1 Introduction

The importance for the climate of the combined effect of cloud occurrence and the various properties of clouds is well appreciated in the scientific community. Clouds are important for at least two major reasons: their crucial role in the global water cycle and their moderating role for the global radiation balance (as outlined by e.g. Peixoto and Oort, 1992, Slingo, 1990, Arking, 1991, Liou, 1992 and Mason, 2002). However, the knowledge of global cloud conditions and, especially, its historic evolution in relation to associated changes of radiation conditions and surface temperatures is limited. The reason is mainly the spatial and temporal scarceness of global cloud observations and the fact that cloudiness is known to exhibit substantial variability on both the global and the local scale.

Observations from space were introduced during the 1960s and this opened up new opportunities. Finally, the goal of achieving homogeneous global coverage of cloud observations and radiation measurements with a high horizontal and temporal resolution seemed to be within reach. However, it lasted another 20 years before reliable quantitative observation datasets started to be systematically collected for monitoring of global cloud amounts as well as of top-of-atmosphere radiation budget components (Schiffer and Rossow 1983, Barkstrom, 1984.). The development of satellite-based observations and applications has accelerated since then and today it is possible to study several microphysical properties of clouds as well as details on the vertical profiles of cloud water and cloud ice (Stephens et al., 2002).

Concerning the description of clouds in climate models, it is clear that substantial improvements have been achieved during the last decades. Despite this fact, it is believed that deficiencies in cloud descriptions are still major sources of error in model scenarios of the future climate and in simulations of the present climate (IPCC3, 2001). In particular, the large sensitivity to the exact formulation of several specific cloud parameters or cloud features has been emphasised in many papers (e.g., Zhang et al., 1995, Lee et al., 1997, Wild and Ohmura, 1999, Chen et al., 2000, Hu and Stamnes, 2000, Colman et al., 2001 and Murphy et al., 2004). From this it naturally follows that high requirements must also be set on cloud observations that are used for validating models or for making diagnostic cloud studies.

The new satellite-derived cloud datasets enable studies never possible before, mainly because of the horizontally homogeneous coverage and the use of more objective (or at least more consistent) methods. This thesis presents two such new applications. Primarily, the thesis addresses the topic of developing, improving and adapting a satellite-derived cloud dataset for enabling an adequate use in a climate model evaluation experiment. A method is described for evaluating cloud simulations from a regional climate model (the SMHI Rossby Centre Atmospheric model – hereafter denoted RCA) using observed cloud parameters retrieved from satellite imagery. Results are discussed with the purpose of assessing the impact on RCA climate scenarios. Secondly, the thesis also includes a short study on one potential cloud-aerosol link to the climate system. This link involves aspects of both large-scale and microphysical scale processes where the latter are not yet explicitly described in current climate models.

Section 2 gives first some general background concerning the cloud observation task from space and points to some basic characteristics of the particular satellite observations utilised in this study. Section 3 describes then the used satellite climatology dataset in more detail.
including the methods for its extraction and the achieved validation results. This is followed in Section 4 by results from the cloud-aerosol study. Here, the dataset has been used to investigate the indirect connection between cloudiness, cloud-aerosol microphysics and large scale atmospheric transport processes as being manifested by measured variable concentrations of the cosmogenetic isotope $^{7}\text{Be}$ in near-surface aerosol samples. The remaining major part of this thesis (Section 5) is then dealing with the climate model evaluation task.

2 Satellite observations of clouds

Clouds appear mainly in the troposphere with cloud altitudes varying from near-surface to about 10-15 km with the highest altitudes in the tropics. They form as a result of atmospheric water vapour becoming supersaturated with respect to ice or water through cooling processes. The cooling is either caused by dynamical processes (e.g. by vertical ascent or by advection and subsequent mixing of air masses) or by radiation processes. Since these processes can occur at very different spatial scales, ranging from scales associated with individual turbulent convective air parcels to scales typical for planetary waves, we consequently find clouds with the same range of horizontal scales. In the vertical we find a variable cloud thickness from very thin clouds (for example fog or shallow clouds trapped in vertical temperature inversions) with a geometrical thickness less than 100 m to very thick convective clouds (Cumulonimbus clouds) with dimensions filling up almost the entire troposphere. If convection is deep and intensive, clouds may even penetrate into the lower Stratosphere ("overshooting cloud tops"). A limiting factor for cloud formation and cloud growth is naturally the amount and the supply of atmospheric water vapour which to a certain extent is positively correlated with the ambient air temperature. However, another important factor for the properties of clouds (especially the optical properties) is the availability of aerosol particles that could serve as cloud condensation nuclei (CCN). For example, a large CCN supply may give clouds with many cloud particles but small particle dimensions. This explains the optically thicker clouds often found over continental and industrial areas compared to clouds over remote oceanic areas (Sekiguchi et al., 2003). Consequently, we understand that also atmospheric aerosols-cloud processes are important for the climate due to its direct link to the optical properties of clouds. More details on cloud microphysics can be found in Pruppacher and Klett (2000).

The advantage of observing clouds from a space-borne platform is obvious. The task of observing global cloudiness based solely on surface observations is immense and practically impossible if aiming at getting coverage of the full spatial and temporal variability of clouds. Furthermore, the space view offers the ability to measure radiation budget components at the top of the atmosphere - the latter being the ultimate forcing factors for the climate on Earth. The first satellite program offering global long-term high-quality and high spatial resolution observations, suitable for both operational meteorological weather monitoring and global climate monitoring, was the Tiros-N series of polar satellites which was introduced in 1978 with the first experimental Tiros-N satellite. The first operational satellite was named NOAA-6 and it was launched in 1979. Since then more than 10 satellites have been launched with almost the same set of sensors and the program is guaranteed to continue until at least 2016 when the next generation of polar satellites (the NPOESS program – see http://www.ipo.noaa.gov/) will take over. Since this work is exclusively based on satellite measurements from the Advanced Very High Resolution Radiometer (AVHRR) carried by the NOAA satellites, a relatively thorough description of these measurements and their relevance for cloud observations follows here as well as some basic definitions.
Table 1. Spectral channels (μm) of the AVHRR instrument. Notice that wavelength intervals may vary slightly between different NOAA satellites, either due to problems to exactly reproduce the spectral response function or due to small adjustments in later AVHRR versions. In addition, a new channel centred at 1.6 μm was introduced on NOAA-15 in 1998 to be used on subsequent satellites during daytime conditions.

<table>
<thead>
<tr>
<th>Channel</th>
<th>Wavelength interval (μm)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.58 - 0.68</td>
<td>Visible</td>
</tr>
<tr>
<td>2</td>
<td>0.725 - 1.1</td>
<td>Near Infrared</td>
</tr>
<tr>
<td>3</td>
<td>3.55 - 3.93</td>
<td>Short-wave infrared</td>
</tr>
<tr>
<td>4</td>
<td>10.3 - 11.3</td>
<td>Infrared (split window 1)</td>
</tr>
<tr>
<td>5</td>
<td>11.4 - 12.4</td>
<td>Infrared (split window 2)</td>
</tr>
</tbody>
</table>

The AVHRR instrument is a cross-track scanning radiometer measuring radiation in five spectral channels (Table 1). Measurements cover a swath width on the Earth of approximately 2800 km. The instrument field of view (FOV, determining the horizontal pixel resolution) gives a resolution of 1.1 km at the sub-satellite point while at the swath edges it degrades to approximately 5 km and here the satellite zenith viewing angle is approximately 60°.

The NOAA satellites circle the Earth in sun-synchronous orbits with an altitude of about 850 km. With an orbital period of about 102 minutes they consequently make 14.1 orbits per day which means that the repeat cycle (i.e. the frequency of repeating exactly the same orbit with the same area coverage) is 10 days. Due to the converging orbital tracks towards the poles the use of NOAA AVHRR data is naturally most favourable at high latitudes and in the polar region. With a constellation of two operational NOAA satellites (one in a so-called morning orbit – a.m. orbit - and on in afternoon orbit – p.m orbit) we get typical near-zenith overpasses at the Norrköping satellite receiving station in Sweden (latitude 58.6°N, longitude 16.1 °E) as exemplified in Table 2.

All AVHRR channels are defined in so called window regions where radiance contributions from atmospheric gases and aerosols are small. Only minor contributions from thermal emission of atmospheric water vapour affect the split-window infrared channels and some marginal reflection of solar radiation from aerosols can be sensed in the visible channel. This means that measurements are very suitable for observation of clouds and the Earth’s surface.

Measured radiances in the infrared channels (3, 4 and 5) are calibrated continuously using an onboard blackbody reference measurement while for the visible channels a static use of calibration constants determined from pre-launch calibration activities has to be used in operational applications. For applications with a more long-term aim, post-launch calibration methods have been developed (e.g., Rao and Chen, 1999 and Tahnk and Coakley, 2001).

In the visible and near-infrared channels there is a strong relation between the reflected radiation and the optical thickness of clouds; $\tau_{\text{cloud}}$. The optical thickness describes the total attenuation of radiation due to the combined effect of scattering and absorption of radiation. In the visible region the absorption of radiation in clouds is negligible which means that there is a direct relation between cloud reflectance $R_{\text{cloud}}$ and $\tau_{\text{cloud}}$. However, this relation is non-linear since the cloud reflectance is given by the relation

$$R_{\text{cloud}} = 1 - \exp(-\tau_{\text{cloud}})$$ (1)
Table 2 Approximate zenith passage times ($T_{\text{zenith}}$) for the a.m. and p.m. NOAA satellites representative for the central part of the Scandinavian region in summer 1997. Values are given for both the ascending (northward) and the descending (southward) passage.

<table>
<thead>
<tr>
<th>Orbit</th>
<th>Node</th>
<th>$T_{\text{zenith}}$ (UTC)</th>
</tr>
</thead>
<tbody>
<tr>
<td>a.m.</td>
<td>descending</td>
<td>06:45</td>
</tr>
<tr>
<td>a.m.</td>
<td>ascending</td>
<td>16:45</td>
</tr>
<tr>
<td>p.m.</td>
<td>descending</td>
<td>02:15</td>
</tr>
<tr>
<td>p.m.</td>
<td>ascending</td>
<td>12:15</td>
</tr>
</tbody>
</table>

where the factor $\exp(-\tau_{\text{cloud}})$ is the cloud transmittance $T_{\text{cloud}}$. Thus, at high reflectances a small change of the reflectance is associated with a large change in optical thickness while the opposite condition is seen at low reflectances. Notice that Eq. 1 deals exclusively with reflection from clouds. For real measurements from space, also a contribution from the Earth’s surface will be added for clouds that are not optically thick. For better understanding the dependence on cloud microphysics we introduce the effective radius of cloud particles $r_{\text{eff}}$ defined as (Hansen and Travis, 1974)

$$r_{\text{eff}} = \frac{\int r^3 n(r)dr}{\int r^2 n(r)dr}$$

(2)

where $r$ is cloud particle radius and $n(r)$ is the particle size distribution. The effective radius is the area weighted mean radius and it can thus be interpreted as a bulk average size of cloud particles. According to Stephens (1978), we can approximate the optical thickness $\tau_{\text{cloud}}$ using Mie scattering theory by the relation

$$\tau_{\text{cloud}} \approx \frac{3}{2} LWP / r_{\text{eff}}$$

(3)

where $LWP$ is the liquid water path (equivalent to the vertically integrated cloud water amount) of the cloud. If introducing the liquid water content ($LWC$, i.e., the cloud water content per volume unit), the density of water $\rho$ and the geometrical thickness of the cloud $\Delta z$ we can reformulate this as

$$\tau_{\text{cloud}} \approx \frac{3 LWC \Delta z}{2 \rho r_{\text{eff}}}$$

(4)

This relation is not fully applicable to ice clouds. In the derivation of Eq. 3 there is a dependence on an extinction efficiency factor (summing up effects from both scattering and absorption processes) which may differ substantially between ice and water particles, also depending on the spectral wavelength. For this reason, Eq. 3 is an approximation valid for spherical water droplets and for Mie scattering processes. The relation will deviate for ice cloud particles due to different extinction efficiency factors and non-spherical geometries. Nevertheless, in at least a qualitative sense we can consider using Eq. 3 also for ice clouds if transforming particle size parameters of ice particles into their spherical equivalences (e.g. as suggested by Wyser, 1998).

We conclude from Eqs. 1-4 that clouds with high total cloud condensate amounts generally give high reflectances in AVHRR channel 1. Also, clouds with less total cloud condensate
may reflect considerably but only if the effective radius is small. For example, this explains the high reflectivity of small cumulus cloud elements as well as for stratus or fog. If the condensate amount is small and the effective radius is large we get very low reflectivities which are typical for thin ice clouds (thin Cirrus). Land and ocean surfaces appear very dark (low reflectance) in AVHRR channel 1 and clouds are consequently easily detected. Exceptions occur for snow- or ice-covered surfaces, over desert surfaces and in case of ocean sunglints.

In AVHRR channel 2 cloud reflection characteristics do not change drastically compared to in AVHRR channel 1. However, land surface reflectances increase due to a larger contribution from vegetated areas which means that this channel is less suitable for cloud detection. Nevertheless, this channel is often used instead of channel 1 over ocean surfaces due to the reduced reflection contribution from aerosols in this channel.

The two split-window infrared channels 4 and 5 measure thermally emitted radiances from the Earth surface and from clouds with only a small contribution from atmospheric water vapour. If transforming the measured radiances to their equivalent Planck radiation temperatures (hereafter denoted Brightness temperatures) these correspond quite closely to the true cloud top temperature of optically thick clouds. However, most clouds are at least partly transparent (semi-transparent) which means that brightness temperatures are normally warmer than true cloud top temperatures. An interesting difference between the two split-window channels is that the cloud transmissivity of thin ice clouds is higher in AVHRR channel 4 than in AVHRR channel 5 due to differences in the imaginary part of the refractive indices of water and ice (influencing cloud particle absorption – see Liou, 1992). This normally results in a negative brightness temperature difference for semi-transparent Cirrus clouds if subtracting channel 5 from channel 4. However, this possibility for identifying thin Cirrus clouds disappears if the underlying surface is too cold and comparable with the true cloud temperature (e.g., as often found in the Arctic region in the polar winter).

Perhaps the most interesting AVHRR channel from the cloud analysis perspective is channel 3. Here, measurements contain contributions from both reflected solar radiation and emitted terrestrial radiation. Furthermore, cloudy radiances are more influenced by the microphysical properties of the clouds than at shorter visible wavelengths. Thus, unlike in AVHRR channel 1 where radiances are largely determined by cloud optical thicknesses, radiances are now influenced also by the cloud phase (water or ice) since ice clouds absorb radiation more efficiently than water clouds at this wavelength (Liou, 1992). This means that if converting channel 3 radiances to brightness temperatures and subtracting corresponding values at channel 4 (where clouds do not reflect radiation) we get a measure of the degree of reflection from a cloud in AVHRR channel 3. For water clouds this difference is large (strongly reflecting – showing a large positive temperature difference) while for ice clouds the difference is small (weakly reflecting). However, for semi-transparent Cirrus we will still find a substantial positive temperature difference due to differences in cloud transmission in the two spectral channels. A further consequence of the reflecting capability of water clouds during daylight hours in this channel means that they do not behave as blackbody radiators at night. Thus, we instead get a negative brightness temperature difference for water clouds. Since cloud free surfaces generally show no such brightness temperature difference (i.e., surface emissivities are normally close to the value 1.0 except for desert surfaces) we understand that this AVHRR channel is very valuable for cloud detection both during day and night. An associated problem is unfortunately that the brightness temperature difference vanishes in conditions of twilight. A more detailed illustration of the spectral characteristics of AVHRR channels is given by Karlsson (1996, 1997).
Finally, it should be said that the most recent NOAA satellites have a dual channel 3 capacity so that the 3.7 μm channel is still used at night while a new channel at 1.6 μm is used during daylight hours (replacing the 3.7 μm channel). The strength of this channel is the capability to separate between water clouds and snow- and ice surfaces due to the very large reflection from water clouds while snow- and ice surfaces appear more or less black (i.e., non-reflecting). However, the cloud climatology used here has not used data from this new AVHRR channel.

3 The SCANDIA 1991-2000 cloud climatology (Paper I and Paper II)

Inspired and encouraged by the pioneering work by Liljas (1982 and 1984), the SMHI Cloud ANalysis model using DIgital AVHRR data (hereafter SCANDIA) was developed in the period 1986-1989 with the purpose of providing cloud and surface type information for supporting the operational weather forecasting work. Operational cloud classifications were introduced to the forecasters at SMHI in 1989 and after some modifications the method was frozen in 1991, the method remaining basically unchanged since then. The development was carried out by making extensive multispectral signature studies with the purpose of learning the spectral signatures of clouds and Earth surfaces, i.e., trying to quantify the characteristics of AVHRR radiances and the typical spectral signatures of clouds (briefly outlined in the previous section). SCANDIA is a multispectral thresholding method where the labelling of cloud and surface types results from a sequential analysis of AVHRR radiances. The first attempt was based on a Bayesian approach using a statistical Maximum-Likelihood classification method (Karlsson, 1989). However, it turned out to be very challenging to prepare the necessary statistical class descriptions due to the need of very large training datasets. Consequently, a multispectral thresholding algorithm was finally developed (Karlsson and Liljas, 1990) which enabled a simpler way of interpolating class signatures for defining thresholds as a function of varying illumination and viewing conditions. More clearly, for the labelling of a specific cloud type, a set of thresholds applied to all or a subset of the AVHRR channels has to be simultaneously passed (either by exceeding or falling below corresponding thresholds). All thresholds are defined in 12 specific sun elevation intervals and some thresholds are also changed according to four specified seasons. The vertical separation of cloud groups is accomplished through comparison with mean temperatures over the Scandinavian area taken from operational Numerical Weather Prediction (NWP) analyses of the 700 hPa and 500 hPa levels. No further compensation for varying viewing geometry is applied which means that the method is not adapted to the rather extreme conditions often occurring close to the swath edges. More clearly, the enhanced forward- and backward scattering seen here - typical for Mie phase functions (e.g. see Liou, 1992) - is not fully compensated for by SCANDIA. No further details of the method are given here but these are well documented by Karlsson and Liljas (1991) and Karlsson (1996).

In Paper I (based on an original idea outlined by Karlsson, 1995), a method is introduced to compile results from the SCANDIA method into monthly cloud climatologies and results are presented for a period covering one entire year (1993). It is based on a selective use of the most optimal overhead NOAA passages over Scandinavia each day, thus utilising four cloud observations per day (compare with Table 2). In this way, errors are largely avoided or minimised due to the exclusion of AVHRR scenes with high viewing angle conditions. Results, being defined in a standard polar stereographic map projection, are also resampled from the original 1 km horizontal resolution into a resolution of 4 km. The reason for this is mainly for taking into account that the navigation accuracy of individual AVHRR pixels (if using standard operational navigation methods) is normally slightly worse than the pixel reso-
Table 3. List of cloud parameters described by the SCANDIA cloud climatology.

<table>
<thead>
<tr>
<th>Cloud parameter</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Total cloud cover</td>
<td>Fraction of all pixels being cloudy</td>
</tr>
<tr>
<td>2. Low-level clouds</td>
<td>Opaque clouds below (warmer) than 700 hPa level</td>
</tr>
<tr>
<td>3. Medium-level clouds</td>
<td>Opaque clouds between 700 and 500 hPa levels</td>
</tr>
<tr>
<td>4. High-level clouds</td>
<td>Opaque clouds above 500 hPa level</td>
</tr>
<tr>
<td>5. Opaque clouds</td>
<td>Sum of categories 2, 3 and 4</td>
</tr>
<tr>
<td>6. Semi-transparent clouds</td>
<td>Cirrus clouds detected with semi-transparency test</td>
</tr>
<tr>
<td>7. Fog and Stratus</td>
<td>Homogeneous low clouds (low texture or variance)</td>
</tr>
<tr>
<td>8. Fractional clouds</td>
<td>Clouds close to detection thresholds (sub-pixel)</td>
</tr>
<tr>
<td>9. Precipitating clouds</td>
<td>Coldest and brightest part of High-level clouds</td>
</tr>
</tbody>
</table>

Table 4. Approximate time-windows (Central European Time = CET = UTC+1 hour) for used AVHRR scenes 1991-2000. The NOAA-satellites during the period are also indicated.

<table>
<thead>
<tr>
<th>Time of day</th>
<th>Time-window (CET)</th>
<th>Satellites</th>
</tr>
</thead>
<tbody>
<tr>
<td>Night</td>
<td>02:30 – 04:30</td>
<td>NOAA-11, NOAA-14</td>
</tr>
<tr>
<td>Morning</td>
<td>07:30 – 09:30</td>
<td>NOAA-10, NOAA-12, NOAA-15</td>
</tr>
<tr>
<td>Afternoon</td>
<td>14:00 – 16:00</td>
<td>NOAA-11, NOAA-14</td>
</tr>
<tr>
<td>Evening</td>
<td>17:30 - 19:30</td>
<td>NOAA-10, NOAA-12, NOAA-15</td>
</tr>
</tbody>
</table>

If no particular correction method is used (Brunel and Marsouin, 2000). Cloud climatologies for parameters defined in Table 3 are extracted and results are expressed in percentages of all observations for each studied pixel position.

In the subsequent paper, Paper II, the AVHRR-based cloud climatology is extended to cover one entire decade (1991-2000). When doing this, some specific and typical problems regarding the temporal observation frequency achieved from polar sun-synchronous satellites become evident. Instead of having relatively fixed observation times close to those given earlier in Table 2, a considerable variation is found depending on the studied year and the used satellite. This is explained by the instability of the sun-synchronous orbit (Price, 1991).

Consequently, instead of fix observation times it is more reasonable to describe observations within larger time windows such as those described in Table 4. Paper II provides also additional details as concerns the lifetime and availability of individual satellites and the evolution in time of the departure from the original sun-synchronous overpass time after the launch of each satellite.

Paper II also explores some new aspects which are now being allowed by this novel cloud dataset. For example, time series of cloudiness are extracted at selected locations where conventional surface observations are normally not available regularly. This is illustrated in Figure 1 where the annual cycle of cloudiness is shown for a position in the Baltic Sea.
Figure 1. The annual course of cloud cover (%) for two selected positions in the southern part of the Baltic Sea and in the Norwegian Sea (see text and Paper II for further explanation).

between the island of Gotland and the coast of Poland (position at 56°N 18°E - called Baltic Proper in Figure 1) as well as for a position in the Norwegian Sea well off-shore of the Lofoten area (position at 68°N 6°E - called Norwegian Sea in Figure 1). The mean cloudiness shown here has been calculated in a surrounding of 9-by-9 pixels which then covers a quadratic area of size 36-by-36 km. A tremendous difference in the appearance of the annual cycle of cloudiness for the two locations is evident. Remarkable is the pronounced minimum in cloudiness in summer in the Baltic Sea region while the corresponding location in the Norwegian Sea instead experiences a summer maximum. Interesting is also that cloud conditions over the Scandinavian mountain range (illustrated further in Paper II) appear very similar to conditions over the Norwegian Sea while a region very close to the Norwegian coast experiences an annual cloud pattern more resembling conditions in the Baltic Sea. This latter feature indicates the importance of the supply of cold freshwater from melting snow during spring and early summer for reducing sea surface temperatures, thus suppressing cloud formation.

Paper II also presents comparisons with surface observed cloudiness and with cloud climatologies derived by the International Satellite Cloud Climatology Project (ISCCP-Rossow and Schiffer, 1999) and the ECMWF Re-Analysis project ERA-40 (Uppala et al., 2005). In general, cloud amount deviations from corresponding surface observations were smaller than 10 % in cloud cover units except for some individual winter months when the separability between medium- and high-level clouds and snow-covered cold land surfaces was often poor. This introduced a considerable overestimation of SCANDIA cloud amounts (close to 10 % in cloud cover units) but consequently restricted to the in-land and most northern portions of the region. Some problems in detecting boundary layer clouds during night and at twilight were also noticed, especially in cold winter situations with presumably mixed cloud phases (i.e., both water droplets and ice crystals). A large part of these problems is related to the non-separability problem for water clouds and cloud-free surfaces occurring at twilight. The mentioned problems in the winter season also affected basic cloud type discrimination seriously. Especially the ability to correctly separate low-level and mid-level clouds was poor due to the existence of near-surface temperature inversions. Also some serious confusion between thin cirrus clouds and fractional (sub-pixel) convective cloud elements over the Norwegian Sea was found during winter.

The comparison with the ISCCP cloud climatologies revealed much larger seasonal cloud amount variability in the SCANDIA climatology as well as on the average lower cloud
amounts for SCANDIA. However, a substantial part of this difference is most likely explained by differences in the detectability of very thin clouds due to differences in satellite viewing geometries. The agreement with the ERA-40 cloud dataset was remarkably good except for the winter season when ERA-40 cloud amounts were found to be even higher than SCANDIA over mountainous land areas (thus, ERA-40 wintertime cloudiness appears here to be even more overestimated than as previously indicated for SCANDIA).

Interesting is that SCANDIA was found to underestimate cloud amounts by 5-10 % in cloud cover units in the summer season when compared to surface observations. However, this deficiency is not believed to be a true weakness of the SCANDIA observation but instead an indication of problems for the surface observer to correctly estimate cloud amounts (i.e., the “sky cover” of clouds) in cases with convective cloud elements with large vertical dimensions. Experiments based on the use of additional satellite observations with large satellite zenith angles supported the view that this deviation is probably caused by differences in geometrical viewing conditions (Karlsson, 1996).

Finally, it must be mentioned that the SCANDIA dataset has not been corrected for effects caused by AVHRR sensor degradation (i.e., reduced post-launch sensitivity of AVHRR visible channels – Rao and Chen, 1999 and Tahnk and Coakley, 2001). However, since the initial multispectral training process for determining SCANDIA thresholds included data from both recently launched satellites and satellites being in service since many years it is hoped that such effects are being minimised (i.e., being largely averaged out). Since there are no signs of deviating trends in cloudiness if comparing the SCANDIA climatology and corresponding climatologies based on the surface observation this is basically confirmed. However, the relatively high RMS differences may be attributed to this problem.

4 A study of the link between large-scale atmospheric transport processes, cloudiness and the occurrence of cosmogenetic isotopes in near-surface aerosol samples (Paper III)

Paper III describes an application of the SCANDIA climatology resulting from cooperation with scientists at the Uppsala University and the Defence Research Agency in Stockholm. The study partly addresses the scientifically debated climate issue whether there is an indirect connection between solar activity and global cloudiness apart from the more direct and apparent influence caused by changes in the radiative solar flux at the top of the atmosphere. This indirect effect is suggested to be linked to variations in galactic cosmic ray (GCR) intensities, the latter being negatively correlated with the solar activity. Several previous studies (e.g. Svensmark and Friis-Christensen, 1997, Pallé, 2005 and Harrison and Stephenson, 2006) have found a positive correlation between cosmic ray intensities and cloudiness, in particular concerning the amount of low-level clouds over low- and mid-latitude oceanic areas. However, other studies did not find any specific connection (Kristjansson et al., 2002) or they claimed that there is instead a weak correlation between cloudiness and solar irradiance variations. The relevance for the climate change discussion is that even if the net effect of potential cloudiness changes could be expected to cancel out during a full solar cycle, it is well-known that there are large variations in GCR intensities with much longer time-scales than the period of individual solar cycles (Carslaw et al., 2002 and Lockwood, 2002). More specifically, a decreasing trend in GCR intensities has been observed since 1964 (Lockwood et al., 1999) implying that a climate warming could have occurred due to a hypothetically GCR-associated decrease of low level cloud cover (i.e., implying a decrease in the planetary albedo).
In contrast to previous studies, Paper III is not specifically oriented towards correlating cloud amounts and GCR intensities on a global scale. Instead, it explores the regional-scale variations of near-surface concentrations of the cosmogenetic isotope $^{7}\text{Be}$ in aerosol samples and its relation to a range of meteorological parameters. One such studied meteorological parameter is cloudiness as here being provided by the SCANDIA cloud climatology in the 1991-2000 period. Thus, in this case we could only claim to have an indirect link to GCR intensities under the assumption that higher cosmic ray intensities leads to higher concentrations of $^{7}\text{Be}$ isotopes in the atmosphere. On the other hand, the link is interesting in the light of the ongoing discussion about the role of GCR intensity variations and its influence on the formation of cloud condensation nuclei (CCN) - see Yu (2002) and Pallé et al. (2004).

The radioactive isotope $^{7}\text{Be}$ ($t_{1/2}=53$ days) together with a few other beryllium isotopes (e.g. $^{10}\text{Be}$) and the more famous $^{14}\text{C}$ (used in carbon dating methods) and $^{3}\text{H}$ (tritium) isotopes are the most well known cosmogenetic isotopes in the atmosphere. They are called cosmogenetic since there is no other natural source for their creation. Very soon after $^{7}\text{Be}$ production through spallation processes (i.e., splitting of oxygen and nitrogen atoms through interaction with high energy neutrons and protons), which occurs with a maximum close to or above the 200 hPa level in the stratosphere and the upper troposphere (Lal and Peters, 1967), the isotopes attach to available aerosol particles. Thus, they can later potentially be involved in the transformation into CCN.

Aerosol sample measurements at three sites in Sweden (Ljungbyhed at 56ºN, Grindsjön at 59ºN and Kiruna at 68 ºN) were compared with standard meteorological parameters such as temperature and precipitation at each site but also with monthly mean cloudiness over an approximately 400 km quadratic area centered at the measurement site position. Results show a striking anti-correlation with mean cloudiness and a much weaker relation with the other parameters. Figure 2 summarises the results for one of the stations (Grindsjön). Results from the other two stations show generally the same features except for some natural variation with latitude that could be expected (i.e., some differences exist in amplitudes of actual values but the same temporal correlation between measurements and cloudiness is seen).

A strong peak-trough coherence correlation is seen in Figure 2 for cloudiness and $^{7}\text{Be}$ concentrations. Thus, in periods with low cloudiness we see high $^{7}\text{Be}$ concentrations and vice versa. This behaviour is significantly different from the positive correlation between cloudiness and GCR intensities that has been found previously in above mentioned papers. However, since there is only an indirect relation between GCR intensities and near-surface $^{7}\text{Be}$ concentrations it is likely that other modulating factors are involved. In particular, the influence of large-scale atmospheric transport processes will become very important since the measurements are made at the surface and not near the regions where the isotopes are produced. This is illustrated by Koch et al. (1996) where $^{7}\text{Be}$ is used as an atmospheric tracer for air with stratospheric origin.

Paper III is only able to point at a likely existence of a link between cloudiness and near-surface $^{7}\text{Be}$ concentrations and it cannot provide a full physical explanation. Nevertheless, two possible hypotheses are suggested which have to be considered further in potential follow-on studies. The first is a mechanism based on an involvement in cloud microphysical processes via the potential link to CCNs and CCN formation. Hence, it is proposed that in periods of high cloudiness, $^{7}\text{Be}$ isotopes are absorbed in clouds and thus being “removed” from clear air portions of the troposphere which could explain the decrease in near-surface concentrations. This effect can be further enhanced by the additional washout (wet scavenging) of $^{7}\text{Be}$ due to precipitation occurring in cloudy periods. The other hypothesis is that near-surface $^{7}\text{Be}$ variations could be explained by natural fluctuations in dominating large
Figure 2. Monthly data (a) versus smoothed data (b) of total cloud cover (TFCC) and $^7$Be concentrations using a running mean with a span of 7 months for the Grindsjön site (upper panels) compared with cosmic ray intensities and sunspot numbers (lower panel). The axis scale for $^7$Be concentrations has been modified to bring out comparable numbers between TFCC and $^7$Be. The QPF-curves simply illustrates a Quadratic Polynomial Fit of the data.

scale transport processes on a monthly and yearly time scale (e.g., as discussed by Koch and Mann, 1996). Thus, in this case there would only be an indirect dependence on cloudiness without any direct microphysical link, suggesting that cloudiness as well $^7$Be aerosol concentrations are basically determined by large scale atmospheric flow patterns. For example, in periods of summertime long-lasting anticyclonic (blocking) situations we will experience a steady but weak transport of air from the upper troposphere downward by the dominating mid-tropospheric subsidence. The transport further down to near-surface layers through the lower tropospheric subsidence inversion can be accomplished by turbulent mixing processes (dry convection). However, in winter the atmospheric circulation pattern is not the same on the local scale. In particular, we know that wintertime anticyclones are much cloudier in this region due to the formation of extensive boundary layer clouds being trapped in strong near-surface temperature inversions. These are often particularly strong since the radiative cooling of continental surfaces further enhances the lower tropospheric subsidence inversion in anticyclones. This will effectively prevent further downward transport of air to the surface. The subsidence inversion is much weaker in summer mostly as an effect of the more efficient convective mixing in the surface layer due to the strong radiative heating of the surface. Consequently, an efficient transport of air from the upper troposphere or lower stratosphere down to the surface can only be accomplished in a systematic and long-lasting manner by blocking anticyclonic situations in the summer season and hardly during the winter season. For the latter, downward transporting events are more likely to be associated with dynamic processes with shorter time-scales (e.g. tropopause folding processes associated with strong baroclinic developments – intense cyclogenesis) mixed in with periods of upward
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ascent of air. Due to this we may not see high $^7$Be concentrations near the surface in monthly averages during the winter season. An additional modulating factor could be the more frequent occurrence of precipitation in the winter half of the year which could contribute further in removing tropospheric aerosols through wet scavenging processes.

A circumstance that favours the second hypothesis and suggests it as the most plausible explanation is that there is currently no physical mechanism presented that explicitly describe how the $^7$Be isotope might play an active part in the formation of CCNs. The isotope is rather considered as being chemically non-reactive and evenly distributed among all available aerosol particles after its creation (Koch et al., 1996), thus implying that it should not be more frequently occurring among those hygroscopic aerosols being involved in CCN formation. This explains also its suitability for use in atmospheric aerosol tracer experiments. Instead, the currently proposed physical mechanism for a potential GCR involvement in CCN formation is that GCR-produced ions (i.e., electrically charged particles, for example formed in parallel to $^7$Be during the spallation process or formed in connection with other GCR-induced ionization processes without any $^7$Be connection) could act either to increase catalytically the rate of production of sub-critical 3-10-nm-diameter condensation nuclei (pre-stage to CCN) or act to modulate the magnitude of aerosol charges around clouds with possible influences on microphysical cloud processes (Yu and Turco, 2001, Yu, 2002 and Carslaw et al., 2002). It could be doubted whether $^7$Be could really take an active part in the above mentioned processes. Firstly, it is uncertain if the isotope might exist in ionized form sufficiently long for taking part in such atmospheric processes before being absorbed to aerosols or transformed to other stable molecular constellations. Secondly, it is much more likely that other GCR-induced ions would be involved since being much more numerous in the atmosphere than $^7$Be isotopes. Studies of $^7$Be occurrences in cloudwater and rainwater samples have not been able to give further clues here, even if an interesting observation is that $^7$Be occurrences appear to increase with increasing acidification of cloud water (Su and Huh, 2006).

Both presented hypotheses in Paper III should to be studied further in follow-on studies. Especially, we need to study this behaviour over several solar cycles. For example, the results for the current period (1991-2000, covering approximately one solar cycle) do not show a convincingly high positive correlation between cloudiness and solar activity (see QPF curves in Figure 2 b). This would rather indicate that the vertical transport hypothesis is as likely as the CCN hypothesis.

A final conclusion in Paper III is that regardless of the exact mechanism for the near-surface variation of $^7$Be we cannot easily use beryllium isotope concentrations from e.g. ice cores (although here $^{10}$Be is used due to a longer lifetime than $^7$Be) to make historical reconstructions of the solar activity. The indicated link to regional climate conditions makes this task quite complicated and suggests that some compensation could be necessary to account for inherent climate fluctuations on both annual and seasonal time scales.

5 Evaluation of cloud simulations of the Rossby Centre regional climate model (Paper IV and Paper V)

5.1 Characteristics of the RCA model

Due to the specific character of clouds being present in the atmosphere at almost all spatial scales, general circulation models (GCM) must be capable of describing clouds both explicitly (i.e., if being resolved in the used model grid) and implicitly (if being of sub-grid...
scale character) through various cloud parameterizations. Due to the constantly improving
(i.e., increasing) spatial resolution of models, it is a challenging task to adjust model
parametrisations so that they are in good balance with the current spatial resolution of the
model grid. This balance exists when parameterized processes remain truly on the sub-grid
scale instead of being completely or partly resolved by the model grid representation. In this
respect one must also take into account the fact that on cloud scales of a few km we will
experience an increased sensitivity to 3-dimensional radiative cloud effects such as multiple
reflection processes between neighbouring grid columns (e.g. as emphasized by Scheirer and
Macke, 2001). More specifically it means that the often used plane-parallel assumption when
calculating cloudy radiances (i.e., clouds being assumed horizontally plane-parallel and to
basically scatter and emit radiation as being of infinite horizontal extension) will show large
deviations from true cloudy radiances on very fine scales. At the same time, it is well known
that the plane-parallel approximation gives a large deviation (overestimation) of the cloud
reflectance if clouds are inhomogeneous (broken or scattered). This error is actually more
serious at coarse horizontal resolutions since at higher resolutions cloud inhomogeneities are
potentially better resolved. All these factors imply that great care has to be taken to the
horizontal scale factor both for effects on dynamical processes through convective and
precipitation parameterization schemes and for effects on radiation calculations through the
used radiation parameterizations every time when increasing the spatial resolution of models.

The SCANDIA-RCA comparison is based on results from the third version of the RCA model
(hereafter denoted RCA3). RCA was originally developed from the high resolution weather
prediction model HIRLAM (Källén, 1996, and Undén et al., 2002). While basically retaining
the dynamical core of the model, the physical parameterization schemes have been replaced
or further developed to allow an accurate operation in climate mode. After the initial
adaptation and definition of the model, a second version, RCA2, was developed which is
described in detail by Jones et al. (2004). RCA3 builds to a large extent on RCA2 but
includes also substantial modifications. RCA3 is a hydrostatic grid point model using a
terrain-following vertical coordinate system. The model uses a 2 time level semi-lagrangian,
semi-implicit dynamical core with 6th order horizontal diffusion. The vertical mixing as
described by the turbulence scheme in RCA3 is based on a prognostic turbulent kinetic
energy scheme with a diagnostic length scale (Cuxart et al., 2000). Compared to RCA2 it has
been updated giving a smoother transition between stable and unstable conditions and
becoming more numerically stable (Lenderink and de Rooy, 2000, and Lenderink and
Holtslag, 2004). Perhaps the most significant RCA-upgrade in RCA3 is the introduction of a
new land-surface scheme. The land portion in each grid point is sub-divided into three tiles
with three surface types: forest, open land or snow. A further specification is made for the
forest tile to cope with forest canopy, forest floor soil and snow on forest floor. Each tile is
thus basically differently treated so that the final grid value is received as a weighted mean of
all fluxes in the tiles according to their fractional coverage. More details (e.g., the coupling to
soil moisture changes) are given by Kjellström et al. (2005) and Samuelsson et al. (2006).

In RCA3 a prognostic process-oriented cloud parameterisation is applied meaning that cloud
condensate (or here cloud condensate mixing ratio $q_c$) is carried as a prognostic variable with
a separate equation for its temporal evolution following Rasch and Kristjánsson (1998)

$$\frac{\partial q_c}{\partial t} = A_{q_c} + S_{q_c} + C_{q_c} + Q - E - P$$

where $A$ is the tendency term associated with convection (i.e., for parameterized sub-gridscale
transports) and $S$ represents the sum of tendencies associated with all other subgrid-scale
parameterisations (e.g., vertical diffusion and boundary layer flux convergences/divergences and parameterisations). Further, \( Q \) is a source term from condensation processes and \( E \) and \( P \) are sink terms due to evaporation and precipitation processes. The \( Q \) and \( E \) terms appear also in tendency equations for temperature \( T \) and water vapour mixing ratio \( q_v \). This is the basic link between cloud processes and other dynamical processes together with the very important forcing term from radiation processes included in the tendency equation for temperature \( T \).

Convective clouds and convective processes (term \( C \) in Eq. 5) are described using the approach of Kain and Fritsch (1990). Meso-scale circulations are assumed to be basically resolved (or being of large-scale type) and only convective cloud scale fluxes have to be parameterised. A consequence here is that Cirrus anvils in deep convective systems can in some sense now be seen as large-scale clouds and that only the inner cores of the clouds with areas of active vertical ascent are parameterised. The treatment of shallow convective clouds has been radically changed in RCA3 (described in detail by Jones and Sanchez, 2002 and Albrecht, 1981). The main impact is reduced precipitation from shallow convective clouds, introduction of a shallow convective cloud fraction and that shallow convective clouds now contain more cloud water (thus being more reflective at visible wavelengths). The diagnosis of the cloud fraction \( f \) in a grid volume is using relative humidity thresholds, i.e., the relative humidity grid value must exceed critical values before clouds are allowed to form, following the ideas of Slingo (1987). To be noticed here is that these thresholds must be set below the humidity saturation level to allow for the description (i.e., parameterisation) of sub-grid scale cloud processes. Various modifications of the methods for treating the large-scale and convective cloud types have been introduced in order to further detail the process-oriented approach. In the RCA3 closure of cloud parameterization equations a cloud fraction tendency factor \( \delta f/\delta t \) is used. However, since \( f \) is only diagnosed this term is estimated in finite difference form based on \( f \) values from previous time steps.

Important for the modelled cloud fields is naturally the exact formulation of factors \( Q, E \) and \( P \) in Eq. 5. The \( Q-E \) term is specified by a sequence of assumptions telling how a surplus (deficit) in water vapour amounts will cause a moistening (drying) of cloud-free portions and a generation (reduction) of clouds. The basic idea here is that moisture forcing will act homogeneously on cloudy and cloud-free portions of a grid volume. More specifically, the forcing of cloudy portions will act to condense or evaporate cloud water and in the case of further condensation the new condensate will always lead to an increasing cloud fraction \( f \). The cloud water or ice content of the newly formed cloud will match that within the originally cloudy part of the grid volume, i.e., the in-cloud mixing ratio is kept constant in this first calculation step. However, some redistribution between the ice and water parts of the cloud condensate can occur as an effect of temperature changes in the grid volume which will be of importance later for description of precipitation and radiation processes. In the cloud-free part, the forcing will give either a moistening or a drying. In a second step, the in-cloud mixing ratio can again be altered by the effect of precipitation processes (described by the \( P \) term) which are specified as the sum of the following contribution factors:

- **Conversion of liquid water to rain**
- **Collection of cloud water by rain (rain coalescence)**
- **Conversion of ice to snow**
- **Collection of ice crystals by snow (snow coalescence)**
- **Collection of cloud water by snow (wet coalescence on snow)**.
In RCA3 some changes were made to the conversion calculations above in order to reduce the occurrence of weak precipitation which was found to be too frequent in RCA2. Also some changes were made in the diagnostic split of total cloud water into liquid and ice fractions. Implicitly involved in the previous cloud-related calculations and directly affecting the tendency equation for temperature $T$, the formulation of the cloud forcing of radiation calculations is perhaps the most critical aspect of the RCA3 cloud description. The RCA3 radiation scheme is based on the original formulation by Savijärvi (1990) but slightly modified by Räisänen et al. (2000). To compensate for some deficiencies of the plane-parallel homogeneous cloud approach (not truly representative of real clouds, especially concerning the case of broken or inhomogeneous clouds) some modifications were introduced according to Cahalan et al. (1994) leading to a reduction of the interpreted cloud optical thicknesses and the cloud albedo for short-wave calculations. A separate treatment of ice and water clouds is made. For water clouds, the effective radius is diagnosed directly from the total water content of clouds (with some consideration taken to if being exposed to marine or continental conditions) while for ice clouds the effective radius is diagnosed from the local air temperature (Wyser et al., 1999). The effective radius and cloud water amounts are then used in the calculation of cloud emissivity, cloud reflectivity and cloud transmissivity. The partitioning of cloud water into liquid and solid fractions is varied non-linearly in the temperature range 273.2 K to 250.2 K.

Noteworthy and of great importance for this study is that in the radiation scheme cloud information is principally treated with a Maximum cloud overlap between vertical cloud layers in both short-wave and long-wave calculations. Also, the basic approach follows a plane-parallel approximation of cloud layers (but again, with some attempt to compensate for the weaknesses of this approach). This formulation was chosen partly for computational efficiency reasons with the desire to be able to perform radiation calculations as accurately as possible while still passing only once through the vertical loop of calculations for an atmospheric column.

The RCA3 model studied here has a horizontal resolution of approximately 49 km and the number of vertical levels is 24. Thus, model simulations are made on a relatively coarse scale meaning that the previously mentioned cloud effect problems that could be encountered at very fine scales will not be seen. Nevertheless, the question still remains whether we have a balance between parameterised and resolved processes and whether the cloud forcing of the radiation scheme appears to be realistically described or not. This study addresses both of these aspects but in particular the second.

5.2 The importance of observational limitations of SCANDIA for evaluation of modelled cloudiness (Paper IV)

Paper IV gives first a thorough introduction to the satellite-to-model comparison problem. It discusses two commonly used but rather different model evaluation approaches namely the radiance-to-radiance comparison, utilising so-called forward-modelling methods (i.e., converting model states to equivalent satellite-measured radiances using a radiative transfer model), and the cloud field-to-cloud field comparison where modelled cloudiness is compared to satellite-retrieved cloud parameters. Both approaches are subject to various weaknesses and problems but it is claimed that their strengths are such that a combined use could be the most efficient way for reaching progress. More clearly, we need detailed evaluation of specific cloud properties but also an overall estimation of how the sum of all cloud-related and other essential influencing factors is reflected in the calculation of crucial radiation budget components. With some further development of the forward-modelling
approach it is also possible that methods can even merge in the future, i.e., so that traditional satellite-retrieval methods can be applied on forward-modelled radiances and then compared to the original cloud products based on satellite retrievals. Some first attempts of applying such an approach have already been demonstrated, e.g., by Chevallier et al. (2001).

Paper IV makes clear that this particular study is of the second type (i.e., comparing retrieved cloud parameters to modelled cloud parameters) and it then addresses the question whether the cloud parameters of the SCANDIA climatology are well suited or not for a direct comparison with corresponding RCA3 model parameters. It is found that there are many differences that can be compensated for by making appropriate model-to-satellite and satellite-to-model adaptations (especially concerning the space-based observation geometry) but that at least one critical uncertainty remains:

- Is the satellite observation method capable of detecting all clouds that are modelled?

In this context, it is claimed that it is very important to try to determine the cloud detection limit of the SCANDIA method. Previous studies (e.g. Wyser and Jones, 2005 and Intrieri et al., 2002) have pointed to the fact that many satellite-based cloud observations appear to match better to manual surface-based cloud observations than to observations based on measurements from high-sensitive cloud radars and cloud lidars. The latter group of observations appears to report higher cloud amounts than the former. Consequently, it is suspected that the existence of optically very thin clouds is not properly captured by either satellite-retrievals or by man-made surface observations. In principle, models generally have a capacity to model infinitely thin clouds since any cloud condensate amount can indeed be modelled in each grid volume. However, this capability is naturally constrained by the fact that the vertical grid dimension is normally too coarse for realistically describing very thin clouds, especially if these clouds also have sub-grid scale horizontal dimensions. A model representation of such cloud layers would instead hypothetically give a cloud with a higher optical thickness than in reality but with a cloud fraction that is much smaller than in reality. This can be understood from the fact that the modelled cloud fraction is rather the volume fraction than the horizontal cloud fraction. Nevertheless, one step towards resolving this uncertainty would be to know more in detail which clouds that are actually observed by the SCANDIA method. Paper IV describes a method to investigate this by utilisation of complementary radiative transfer calculations. The used radiative transfer model (RTM) is the Signal Simulator for Cloud Retrieval (SSCR) radiative transfer code described by Nakajima and King (1992) and Nakajima and Nakajima (1995) which is capable of simulating the spectral radiances of the AVHRR sensor. Radiance simulations are performed for measurements at the non-(cloud)absorbing 0.6 μm channel of the AVHRR instrument (channel 1 in Table 1) where a direct relation between measured reflectances and optical thicknesses of clouds exists (Nakajima and King, 1990). The idea is to later use results about the minimum optical thickness of SCANDIA-detected clouds for filtering of model datasets using the relation between modelled cloud condensate amounts and the combination of optical thickness and effective radius parameters (Eq. 4).

Simulations are made for one typical water cloud (effective radius 10 μm) and one typical ice cloud (effective radius 40 μm) over land or ocean surfaces and for a wide range of solar zenith angles, satellite zenith angles, sun-to-satellite azimuth difference angles and standard atmospheres. These reflectances are then compared to the lowest values of the SCANDIA reflectance thresholds for the assumed thinnest ice and water clouds and matching of results enables then the determination of the corresponding minimum optical thickness for the detected clouds. Figure 3 summarises these results for the thinnest ice cloud over land surfaces (a similar appearance is also seen for water clouds). Results vary considerably since
Figure 3. Extracted values of the minimum optical thickness (tau) for the thinnest Cirrus cloud category of SCANDIA as simulated by SSCR. Simulations were made for an ice cloud over a land surface having a surface reflectance of 5%. Results for three different satellite zenith angles (satzen - in degrees) are highlighted.

they depend strongly on the chosen simulation conditions. Minimum optical thicknesses vary between 0.5 and 2.5 if disregarding conditions for the highest solar zenith angles where calculations are uncertain (i.e., due to division of very small quantities). A clear sign of a higher detectability (i.e., lower minimum optical thicknesses) for the highest satellite viewing angles is seen as well as the opposite condition for low viewing angles. Since the chosen AVHRR overpasses for the SCANDIA climatology have viewing angles restricted to be between 0 and 45 degrees in the vast majority of cases (i.e., if not being too far away from the reception site in Norrköping) it is suggested to use a representative minimum optical thickness value of 1.0 as a starting point for the study but to also test the two mentioned extreme values of 0.5 and 2.5 in sensitivity tests. However, simulation of night-time and twilight conditions revealed a lower detectability for lower tropospheric water clouds. This is mainly explained by the overlapping spectral signatures of cloudy and cloud-free surfaces close to the terminator and to the restriction of the threshold of the 3.7 μm channel due to unfortunate channel noise problems with maximum influence for the coldest temperatures. Consequently, the minimum optical thickness at night for these particular clouds is adjusted to 3.0 while keeping the default value 1.0 for mid- and high-level clouds.

A final conclusion of Paper IV is the acknowledgement of the fact that it is very difficult to make a full judgement of all influencing error sources for the used radiance simulations. These factors are numerous and some of them (e.g. the use of the plane-parallel assumption) are also scale-dependent. Especially, we notice here that we apply the simulations at a pixel scale of 4 km which is indeed close to the previously mentioned (section 5.1) finer scale where serious deviations of simulated and true radiances start to appear due to the influence
of multiple reflection effects not properly taken into account by the plane-parallel assumption. Consequently, it seems more reasonable to apply filtering using a wide range of possible minimum optical thickness values than to try using just one interpreted average value.

5.3 Evaluation of cloud simulations of the Rossby Centre regional climate model (Paper V)

5.3.1 Details of the validation experiment

Paper V evaluates RCA3 results covering the SCANDIA 1991-2000 period produced in a perfect boundary climate simulation experiment, i.e., a simulation of the present climate with analysed boundary fields. The model was run using ECMWF re-analysis fields (ERA-40) for specifying the lower and lateral boundary conditions.

The treatment of cloud overlap in model grid columns is of great importance in the radiation scheme of climate models (e.g. as emphasized by Chen et al., 2000). Consequently, the three different alternatives Random, Maximum and Maximum-Random cloud overlap are all examined here, although with some specific focus on results for the Maximum overlap due to its direct use in the RCA3 radiation scheme. The sensitivity to the cloud overlap approach is obvious since it will greatly influence model-calculated cloud optical thicknesses for later use in the filtering of modelled cloud datasets. However, it is also clear that we can never estimate this quantity perfectly for the Maximum and Maximum-Random overlap approaches since this would require knowledge of the exact location of individual cloud layers within one grid volume. Consequently, we are in these cases forced to make approximations and a linear averaging of optical thicknesses is used here for cloud layers within the cloudy part of a grid column as proposed by Stubenrauch et al. (1997). This will again lead to some overestimation of cloud optical thicknesses (Cahalan et al., 1994) but since the averaging is made only in the cloudy part of the grid column the error is at least smaller than if averaging over the whole grid square. Thus, we have the following expressions for the calculation of the total cloud amount $f_{TOT}$ and the effective optical thickness $\tau_{eff}$ from modelled cloud quantities (layer contributions denoted with index $i$) for the case of Maximum cloud overlap:

$$\tau_{eff,Max} = \frac{\sum_{i=1}^{N} f_i \tau_i}{f_{TOT,Max}} \quad (6)$$

$$f_{TOT,Max} = \max(f_i) \quad (7)$$

Layer contributions $\tau_i$ are all in-cloud quantities calculated using Eq. 4 but applied in a modified form to handle the mixed contribution from both ice and water clouds in individual vertical layers. Corresponding relations for the other cloud overlaps are more complicated and based on expressions for the total vertical transmission. Further details are given in Paper V.

To get a more complete view of the modelled cloud situation, e.g., for assessing the impact on radiation budget parameters like the outgoing longwave radiation (OLR), also modelled cloud amount contributions from the three vertical cloud groups Low-level, Mid-level and High-level were defined. Here, the top-down approach was used in order to simulate the space-view and the same definition of the vertical cloud groups as for SCANDIA was applied. Thus, the following relation between the contribution from the three cloud groups (index $L$, $M$ and $H$) is used:

$$f_{TOT} = f_L + f_M + f_H \quad (8)$$
In addition, to even further sharpen the cloud analysis and to provide more information on the effect on crucial radiation budget parameters, results were sorted according to co-occurring vertical cloud group and cloud optical thickness categories following the concept introduced by Tselioudis and Jakob (2002). The resulting cloud top - cloud optical thickness histograms have its main use in visualising in a compact form how clouds are distributed in the vertical as well as in the optical thickness domain. SCANDIA thresholds are converted to optical thickness categories using additional SSCR simulations. Corresponding RCA3 quantities are calculated, again using the top-down approach and thus mixing contributions from multiple layers depending on cloud overlap. For achieving the final RCA3 grid average of in-cloud properties from pixel information and for each of the three vertical cloud groups the following relation is used to ensure consistency in the radiative transfer sense (i.e., avoiding a linear averaging of optical thicknesses):

$$\bar{\tau} = -\ln\left[\frac{1}{N} \sum_i \exp(-\tau_i)\right]$$

Index \(i\) means here individual cloudy pixels. However, it should be noted that this expression only accounts for effects of direct cloud transmission and not for additional effects due to multiple scattering from neighbouring cloud elements. Consequently, we are again forced to use approximations typical for the plane-parallel cloud concept (as being used by the SSCR RTM model as well as in the RCA3 radiation scheme) which could be critical at these pixel scales.

An important additional model-to-satellite adaptation is that the selected model-simulation times are matched to be within 30 minutes from satellite overpass times. This is important for avoiding artificial differences coupled to the diurnal cycle of cloudiness because of the observed drift of sun-synchronous orbits of the NOAA satellites.

As a final comment on the methods used for adapting model and satellite datasets, it should be said that the methods used here have many similarities with those used by the so-called ISCCP simulator (introduced by Klein and Jakob, 1999). However, in this case we do not simulate the corresponding satellite-derived high resolution cloud fields by stochastic processes using modelled fields and specific overlap assumptions as in the ISCCP simulator. Furthermore, in the use of the SCANDIA dataset we rely more confidently on a proper identification of thin semi-transparent high clouds and do not compensate for the satellite misinterpretation of high thin clouds as lower level cloud types which is done in the ISCCP simulator.

5.3.2 General results

Figure 4 shows an example of the unfiltered and filtered RCA3 results compared to SCANDIA results over the Scandinavian region for the Maximum cloud overlap approach.

Table 5 summarizes results for all cloud overlaps.

Generally quite good agreement is found for seasonal and yearly averages with only a few percents difference from corresponding SCANDIA results. The only exception here is naturally for the RCA3 Random cloud overlap approach where results indicate excessive cloud amounts, especially for the summer and autumn seasons. When considering the geographical distribution of clouds, it is found that RCA3 cloud amounts are always (i.e., for all seasons) in excess of SCANDIA amounts over the Scandinavian mountain range. Simultaneously, a pronounced deficit in cloud amounts is found lee-ward (i.e., closely to the east) of the Scandinavian mountain range.
Figure 4. Seasonal means (winter, spring, summer and autumn from top to bottom) 1991-2000 of total cloud cover (%) for SCANDIA (left column) compared with original (central column) and filtered (right column) RCA3 results using Maximum cloud overlap.
Table 5. Area mean of seasonal and yearly differences in total cloud cover (%) between RCA3 and SCANDIA for filtered and unfiltered (in brackets) results and for different cloud overlap approaches.

<table>
<thead>
<tr>
<th></th>
<th>Max-Ran</th>
<th>Max</th>
<th>Ran</th>
</tr>
</thead>
<tbody>
<tr>
<td>WINTER</td>
<td>-6.09 (-0.34)</td>
<td>-7.16 (-2.72)</td>
<td>-2.29 (8.82)</td>
</tr>
<tr>
<td>SPRING</td>
<td>-4.27 (1.80)</td>
<td>-4.93 (-1.40)</td>
<td>0.49 (12.00)</td>
</tr>
<tr>
<td>SUMMER</td>
<td>5.41 (8.90)</td>
<td>2.83 (3.88)</td>
<td>14.7 (21.20)</td>
</tr>
<tr>
<td>AUTUMN</td>
<td>1.97 (6.16)</td>
<td>-0.17 (2.09)</td>
<td>8.19 (15.90)</td>
</tr>
<tr>
<td>YEAR</td>
<td>-0.74 (4.30)</td>
<td>-2.36 (0.46)</td>
<td>5.27 (14.48)</td>
</tr>
</tbody>
</table>

Table 6. Area mean of seasonal differences (excluding Winter season) between RCA3 and SCANDIA for filtered and unfiltered (in brackets) high, medium and low cloud amount contributions (%) using the Maximum cloud overlap assumption.

<table>
<thead>
<tr>
<th></th>
<th>Difference high-level clouds</th>
<th>Difference medium-level clouds</th>
<th>Difference low-level clouds</th>
</tr>
</thead>
<tbody>
<tr>
<td>SPRING</td>
<td>3.39 (5.69)</td>
<td>-7.01 (-6.55)</td>
<td>0.61 (1.21)</td>
</tr>
<tr>
<td>SUMMER</td>
<td>5.58 (6.39)</td>
<td>-5.71 (-5.70)</td>
<td>4.92 (5.04)</td>
</tr>
<tr>
<td>AUTUMN</td>
<td>-1.84 (-0.54)</td>
<td>-2.86 (-2.70)</td>
<td>6.48 (6.87)</td>
</tr>
<tr>
<td>YEAR excl. Winter</td>
<td>2.38 (3.85)</td>
<td>-5.19 (-4.98)</td>
<td>4.00 (4.37)</td>
</tr>
</tbody>
</table>

For the RCA3 contributions from the three vertical cloud groups (results for Maximum overlap are shown in Table 6) smaller contributions compared to SCANDIA are found for all cloud overlap approaches concerning the group of Medium-level clouds. As a contrast, contributions from Low-level clouds are generally larger and contributions from High-level clouds are slightly larger or for Random overlap considerably larger than SCANDIA contributions. This result is largely supported by Illingworth et al. (2006) and Willén et al. (2005) comparing model results with ground-based measurements from the CLOUDNET network of cloud radars and cloud lidar. Concerning the excessive total cloud amounts over the Scandinavian mountain range, it is interesting to notice that this feature could be seen for all the vertical cloud groups (including high-level clouds). Notice that no results for the Winter season are presented here due to the reduced capability of SCANDIA during winter concerning the vertical separation of cloud types (as mentioned in section 3).

In the investigation of frequencies of co-occurring vertical cloud groups and optical thickness categories, it was convincingly illustrated how sensitive results are for the chosen cloud overlap approach. For Maximum cloud overlap (Table 7) RCA3 gives much higher relative frequencies than SCANDIA of the THICK (optical thickness range 14.8-28.6) and VERY THICK (optical thicknesses above 28.6) categories for all three vertical cloud groups. Remarkable is the very high frequency of 28 % for the VERY THICK category of low-level
Table 7. Summary of seasonal results (excluding Winter) of the relative distribution (%) of clouds among cloud altitude-cloud optical thickness categories for the Maximum overlap approach. Corresponding interpreted categories from SCANDIA results are given in brackets. See text for further explanation.

<table>
<thead>
<tr>
<th>CATEGORY</th>
<th>SPRING</th>
<th>SUMMER</th>
<th>AUTUMN</th>
</tr>
</thead>
<tbody>
<tr>
<td>HIGH – THIN</td>
<td>37.2 (28.8)</td>
<td>29.6 (40.7)</td>
<td>33.2 (35.5)</td>
</tr>
<tr>
<td>HIGH - THICK</td>
<td>3.7 ( 3.5)</td>
<td>2.7 ( 2.6)</td>
<td>3.5 ( 3.3)</td>
</tr>
<tr>
<td>HIGH - VERY THICK</td>
<td>6.1 ( 1.5)</td>
<td>13.1 ( 0.5)</td>
<td>10.1 ( 1.8)</td>
</tr>
<tr>
<td>MEDIUM - THIN</td>
<td>10.2 (33.9)</td>
<td>3.2 (29.4)</td>
<td>7.2 (27.0)</td>
</tr>
<tr>
<td>MEDIUM - THICK</td>
<td>1.8 ( 1.4)</td>
<td>1.5 ( 0.1)</td>
<td>1.6 ( 0.9)</td>
</tr>
<tr>
<td>MEDIUM – VERY THICK</td>
<td>3.0 ( 0.2)</td>
<td>6.7 ( 0.0)</td>
<td>4.3 ( 0.0)</td>
</tr>
<tr>
<td>LOW - THIN</td>
<td>15.8 (30.1)</td>
<td>4.4 (26.2)</td>
<td>10.4 (29.7)</td>
</tr>
<tr>
<td>LOW - THICK</td>
<td>6.8 ( 0.6)</td>
<td>5.5 ( 0.3)</td>
<td>7.0 ( 1.3)</td>
</tr>
<tr>
<td>LOW - VERY THICK</td>
<td>10.1 ( 0.1)</td>
<td>28.0 ( 0.0)</td>
<td>17.8 ( 0.4)</td>
</tr>
</tbody>
</table>

Clouds in the summer season in comparison to the corresponding SCANDIA frequency which is practically zero. Results for the Maximum-Random overlap (not shown here) are only slightly improved. Instead, a much better fit to the SCANDIA-interpreted frequencies is found for the Random overlap approach. These results are indeed interesting since we know from observational evidence that true clouds are often seen to cluster together if being present in adjacent layers (Maximum overlap) while they appear more in a random manner if being separated by cloud-free layers (e.g. as reported by Tian and Curry, 1989, and Willén et al., 2005). We interpret this deviation as a sign of a true excess of cloud condensate for RCA3-simulated clouds. More clearly, we can only get a good agreement with the observed optical thickness categories if we distribute the excessive cloud condensate amounts in an unrealistic way using the Random overlap (spreading out or diluting cloud water amounts horizontally as much as possible).

5.3.3 Important error sources and sensitivity studies

Paper V discusses the possible influence on the results of a number of error sources and the sensitivity to some particular parameters. The most critical remaining deficiency of the SCANDIA cloud climate dataset that still could affect results significantly is explained to be the wintertime overestimation of cloud amounts over very cold land surfaces. This deficiency is likely to partly explain the large negative biases of RCA3 cloud amounts in the Winter season (e.g. see Table 5). Some spill-over effects of this weakness could also be anticipated for the Spring season. Consequently, there is a high likelihood that results for the Maximum overlap approach in Table 5 actually have the best matching to observations on a yearly basis instead of the currently deduced best matching given by the Maximum-Random results.

Concerning the sensitivity tests, it is shown that the adjustment to a smaller minimum optical thickness value of 0.5 for filtering has a very small impact. Most interesting here is that the
small additional cloud contribution comes predominately from high level clouds. Concerning the use of the maximum value of 2.5, it is clear that a large fraction of modelled clouds are now filtered out leading to a general underestimation of cloud amounts for almost all cloud overlaps and all seasons. Despite this, the previously found imbalance between contributions from Low, Medium and High clouds basically remains unchanged indicating that this is a very robust result.

Finally, an experiment with an increased horizontal resolution (changed from 49 km to 24 km) showed slightly reduced total cloud amounts and lead to the removal of most of the cloud amount minima seen leeward of the Scandinavian mountain range, although still keeping excessive maxima at mountain tops. Obviously, here we find evidence of the large sensitivity to the spatial resolution which has been discussed previously. In this case one can suspect that the finer resolution of topographic features alters vertical mixing processes and the way the dynamic flow surpasses the mountain range which obviously changes the cloud pattern. Even if this circumstance is interesting we cannot draw more firm conclusions here because of the lacking knowledge of how well adapted the used RCA3 parameterisations are for the use at this finer scale. New experiments, including also an improved vertical resolution, have to be conducted to investigate this further.

5.3.4 Reference to validation experiments with global climate models

As already mentioned, the results of this model validation experiment agree with many of the features that has been seen when comparing modelled cloud fields to ground observations made by cloud-profiling radars and lidars (e.g., Willén et al. (2004) and Illingworth et al. (2006)). Thus, we can state that this study has given us higher confidence in that e.g. the problem of underestimating mid-level clouds and the overrepresentation of optically thick low level clouds is a general feature and not a local one (typical only for specific cloud radar and lidar observation sites).

Even more interesting is that very similar results are also seen in validation experiments dealing with global climate models. Especially, the comprehensive study by Zhang et al. (2005) must be mentioned in this context. Here, 10 global climate models were compared to global satellite datasets using the ISCCP simulator (Klein and Jakob, 1999). The study indicates that RCA3 seems to share the problem of correctly describing middle level clouds and the feature of excessive optical thicknesses of clouds with several of the studied global climate models. The only evidently unique feature of RCA3 results compared to the studied global models seems to be the overestimation of the amount of low-level clouds. Global models rather tend to underestimate the amount of low-level clouds.

5.3.5 Discussion

Let us now try to discuss the results in the light of the following two questions:

- What are the implications on RCA3 model dynamics and model physics?
- What is the importance of the identified cloud description weaknesses for RCA3 climate scenarios, i.e., how do they affect their reliability?

The most direct and understandable impact is believed to be seen for the surface radiation budget components calculated by the RCA3 radiation scheme. The overrepresentation of optically thick clouds while still reasonably well simulating the total cloud amounts should lead to excessive reflection of incoming solar radiation from clouds at visible wavelengths. Furthermore, at long-wave terrestrial wavelengths radiation emitted by the surface would be
more efficiently absorbed by clouds and re-radiated back to the surface due to the resulting higher effective cloud emissivities. The resulting excessive cooling of the surface during day and corresponding warming during night would imply a reduced diurnal cycle of surface temperatures. Such behaviour has also recently been verified for RCA3 simulations (Kjellström et al., 2006). However, we can still not be absolutely sure that this comes entirely from this apparent deficiency of modelled cloudiness. There are still many factors that could give a similar appearance (e.g., incorrect description of surface albedo or vegetation characteristics and incorrect treatment of cloud-free radiative transfer conditions) and further investigations are consequently needed to get a complete understanding. Nevertheless, the current study has pointed at one serious weakness of the current cloud description that needs to be taken into account in further model development.

Paper V also discusses the potential impact on RCA3 radiation budget components at the top of the atmosphere. However, here it is much more difficult to make firm conclusions. Even if we do suspect some excessive reflection back to space of solar (visible) radiation due to the overestimation of optical thicknesses, we cannot easily interpret the impact on the resulting outgoing long-wave radiation (OLR). The reason is that the RCA3 imbalance in the vertical distribution of clouds evidently creates some compensating effects, i.e., increasing low-level cloud amounts give increasing OLR while increasing high-level cloud amounts give decreasing OLR. However, since low-level clouds appear to increase more than high-level clouds the net effect should be an increase in OLR.

Thus, in addition to the suggested effect on the diurnal cycle of surface temperatures we may propose: The overestimated reflection of solar radiation by clouds at visible wavelengths and the overestimation of OLR caused by excessive low-level cloudiness will give a total cloud radiative forcing on the Earth-atmosphere system for RCA3 resulting in a too strong cooling. Attempts to compare RCA3 results to Earth Radiation Budget Experiment data (Barkstrom, 1984) covering the period 1985-1990 have largely confirmed the cooling effect in the shortwave region but not in the long-wave region where results are more neutral (Willén and Wyser, 2006 – personal communication). This could mean that the indicated difference in this study of the increments of low-level and high-level cloud amounts is too small to be significant or alternatively, influenced too much by remaining uncertainties of the SCANDIA climatology retrieval method. A third possibility is that there are further compensating effects in the long-wave region that could mask this effect. A final remark here is that several global climate models also seem to have a tendency of producing excessive cooling from clouds (Zhang et al., 2005).

Whereas the effect on the diurnal cycle of surface temperatures has been confirmed by Kjellström et al. (2006) there is no sign of an overall and significant negative surface temperature bias. Rather, a seasonal variation from a warm bias in winter to a cold bias in summer can be seen (Kjellström et al. (2006). This could indicate a higher importance of the incorrectly described surface radiation budget components (dominated by long-wave processes in winter and by shortwave processes in summer) than of the deficiencies of top-of-atmosphere radiation budget components. The most plausible explanation here is that the exchange of air masses through the lateral boundaries might reduce the chances for achieving a noteworthy overall cooling within the RCA3 domain. Thus, the forcing from either the ERA-40 dataset (used in perfect boundary simulations) or from the global climate models (used in climate transient simulations and climate scenarios) is most probably dominating in comparison to the excessive cooling due to erroneous RCA3 cloudiness. Regarding the overall effect on RCA3 climate scenarios, we may thus conclude that average surface temperatures over the whole RCA3 domain are not expected to deviate significantly from the
corresponding temperatures of the used global model (i.e. significant deviations will only be seen on the regional and local scales not described by the global model). However, deficiencies in the cloud description will negatively affect the diurnal cycle of surface temperatures and follow-on studies using RCA3 scenarios for assessing consequences of a changed climate (e.g., changes in vegetation season, number of days with frost, etc.) have to take this into account.

One comment on the model evaluation experiment as such must also be given and this concerns the approach of filtering model datasets for the optically thin clouds. It is clear that modelled cloud amounts are not drastically changed after filtering. Thus, this study has shown that the suspected difference between modelled and satellite-observed cloud climatology, due to the lack of contribution from optically very thin clouds not being detected, can be considered as relatively small. Consequently, the general results are largely similar to the results that were achieved also for unfiltered results. Nevertheless, the filtered cloud amounts are by no means negligible and they strongly depend on the chosen cloud overlap approach. For example, they become quite large for the case using random cloud overlap. Consequently, the filtering aspect could still be of importance and it should be taken into account also in similar validation studies in the future.

A final very important factor for this study has to be addressed before closing the discussion. This is the well-known problem for coarse resolution GCMs and climate models in correctly describing the average optical thickness of clouds based on grid average quantities (e.g., as discussed by Cahalan et al., 1994). The resulting positive bias of the cloud albedo (i.e., overestimation of cloud optical thicknesses) is exactly what has been found also in this study. However, what is more important here is that it appears as the measures that have been taken in the RCA3 radiation scheme to try to compensate for this defect of the plane-parallel cloud description are largely insufficient. More specifically, the currently applied reduction of cloud optical thicknesses by 20 % in the RCA3 radiation scheme is hardly able to more than marginally improve results if comparing to results previously shown in Table 7. It appears that more drastic changes or improvements are needed. A natural step, together with improving the spatial resolution to resolve clouds better, would probably be to try to adapt the cloud description in the radiation scheme towards a more physically realistic Maximum-Random cloud overlap approach (e.g. according to results presented by Willén et al., 2005) while still trying to retain the computational efficiency of the scheme.

6 Main results and outlook

This thesis has demonstrated how a satellite-derived cloud climatology can be used in studies of cloud-aerosol processes and for the evaluation of a regional climate model.

The cloud-aerosol study revealed an intriguing anti-correlation between monthly mean cloudiness and the concentration of the cosmogenetic isotope $^7$Be in near-surface aerosol samples. Whereas it is pointed out that the most likely physical mechanism for this behaviour could be attributed to large-scale transport processes (affecting general cloud conditions), it is not ruled out that also a more complex relation to cloud microphysical processes could exist. Further studies are needed here to increase the understanding of the underlying physical mechanisms.

The major part of the thesis has dealt with the regional climate model evaluation task. A complete procedure for evaluation of modelled cloud fields by use of satellite-retrieved measurements has been demonstrated, i.e., starting from the initial extraction of cloud parameters from satellite imagery and ending with the final interpretation of modelled cloud
parameters and the analysis of its impact on model physics. Special emphasis has been made on using a large number of model-to-satellite and satellite-to model adaptations in order to avoid artificial biases of the results. In particular, the study has stressed the necessity to establish a deeper knowledge about the initial differences between the observed and modelled clouds as explained by cloud detection limitations of the satellite retrieval method. The study has also highlighted the importance of studying individual cloud features as well as the combined effect of several cloud parameters. The latter is in fact absolutely necessary if being able at all to make conclusion about net effects of clouds, for example considering their impact on crucial radiation budget parameters.

The validation experiment found the following characteristic behaviour of RCA3 cloudiness:

- Reasonable total cloud amounts on yearly basis
- Underestimated amounts of Medium level clouds and overestimated amounts of Low and High level clouds
- A large overrepresentation of optically thick or very thick clouds for all vertical cloud group categories

The overall effect on model physics and on RCA3 climate scenarios are concluded to be most serious for the description of surface radiation balance conditions. Here, an excessive cooling is expected during daytime while at night-time the opposite will prevail. Consequently, the diurnal surface temperature range will be underestimated which is a feature that must be taken into account in further climate consequence studies based on RCA3 climate scenarios. Furthermore, the RCA3 cloud radiative forcing at the top of atmosphere is expected to lead to excessive cooling. However, this effect is believed to have only minor importance due to the strong forcing at lateral boundaries from the used global climate model.

The study has an obvious weakness in the systematic use of radiative transfer simulations based on the horizontally homogeneous and plane-parallel cloud assumption. Apparently, future studies with an improved simulation of cloudy radiances using e.g. cloud resolving models and more accurate RTM tools are greatly needed and encouraged. Another weakness of importance is the lacking compensation for calibration defects of the satellite-measured radiances. However, to enable such corrections of the data a substantial support from major satellite operators (e.g. NOAA and EUMETSAT) or other satellite study institutions is necessary. Such a support has not been available or been applicable here. At last, an important caveat to remember is that the results found here are specific for the Scandinavian region which is only a small part of the full RCA3 domain. Thus, further studies are still needed to confirm if results can be generalised.

For the future, the utilisation of enhanced satellite-derived datasets (i.e., including a larger number of cloud parameters) with high horizontal resolution and with a much larger spatial coverage than the current SCANDIA dataset (e.g., as introduced by Schulz et al., 2005) will be possible for model evaluation purposes. This includes also the use of top-of-atmosphere radiation datasets which are needed for assessing the relative importance of the description of various cloud features in relation to other influencing factors. Furthermore, an improved support from satellite operators regarding calibration issues, the availability of even more advanced satellite measurements and the use of more sophisticated RTM models for simulation of cloudy radiances will significantly improve the prospects for achieving a better understanding of modelled and observed clouds on both the regional and global scale.
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Thesis Summary


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Thesis Summary


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