements for ATLID measurements

The backscatter Lidar will complement cloud observations from the radar, allowing the detection of small ice particles and water droplets, differentiating water phases. It will further provide information on the aerosol vertical structure and optical properties and allow identification of aerosol type and cloud particle size and type in synergism with the MSI.

Table 4.7 lists the parameters that need to be obtained. Separation of Rayleigh and Mie signals is required to distinguish molecular and particle signals and determine aerosol and cloud extinction from transmission variation.

The required lidar sensitivity should allow to detect radiatively significant cloud and aerosol layers corresponding to the minimum detectable optical depth. This is done here assuming an average extinction-backscatter ratio of 60 corresponding to high level clouds or elevated dust layers which may have a significant radiative impact. Indeed this requirement leads to detect an optically thin cirrus cloud (or aerosol layer) having an optical depth of 0.05 uniformly distributed over 1 km on the vertical, corresponding to a change of 10 Wm^{-2} on the OLR (Brown et al., 1995). Cirrus clouds corresponding to these characteristics may be as high as 20 km in altitude in the tropics.

Parameter	Requirement	Reference/Comments
Lidar Sensitivity (Minimum detectable backscattering coefficient)	$8x10^{-7}m^{-1}sr^{-1}$	For a 100 m vertical, 10 km horizontal resolution and an accuracy of 50 %.
Lidar Sensitivity for cross-polar channel (note 5)	$2.6x10^{-6}m^{-1}sr^{-1}$	For a 100 m vertical, 10 km horizontal resolution and an accuracy of 50 %. (Shimizu, 2004)
Cloud optical depth τ_C detection sensitivity and range (note 1)	0.05 - 2	For a single layer 1 km thick cirrus cloud and 10 km horizontal resolution or better. The required sensitivity (0.05) corresponds to the minimum detectable optical depth with an accuracy of 100%. The required accuracy is 0.05 or 20 % whichever is largest.

Aerosol optical depth τ_A detection sensitivity and detection range (note 1)	0.05 - 2	This applies to the PBL for a thickness of 1 km and a 100 km horizontal resolution The required sensitivity (0.05) corresponds to the minimum detectable PBL aerosol optical depth with an accuracy of 100%. The required accuracy is 0.05 or 20 % whichever is largest.
Other requirements		
Altitude range	0-30 km	This range is defined by the vertical extension of clouds (0-20 km) complemented to the lower stratosphere to allow volcanic aerosol detection.
Accuracy on altitude determination	100 m or better at horizontal sampling	Note 3
Vertical sampling requirement	100 m or better for altitudes up to 20km, 500m for 20km-30km altitude range	High resolution is needed for structure analysis. (Altitudes >20km, up to 40km, to be considered for calibration.)
Horizontal sampling requirements	100m	Higher resolution along-track than other instruments to properly analyse surface and atmosphere heterogeneity and sample radar and MSI footprints (Hogan et al., in ESA SP1257-1). See also Note 4.

Table 4.7: *Observational requirements for ATLID measurements. The cirrus requirements apply to daytime measurements with worst case background condition for a cirrus cloud at the altitude of 9-10km.*

Vertical and horizontal resolution of products should correspond to what is needed in terms of analysis, combining observations of the EarthCARE platform. It should also be commensurate with model outputs. This typically corresponds to a vertical resolution of 100 to 500 m on backscattering and extinction coefficients, as well as an horizontal resolution of about 100m to 100 km for low cloud altitude and aerosol optical depth, for example.

To minimise multiple scattering, the lidar receiver field of view on ground level should be 30m or smaller.

Note 1:

- The corresponding accuracy on the Rayleigh signal S_R can be derived from the constraints given on optical depth in Table 4.7. $\langle S_R \rangle$being the average attenuated Rayleigh signal, the overall accuracy on the Rayleigh signal S_R is defined by the error on the retrieved optical depth τ estimated from $[d<S_R>/<S_R>] \sim \tau.[d\tau/\tau]$.

- As a reference case, a cloud between 9 and 10km altitude height and optical depth of 0.1 and 1 shall be considered with a vertical averaging of the Rayleigh signal of 1.2km above and below the cloud.

Note 2: The Mie signal shall be measured in two polarisations to further analyse particle shapes determining the aspect ratio (Del Guasta, 2001 ; Noel et al.,2005) and contribute to identify aerosol type (Ansmann et al, 2004).

Note 3: An accuracy of 300 m on cloud top altitude would correspond to an error of 10 W/m^2. However the accuracy required here corresponds to an error in temperature compatible with the one of MSI measurements assuming a saturated adiabatic lapse rate of 6.5 $K.km^{-1}$.

Note 4: 100m horizontal sampling (separation of lidar pulses) is considered ideal, 200m would be acceptable (providing the sensitivity requirements over 10km integrations are maintained). The individual lidar shots should preferably be processed on-ground.

Note 5: The use of circular polarized light would be preferred. Multiple scattering will give rise to depolarized lidar returns even in the case of spherical targets (Hu et al., 2001). The amount of multiple-scattering induced depolarization is, in general, reduced when circular polarized light is employed (Hu et al., 2003). This will simplify the process of cloud phase discrimination using the depolarization ratio.

4.7.4 Requirements for MSI Measurements

The MSI measurements shall, as a minimum requirement (threshold requirement), provide information on the horizontal structures of clouds, like cloud cover and specifically cloud type, cloud optical and microphysical properties over sea and land surfaces.

As a goal requirement, MSI shall furthermore be able to provide measurements of aerosol optical properties and aerosol type over sea surfaces to supplement the ATLID aerosol measurements in the across-track dimension. The goal requirements is that the VIS and NIR channels are able to measure aerosol optical depths over ocean with a detection threshold of 0.05 and to an absolute accuracy of 0.02 for an integrated area of 10km×10km.

Tables 4.8 give an overview of the observational requirements.

Parameter	Requirement	Reference/Comments
Band definitions	see Table 4.8b	all bands are defined considering cloud and aerosol spectral characteristics and other the bands from routinely operating meteorological satellite systems, like MSG-SEVIRI or NPOESS AVHRR, to ensure a close link to temporal higher resolved satellite data respectively products
Radiometric requirements	see Table 4.8c and 4.8d	are based on various scenarios: • clear-sky conditions to infer aerosol properties • clear-sky conditions to infer surface spectral reflectances and surface temperatures • cloudy skies to infer cloud properties (from dark to bright surfaces)
Horizontal resolution	500 m	to resolve mesoscale cloud properties incl. scale break for clouds (Schröder, 2004)
Horizontal coverage	- continuous along-track - 150 km across-track	to resolve meso-β cloud field structures (Orlanski, 1975)

Channel co-registration	Between band 1 and 2: 0.15 (0.1 goal) pixels All channels 0.5 pixels for altitude range 0-20km	Necessary for meso-γ cloud properties Applicable to full swath.

Table 4.8a: *Observational requirements for MSI measurements*

Mesoscale variations (Orlanski, 1975), such as mesoscale cellular convection or cloud streets, are often observed and can be simulated considering large turbulent eddies using numerical weather prediction models as well as regional climate models. Such simulated flow fields can than be used to examine the effect of mesoscale variations on the cloud properties, e.g. for marine stratocumulus clouds (GEWEX GCSS, 2000).

		centre wavelength (μm)	50 % transmittance (μm)	5 % transmittance (μm)
Band 1	VIS	0.67, +/- 0.01	0.02	0.03
Band 2	NIR	0.865, +/- 0.01	0.02	0.03
Band 3	SWIR 1	1.65, +/- 0.015	0.05	0.08
Band 4	SWIR 2	2.21, +/- 0.015	0.1	0.15
Band 5	TIR 1	8.8, +/- 0.05	0.9	1.1
Band 6	TIR 2	10.8, +/- 0.05	0.9	1.1
Band 7	TIR 3	12.0, +/- 0.05	0.9	1.1

Table 4.8b: *MSI observation band requirements. Note: The TIR channel requirement at 5% transmittance is not a hard requirement, as long as a 1% out-of-band rejection is achieved.*

		Dynamic range [%]	SNR at 100% TOA reflectivity	*Goal* values at low TOA reflectivity		Absolute radiometric accuracy
				SNR	Reference signal [$Wm^{-2}sr^{-1}\mu m^{-1}$]	
Band 1	VIS	0 – 110	500	75 (*) (**)	30	2% goal 10% threshold (***)
Band 2	NIR	0 – 110	500	65 (*) (**)	17	
Band 3	SWIR 1	0 – 102	250	18 (*)	1.5	
Band 4	SWIR 2	0 – 100	250	21 (*)	0.5	

The radiometric stability shall be better than 1% of the estimated reflectance value over one year. The inter-band accuracy shall be <1%.
(*)The goal values for SNR values and the corresponding signal levels in the solar channels shall be based on the cloud optical thickness (δ_c) for a low water cloud and the requirement to retrieve its variation ($\Delta\delta_c$), with $\delta_c = 0.5$ and $\Delta\delta_c = 0.05$. For SWIR channels, the sensitivity criterion shall be based on a retrieved water droplet size uncertainty of $\Delta r_e = 0.5\mu m$ and ice particle size uncertainty of $\Delta d_e = 5.0\mu m$ for a water cloud with 1km cloud base and an ice cloud with 10km cloud base and an optical thickness of OD=0.5. Requirements shall refer to the individual MSI pixel level.
(**) In addition, the SNR at low reflectivity for band 1 (VIS) and band 2 (NIR) shall also be derived from the aerosol accuracy requirement of 0.02 optical depth over open sea for 10km x 10km area. These values are goal requirements for band 1 and 2.
*(***) As a minimum scientific requirement, or threshold requirement, an absolute radiometric accuracy of <10% would allow for scene identification (which is the primary objective of MSI, needed as support to BBR flux retrievals) and retrieval of parameters of optically thick clouds. However, this accuracy would compromise aerosol retrieval capabilities and render the retrievals of thin clouds impossible. Therefore, **5% is considered the scientific breakpoint value**, which would significantly improve aerosol and thin cloud retrievals. The design goal requirement should be 2%, which would be the ideal value for accurate cloud and aerosol retrievals. The accuracy shall be achieved over a dynamic range reaching from lowest to the highest TOA spectral radiances that can be expected during daylight.*
Table 4.8c: *Dynamic range, SNR and radiometric calibration for solar channels.*

The MSI bands 1 to 4 will also be used for scene identification, which is needed for the BBR processing. The scene identification is necessary for the unfiltering process, in order to obtain absolute radiances from the BBR measured radiances which are filtered by the BBR spectral response. Furthermore, the scene identification is necessary for the selection of the ADM value to account for anisotropic effects.

		Dynamic range [K]	Threshold requirements NEDT at 220K	NEDT at 293K	Goal requirements NEDT at 220K	NEDT at 293K
Band 5	TIR 1	170 – 350	0.8	0.25	0.6	0.1
Band 6	TIR 2	170 – 350	0.8	0.25	0.7	0.15
Band 7	TIR 3	170 – 350	0.8	0.25	0.8	0.15

Table 4.8d: *Dynamic range and NEDT for longwave MSI bands 5 to 7. The goal requirement of 0.1K NEDT at 293K is required for the retrieval of cirrus particle sizes which would not be possible with 0.25K. The required radiometric accuracy shall be better than 1K at 280K. The interband accuracy shall be better than 0.25K (goal 0.1K) over one year. The radiometric stability shall be better than 0.3K over one year.*

The justification for the band selection can be found in [ESA, 2001, pp. 67]. The VIS and NIR channels are comparable to AVHRR (NOAA) and essential for cloud detection, scene identification and aerosol. Furthermore, they can be used for land surface and vegetation monitoring. The SWIR 1 channel is important for snow and cloud discrimination and essential to distinguish ice and water clouds. It furthermore provides additional aerosol information. The SWIR 2 channel improves the determination of cloud optical properties (cloud optical thickness and microphysical properties, see King et al., 2004). The three TIR channels are all window channels for split-window, dual window and triple window techniques to infer surface temperatures (sea and land surface as well as cloud tops, Qin and Karnieli, 1999). TIR1 provides quantitative information on thin cirrus and supports the discrimination between ice and water clouds. TIR 2 and 3 can also be used for cirrus cloud detection (e.g. Inoue, 1987) and volcanic ash clouds (Prata, 1989).

Cloud optical thickness for the atmospheric column and cloud droplet radius or crystal size diameter at cloud top can be determined applying visible and near-infrared reflectances relationships (Nakajima et al, 1990, 1991, 1995, Berger, 2001). To ensure a sufficient accuracy the cloud reflectances has to be determined based on the given signal-to-noise ratios.

The sunglint influences mainly sea surface reflectances and therefore the determination of the aerosol optical thickness (in the order of 0.02). To minimize this effect on the MSI swath, a satellite equatorial crossing time later than 13:00 LT or earlier than 11:00 LT would be desirable. At 13:00 LT and considering a swath of 150 km, the sunglint peak will be out of the swath, but will still influence some parts.

In order to minimise the number of pixels affected by sun glint, the field of view of the MSI shall be tilted in across-track direction. The 150 km FOV should be positioned in a way that one edge is located in a distance of 35 km from nadir and the other side accordingly at approximately 115 km away from nadir[5].

[5] This is applicable for a sun-synchronous orbit with 13:30 hours equator crossing time.

The MSI swath shall extend beyond the BBR footprint in across-track direction by at least three times the size of the BBR (across-track) footprint size.

Vicarious calibration shall be used for improvement of the MSI calibration towards the radiometric accuracy goal values

4.7.5 Common Measurement Requirements

Equator crossing time

From the science requirement viewpoint, the choice of orbit equatorial crossing time is a compromise depending on both instrument performance issues and diurnal variations of atmospheric properties and radiation fields.

Firstly, afternoon orbits are generally preferred over morning orbits as convection is mainly initiated in the early afternoon over land. Secondly, near-noon retrievals of TOA broadband reflected SW fluxes are most representative of the diurnal average, and *relative* errors of SW radiance measurements and flux retrievals are smallest when the SW signal is largest. Therefore, the optimum observing time is in the early afternoon. To the extent that the EarthCARE mission focus on cloud-aerosol-radiation interactions is more on convective than on boundary-layer aerosol and cloud, these arguments go in the direction of an optimum equator crossing time between 13:15 and 14:15. Note that focus on low-level aerosol and clouds un-obscured by higher-level cloud would however be in favour of equator crossing time between 09:45 and 10:45.

A third factor involves the MSI aerosol retrievals. Over cloud-free ocean, the often severe sun-glint effects can be reduced by increasing the satellite-sun angle (implying equator crossing time later than 14:30 or earlier than 09:30), and this would also reduce stray light. Alternatively or in addition, the MSI FoV could be shifted sidewards to reduce sun-glint, maintaining however at least 35 km of coverage on either side of nadir to characterize the environment of the BBR pixels.

Finally, if EarthCARE is flown at the same equator crossing time as CloudSAT/CALIPSO (13:30), analysis of EarthCARE data can benefit from the additional types of data available from the A-train satellites and at the same time shed light on issues irresolvable without the superior EarthCARE observing capabilities (precise simultaneity and co-location, better signal/noise, Doppler capability, etc.).

The equator crossing time should be close to 13:30, but that afternoon crossing times between 13:15 and 14:00 could be acceptable if they strongly simplify technical issues.

Instrument co-registration

The instrument suite has been selected in order to simultaneously measure cloud and aerosol properties together with TOA radiance. Therefore, the centres of the instrument footprints have to be located as closely together as possible to ensure good co-registration. The co-registration requirements have been discussed in ESA (2001) and have not changed since. The scientific requirements for instrument footprint co-registration, however, depends on the instrument's engineering, namely, their footprint sizes. Should the footprint sizes be significantly changed in on-going or future instrument design and development, the co-location requirements will have to be reviewed.

Whilst all observations should be made in the nadir direction, there are system design considerations requiring different observation angles (ESA, 2001). The lidar needs to be pointed slightly off nadir to avoid specular reflection. Some of the MSI chanels, which are obtained by means of in-field separation in the detector planes, will also be offset from nadir. Furthermore, the forward and backward views of the BBR are offset by an appropriate angle (55º according to section 4.7.1) for optimisation of radiance-to-flux conversion.

Across-track, the distance between the centres of the CPR and ATLID footprints shall have an RMS value of not more than 350m (goal 200m). An MSI pixel shall be identifiable with a RMS value of the knowledge of its distance towards ATLID and CPR with not more than 350m (goal 200m). The distance of the centre of the BBR footprint to the CPR and ATLID shall not be larger than 1000m (RMS).

Along-track, the footprints of the four instruments shall be collocated (in space) and the positions between the footprint centres of any integrated CPR, any ATLID and any MSI centre pixel shall be known to an RMS accuracy of 350m (goal 200m). The position of any of these instruments footprint centres and the BBR footprint centre shall be known to an RMS accuracy of 1000m.

In principle, it would be desirable to know the relative position between ATLID and CPR to half of the ATLID sampling distance, in order to accurately assign the individual ATLID shots to the respective integrated CPR sampling interval (of 500m). However, considering that this would put very unrealistic engineering demands on angular knowledge accuracy, this should not be considered as a requirement.

Table 4.9 summarises the co-registration requirements, applicable to Level 1b data.

		ATLID	BBR	MSI
CPR	across track	**350 (200)**	**1000**	*350 (200)*
	along track	*350 (200)*	*1000*	*350 (200)*
ATLID	across track		**1000**	*350 (200)*
	along track		*1000*	*350 (200)*
BBR	across track			*1000*
	along track			*1000*

Table 4.9: Instruments' relative pointing requirements. All value are given in meters. Numbers in brackets are goal values. ***Bold numbers*** *refer to an absolute alignment requirement between the instruments' respective footprint centres. Italic numbers refer to a required knowledge of the relative position of the instruments' respective footprint centres. BBR relates to the centre of the nadir view; the relative alignment of the three BBR fields of view with each other are defined under BBR requirements. A sampling of the BBR along-track of 1km is assumed. MSI relates to MSI pixel centred closest to nadir. The values given are RMS values over one orbit.*

Reference altitude for BBR collocation with the other instruments

The reference altitude for the BBR observations can be at any location between 0 and 20km. This requirement and the co-registration requirement require the scene altitude be accurately defined. If the altitude is not known when processing the data, it will not be possible to obtain good co-registration of the views. An error of 1 km on the altitude's scene can lead to a de-registration between the forward and the nadir view of about 1.4 km. See also Loeb et al., 2002

5. Data Processing Requirements

5.1 Geophysical data products

The scientific mission objectives can only be fulfilled if geophysical products derived from EarthCARE observations are available. Furthermore, the synergy of the EarthCARE instruments could only be fully exploited on the level of geophysical products. It is therefore required that geophysical (Level 2) retrieval algorithms will be developed and Level 2 data products will be produced as part of the EarthCARE mission.

Many of the algorithms for retrieving the required cloud parameters such as profiles of cloud extinction, ice water content, particle size and radiative flux profiles exist at present and will be further developed by the time EarthCARE is launched. The techniques that will be employed in the analysis of EarthCARE data range from well established methodologies (such as the retrieval of cloud optical depth and effective radius from multi-spectral radiances (Nakajima and King, 1990) to recently developed synergetic algorithms involving multiple instruments (i.e. Austin and Stephens, 2001, Cooper et al. 2003, Donovan, 2003, Okamoto et al. 2003). Some of the more recent developments in the applied use of multisensor techniques have been described in chapter 4. This chapter outlines how the global data set to be obtained from the EarthCARE satellite could be used to evaluate the representation of clouds in operational models and the work being carried out to set up a system for assimilating the cloud data into operational models.

5.2 Data delivery requirements

For scientific data quality analysis and feedback it would be very important to access level 1 and level 2 data products within the shortest possible time, in order to enable fast detection of possible data degradation. Otherwise, corrective actions could have a corresponding time delay that might lead to extended periods of degraded data quality. It is therefore strongly recommended to keep the delivery of level 1 and level 2 products as short as possible, and not longer than a few days.

Nearly real time (NRT) access to the Level 1 data or, for research satellites, often to Level 2 data, is necessary if the data are to be monitored or assimilated into operational NWP models. Data monitoring would provide a very valuable continuous and near-real time science data quality control, while the assimilation itself would be a unique opportunity to exploit EarthCARE's synergistic measurements of clouds, aerosols and radiation. The required Level 1 and/or Level 2 data should be available on the time scale comparable to the current cut-off times applied in operational NWP centres. At ECMWF, two 4D-VAR assimilation windows are operational, a 6-hour period and a time-shifted 12-hour period. The 12-hour 4D-VAR window assimilates data acquired within the time period from 21:00 to 9:00 hours (or 9:00 to 21:00 hours, respectively). The assimilation run is executed starting at 14:00 hours (and 2:00 hours, respectively). By this time, all data to be considered for the run must have been delivered to ECMWF. Data delivered later will be discarded. Research satellite data are typically assimilated into the 12 hours runs. In order to assimilate EarthCARE data, they have to be delivered accordingly, i.e., data acquired at 9:00 hours (21:00 hours, respectively), must be delivered within 5 hours.

It is worth mentioning that by the time EarthCARE is launched, the quality of numerical models in their representation of clouds and aerosols will have certainly reached a degree of maturity that may allow a direct assimilation of EarthCARE cloud and aerosol products, directly providing benefit to the forecasts.

It is important to realise that NRT availability of the data will be very important for the mission as the complex data processing systems of NWP centres can in return provide continuous monitoring and immediate feedback on the quality of the observations. This real time feedback mechanism between data users and producers has already been implemented at ECMWF (in collaboration with ESA) for a number of ENVISAT instruments (see for example:
http://www.ecmwf.int/products/forecasts/d/charts/monitoring/satellite/o3/sciamachy/). This mechanism allows direct alert procedures in case of instrument anomalies or problems in data processing.

5.3 Ground based evaluation of clouds in operational models

Analysis of ground based vertically pointing radar and lidar profiles has been compared with the representation of clouds in the operational mode grid box above the ground-based station. An example of one day's cloud radar and lidar profiles with 30 second temporal and 60 m vertical resolution is provided in Figure 5.1. The operational model holds values of cloud fraction and IWC each hour with a vertical resolution of between 200 and 500 m, so cloud fraction can be derived for each model grid box from over 100 observations of cloud/no cloud and grid box IWC from the mean reflectivity. Figure 5.1.a, c and d display mean observed profiles of cloud fraction and IWC for the period May 1999 to May 2000 together with the corresponding values from the UK Met Office and ECMWF models. Figure 5.1.a shows that the mean cloud fraction is in fairly good agreement but the ECMWF model is overestimating the occurrence of cloud below 2 km by a factor of two. Figure 5.1.c reveals that for mid-level clouds there is an underestimate of the mean fraction of cloud when present by both models. The horizontal blue lines in Figure 5.1.d are the uncertainty in the retrieval of IWC from the observed Z. The errors shown here are larger than those using the EarthCARE algorithms. They show that although the mean value of IWC in the model for mid-level clouds is rather less than observed the values are within the observed error bars. This agreement is much more reassuring than large spread in IWC from AMIP shown in Figure 2.3. More recent comparisons of cloud fraction for April-October 2003 are shown in Figure 5.1b and confirm that the modification to the clouds scheme in ECMWF has successfully removed the excess low level cloud apparent in Figure 5.1a.

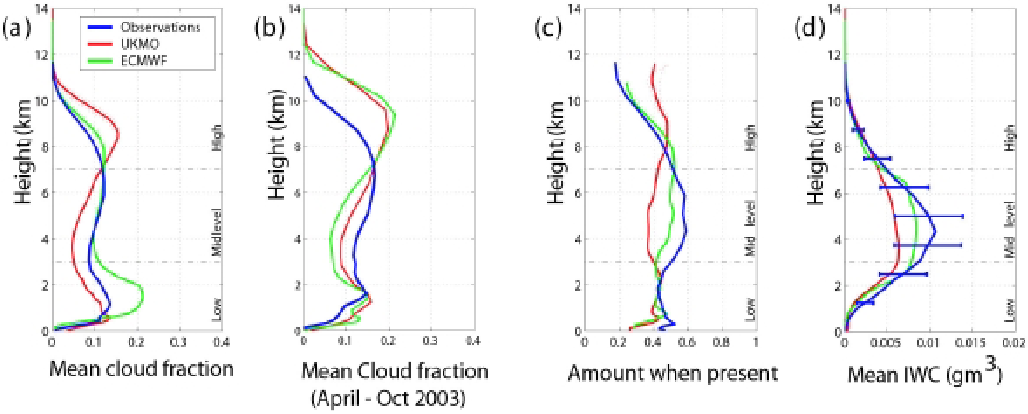

***Figure: 5.1:** Mean profiles, as observed by cloud radar and lidar at Chilbolton and output form the Met Office and ECMWF forecast models of a) Cloud fraction for May 1999 to May 2000, and b) for April-October 2003, and for the 1999-2000 period: c) Mean cloud fraction when cloud is present, and d) mean ice water content.*

The improved sensitivity of the radar in 2003, -31 dBZ at 10 km, means that it is no longer necessary to remove the high altitude low IWC cloud from the model which the radar could not detect (Figure 5.1a). This new data does however confirm that both models have too much cloud above 8 km by as much as a factor of two. EarthCARE will extend such comparisons to a global scale.

5.4 Assimilation of cloud data

Despite the importance of clouds and precipitation in the atmosphere there is no explicit analysis of clouds and precipitation in global data assimilation systems in current operational models. The first steps in preparation for the assimilation of cloud and radiation observations have been taken at ECMWF, where incremental four-dimensional variational assimilation (4D-Var) is used as an operational data assimilation system. As a part of this preparation, one-dimensional variational (1D-Var) assimilation experiments have been carried out using observations from the Department of Energy's Atmospheric Radiation Measurement (ARM) Program. Measurements of both cloud properties and radiative fluxes have been used in these feasibility studies. The ARM data sets also offer an ideal platform to study the impact of assimilation of new observations from cloud radars or lidars in preparation for upcoming satellite missions. The aim of these studies has been to show the ability of 1D-Var to modify dynamical variables, namely temperature and humidity, through employing different parametrization schemes. This is important for consistency between cloud parameters and dynamics in order not to lose the analysed cloud information within a few model time steps. Several combinations of observations were assimilated to explore the benefits deriving from profiling versus integrated measurements and from the synergy of both. Specifically the following set of experiments were performed: using cloud radar reflectivity profiles only (experiment AN1), using observations of the total column water vapour (TCWV) and the downward longwave radiation (LWD) flux at the surface only (experiment AN2) and using a combination of all the above mentioned observations (experiment AN3). The combination of observations used in AN3 was chosen to reproduce an ``upside-down'' satellite configuration such as that of EarthCARE, where the observations from the active sensors will be perfectly co-located with the passive radiation measurements. Figure 5.2 shows a comparison of the first guess (b) and the 1D-Var retrieved radar reflectivity for the AN1 experiment (c) versus the MMCR (35 GHz Millimetre-wavelength Cloud Radar) observations (a). The 1D-Var analysis is closer to the observations for most of the profiles. The best fit to cloud radar reflectivity observations is achieved when assimilating reflectivity only or reflectivity together with the surface LWD radiation flux and TCWV (not shown). Comparisons with independent radio-sonde observations conducted to examine the impact of the different assimilated observations show that the assimilation experiment when combining passive and active observations (experiment AN3) gives the lowest analysis bias at all levels with respect to radio-soundings. This indicates that the synergy of different cloud observations seems to be the most beneficial towards improving the performance of the cloud assimilation system.

ECMWF has expressed strong interest in the assimilation of EarthCARE data into its operational model. Assimilation in operational models, such as ECMWF, would require the delivery of the satellite measurements in near-real time (see section 5.2).

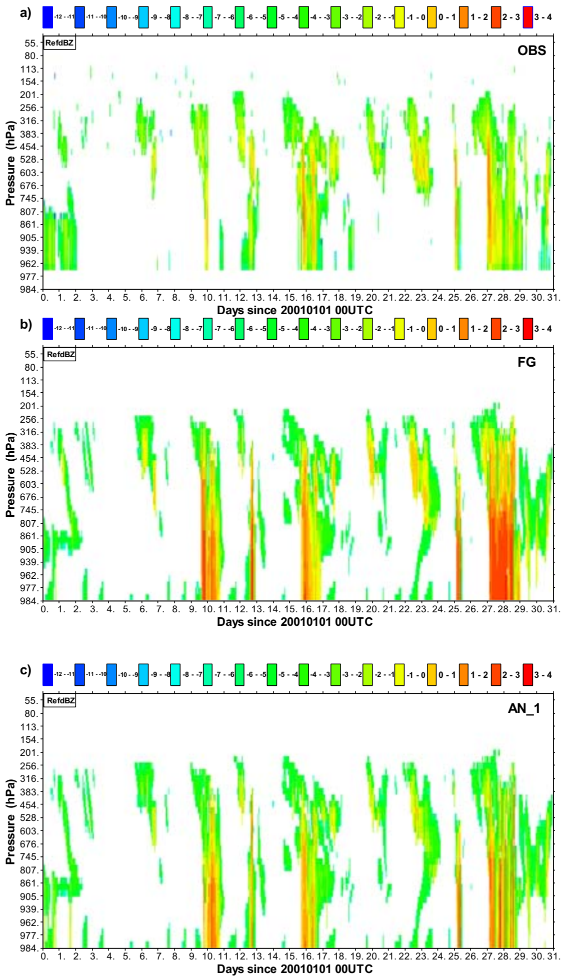

Figure 5.2*: SGP site radar reflectivity (dBZ/10) for January 2001: (a) MMCR observations, (b) model first guess and (c) 1D-Var retrieval, when assimilating reflectivity only.*

6. Summary

The observational requirements of the mission discussed above, are based on the overall scientific mission objective to quantify aerosol-cloud-radiation interactions. This leads to the described observational requirements based on the fundamental requirement to measure and being able to model the outgoing flux at the TOA with an accuracy of 10 Wm^{-2}. This again leads to the need of collocated measurements of broadband TOA radiances and detailed, global and vertically resolved measurements of cloud and aerosol abundances and composition, cloud-precipitation interactions and characterisation of vertical motions within clouds.

The required measurements can be achieved by embarking on a single platform of a high spectral resolution lidar and a Doppler millimetre wave radar complemented by passive instruments to synergistically provide:

- Direct determination of the optical depth of the aerosol in the boundary layer.
- Independently derived vertical profiles of both the backscatter and the extinction coefficient of aerosols and clouds.
- Characterisation of the shapes and properties of aerosols and ice particles in clouds from the observed ratio of lidar extinction to backscatter and the lidar depolarisation ratio.
- Accurate profiles of ice water content and ice particle size from radar/lidar and imager synergy.
- Improved quantification of liquid water clouds including determination of drizzle fluxes and detection of supercooled water layers.
- Cloud inhomogeneities on the km scale.
- PDFs of vertical velocity including sub-grid scale motions.
- Observations of condensed water mass flux in convection including that penetrating the tropopause.
- Direct observations of ice particle sedimentation velocities in cirrus.
- Accurate drizzle and precipitation rates.
- Estimation of radiative flux gradients and hence heating rates at different vertical levels.
- The ability to verify the radiative self-consistency of the derived products.

The objectives of the EarthCARE mission are at the heart of the World Climate research programme (WCRP) whose aims include the improvement of climate prediction models. In this context observations of cloud and aerosols are of utmost importance.

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Acronyms

ADM	Angular Dependence Models
AERONET	AErosol RObotic NETwork
AMIP	Atmospheric Model Intercomparison Project
ARM	Atmospheric Radiation Measurement
ATLID	Atmospheric Backscatter Lidar
BBR	Broadband Radiometer
BRDF	Bidirectional Reflectance Distribution Functions
CERES	Clouds and the Earth's Radiation Energy System
CPR	Cloud Profiling Radar
ECMWF	European Centre for Medium-Range Weather Forecasts
ERBE	Earth Radiation Budget Experiment
ESA	European Space Agency
GCM	General Circulation Model
HSRL	High-Spectral Resolution Lidar
ISCCP	International Satellite Cloud Climatology Project
IWC	Ice Water Content
LW	Longwave
LWC	Liquid Water Content
MAG	Mission Advisory Group
MERIS	Medium Resolution Imaging Spectrometer
MISR	Multi-angle Imaging Spectro-Radiometer
MRD	Mission Requirements Document
MSI	Multi-spectral Imager
NEDT	Noise equivalent delta temperature
NWP	Numerical Weather Prediction
PDF	Probability Distribution Function
POLDER	Earth Radiation Budget Scanning Radiometer
ScaRaB	Scanner for Radiation Budget
SNR	Signal-to-noise ratio
SW	Shortwave
TOA	Top of the Atmosphere
WCRP	World Climate Research Programme