Evaluation of cloud systems in the Met Office global forecast model using CloudSat data

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Abstract. Clouds and their radiative properties are still not sufficiently well represented in numerical weather prediction (NWP) and climate models. Improving their representation is a key priority, as clouds play a main role in the Earth’s radiation budget and are a key uncertainty in predictions of climate change. CloudSat is the first spaceborne millimetre wavelength radar (94 GHz) that provides estimates of cloud condensate and precipitation globally. We have developed a system to simulate CloudSat data in the Met Office Unified Model (MetUM) that is consistent with the observations. We apply this simulator to evaluate the MetUM global forecast model, as previous studies have shown that the global forecast model is a useful framework in which to analyse cloud representation errors that are relevant for both numerical weather prediction and climate time scales. These comparisons are presented from two different points of view. Firstly, we analyse the direct comparisons of CloudSat passes over a mid-latitude system and tropical convection. Secondly, a more statistically-based approach is followed by computing two dimensional height-reflectivity histograms integrated over time. This second approach allows us to study how well the vertical distribution of clouds is represented in the MetUM.

1. Introduction

The most recent work on cloud feedbacks [Bony and Dufresne, 2005; Ringer et al., 2006; Soden and Held, 2006; Webb et al., 2006] continues to show that they represent the largest source of variation in model predictions of climate sensitivity. Although some progress has been made - for example, the role of low clouds has recently been shown to be of key importance in explaining the variation of climate sensitivity across contemporary models [Bony and Dufresne, 2005; Webb et al., 2006] - it is still fair to say that this range of uncertainty has not narrowed appreciably over the last decade or so. A key reason for this is that clouds and their radiative properties are still not sufficiently well represented in climate models. As clouds have such a large impact on the Earth’s radiation budget even relatively small changes in cloud coverage, distribution, or properties can have significant impacts on the climate response to, say, greenhouse gases. Another potentially important factor is the interaction of clouds and the hydrological cycle and the extent to which cloud feedbacks are influenced by the response of the hydrological cycle to climate forcing [Stephens, 2005].

The CloudSat mission [Stephens et al., 2002] has the specific aim of providing observations necessary to improve our understanding of these issues - key goals are to evaluate clouds and cloud processes in global climate models (GCMs), and the relationship between the vertical distributions of cloud water and ice and cloud physical and radiative properties. CloudSat provides the first direct global survey of the vertical structure of cloud systems and thus provides data which allows us to evaluate the representation of clouds in climate models in much greater detail than has previously been possible. This should lead to both improved cloud parameterizations and, given the importance of cloud feedbacks, more reliable climate change predictions.

For these reasons we have developed a system which allows us to make full use of CloudSat data for the evaluation of clouds in the UK Met Office model. We adopt the model-to-satellite approach, i.e. we simulate as closely as possible the data produced by CloudSat, including taking into account the orbital path of the satellite. The benefit of this approach is that it allows a like with like comparison because it takes into account some processes that impact the radar signal, and also removes the effects of inconsistent assumptions in model and cloud retrievals. This is the approach commonly taken in numerical weather prediction and data assimilation, and is now increasingly used in climate model evaluation [Webb et al., 2001; Ringer et al., 2003].

Williams and Brooks [2007] show that cloud regime properties are found to be similar in the UK Unified Model at all forecast times, including the climatological mean, which suggests that weaknesses in the representation of fast local processes are responsible for errors in the simulation of the cloud regimes. This makes the global forecast model a useful framework in which to analyse cloud representation errors that are relevant for both numerical weather prediction and climate time scales. We apply this approach to evaluate the three-dimensional structure of hydrometeors of the UK Met Office global forecast model from December 2006 to February 2007. This paper is organised as follows. Section 2 describes the algorithm developed to simulate radar reflectivities. Section 3 gives a brief description of the model, the experimental design and the observations used in the study. Section 4 compares the simulation of a mid-latitude system and tropical convection case studies against observations. We focus on a more statistical approach in the comparisons against observations in Section 5. Conclusions are presented in Section 6.

2. Simulation of radar reflectivities in the MetUM model
This section details the methods employed to compute the different contributions to the total reflectivity and attenuation terms of the radar signal. The gamma distribution function is used in the UM for the representation of the cloud particle size distribution, its mathematical expression being:

\[ n(D) = N_0 D^\nu \exp(-\Lambda D) \]  

(1)

where \( D \) is the diameter of the particle, \( N_0 \) is the intercept parameter, and \( \nu \) and \( \Lambda \) are free parameters.

This form of the size distribution allows us to obtain the \( i^{th} \) statistical moment of the distribution analytically:

\[ \mu_i = \int_0^\infty D^i n(D)dD = N_0 \frac{\Gamma(\nu + i + 1)}{\Lambda^{\nu+i+1}}. \]  

(2)

2.1. Reflectivity due to liquid clouds

The radar effective reflectivity factor, \( Z_e \), is given by:

\[ Z_e^l = \frac{|K_i(f, T)|^2}{\rho_{\text{e}} \rho_3} \mu_0 = \frac{|K_i(f, T)|^2}{\rho_{\text{e}} \rho_3} N_0 \frac{\Gamma(\nu + 7)}{\Lambda^{\nu+7}}, \]  

(3)

where a Khrgian-Mazin distribution is used for water droplets (\( \nu = 2 \)). The factor \( |K_i(f, T)|^2 \) is a calibration factor, \( K_i(f, T) \) being the dielectric factor of liquid water at temperature \( T \) for a frequency \( f \). This calibration factor ensures that at centimetre wavelengths the reflectivity reduces to the familiar form \( Z = \mu_0 \) [Hogan et al., 2006].

Two additional equations are needed to obtain \( Z_e^l \), namely expressions for the total cloud droplet number concentration, \( N \),

\[ N^l = \mu_0 = \frac{\Gamma(\nu + 1)}{\Lambda^{\nu+1}}, \]  

(4)

and the liquid water content (LWC),

\[ \text{LWC} = \frac{\pi}{6} \rho_{\text{e}} \rho_3 = \frac{\pi}{6} \rho_{\text{e}} N_0 \frac{\Gamma(\nu + 4)}{\Lambda^{\nu+4}}. \]  

(5)

Using (4) and (5) in (3), the radar reflectivity of water clouds can be obtained as:

\[ Z_e^l = \frac{|K_i(f, T)|^2}{\rho_{\text{e}} \rho_3} \frac{201.6 \text{LWC}}{N^l} \left[ \frac{\Gamma(5)}{\Lambda^5} \right]^2 \text{[m}^3\text{]} \]  

(6)

In order to express \( Z_e \) in the more common units of dBZ, it first has to be converted to \([\text{mm}^3\text{m}^{-3}]\).

2.2. Reflectivity by ice clouds

In the case of ice clouds, using the Debye approximation [Debye, 1929] the radar reflectivity factor is expressed as:

\[ Z_e^i = \frac{|K_i(f, T)|^2}{\rho_{\text{i}} \rho_{\text{p}}} \frac{6}{\rho_{\text{i}} \rho_{\text{p}}} 2 \int_0^\infty m_i(D) n(D)dD = \frac{|K_i(f, T)|^2}{\rho_{\text{i}} \rho_{\text{p}}} \frac{6 a}{\rho_{\text{i}} \rho_{\text{p}}} N_0(T) \frac{\Gamma(5)}{\Lambda^5}, \]  

(7)

where the mass is parameterized as \( m_i(D) = a D^2 \) (\( a = 0.069 \) kg m\(^{-2}\)), and a dependence of temperature in \( N_0(T) = N_0 \exp(-0.122T) \) has been introduced [Wilson and Ballard, 1999]. Therefore, only one additional moment of the distribution is needed to compute \( Z_e^i \). This is obtained from the definition of ice water content (IWC):

\[ \text{IWC} = \int_0^\infty m_i(D) n(D)dD = a \mu_2 = a N_0(T) \frac{\Gamma(3)}{\Lambda^3}. \]  

(8)

Eliminating \( \Lambda \) between (8) and (6),

\[ Z_e^i = \frac{|K_i(f, T)|^2}{0.93 \rho_{\text{i}} \rho_{\text{p}}} \frac{\rho_{\text{i}} \rho_{\text{p}}}{N_0(T)^2} \left[ \frac{\alpha}{N_0(T)^2} \right]^{1/3} \text{IWC}^{5/3}. \]  

(9)

The dielectric factor of solid ice, \( |K_i|^2 \), is independent of the frequency and temperature with a value of 0.174. A non-Rayleigh scattering correction is applied following the methodology proposed by Benedetti et al. [2003].

2.3. Reflectivity by precipitation

For large particles, the radar reflectivity factor no longer follows the sixth power law in (3), and therefore a different approach must be followed. For a given frequency, the reflectivity factor is proportional to the reflectivity \( \eta \), which is the total backscatter cross-section of all scatterers in unit volume. We have used a lookup table of scattering cross-sections for spherical particles to develop a parameterization of \( \eta = \eta(P; T) \), where \( P \) is the precipitation water content (liquid or ice). The lookup table for \( f = 94\text{GHz} \) was computed using the Mie codes of Barber and Hall [1990], integrated over a Marshall-Palmer distribution for different rainrates and temperatures [M. Rogers, personal communication]. The parameterizations have been obtained from non-linear least squares fits to equations of the form \( \eta = E_1(T) P^{E_2(T)} \). Temperature was sampled at 19 values, representative of the values observed in the troposphere. This approach is followed both for rain and snow precipitation.

2.4. Attenuation

The reflectivity factors from the previous subsections were derived under the assumption that the space between the radar antenna and the target was a vacuum. However, at mm wavelengths the radar signal can be attenuated due to the interaction by several atmospheric components: absorption by gases (\( \text{H}_2\text{O} \) and \( \text{O}_3 \)), and extinction by water droplets and precipitation. Therefore, the reflectivity factor corrected by attenuation will be:

\[ Z_e = Z_e \exp(-2 \sum_i A_i), \]  

(10)

where \( A_i = \int_{\text{path}} \sigma^\text{e}(s)ds \) is the integral extinction along the portion of atmosphere between the antenna and the target for component \( i \). The factor 2 accounts for the two-way attenuation as the signal travels twice through the same atmospheric mass. Gaseous attenuation is accounted for by using the model proposed by Liebe [1985]. In the lower troposphere, attenuation by gases ranges from \( \approx 1 \text{dB} \) in cold and dry atmospheres to \( \approx 6 \text{dB} \) in tropical atmospheres.

For the computation of the attenuation by precipitation, we have followed an approach similar to that used in the computation of the reflectivity. The extinction coefficient \( \sigma^\text{precip} \) is modelled from the data of lookup tables. Attenuation by rainfall is very strong, and can be of the order of tens of dBZ in heavy precipitation. Attenuation by snowfall is negligible.
Attenuation by clouds is mainly caused by absorption by liquid droplets, and can be of the order of 5 dBZe. The scattering effect is much smaller at radar wavelengths. The extinction coefficient is obtained by its Rayleigh approximation, neglecting scattering,

\[ \sigma_{\text{cloud}} = \frac{6\pi Im(K)}{\lambda^2} LWC. \] (11)

2.5. Sub-grid sampling

Cloud structure at the sub-grid scale is parameterized by large-scale models. Models show large biases if only vertical profiles of grid-box mean cloud values are used in radiation calculations [Cabalan et al., 1994; Barker and Räisänen, 2005]. To overcome this, large-scale models employ cloud overlap assumptions, which influence the fluxes of radiation and precipitation through the atmosphere [e.g. Ritter and Geleyn [1992]]. When simulating satellite data in large-scale models, this sub-grid scale variability has to be included and treated in a consistent manner.

The scheme that we use to introduce subgrid variability in the computation of the radar reflectivity is the Subgrid Cloud Overlap Profile Sampler (SCOPS), a technique developed for the International Satellite Cloud Climatology Project (ISCCP) simulator [Klein and Jakob, 1999; Webb et al., 2001]. Each grid box is divided into a number of vertical columns (of the order of tens to a few hundreds) and a subgrid distribution of clouds is generated within the model grid box [Klein and Jakob, 1999]. Webb et al. [2001] developed SCOPS which uses a pseudo-random sampling process, fully consistent with the max, random and max/random cloud overlap assumptions used in many models. As rain drops and snow particles produce a strong signal at radar frequencies, they also have to be taken into account in the sub-grid sampling algorithm. We have developed a simple algorithm that provides sub-grid distribution of precipitation fluxes compatible with the cloud distribution output by SCOPS and the grid box mean precipitation fluxes simulated by the model.

Once this sub-grid sampling has been carried out, the outputs can be aggregated to produce a final product at the original (grid-box) resolution for visualization. Because of this sub-grid vertical structure and the fact that radar reflectivity is a non-linear function of cloud/precipitation condensate, the aggregated reflectivity will not be the same as the one calculated from grid-box mean profiles. All the statistical analysis carried in this paper uses the reflectivities at sub-grid resolution.

3. Model description, experimental design and observations

The model used in this study is the MetUM global forecast model at cycle G42, operational from December 2006 to May 2007. The horizontal resolution is N320, 0.5625 degrees longitude by 0.375 degrees latitude (40km in mid-latitudes) and 50 levels in the vertical with the model top at around 63km (0.01 hPa). The dynamical core is a two-time level semi-implicit, semi-Lagrangian (SISL) formulation and is also non-hydrostatic [Davies et al., 2005]. The physical parameterizations are as follows: radiative transfer is based on the two-stream equations in both the long-wave and short-wave spectral regions [Edwards and Slingo, 1996] and includes a treatment of the effects of non-spherical ice particles, multiple scattering between cloud layers, and the effect of CO2, H2O and O3 as well as the trace gases O2, N2O, CH4 and CFC11 and a background aerosol climatology; the cloud microphysical scheme [Wilson and Ballard, 1999] is based on Rutledge and Hobbs [1983] and models the transitions between water vapour, liquid, ice and rain with the cloud ice content a prognostic variable within the model, rather than a diagnosed quantity from the cloud scheme. The cloud scheme based on Smith [1990] has been modified to diagnose only the cloud liquid water contents. In addition a revised cloud area parameterization is also used which allows clouds to fill only part of the vertical thickness of a model layer [Brooks et al., 2005]; boundary layer turbulence allows for non-local mixing in unstable regimes [Lock et al., 2000, Martin et al., 2000] and is based on a similar approach by Holtslag and Boville [1993]. An explicit boundary layer top entrainment parameterization is also used [Lock, 1998]; convection is based upon the mass flux scheme of Gregory and Rowntree [1990] including convective downdrafts [Gregory and Allen, 1991], a parameterization to treat the radiative effects of anvil cirrus in deep convective systems [Gregory, 1999] and convective momentum transports. The cloud base closure for shallow convection is based on Grant [2001] and that for deep uses the CAPE closure of Frischt and Chappell [1980]; gravity wave drag is described in Webster et al. [2003] and a parameterization of orographic roughness is also applied [Milton and Wilson, 1996].

The forecast methodology is similar to that used in previous comparisons of the NWP model with Geostationary Earth Radiation Budget (GERB) data [Allan et al., 2005]. Forecast model diagnostics were produced every 3 hours, by a two-timestep forecast run from each of the four analyses per day (00,06,12, and 18 UTC) generated from the four dimensional variational assimilation scheme of Rawlins et al. [2007] and the T+3 forecast states. Using short-range NWP forecast has the advantage, for the purpose of evaluating the model parameterisations against observations, that the large-scale atmospheric circulation is represented as accurately as possible within the limits of a modern data assimilation system. This enables a less ambiguous attribution of any discrepancies found to the model parameterisations rather than their inputs. Model outputs are sampled along the orbit track, choosing the model time that is closest to the observation time. As the model output frequency is 3-hourly, the time mismatch between model and observations is smaller than 1.5 hours. A sensitivity test was carried out by increasing the output frequency of model data to minimise the time mismatch, and it did not produce any significant impacts on the results.

CloudSat and the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIOP) were successfully launched on April 28, 2006. Both satellites were put in nearly identical sun synchronous orbits at 705 km altitude within the A-Train [Stephens et al., 2002]. CloudSat carries the first millimeter wavelength cloud profiling radar (CPR) in space, which operates at a frequency of 94 GHz [Im et al., 2005]. The CPR points in the nadir direction, and its pulses sample a volume of 480 m in the vertical, with a horizontal resolution of 1.4 km across-track. The first CloudSat products [Stephens et al., 2002] were released on October 16th, 2006. Among these products is the CloudSat 2B-GEOPROF dataset, which provides the radar reflectivity, in dBZ, and identifies where hydrometeors occur (‘cloud-mask’) [Mace et al., 2007]. The primary instrument on board CALIPSO is the Cloud-Aerosol Lidar with Orthogonal Polarization (CALIOP), the first polarization lidar in space, operating at 532 nm and 1064 nm [Winker et al., 2007]. CALIOP...
is nadir-pointing with a beam diameter of 70 meters at the Earth’s surface, and a pulse repetition frequency that produces 533 m in the along-track direction. CALIPSO is able to detect thin cloud layers with optical depths of 0.01 or less, provided that the signal is averaged along-track [McGill et al., 2007]. CALIPSO products are described by Vaughan et al. [2004]. Among these products, the Vertical Feature Mask provides a target classification that gives information on the location and properties of aerosol and cloud layers.

4. Case studies


Figure 1a shows the Met Office surface analysis chart for the North Atlantic and part of Europe for February 26th, 2007 at 1200 UTC. A depression (966mb) is located East of Labrador and south of Greenland, between 50°N and 60°N. The system is in its mature stage, with an occluded front in the core, the cold front extending southward from 60°N with the warm front ahead reaching the Bay of Biscay. There is an occluded front that wraps the system from the north, crossing the Atlantic from Terra Nova, passing near the south coast of Greenland and finishing in the south west coasts of Ireland. This occluded front is a remnant of a depression that was located off the coast of Labrador 36-24 hours before. The A-Train passed over this system at 14:15 UTC, and the red line in Figure 1a shows the approximate track followed by CloudSat in its ascending node, travelling from South to North. CloudSat passed over the warm front of this depression and over the occluded front remaining from the previous system. Figure 1b shows the RGB composite from MODIS which shows the large scale features of this system. MODIS observes cloud most of the time, except at the ends of the transect. During the first third of the transect the cloud is thinner than in the middle of the transect, when CloudSat crosses the frontal systems.

The radar reflectivities as simulated from the model variables along that transect are shown in Figures 2a (sub-gridbox resolution) and 2b (gridbox resolution). Figure 2a illustrates the impact of the sub-grid sampling. Generally, in mid-latitude systems the greatest impact occurs in the core of the system, where convection is more likely to occur. The dotted contour lines show the atmospheric temperature as simulated by the model, with the 0°C isotherm plotted with a solid line. As expected, a South-North gradient in temperature is observed with warmer temperatures in the South. On top of this background gradient, some discontinuities in the slope of the isotherms indicative of transitions between warm and cold sectors can also be identified. Overall, the model represents the 3D structure reasonably well, although there are noticeable differences from the observations (Figure 2c). We divide this transect in three sectors.

The first sector goes from the origin of the transect to the warm sector at around 1400 km from the origin. The first part of this sector (0-500 km) is dominated by high cloud in the model simulations, with ice cloud with top at around 11 km height and initially ≤3 km depth that deepens as we move towards the warm front. This cloud deck in the first part of the transect is not seen by CloudSat.

Figure 2d shows the target classification provided by the Vertical Feature Mask product from CALIPSO. As CALIPSO is able to detect thinner clouds than CloudSat, this helps to identify whether these cirrus clouds are spurious generated by the model or present in reality. Although CALIPSO (Fig. 2d) sees high cloud in this first part of the transect, it is thinner than the cloud simulated by the model and higher, above 10 km. The second part of the first sector (500 to 1000 km) is characterised by a multilayered cloud structure, with a high cloud layer between 9 and 12 km that is captured by CALIPSO but mostly missed by CloudSat. Underneath this layer, there exists a second layer, between 5 and 7 km that is clearly visible by both CloudSat This mid-level cloud shows reflectivity factors between -20 and 0 dBZ, and it attenuates the lidar signal most of the time. These two effects combined are an indication of mixed phase cloud [Hogan et al., 2003]. The model produces cloud in this part of the transect, continuously from around 5 km up to 11 km, therefore not capturing the multilayered structure seen in reality. The model also shows a strong signal in the radar reflectivity factor below the freezing level that is not seen in the CloudSat measurements. This signal comes from drizzling boundary layer clouds. We cannot be certain whether or not these clouds are present in reality. The ability of CloudSat to detect boundary layer clouds is limited, as these clouds consist of small droplets and the radar reflectivity depends on the sixth power of the radius. In this case, CALIPSO does not provide too much help as the signal at these levels has been attenuated by the mixed phase cloud above. In those regions where the lidar signal is not attenuated, CALIPSO finds no evidence of cloud, except in the first hundreds of meters above the surface. In any case, the production of cloud by the model is spurious as it would be detected by CloudSat. The third part of the first sector (1000 to 1400 km) is characterised by the presence of hydrometeors from around 13 km (seen by CALIPSO) down to the ground. The CloudSat signal below 7 km seems to be dominated by precipitation, which sometimes causes strong attenuation at low levels. The freezing level can be clearly seen in the CloudSat signal in this part of the transect dominated by precipitation, with an abrupt increase in reflectivity with respect to the values just above the freezing level (bright band). The origin and characteristics of the 94 GHz radar bright band (or its non-existence in non-precipitating clouds) is a topic of research currently under discussion in the literature [Sassen et al., 2005; Kollidas and Albrecht, 2005; Sassen et al., 2007].

The second sector corresponds to the stretch between the two fronts, from 1400 to 1800 km. The vertical structure of the system is well represented, although with reflectivity values that are typically 5-10 dBZ smaller than the observations. Reasons for this are unclear, although several factors might contribute:

- lack of cloud cover: the large-scale scheme produces cloud fraction 1 (or close) most of the time in these regions, so this cannot explain the differences.
- simulation of radar reflectivity: typical values of simulated T and IWC in that region are -30°C and 0.15 gm⁻³, respectively. Using the empirical formula given in Table 2 of Hogan et al. [2006] for a 94 GHz radar, a reflectivity factor of -0.6 dBZ is obtained. This is quite close to the simulated values, which are around 0 dBZ, and therefore gives confidence in the non-Rayleigh scattering parameterization of Benedetti et al. [2003].
- lack of IWC: Hogan et al. [2006] show that the simulated ice water content by the mesoscale version of the model agrees quite well with radar retrievals from the Chilbolton 3GHz radar, particularly in the region between -30°C to -10°C. For clouds colder that -30°C, Hogan et al. [2006] find that the mesoscale model underestimates the ice water content by a factor of 2. However, the global forecast model seems to behave differently to the mesoscale in the simulation of IWC. The Cloudnet project provides systematic evaluation of cloud profiles in seven forecast models using ground observations (radar, lidar and microwave radiometer) [Hillier et al., 2007]. Cloudnet results for December 2006 over Chilbolton, UK show that the MetUM global forecast model generally underestimates IWC by approximately a factor of 2 or more. This underestimation of simulated IWC can account for more than 6 dBZc, which supports the interpretation that the lack of IWC is the most likely cause, in contrast with the results of the mesoscale model.
In the third sector of the transect, from 1800 km to the end, the CloudSat observations and model forecasts agree quite well in the height and extent of the cloud deck, with CALIPSO showing thin cirrus extending a bit northward of those simulated by the model. CALIPSO also sees an aerosol layer near the surface and boundary layer cloud, the latter being properly simulated by the model. This low cloud is present in part of the second sector as well.

4.2. Tropical system: 23 December 2007

We now focus on the analysis of a transect along a region of tropical convection over the Indian Ocean. Figure 5 shows the MODIS RGB composite simultaneous with the CloudSat overpass (0820 UTC) along a transect across a region of tropical convection in the Indian Ocean in which clouds are spatially organised in convective systems with varying scales. During the first half of the transect (bottom of figure), CloudSat passes over several small-scale convective cells, and in the second half of the transect it crosses a large-scale system with a typical scale of ≈ 500 km. Before crossing over this large-scale system, CloudSat passes over a small region of very intense convection, visible as a bright white region in the center of the picture. Allan et al. [in press] show that convection in the model is less spatially organised in the model than in reality. This contrasts with the good agreement in the two-dimensional spatial structure of mid-latitude systems found in previous studies [Allan et al., 2005], which is extended to the vertical dimension in this analysis.

The comparison of the simulated reflectivities against the observed reflectivities is shown in Figure 4. The impact of the sub-grid sampling is even more evident in this example as convective cloud plays a greater role. It can be seen that the model has difficulty capturing the spatial structure of the tropical convection. It partially captures the position of the large-scale convective system between 1500 and 2000 km from the origin, but this extends too far to the South in the model simulations. The CloudSat observations show the presence of a region of very intense convection that was suggested by MODIS imagery, between 1400 and 1500 km from the origin of the transect. The cloud extends above 17 km (it is labelled as stratospheric in the CALIPSO mask), and the whole vertical structure is characterised by a strong radar reflectivity at all levels above 5 km, with strong attenuation by rainfall below the freezing level. The model simulates a continuous cloud layer between 5 and 12 km which is unrealistic. This cloud deck has contributions from both the large-scale convective cloud. The observations show shallower and more broken layer. The observed high-level cloud often shows a multilayered structure which is completely absent in the model forecasts. These structures are characterised by small-scale intense precipitation events generated by mid-level convection, with cloud tops between 3 and 7 km. This type of structure is consistent with the mid-level mode of tropical convection reported by Johnson et al. [1999]; Haynes and Stephens [2007].

5. Statistical analysis

We now focus on the vertical structure of hydrometeors from a statistical perspective, first globally and then in four oceanic regions. To do this, we construct joint height-reflectivity histograms in 2.5° by 2.5° regions. We divide the vertical axis of these histograms in bins of 1 km, and the horizontal axis in bins of 2.5°. For the case of CloudSat data, the counter is increased when the CPR cloud mask of the 2B-GEOPROF product is greater or equal than 20, which gives an estimated false detection smaller than 5%. Cloud fraction in each layer is then defined by the number of positive identifications with reflectivity greater or equal than -27.5 dBZ divided by the total number of measurements in that layer.

Figure 5 shows observed and simulated hydrometeor fraction curves. We show every other 1 km layer, starting above 1 km as CloudSat measurements within the lowest kilometer are affected by ground contamination [Mace et al., 2007]. The geographical pattern of hydrometeor fraction changes significantly with height. At low levels, the midlatitude regions associated with the storm tracks show high hydrometeor fraction, with typical values greater than 0.5. At these levels, the hydrometeors detected by CloudSat are concentrated in the ITCZ in the tropics, and in the subtropical oceanic regions characteristic of transition between Cu and Sc cloud. Although there are large regional differences in the hydrometeor amount, there are only a few regions with hydrometeor fraction close to zero, and some of these regions correspond to the Sc regions of the eastern oceanic basins, with non-precipitating clouds that are below the sensitivity of CloudSat. At higher levels (≈ 3-9 km), the maximum hydrometeor fraction decreases, with typical values in the storm tracks smaller than 0.4. At mid-levels, the absence of cloud above the inversion in the subtropical subsidence regions is noteworthy. The geographical pattern of these regions with no cloud is very different in the southern and northern hemispheres. Whereas in the northern hemisphere the lack of cloud is almost ubiquitous between 10°N and 90°N, in the southern hemisphere the cloud cover is located primarily in the eastern basins of the subtropical oceans. Above 10 km, hydrometeors concentrate in the deep convective regions: the ITCZ, the south Pacific convergence zone (SPCZ), the west Pacific warm pool, central Africa, the Indian Ocean and South America. A small but non-negligible amount is observed above 15 km in central Asia, the west Pacific and South America.

The pattern of the vertical distribution of hydrometeors simulated by the model is similar to the observations, although there are notable differences. Less hydrometeor is found at midlatitudes below 5 km, and the presence of hydrometeor in the lower levels of tropical deep convection regions is clearly underestimated compared to the observations. This underestimation at lower levels switches to overestimation above 8 km. Deep convective cloud does not reach levels as high as observed, with a very small hydrometeor fraction above 14 km.

Figures 6a and 6b show the zonally averaged hydrometeor fractions. The lack of hydrometeor in the storm tracks below 6 km is marked. This contrasts with the comparisons between HadGAM1 and ISCCP-D2 from 45°S to 60°S, where HadGAM1 produces more low cloud than the observations [Martin et al., 2006]. CloudSat shows a clear asymmetry in hydrometeor fraction between the two descending branches of the Hadley circulation. In the winter hemisphere descending branch, where the Hadley circulation and its associated subsidence strength is stronger [Peixoto and Oort, 1992; Källberg et al., 2005], the presence of hydrometeor is negligible in almost the entire vertical column, with only values near 0.1 below 3 km. The model seems to capture this asymmetry in the circulation, although due to its inability to produce mid-level cloud, this asymmetry is not so evident in the cloud distribution. On the other hand, the convection scheme detains too much moisture at high levels, therefore producing too much cloud at these levels in the tropics.

Area-weighted joint height-reflectivity histograms are shown in Figures 6c and 6d. The observations show a triangular region in the 2D space. This region is limited in its left-hand side by the sensitivity limit of the CPR (approx. -30 dBZ). There seems to be a linear relationship between maximum reflectivity and height, with increasing maximum reflectivity with decreasing height, with an approximate slope of -0.25 km/dBZ. Low levels, below 3 km,
seem to show a slightly bimodal distribution, with a peak around -25 dBZ, and a second maximum near 5 dBZ. The trend of these two maxima is smooth (low cloud - drizzle - rainfall), with highly populated bins in between. The lowest 1-km layer is not shown in the CloudSat plot due to the effect of the contamination by ground clutter. The model shows a completely different picture, sampling three main preferred regions of the height-reflectivity space. In the mid and high levels, the model explores a much smaller main preferred region of the height-reflectivity space. In model shows a completely different picture, sampling three regions, with a peak around a much tighter height-reflectivity relationship. At lower levels, the model seems to operate in two regimes, one with non-precipitating cloud (reflectivities ≈ -30 dBZ), and the other for precipitating cloud (reflectivities ≈ 5 dBZ).

5.1. North Atlantic

In the zonally-averaged hydrometeor fraction for this region (Figures 8a and 8b), CloudSat shows a high hydrometeor fraction in the layer between 1 and 2 km at all latitudes, although a positive gradient with latitude is observed, with more hydrometeor frequency at high latitudes. Between 2 and 6 km there are two regimes, with a boundary a little north of 50°N. At lower latitudes the hydrometeor fraction has values close to 0.2, whereas at higher latitudes the fraction is typically greater than 0.3, reaching values above 0.4 in some areas. This region of high hydrometeor fraction north of 50°N is located in the region where the Icelandic low sits in the cold season Sterrce et al. [1997]. Cloud tops reach altitudes up to 11 km on the southern side of this region and up to 10 km on the northern side. Above 6 km, the model shows a latitudes-height hydrometeor distribution similar to the observations, with the cloud top height well modelled, both in the maximum height and in the south-north gradient. The hydrometeor fractions are also similar to those observed at these heights. However, below 6 km, there are clear differences. In the layer between 1 and 2 km the model shows less hydrometeor occurrence and there is no latitudinal dependence. The influence of the core of the Icelandic low is only seen between 2 and 6 km, where two regimes exist: one south of 50°N, with small values of hydrometeor occurrence, and another north of 50°N with high hydrometeor occurrence.

The overall structure of the area-weighted joint height-reflectivity histograms simulations is similar to that of the observations (Figures 8c and 8d), with three main clusters. Both the observations and the model show a height-reflectivity relationship for ice clouds, with the model showing less spread in the reflectivity values. The split into two regimes at lower levels is again clearly visible in the model histogram. Analysis of case studies suggests that the lack of overlapping between the cloud and precipitation distribution is caused by excess of drizzle production, particularly in the warm sector of mid-latitude systems.

Figure 8e shows the area average of hydrometeor fraction as a function of height. The grey shading has been obtained by using a threshold in Z, modified by +/−1 dBZ, to test the sensitivity of the profiles to the reflectivity threshold used. This also serves as a measure of the impact of the calibration error in the observations. The observations show a monotonic decrease of hydrometeor occurrence between 3 and 5 km where the profile is almost constant with height. The model agrees well with the observations above 7 km, and underestimates the occurrence below that level.

5.2. Californian stratocumulus

We now focus on the stratocumulus region off the coast of California. The zonal cross sections in figures 9a and 9b display a similar height-latitude structure in the observations and model simulations. They show the intrusion of high-level cloud from the south, likely cirrus cloud advected from the deep convective systems in the ITCZ. This seems to be the main source of cloud of the cluster observed above 5 km in the height-reflectivity histograms. Both figures show that the low-level cloud deck is present at all latitudes in that region, with frequency of occurrence decreasing when latitude increases. The model seems to underestimate the hydrometeor occurrence between 1 and 2 km, particularly in the north half of the region.

The observations show a reflectivity-height histogram with only two clusters (Figure 9c), one that comprises clouds above 5 km height, and another for hydrometeors [cloud and...
drizzle) below 3 km. The upper cluster corresponds to cirrus cloud, which are common in that region during DJF [Wylie and Menzel, 1999], while the clusters at higher levels corresponds to the stratocumulus deck and associated drizzle characteristic of the eastern basins of the subtropical oceans. The model also shows the upper level cluster (Figure 9d), although with less spread in the reflectivities for a given height, showing a very linear relationship between height and reflectivity, as mentioned above. The model produces two clusters in the lower levels, probably indicating an abrupt transition between non-drizzling cloud (reflectivities \( \approx 30 \text{ dBZ} \)) and drizzling cloud (reflectivities \( \approx 5 \text{ dBZ} \)). The fact that the reflectivities for drizzling cloud span a small range of high values implies that the model is producing too much drizzle. O'Connor et al. [submitted] show that the overestimation of drizzle flux at cloud base is a common feature in several models.

As in the north Atlantic region, the model reflectivity against height histogram shows a strong relationship between reflectivity and height for the higher cloud. This is a constant that we observe in all the histograms, as we shall show below. The main cause of this relationship may be the \( N_{occ} = N_{occ}(T) \) relationship that the particle size distribution function uses. In reality there is considerable variation in the particle size distribution for a given ice water content and temperature [Field, 1999; Field et al., 2005], and this would diffuse out the relationship seen in the model. In order to get this variation from the model a dual moment scheme would be needed, that is, a scheme that predicts ice number concentration as well as ice water content. However, the general agreement in the ice cloud is very good.

This region is an appropriate example to show the complementary information given by the cross-sections and the height-reflectivity histograms. The zonal cross-sections look very similar, indicating that the model is putting clouds at the correct places, which may indicate that the meteorology is well captured. However, the height-reflectivity histograms look much more different, suggesting that the physical properties of those clouds are not simulated so well.

The average profile of hydrometeor fraction as function of height (Fig. 9e) clearly shows the two modes observed in this region. A mode of high cloud with maximum frequency is present at \( \approx 10 \text{ km} \). The second mode occurs below 2 km, with its maximum in the first kilometer. The model captures the high-level mode reasonably well, although it is a little too shallow, which causes an underestimation of hydrometeor at mid-levels. The frequency of occurrence of the lower mode is underestimated in the model simulations.

5.3. Hawaii trade-cumulus

The zonal cross sections in Figures 10a and 10b seem to indicate that the origin of the high-cloud cluster in the height-reflectivity histograms is of different origin than in the Sc region. In this case, the main proportion of cloud at those levels comes from intrusion from the north side of the region, suggesting that the origin of this cluster is the intrusion of dissipating frontal cloud from extra-tropical cyclones in the north Pacific. There are also some differences in the low-level cluster. The model underestimates the hydrometeor occurrence by 5-10% between 2 and 3 km. The height-reflectivity histograms (Figures 10c and 10d) display very similar characteristics to the ones obtained for the stratuscumulus region, with two clusters in the observations, one below 3 km height and another between 6 and 13 km approximately. Again, the model shows a linear relationship height-reflectivity for the high cluster, and splits the low-level cluster into two distinct regimes, namely, non-drizzling and drizzling/precipitating cloud.

The mean vertical profile of hydrometeor occurrence (Figure 10e) presents a similar picture to the one observed in the Sc region, with two modes and the model underestimating mid-level cloud and the low-level mode. A significant difference with respect to the Sc region is the abrupt decrease in hydrometeor occurrence in the lowest level of the model simulations. This decrease is consistent with the fact that the boundary layer deepens as the stratuscumulus are advected by the trade winds from the coasts of California into the Hawaii trade-cumulus regions [Wood and Brehterton, 2004]. This deepening is associated an increase in SST and decrease in subsidence, that causes the stratuscumulus layer to become decoupled from the surface. The stratuscumulus layer eventually dissipates as the trade wind cumulus boundary layer develops, causing an increase in the cloud top and bottom heights [Lock et al., 2000; Martin et al., 2000]. This reduces the cloud frequency of occurrence in the first kilometer, moving the maximum upwards, between 1 and 2 km.

The evolution in hydrometeors linked with changes in the characteristics of the marine boundary layer is observed in the meridional mean frequency of occurrence for the joint Californian Sc and Hawaii trade-cumulus regions, displayed in Figure 11. Cloud in the eastern part of this region, near the coast of California is more frequent in the bottom kilometer. This maximum in frequency of occurrence changes to higher levels as the trades advect clouds to the west. The model represents very well the location of the maximum of occurrence in the second kilometer, although the observations show a greater frequency of hydrometeors between 150°W and 180°W than the model simulates. Due to data contamination by ground clutter, we cannot evaluate the representation of hydrometeors in the first kilometer. The frequency and spatial structure of high-level cloud is very well represented by the model. The Californian stratuscumulus region (right-hand halves of each plot) shows a greater frequency of occurrence of cirrus cloud. The differences in the frequency of occurrence of cirrus between both regions seems to be related with the origin of these cloud. As it has been mention above, the zonal mean cross-sections (Figs. 9a, and 10a) suggest different origin of the cirrus in these regions.

5.4. Tropical warm pool

The zonal mean cross-sections for the tropical warm pool (Figures 12a and 12b) emphasize what was mentioned in the global picture. The observations show a maximum at 11°N and 5°S, and show a high frequency of occurrence all the way down to the surface. North of this region dominated by deep convection, the impact of cirrus cloud with less coverage underneath is also observed. The model produces the maximum at \( \approx 11 \text{ km} \) height and the cirrus cloud extending North. However, it produces too little cloud below 7 km.

The reflectivity-height histograms for the tropical warm pool region are shown in Figures 12c and 12d. These histograms share some common characteristics with the ones shown above, but they show some interesting and specific features. In this case, the observations show two different modes at low levels with two distinct maxima for reflectivities \( \approx -27 \) and \( 0 \text{ dBZ} \), although the disjoint character of these two low-level distributions is still more pronounced in the model simulations. The inherent lack of spread in the height-reflectivity linear relationship for ice clouds is also present in the model simulations. The observations seem to connect this high-level branch with the cluster of precipitation at low-levels. For the model, there are two modes of reflectivity at around 5 km. The first mode is high reflectivity (\( 5 \text{ to } 10 \text{ dBZ} \)), this is seen in the observations to a much greater extent than in the model. Presumably this relates to the lack of mid level cloud in the model. The second mode is the rapidly reducing reflectivity as we descend from about 6-7 km to 5 km. We interpret the high cloud
region of the histogram as precipitating ice. The strength of the signal grows as height decreases due to aggregation that produces larger particles. Once the falling ice reaches the freezing level, located at ~5 km in the tropical atmosphere, it melts and falls as rain. Attenuation in heavy rain-fall is significant, and explains the decrease in reflectivity as we move downwards below the freezing level. The frequency of occurrence increases below 5 km, which seems to be consistent with the existence of a second precipitation mode from clouds with tops below 5 km [Hagymás and Stephens, 2007]. The model does not seem to capture this process. Instead, there is a branch in the model histogram that connects cloud at around 9-10 km with low-level cloud in the bottom-left corner of the histogram. The temperature at the levels where this branch bifurcates is between -30°C and -40°C. The model incorporates a parameterization that partitions the convective cloud condensate as function of temperature [Bower et al., 1996], allowing for supercooled water in the temperature range between 0°C and -40°C. The relative fraction of liquid water with respect to ice decreases as temperature decreases. A given amount of liquid water will produce a smaller reflectivity than the same amount of ice because the liquid droplets are smaller. This provides a plausible explanation for the existence of this spurious branch in the model simulations, which would imply that the Bower et al. [1996] parameterisation underestimates the amount of ice at high temperatures in the mixed-phase region. A second and minor contribution to this branch comes from the large-scale cloud: ice cloud that is evaporating as it falls into dry air beneath an anvil cloud. Evaporation of ice that falls into dry air is not fast enough in the model, and this is due to the vertical resolution of the model being insufficient to maintain a dry enough layer beneath the cloud [Wilkinson, 2007].

The mean vertical profile of hydrometeor fraction is shown in Figure 12c. The lack of hydrometeors below 7 km is clearly seen in this plot, underestimating their occurrence by a factor of 3. Between 8 and 10 km height, the model overestimates the hydrometeor occurrence, although the location of the maximum (~11.5 km height) is well captured. Above 13 km, the model hardly produces any cloud, whereas CloudSat still observes a significant amount of cloud.

6. Conclusions

We have developed a system allows us to make full use of CloudSat data for the evaluation of clouds and precipitation in numerical models, and applied it to the UK Met Office global forecast model. We simulate as closely as possible the data produced by CloudSat, being consistent with the microphysical assumptions used in the model, and taking into account the spatio-temporal sampling of the satellite orbit.

This work illustrates this approach and documents some aspects of the behaviour of UK Met Office global forecast model:

- The model shows a good overall representation of vertical structure of mid-latitude systems, with high-cloud top height very well captured.
- These results suggest that the model underestimates the ice water content in frontal systems by a factor of 2 or more.
- Tropical convection case studies show that tropical convection is not well simulated. A possible interpretation for the differences in the performance between mid-latitudes and the tropics is that, in the tropics, the data assimilation system does not provide a hard constraint as convection is sub-grid scale and therefore difficult to initiate in the right place at the right time [R. Allan, personal communication]. This seems to be supported by the fact that large-scale tropical systems, like typhoons, seem to be better captured (not shown).

- The low-level distribution of hydrometeor reflectivities is strongly bimodal, with a non-drizzling cloud mode and a drizzling mode clearly separated, independently of the geographical region in the tropics. This suggests that the model is producing too much drizzle, confirming on a global basis what recent ground-based measurements have shown [O'Connor et al., submitted].

- Global underestimation of hydrometeors below ~7 km. The underestimation is greater at mid-levels in deep convective cloud, consistent with the lack of detrainment of moisture of the convection scheme at mid-levels. This probably has the side effect of producing too much cloud at high levels, where most of the moisture is detrained [Maidens and Derbyshire, submitted].

- Underestimation of deep convective cloud top height in the tropics, where the model shows cloud top heights that are 1-2 km lower on average than CloudSat.

The evaluation of clouds in the global forecast model is not only relevant in the numerical weather prediction framework, but also for climate studies. Williams and Brooks [2007] show that cloud regime properties are found to be similar in the UK Unified Model at all forecast times, including the climatological mean, which suggests that weaknesses in the representation of fast local processes are responsible for errors in the simulation of the cloud regimes.

This approach can be extended to the evaluation of climate models. In the context of the Cloud Feedback Model Intercomparison Project (www.cfmip.net), several climate modeling centers (Hadley Centre, LMD/IPSL, LLNL, CSU) have joined together to develop a community ISCCP/CloudSat/CALIPSO simulator designed to be easily plugged in to numerical models, from high-resolution models to climate models. This community simulator has a similar structure to the one presented here, but QuickBeam [Hagymás et al., in press] is the subcomponent that simulates the radar signal. The code used to simulate the lidar signal is an evolution of the one developed by Chiriaco et al. [2006], and subsequently used by Chester et al. [submitted]. The development of a new precipitation overlapping algorithm is being developed by Y. Zhang and S. Klein at LLNL, and A. Bodos-Salcedo.

Acknowledgments. This work was supported by the Joint Defra and MoD Programme, (Defra) GA01101 (MoD) CBC/2B/0417 Annex C5. We thank Richard P. Allan, William J. Ingram and Keith D. Williams for providing comments on drafts of the paper. We thank Matt Rogers, from Colorado State University, for providing the lookup tables used to parameterized reflectivity and attenuation by precipitation. CloudSat data were obtained from the CloudSat Data Processing Center (http://cloudsat.cira.colostate.edu). CALIPSO and CERES data were obtained from the NASA Langley Research Center Atmospheric Sciences Data Center (http://eosweb.larc.nasa.gov). MODIS images were obtained from the NASA Goddard Space Flight Center Level 1 and Atmosphere Archive and Distribution System (LAADS) (http://ladsweb.nascom.nasa.gov). The program sam2p (http://www.inf.lume.hu/ pts/sam2p) was used to convert images to different formats.

References


Figure captions

Figure 1. Large-scale background of the CloudSat pass over the north Atlantic on February 26th, 2007: (a) Met Office analysis chart at 1200 UTC, and (b) MODIS RGB composite. The approximate CloudSat track is shown in red in both pictures.

Figure 2. Example of simulated mid-latitude system in the North Atlantic by the MetUM global forecast model on February 26th, 2007: (a) simulated radar reflectivity (in dBZ) from the model outputs at sub-grid scale (b) simulated radar reflectivity (in dBZ) from the model outputs at sub-grid scale averaged over the model gridbox, (c) radar reflectivity observed by CloudSat, and (d) CALIPSO vertical feature mask.

Figure 3. Large-scale background of the CloudSat pass over the Indian Ocean on December 23rd, 2006. The image shows the MODIS RGB composite with the approximate CloudSat track in red.

Figure 4. Example of simulated tropical system over the Indian Ocean by the UK Met Office global forecast model on December 23rd, 2006: (a) simulated radar reflectivity (in dBZ) from the model outputs at sub-grid scale (b) simulated radar reflectivity (in dBZ) from the model outputs at sub-grid scale averaged over the model gridbox, (c) radar reflectivity observed by CloudSat, and (d) CALIPSO vertical feature mask.

Figure 5. Hydrometeor fraction as observed by CloudSat and simulated by the MetUM global forecast model. Each map represents the global distribution of hydrometeor fraction for a 1 km height layer. The altitude of the centre of the layer is shown in the title of each plot, with rows interleaved to facilitate the visual comparison (CloudSat: 1st and 3rd rows; MetUM: 2nd and 4th rows). Hydrometeor fraction in each layer is defined by the number of positive identifications with $Z_e$ greater or equal than -27.5 dBZe divided by the total number of measurements in that layer.

Figure 6. Comparison of DJF 2006 statistics for the whole globe: (a) zonal mean cross-section of hydrometeor occurrence as observed by CloudSat, and (b) simulated by the MetUM model; (c) joint height-reflectivity hydrometeor frequency of occurrence as observed by CloudSat, and (d) simulated by the MetUM global forecast model.

Figure 7. Longwave cloud radiative forcing from DJF seasonal average of the CERES Edition2D SRBAVG dataset, from 2000 to 2005. The four rectangles show the oceanic regions chosen for the statistical analysis.

Figure 8. Comparison of DJF 2006 statistics for the North Atlantic region: (a) zonal mean cross-section of hydrometeor occurrence as observed by CloudSat, and (b) simulated by the MetUM model; (c) joint height-reflectivity hydrometeor frequency of occurrence as observed by CloudSat, and (d) simulated by the MetUM global forecast model; (e) fraction of hydrometeor occurrence as function of height.

Figure 9. As Figure 8 but for the Californian stratocumulus region.

Figure 10. As Figure 8 but for the Hawaii trade-cumulus region.

Figure 11. As Figure 8 but for the tropical warm pool region.

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North Atlantic - DJF 2006

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Figure 12. As Figure 8 but for the tropical warm pool region.