Cloud Inhomogeneity Effects in Microwave and Sub-mm Passive Sounder Data

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Cloud inhomogeneity effects have impacts on the measured signals (radiance) by a passive microwave sensor. These affect the retrieval results of observations made by the sensor, so in order to correctly interpret the sensor’s data retrieval it is important to determine the cloud inhomogeneity effects. The scope of this thesis is to investigate cloud inhomogeneity effects as a result of beamfilling effects.

The beamfilling effect has been defined as a systematic error which arises as a result of non-uniform trace gases (molecules and/or particles) in the sensor’s field of view (FOV). So in order to determine cloud inhomogeneity a comparison has to be made with radiative transfer (RT) simulation results, of the same atmospheric data, when the atmospheric parameters in the sensor’s FOV is homogenized (i.e. averaging the atmospheric parameters with the sensor’s weighting function), and when the atmospheric parameters in the sensor’s FOV are non-uniform (i.e. without averaging the atmospheric parameters). The difference between the two simulation results is as a result of beamfilling effects.

In order to determine cloud inhomogeneity effects, clear-sky and cloud RT simulations are carried out using Nonhydrostatic ICosahedral Atmospheric Model (NICAM) data. NICAM is a 40 level data set containing atmospheric profiles of atmospheric parameters such as water vapor, temperature, cloud ice, cloud water, snow, etc.. The RT simulations are performed with the Atmospheric Radiative Transfer Simulator (ARTS) model.

The clear-sky RT simulations involve simulation of RT of absorbing gases only without influence of particle scattering. The clear-sky RT simulations are performed to determine clear-sky inhomogeneity which as a result of non-uniform absorption gases in the sensor’s FOV. The purpose is to determine the amount of inhomogeneity coming from clear-sky only before introducing scattering particles in the atmospheric parameters. The clear-sky RT simulation results can be seen as a subset of cloud RT simulation results. This is because the cloud RT simulations are carried with the same atmospheric parameters as clear-sky except that cloudy particles (ice cloud and liquid cloud) are added into the atmospheric parameters.

To fulfill the aim of this thesis, a set of frequency channels in the microwave regions (181.5, 327, 456, 494.5, and 663.5 GHz) are selected for these simulations.
The frequency channels are selected because they have similar temperature Jacobians, which are obtained through clear-sky RT simulations. The results show that clear-sky inhomogeneity is the same at these selected frequency channel for the fact that they have similar temperature Jacobian. However, cloud inhomogeneity effect is different for each frequency channel as they are sensitive to different optical properties (particles’ sizes) of the cloud. The results obtained in this study depend on the frequency channels under investigation and the ground resolution of the space-borne sensor.
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Finally, I want to thank my family and God for being there for me during the course of my study.

This thesis is dedicated to the memory of my dad, Anyairo Collins Iroabuchi, who died at the beginning of this project.
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Chapter 1

Introduction

The aim of this thesis is to study the effects of cloud inhomogeneity on microwave and sub-millimeter passive satellite data. According to Solomon (2007), clouds and their interactions with radiation present the greatest uncertainty in future projections of the Earth’s climate. Cloud inhomogeneity effects present uncertainty in remote sensing retrievals, not just in microwave/sub-mm spectral regions but also in other spectral regions (e.g. infrared, visible etc.). Several studies have been conducted to investigate cloud inhomogeneity effects on different remote sensing retrievals and at different spectral regions.

Cloud inhomogeneity effects in cirrus cloud observation is defined, in Davis et al. (2005), as the difference between the cloud induced radiance and that of the same atmosphere without cloud (which is referred to as clear-sky radiance). According to the paper, cloud inhomogeneity effects can influence the observed radiance of cirrus cloud in three (3) different ways. The first effect is the change in the observed radiance as a result of the presence of cloud in the atmosphere. Cirrus cloud reduces radiance measured by sensor by scattering, so when compared with that of clear-sky will always yield a negative change in radiance. This effect is referred to as beamfilling effect (BFE), in Lafont and Guillemet (2005), where it is defined as a systematic error which is introduced by non uniform trace gases (molecules or particles) in the field of view (FOV) of a sensor. The second effect occurs in limb observation, where the change in radiance changes from negative to positive with increase in tangent height of the sensor direction. The change in sign is a result of the cloud changing role from scattering radiance way from the sensor’s line of sight to into the line of sight. The third effect is when particles’ scattering introduces polarisation and influence from outside atmospheric properties into the sensor’s line of sight. In order to take all these influences into account, this effect requires a 3-D geometry (i.e. 3-D radiative transfer (RT) simulation).

In Davis et al. (2005), cloud inhomogeneity was investigated by making 3-D and
1-D RT simulations in a RT model ARTS-MC, which is described in Section 3.2.2 of this thesis. In order to distinguish any difference between 1-D and 3-D to either BFE or 3-D RT effects, an Independent Pixel Approximation (IPA) simulation was carried out. IPA simulations integrate radiance over the FOV with 3-D heterogeneous cloud field. Any difference between 3-D and 1-D is as a result of BFE, and that between 3-D and IPA is as a result of 3-D RT effects. The results showed that, in all the simulations with different instruments, BFE is a dominant effect of cloud inhomogeneity, and that it also depends on the aspect ratio of the ice particles’ sizes. It is also mentioned that BFE depends on the sensor’s footprint dimension and the frequency under investigation. However there is generally good agreement between IPA and 3-D RT results, which indicates that 3-D RT effects do not have significant impact on the observed radiance. The authors tried to explain that the small impact of 3-D RT effect could be that it produces a positive and a negative effects on the pencil beam results which cancels out when integrated over the sensor’s FOV.

Also in Zhang et al. (2012), the effects of cloud horizontal inhomogeneity and drizzle was investigated on effective radius of cloud droplet retrievals at 2.1 \( \mu m \) and 3.7 \( \mu m \) spectral region. In this paper 3D RT effects were investigated, where the impact of 3-D effects was determined by comparison between 3-D and 1-D RT simulations. The results showed that 3-D RT effects such as illumination and shadowing induced significant differences between the effective radius retrieved as a result of reflectance at 2.1 \( \mu m \) and 3.7 \( \mu m \). The 3-D RT effects here is defined as the difference in effective radius obtained at 3.7 \( \mu m \) and at 2.1 \( \mu m \). In illumination it was realized that 3-D RT effects have a stronger influence on the effective radius observed at 2.1 \( \mu m \) than that of 3.7 \( \mu m \) resulting to a positive difference, while the opposite is the case in shadowing which gives negative difference. This contradicting effects results to an overall agreement between the effective radii observed at these spectral regions, as the 3-D RT effects tend to cancel each other. Hence a very small net impact of 3-D RT effects on effective radius retrieval was observed.

Apparently, cloud inhomogeneity effects on remote sensing retrievals are ongoing research and there are still a lot to learn about them. Especially in the 3-D geometry which takes into account the scattering properties of ice cloud particles, polarisation, and induced influence of outside atmospheric properties into the sensor’s line of sight. The field of 3-D RT is an area of considerable effort, and its simulations are computational demanding.

The motivation of this thesis is to study the three possible ways that cloud inhomogeneity can impact on the measurements in microwave and sub-mm regions. However the scope of this study is limited to determine the cloud inhomogeneity effects as a result of the so-called beamfilling. As can be read from example Lafont and Guillemet (2004); Wilheit et al. (1977), BFE was found to be a major source of error in retrieved rainfall rate with a possible factor of 2, and also in Davis et al. (2005)
BFE was the major source of error in cirrus cloud observation. But with the other effects in mind to be study later, the RT simulations in this study is carried out with ARTS-MC, which has the capability of handling 3-D RT calculations.

In this thesis cloud inhomogeneity effect is viewed as the difference in the signal (radiance but in terms of brightness temperature (BT)) obtained by the sensor when the cloud parameters in the FOV of the sensor is heterogeneous, and when these parameters, of the same atmosphere, are averaged with the sensor’s weighting function in its FOV. Hence a comparison between 1-D pencil beam calculation is carried out with an atmospheric model data (without averaging its parameters), and that of averaged atmospheric parameters (which is referred to as a homogenized scene). However in order to estimate the difference in BT that is from the clear-sky already, a clear-sky RT simulation is performed in a RT forward model, ARTS, to determine the magnitude of clear-sky inhomogeneity. In both cases, i.e. cloudy and clear-sky, the simulations are carried out with down looking sensor at different frequencies. The purpose of this study is to investigate cloud inhomogeneity effects on the observed radiance (in BT [K]) at these frequencies with a defined sensor’s footprint.

The footprint resolution of the sensor used in these simulations are taken to be 15km in respect to Ice Cloud Imager (ICI). ICI is a planned down looking sub-millimeter radiometer to be carried on the MetOp-SG-B satellite that will be launched in 2020. So understanding the effects of cloud inhomogeneity at the spectral regions in which ICI will be operating will help to understand its future data.

The thesis is structured in the following ways: Chapter 2 presents the advantages of passive microwave remote sensing in cloud ice observations, and also discusses passive microwave sensor ICI. Chapter 3 discusses the background of radiative transfer, and the RT model, ARTS, used for simulations in this thesis. It also discusses the theory behind clear-sky and cloudy (scattering) RT. Chapter 4 discusses the atmospheric model data, and the methodology used in this study. Chapter 5 presents the results of the findings. Finally Chapter 6 presents the conclusions.
Microwave remote sensing

“Microwave remote sensing entails the physics of radio wave propagation in and interaction with material media, including surface and volume scattering and emission; the techniques used for designing microwave sensors and processing the data they acquire; and translation of the measured data into information about the temporal or spatial variation of atmospheric or surface and medium parameters or properties” (Ulaby et al., 1981). This chapter discusses the essence of microwave remote sensing and its applications in respect to ice cloud measurements.

Passive microwave remote sensing has some advantages in comparison with other space-borne cloud ice remote sensing techniques. Microwave radiation has the capacity to penetrate clouds, and some degree rain, and it is independent on the sun as a source of illumination (day and night capability i.e. independent of intensity and angle of sun illumination). It is more sensitive to ice water path and particle sizes as the wavelengths are quite similar to cirrus ice crystals’ sizes. Sub-millimeter wave frequencies, when compared to 94GHz Cloud Radar System, are more sensitive than the lower frequencies (Evans et al., 2005). Visible and near infrared remote sensing techniques present difficulties in cloud retrievals over bright surfaces (ice and snow), i.e. cannot measure lower optical depth, Platnick et al. (2001), and only work during daytime (Rolland et al., 2000). Thermal infrared technique can only detect effective radius of small ice particles and goes into saturation for moderate optical depth. According to Stephens et al. (2002), 94 GHz cloud profiling radar can penetrate thick clouds but it is liable to particle size distributions. It also has poor sampling due to lack of horizontal scanning ability. However in mm/sub-mm region, it is possible to measure different sizes of the cloud ice as each frequency is sensitive to different optical properties and can detect most of the cloud ice mass.

Passive MRS is carried out using microwave radiometers or sensors. This technique is possible because all natural materials emit electromagnetic radiation (at non-zero temperature), which is a complex function of physical properties of the emitting sur-
Microwave remote sensing

The passive sensors detect the radiated energy in the microwave spectrum.

Because we will have operational sub-mm sensor in the near future, called Ice Cloud Imager (ICI), the RT simulations on this study focus on the regions this sensor will be operating. Part of the motivation of this thesis is to study these regions, since we will use the data later, in order to understand the sensor’s data. ICI is a planned sub-millimeter radiometer capable of measuring thermal radiance emitted by the Earth, at high spatial resolution in specified spectral bands in the sub-millimeter wave region of the electromagnetic spectrum. EUMETSAT’s (European Organisation for the Exploitation of Meteorological Satellites) decision to include ICI on the MetOP-SG-B satellites is the latest news in the history of proposed down-looking sub-millimeter radiometers.

The primary objective of the Ice Cloud Imaging mission is to support climate (by provision of ice clouds measurements such as ice water path) and present the validation of ice clouds in weather and climate models through the provision of the ice cloud products including microphysical variables (EUMETSAT-STG, 2013). The ICI will have 11 channels and the nominal spectral bandwidth in Table 2.1 in terms of the full width at half maximum (FWHM).

<table>
<thead>
<tr>
<th>Channel</th>
<th>Frequency (GHz)</th>
<th>Bandwidth (MHz)</th>
<th>Utilization</th>
</tr>
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<tbody>
<tr>
<td>ICI-1</td>
<td>183.31±7.0</td>
<td>2x2000</td>
<td></td>
</tr>
<tr>
<td>ICI-2</td>
<td>183.31±3.4</td>
<td>2x1500</td>
<td>Water vapor profile, cloud ice water path</td>
</tr>
<tr>
<td>ICI-3</td>
<td>183.31±2.0</td>
<td>2x1500</td>
<td></td>
</tr>
<tr>
<td>ICI-4</td>
<td>243.2±2.5</td>
<td>2x3000</td>
<td>Quasi-window, cloud ice water path, cirrus clouds</td>
</tr>
<tr>
<td>ICI-5</td>
<td>325.15±9.5</td>
<td>2x3000</td>
<td></td>
</tr>
<tr>
<td>ICI-6</td>
<td>325.15±3.5</td>
<td>2x2400</td>
<td>Cloud ice effective radius</td>
</tr>
<tr>
<td>ICI-7</td>
<td>325.15±1.5</td>
<td>2x1600</td>
<td></td>
</tr>
<tr>
<td>ICI-8</td>
<td>448±7.2</td>
<td>2x3000</td>
<td></td>
</tr>
<tr>
<td>ICI-9</td>
<td>448±3.0</td>
<td>2x2000</td>
<td>Cloud ice water path and cirrus cloud</td>
</tr>
<tr>
<td>ICI-10</td>
<td>448±1.4</td>
<td>2x1200</td>
<td></td>
</tr>
<tr>
<td>ICI-11</td>
<td>664±4.2</td>
<td>2x5000</td>
<td>Cirrus clouds, cloud ice water path</td>
</tr>
</tbody>
</table>

Table 2.1: ICI channels. Table courtesy of (EUMETSAT-STG, 2013)

From the Table 2.1, the channel, for example, ICI-1 is defined to have a central frequency of 183.31 GHz with two (upper and lower) sidebands of 7 GHz from the central frequency. With the sidebands having bandwidths of 1000 MHz at each side of the side bands.

The characteristics of ICI are that the instrument is a conical scanner with an observation zenith angle of $53^\circ \pm 2^\circ$ and a total azimuthal scan of $120^\circ$, will orbit at 820km above the Earth’s surface in a Sun-synchronous orbit, has 9.30am equatorial crossing time, swath width $\approx 1500$ km, and $<15$ km ground resolution.
Figure 2.1 shows the BT spectrum over channels ranging from 160 GHz to 664 GHz. The thick vertical lines show the central frequencies of all the ICI channels, while the vertical dash lines indicate the sizes of the sidebands with their respective bandwidths. The channels at 183.3, 325.15 and 448 GHz are water absorption lines. This can be seen at Figure 2.1 with a sharp deep curve at these frequencies. The channels 234 and 664 GHz are polarized so as to obtain information on cloud orientation and asphericity (Buehler et al., 2012).
3.1 Radiative transfer

Radiative transfer is the transfer of radiation in the form of electromagnetic waves through a medium, which is affected by absorption, emission, and scattering processes. Absorption, emission and scattering are the three major phenomena, at which electromagnetic waves interact with a medium. The medium could be clouds, water vapour, the sun, sea, rain, vegetation etc. In respect to this thesis, radiation is emitted by sea surface (thermal radiation), transferred through the atmosphere, and finally either escaping into space, or absorbed by the sea surface, by a satellite sensor, or the Earth’s atmosphere. Radiative transfer equation (RTE) is used to mathematically describe the interaction of radiation with a medium. For simplicity and yet valid results, the RTE is solved under these conditions:

1. The conditions of local thermodynamic equilibrium (LTE), where temperature is well-defined from collisions between molecules. A valid condition for LTE is that the rate of collisions between molecules should be much higher than the rate of spontaneous energy level changes within the molecules (Thomas and Stamnes, 2002). LTE is also fulfilled if the medium density is sufficiently high, the frequency is sufficiently low, and the intensity is not too high. The photon energy should be small compared to typical kinetic energy of a molecule: $h\nu \ll k_bT$ (breakdown at high frequencies and low temperatures). These conditions are not true in all layers of the atmosphere. For example, they are true for terrestrial radiation in the troposphere and stratosphere of Earth, but invalid in the upper atmosphere and for radiation at UV or higher frequency.

2. The condition of fully elastic scattering, which implies that the frequency of radiation remains unchanged before and after a scattering event. This is valid for the radiative transfer of terrestrial radiation, where elastic scattering is assumed.
If these conditions are fulfilled, the scalar radiative transfer equation SRTE, which assumes that the radiation field is unpolarized, Eriksson et al. (2011b), is given as

$$\frac{dI(\nu, r, \hat{n})}{ds} = -\xi_{11}(\nu, r, \hat{n}) I(\nu, r, \hat{n}) + \alpha(\nu, r, \hat{n}) B(\nu, T(r)) + \int_{4\pi} d\hat{n}' Z_{11}(\nu, r, \hat{n}, \hat{n}') I(\nu, r, \hat{n}')$$

(3.1)

where $I$ is the spectral radiance as a function of frequency $\text{Wsr}^{-1}\text{m}^{-2}\text{Hz}^{-1}$, $Z_{11}$ is the scattering phase function, $r$ represents the atmospheric position, $\hat{n}$ and $\hat{n}'$ are (unit vectors) propagation directions which indicate outgoing and incoming direction respectively, $s$ is the path length through the medium, $\nu$ is the frequency in Hz, $\xi_{11}$ is the extinction coefficient, $\alpha$ is the absorption coefficient, $T$ is the local temperature, and $B$ is the Planck function that describes black body radiation, giving in equation (3.2).

Equation (3.1) describes the change of radiation along the line of sight. The three terms on the right of Equation (3.1) describe physical processes which the radiation through an atmosphere that contains different trace gases and particle types. These three terms include:

- Extinction term, $-\xi_{11}(\nu, r, \hat{n}) I(\nu, r, \hat{n})$. The negative sign before this term shows losses through extinction of radiation. Losses of radiation can occur through absorption and/or scattering of radiation from the line of sight, i.e. $\xi_{11} = \alpha + s_1$, where $s_1$ is the scattering coefficient. For microwave radiation in cloudy atmospheres, extinction is caused by gaseous absorption, particle absorption and particle scattering.

- Thermal (emission) term, $\alpha(\nu, r, \hat{n}) B(\nu, T(r))$. This describes thermal emission by gases and particles in the atmosphere.

- Scattering source term, $\int_{4\pi} d\hat{n}' Z_{11}(\nu, r, \hat{n}, \hat{n}') I(\nu, r, \hat{n}')$. This indicates the amount of radiation which is scattered from all directions $\hat{n}'$ into the direction of propagation $\hat{n}$.

The absorption cross-section $\sigma_a [m^2]$ of a particle or molecule describes how much radiation is absorbed by the molecule or particle. Absorption coefficient $\alpha [m^{-1}]$ is the absorption cross-section $\sigma_a [m^2]$ per unit volume $[m^3]$. The absorption coefficient is related to the absorption cross-section through $\alpha_1 = n \sigma_a$, where $n$ is the number density of the particles (Rees and Rees, 2012). The absorption efficiency $Q_a$ is the ratio of $\sigma_a$ and the physical cross-section, and it describes what fraction of the surface is absorbed. The cross-section $\sigma_s$, scattering coefficient $s_1$, and efficiency $Q_s$ are also defined correspondingly, but in respect to radiation scattering.

The scattering phase function $Z$ describes the distribution of radiation after a beam is scattered or it can be thought of as a probability density function showing the chances of a photon of light being scattered in a specific direction. In Equation (3.1), the scattering phase function is normalized with respect to $s_1$, e.g $\int_{4\pi} Z_{11} dn = s_1$.
Note that we are interested in the SRTE because all the RT simulations in this thesis are performed to solve it.

Thermal radiation is emitted by all objects with non-zero temperature, Rees and Rees (2012), including the Sun, the Earth’s surface, and the Earth’s atmosphere, ocean surfaces etc. Radiation which is emitted by the Sun and then reflected by Earth is referred to as solar radiation, while radiation that is emitted by the Earth is known as terrestrial, long wave, or thermal radiation. In this thesis, we are interested the latter. The electromagnetic radiation inside a closed cavity, held at an absolute temperature $T$ with opaque walls is known as black-body radiation. The energy emitted by this body is described by Planck’s law as a function of frequency ($\nu$):

$$B_{\nu,p} = \frac{2h\nu^3}{c^2(\epsilon e^{\frac{kT}{h\nu}} - 1)}, \quad (3.2)$$

In these equations, $h = 6.626 \times 10^{-34}$ Js is the Planck constant, $k = 1.381 \times 10^{-23}$ J/K is the Boltzmann constant, $c = 2.99792 \times 10^8$ m/s is the speed of light, $\nu$ is the frequency in Hz, $T$ is the absolute temperature in Kelvin [K], and $B_{\nu,p}$ is the spectral radiance as a function of frequency in Wsr$^{-1}$m$^{-2}$Hz$^{-1}$.

Note: real bodies are not blackbodies, but emit energy at a certain emissivity $\epsilon$:

$$B_{\nu} = \epsilon(\nu)B_{\nu,p} \quad (3.3)$$

where by definition, a blackbody has $\epsilon = 1$. When $\epsilon$ is constant as a function of $\nu$; the source is, by definition, a gray body.

However, at a sufficiently long wavelengths or low frequencies (eg. radio- and microwave), equation (3.2) can be approximated as

$$B_{\nu} \approx \frac{2kT\nu^2}{c^2} = \frac{2kT}{\lambda^2} \quad (3.4)$$

This approximation is known as the Rayleigh-Jeans approximation and the condition for this approximation to be valid is

$$\frac{h\nu}{kT} \ll 1 \quad (3.5)$$

This implies the Rayleigh-Jeans approximation is valid at low frequencies and not too low temperatures.

Sometimes it is more convenient, at low frequencies like in microwave region, to define the brightness temperature of a body that is emitting thermal radiation. Brightness temperature (BT) is the temperature of the equivalent blackbody that would give the same radiance at the wavelength under consideration (Rees and Rees,
BT is used because it is closely related to the physical temperature of the radiating body. Also it has no dependency on the choice of spectral parameter, that is, wavelength, or frequency, or wave number. In this study, the signal (radiance) that is obtained by the passive microwave sensor is converted to BT in the RT simulations. So it is used to deduce the effects of cloud inhomogeneity in passive microwave data.

### 3.1.1 Clear-sky radiative transfer

Clear-sky radiative transfer is the radiative transfer in a purely gaseous atmosphere without particles which could cause scattering of the radiation. This is achieved if scattering can be neglected and the atmosphere is assumed to be in LTE. These assumptions are normally valid for the infrared region and longer wavelengths as in the microwave region. In reality, gases and aerosols scatter radiation. For this study, aerosols are not taking into account, and for terrestrial radiation, scattering from gases can be neglected. Aerosols are minute particles which are suspended in the atmosphere, which causes scattering of radiations when they are sufficiently large. Physical processes in clear-sky radiative transfer are still important in cloudy (scattering) radiative transfer and the solution for clear-sky radiative transfer can be considered as a subset of the solution for cloudy radiative transfer. Having neglected scattering, the relevant processes for clear-sky radiative transfer are absorption and emission. Hence Equation (3.1) reduces as

\[
\frac{dI(\nu, r, \hat{n})}{ds} = \alpha(\nu, r, \hat{n})[B(\nu, T(r)) - I(\nu, r, \hat{n})],
\]

where \(\xi_{11} = \alpha\) (Kirchoff’s law), and \(Z = 0\) (scattering phase function). From these conditions the atmospheric absorption and emission are taken to be the same and the basic problem to determine the radiative transfer is to calculate the absorption. Hence neglecting emission, e.g. because the background radiation is very strong, Equation (3.6) becomes the Beer’s law, which relates the attenuation of radiation or light through a medium to it’s optical depth,

\[
I = I_0 e^{-\tau}
\]

where \(I_0\) can represent thermal emission from the surface, solar radiation at the top of the atmosphere or cosmic background radiation depending on the observation geometry, and \(\tau\) is the optical depth. When discussing radiative transfer the quantity optical depth, \(\tau\), is commonly used and it is defined as:

\[
\tau = \int_0^t \alpha ds'
\]

Electromagnetic radiation interacts with different gases present in the atmosphere and each constituent gas absorbs radiation at specific regions in the electromagnetic spectrum. An absorption line is expressed as the corresponding absorption coefficients as a function of frequency \(\alpha(\nu)\) by Goody and Yung (1989) as
\[ \alpha(\nu) = nS(T)F(\nu) \]  

(3.9)

where \( S(T) \) is the line strength which is a function of temperature, \( n \) is the number density of the absorber, \( T \) is the temperature, and \( F(\nu) \) is the line shape function. The line shape function is normalized as \( \int F(\nu)d\nu = 1. \)

### 3.1.2 Scattering

Scattering of electromagnetic radiation from the Earth’s surface is an important phenomenon in most remote sensing situations. The scattering of sunlight which incident on the Earth’s atmosphere results in white clouds, blue skies, and different optical effects like rainbow (which is formed by the refraction and internal reflection of sunlight in raindrops), bright halos (formed by the refraction of light by hexagonal, prism shaped ice crystals in high, thin cirrostratus cloud decks), coronas (which is formed by the diffraction of light in water drops in low or sometimes in middle cloud deck), and glories (Wallace and Hobbs, 2006). The effect of scattering reduces the amount of energy contained in the incident wave.

The scattering characteristics between a particle and incident radiation depend on the particle shape and orientation, refractive index of the material, and on the particle size in respect to the wavelength of the radiation. The size parameter relates the particle size to the wavelength of the incident radiation

\[ \chi = \zeta r = \frac{2\pi r}{\lambda} \]  

(3.10)

where \( \lambda \) is the wavelength of the radiation, \( \zeta = \frac{2\pi}{\lambda} \) is the angular wave number, and \( r \) is the radius of the particle (equation (3.10) assumes a spherical particle). The parameter \( \chi \) indicates how radiation interacts with the particle, and it categorizes the scattering behavior in different regimes:

- **Rayleigh scattering regime** describes the interaction of very small particles with radiation of a much longer wavelengths, resulting in size parameters \( \chi \ll 1 \). This regime normally applies to solar radiation and air molecules and/or small aerosols. These particles are not relatively effective at scattering radiation, and the radiation is in phase across the whole particles. In this situation, the particle will absorb much less power than the power that is carried by the radiation cross section equal to the geometrical area of the particle. For Rayleigh scattering, the scattering efficiency \( Q_s \) is proportional to \( \lambda^{-4} \), hence radiation scattering is weak and even gets weaker with an increasing wavelengths.

- **Mie regime** is used to describe the scattering behavior of a particle, if the particle’s size and the incident wavelength are approximately of the same size (\( \chi = 1 \)). As it can be seen from Figure 3.10, ice cloud particles with radii
up to 1000µm lie within Mie scattering regime. Mie scattering is responsible for white appearance of clouds.

- Geometric optics regime ($\chi \gg 1$) is used to describe scattering of particles with size parameter much larger than 1. For very large particles, the scattering does not depend on the radiation wavelength, unless the refractive index has a significant wavelength dependence. This is known as non-selective scattering.

![Figure 3.1: Scattering regimes with particle radius on y-axis and spectral regions (in wavelength) on x-axis. Image courtesy of Wallace and Hobbs (2006)](image)

Figure 3.1 shows the different scattering regimes, with the particle radius $r[\mu m]$ on the y-axis and the spectral regions $\lambda[\mu m]$ on the x-axis. The diagonal lines distinguish the scattering regimes from each other.

### 3.2 ARTS

The Atmospheric Radiative Transfer Simulator (ARTS) is a radiative transfer model, used in this thesis, for the simulations of atmospheric radiative transfer. ARTS can simulate polarized radiances in up to three spatial dimensions in any geometry above an ellipsoidal planet. It focuses on thermal radiation from the microwave to the infrared spectral range. It is a flexible radiative transfer model and could easily allow for definition of observation geometry (including scanning features), and sensor characteristics. In ARTS, RT simulations can be performed from any position and along any direction, in as much as the simulations make sense with respect to the model atmosphere (Eriksson et al., 2011a).
In order to fulfill aforementioned aim of this thesis, the RT simulations in ARTS are carried out in two ways which include clear-sky RT and scattering (cloudy case, as it will be referred to) RT. The clear-sky simulations are performed to basically determine the magnitude of the clear-sky inhomogeneity. This implies that we could estimate the magnitude of cloudy inhomogeneity effect that we get from the cloud. While of course, the cloudy case RT simulations are performed to determine the cloud inhomogeneity effects on the passive mm/sub-mm data.

3.2.1 Clear-sky case

Clear-sky radiative transfer in ARTS refers to simulation of radiative transfer where the influence of "particles" is ignored, and only "absorbing species" are relevance. Here the term "particles" treats all matters causing significant scattering. However, a clear-sky simulation can involve eg. cloud water droplets, but on the condition that the wavelength of the radiation is such that scattering can be ignored.

The clear-sky simulation is used to solve Equation (3.6) which treats a single frequency and a single direction, at a time, and can be referred to as monochromatic (infinite frequency resolution) pencil beam (infinite spatial resolution) radiative transfer. For this, a sensor model is required because a practical instrument gives consistently spectra deviating from the hypothetical monochromatic pencil beam spectra provided by the atmospheric part of the forward model (Eriksson et al., 2003). So a hypothetical nadir looking sensor is defined in ARTS for the clear-sky simulation.

The clear-sky simulation is used to study the effects of non-uniform trace gases, and atmospheric temperature in the line of sight of the passive mm/sub-mm sensor. In order to determine clear-sky inhomogeneity, clear-sky RT simulations are performed in ARTS in the following ways:

- First, a pencil beam forward calculation is performed with the clear-sky atmospheric data (without averaging the atmospheric parameters in the data over the sensor’s FOV), and the simulation result (which is obtained in terms of BT) can be designated with an acronym $BT_{IPA}$

- Second, the result (BT) obtained in the pencil beam forward calculation above, is then averaged with the sensor’s FOV and can be designated as $BT_{IPAFOV}$

- Lastly, a pencil beam forward calculation is performed with the same clear-sky atmospheric data, but in this case the atmospheric parameters in this data are averaged with the sensor’s FOV (i.e. the sensor’s weighting function). Because the atmospheric parameters are averaged over the sensor’s FOV, it is referred to as homogenized atmospheric scene and the simulation’s result (BT) can be designated as $BT_{ATMFOV}$

The clear-sky inhomogeneity is then defined as the difference in BT [K] given in Equation 3.11.
\[
\Delta BT = BT_{IPAFOV} - BT_{ATMFOV}
\]  

The sensor’s weighting function is a 2D Gaussian function with a FWHM of 1.048°. Here, the original footprint parameter of the atmospheric data receives the highest weight (i.e., having the highest Gaussian value) of the sensor’s weighting function, and the neighboring footprint parameters receive smaller weights as their distance from the original footprint parameter increases.

The sequences at which these simulations are carried out is shown in Figures 3.2 and 3.3:

**Figure 3.2:** The sequence of clear-sky pencil beam RT simulation for the scene inhomogeneous atmosphere.

**Figure 3.3:** The sequence of clear-sky pencil beam RT simulation for the atmospheric homogenized.

### 3.2.2 Cloudy (scattering) case

Cloudy case RT simulation incorporates scattering of radiation. As a rule, scattering is very complex to model in radiative transfer because the shape and size distribution of the scattering particles are highly variable quantities. Particle here can be seen as any liquid or solid object capable of causing scattering of incident radiation. The presence of such an object changes the electromagnetic field that would otherwise exist in an unbounded homogeneous space (Eriksson et al., 2011b).
For the cloudy case, there are two different solvers in ARTS which include the Discrete Ordinate ITerative (DOIT) method, and the ARTS Monte Carlo scattering module (ARTS-MC). Although DOIT module can be implemented for 1D and 3D atmosphere, it is strongly recommended for only 1D as ART-MC is much more appropriate for 3D calculations (Eriksson et al., 2003). Since one of our motivations is to study 3-D RT effects on mm/sub-mm observations in the future, ART-MC is used in this study for the cloudy case RT pencil beam simulations in order to make comparison of results with the same RT model solver.

ARTS-MC implements a reversed MC approach, thus it traces a large number of photons backwards from the sensor in randomly selected multiple scattered propagation paths to either their point of emission, or entry into the scattering domain. It also accounts for dichroism (i.e. change of polarisation of a beam as it passes through particles). The overview and implementation of ARTS-MC is described in (Davis et al., 2005)

Unlike in the clear-sky case, cloudy case is used to study the effects of cloud inhomogeneity in the observed radiance in the line of sight of a passive mm/sub-mm sensor. The sequence of this analysis is the same with that of the clear-sky case, except that since particle scattering is involved, and ARTS-MC is used for the RT simulation. The cloud inhomogeneity effect is shown as the difference between the BT of inhomogeneous atmosphere (as described above in clear-sky case) over the sensor’s FOV, and the BT of the atmospheric homogenized scene as in Equation (3.11). It should be noted that any difference arising between these two ARTS-MC pencil beam simulation results is as a result of BFE which was discussed earlier in Chapter 1.
Chapter 4

Data and Methodology

In this chapter, we will discuss the atmospheric model data, the frequency channels, and the methodology which are used to fulfill the objectives of this thesis. The set up for the clear-sky and cloudy pencil beam RT simulations are also discussed.

4.1 Data

In this study a global atmospheric model, Nonhydrostatic icosahedral atmospheric model (NICAM), is used for all the RT simulations. NICAM is an ultra-high (3.5 km) resolution atmospheric circulation model which is designed to perform “cloud resolving simulations” by directly calculating deep convection and meso-scale circulation that play key roles not only in the tropical circulations but in the global circulations of the atmosphere. NICAM data can be used for both short term numerical predictions for weather systems such as a week, and long term simulations to obtain quasi-equilibrium climate states (Satoh et al., 2008).

Higher resolution computations are continuously demanded for global atmospheric modeling, and NICAM adopts non-hydrostatic governing equations and icosahedral grids in order to drastically enhanced horizontal resolution since the cores of deep convection have a few kilometer in horizontal. NICAM is used since its spectral resolution (3.5 km) of NICAM data is higher than that of the sensor’s ground resolution. For formulation of NICAM as well as the results of its global cloud resolving simulations, the reader should read (Satoh et al., 2008), and for more details on NICAM visit http://nicam.jp/hiki/.

NICAM is a 40 level data set consisting of atmospheric profiles of temperature, water vapour, cloud ice, cloud water, snow, rain e.t.c.. These levels are defined by atmospheric pressure which ranges from 1002 hPa to 3.5hPa, corresponding to altitude of about 80 m above the sea surface, and 38 km respectively.
Since a NICAM is a global data, a specific region of interest was selected for this study. For the clear-sky case, lat(9°N,13°N) and lon(101°W,107°W) region was selected. Because of the complexity of the cloudy case simulations, and with higher computational time in mind, a smaller region lat(10.5°N, 11°N) and lon(101°W,102°W) was selected. This region, for the cloudy case, is small enough that it covers the ground resolution of the sensor.

These regions were selected where there is a tendency of having cloud inhomogeneity in the atmosphere, by looking particularly at the distribution of the ice cloud over the regions. They are chosen over sea surface because we expect the background to be more homogeneous, so that the cloud inhomogeneity effects that we will get is actually coming from the cloud and not from the radiative surface. The reason we do not want to mix a land and sea surface is that land has very different surface reflectivity from ocean, and also surface reflectivity changes with surface type over land. Although ocean reflectivity depends on wind speed, Hollinger (1971), it is neglected in this study.

4.2 Selection of frequencies

The frequencies selected for this study are 181.5, 327, 456, 494.5, and 663.5 GHz. These frequencies were selected because they have the similar temperature Jacobians, i.e. changes in BT with respect to changes in the local atmospheric temperature, in the clear-sky case. Temperature Jacobian is given as:

\[ \text{Jacobian}[K/K] = \frac{\partial(BT)}{\partial T} \]  

where \( \partial(BT) \), which is a change in BT, is a measurement/derived vector (in our case, BT is the simulation result), and \( \partial T \) is the state vector.

The clear-sky temperature Jacobians were used because at the same altitude, they ensure the same clear-sky signal from the different frequency channels. If temperature Jacobians of the cloudy case had been used, they would have looked completely different in each profile case and also in different altitude.

In order to select these frequencies with similar temperature Jacobians, a clear-sky RT simulation was made in ARTS with range of frequency between 160 - 664 GHz, two sensor positions (i.e. nadir looking sensor and observation zenith angle of 53°), and a so-called statistically analyzed atmospheric data (SAAD) are used. SAAD consists of 9 atmospheric profiles which comprises combination of mean and standard deviations (std) of pressure, temperature, and water vapor over the 40 levels of NICAM data. The essence of the SAAD was to create a smaller number of atmospheric profiles from that of NICAM data selected for clear-sky RT simulations which contains about 20000 atmospheric profiles. The idea is that at the selected frequencies, the temperature Jacobian should be quite similar in all the atmospheric
profiles used in the simulation/calculation. So comparison over 9 atmospheric profiles of SAAD is more time efficient than that of NICAM profiles. In order to create the profiles of SAAD, the mean and std of the NICAM profiles’ parameters were calculated over the defined 40 levels of the atmosphere. Table 4.1 shows the arrangement of SAAD used for the temperature Jacobian calculations.

<table>
<thead>
<tr>
<th>Pressure (P) [Pa]</th>
<th>Temperature (T) [K]</th>
<th>Altitude (z) [m]</th>
<th>Water vapor (H20) [kg/kg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean P</td>
<td>Mean T</td>
<td>z</td>
<td>Mean H20</td>
</tr>
<tr>
<td>Mean P</td>
<td>Mean T</td>
<td>z</td>
<td>Mean + std H20</td>
</tr>
<tr>
<td>Mean P</td>
<td>Mean T</td>
<td>z</td>
<td>Mean - std H20</td>
</tr>
<tr>
<td>Mean + std P</td>
<td>Mean T</td>
<td>z</td>
<td>Mean H20</td>
</tr>
<tr>
<td>Mean - std P</td>
<td>Mean T</td>
<td>z</td>
<td>Mean H20</td>
</tr>
<tr>
<td>Mean P</td>
<td>Mean + std T</td>
<td>z</td>
<td>Mean H20</td>
</tr>
<tr>
<td>Mean P</td>
<td>Mean - std T</td>
<td>z</td>
<td>Mean H20</td>
</tr>
<tr>
<td>Mean + std P</td>
<td>Mean + std T</td>
<td>z</td>
<td>Mean + std H20</td>
</tr>
<tr>
<td>Mean - std P</td>
<td>Mean - std T</td>
<td>z</td>
<td>Mean - std H20</td>
</tr>
</tbody>
</table>

Table 4.1: Atmospheric parameters’ arrangement of the 9 profiles of SAAD

Figure 4.1 shows the mean parameters of temperature, water vapor, and pressure in comparison with the clear-sky NICAM data over the altitudes. The mean temperature and water vapor can be seen by thick black lines in the upper panels in this figure. The mean pressure, at the lower panel, is almost the same with the atmospheric pressure profiles as pressure does not vary much within an altitude.

With this SAAD set up, clear-sky temperature Jacobians were calculated over the range of frequencies (160 - 664 GHz). And the five (5) above mentioned frequencies were selected. Figure 4.3 shows that these selected frequencies have similar temperature Jacobian not just in the first profile of SAAD but also in all 9 profiles of SAAD at both sensor’s positions. The left panel shows the temperature Jacobian simulated at nadir direction, and the right panel shows that of observation zenith angle of 53°. For the slant down looking sensor (53°), the peak of the temperature Jacobians of all the selected frequencies is at 9.1544km, and that of the nadir direction is at 8.3439km. This implies that the sensors (at the different viewing angles) are most sensitive to the simulated signals at these altitudes respectively. It can also be seen that the sensor looses its sensitivity at or slightly above 5km in both viewing directions.

Figure 4.2 shows the variation of the SAAD profile parameters over the atmospheric altitude. It can be seen that there is little or no variation in pressure within each altitude.

Figure 4.4 shows the BT spectrum of the first profile of SAAD against the fre-
quency channels, with the positions of selected frequencies. These set of frequencies are selected, apart from the fact that they have similar temperature Jacobians, with the following reasons:

- They are spread apart, so that we can compare the cloud inhomogeneity effects at these frequencies as they are sensitive to different optical properties of the cloud, particularly they are sensitive to different ice particle sizes. As a rule, cloud scattering increases with frequency i.e. with increase in frequency the scattering behavior of the particles move from Rayleigh to Mie scattering. Scattering efficiency increases with size parameters but only to a certain point and then goes in to saturation.

- For the clear-sky case, we want to know if the sensor obtain the same temperature values at these frequencies. If it obtains the same temperature at these selected frequencies, despite the atmospheric conditions, it implies the sensor gets its signal from the same point in the atmosphere.

Figure 4.1: Atmospheric parameters’ profiles of the clear-sky NICAM data in comparison with the mean of these respective parameters. Upper left panel: temperature profiles. Upper right panel: water vapor profiles. Lower panel: pressure profiles
4.2. Selection of frequencies

- There was a restriction (as part of the motivations of this thesis) to have one of the selected frequency channels to be in one of the ICI channels, which is at 664 GHz, which limited us to this set of frequencies.

- Looking at Figure 4.4, it can be seen the selected intersects the BT spectrum at higher values. This implies the sensor gets signal further down the atmosphere enabling to cover as much cloud as possible. Since our objective is to study cloud inhomogeneity effect, we need set of frequency channels that could measure signals at the ice cloud altitude.

In order to verify the second reason above, pencil beam RT simulation was carried with the clear-sky NICAM data and that of the so-called SAAD, and the results are shown in Figure 4.5. Table 4.2 shows the correlation values and the best fit line values (a and b) of the clear-sky RT simulation results obtained at these selected frequency. The high correlation values indicate we can make direct comparison of clear-sky inhomogeneity effects obtained in these selected frequencies.

![Figure 4.2: Plots of the SAAD parameters’ profiles over the atmospheric altitude. Upper left panel: temperature profiles. Upper right panel: water vapor profiles. Lower panel: pressure profiles. The blue lines are the mean, the green lines are the mean + std, and the red lines are the mean - std of each of these respective parameters.](image-url)
The best fit (polyfit) line is given as:

\[ \text{polyfit} = a(BT) + b \]  

(4.2)

where \( a \) is gradient of the line, \( b \) is the intercept on the vertical axis, and \( BT \) is the brightness temperature.

Figure 4.3: Temperature Jacobians of the first profile of SAAD at the selected frequencies over the atmospheric altitude. Left panel: temperature Jacobians obtained at sensor’s nadir direction. Right panel: temperature Jacobians obtained at sensor’s observation zenith angle of 53°.

Figure 4.4: BT spectrum line against frequency channels. The vertical dash lines are the positions of the selected frequency channels.
4.3 Clear-sky simulation ARTS set up

The clear-sky RT simulation was carried out at the selected frequencies with the sensor to be 2-D Gaussian with FWHM of 1.048° and at altitude of 820km. The sea surface reflectivity is taken to be 0.4. It is a rough estimate of sea surface reflectivity in the microwave region.

This simulation was set up to calculate pencil beam radiances (I) in terms of BT (by Rayleigh Jeans approximation) by solving Equation (3.6). Since scattering is ignored in the clear-sky case, absorption becomes the significant phenomenon at which the radiation interacts with atmospheric medium. The model atmosphere is set up to consist of nitrogen (78.08%), oxygen (20.95%), and water vapor as absorbing species. The ozone (O₃) is not included because according to John and Buehler (2004), the impact of ozone lines in this frequency range is very small (about 0.5 K in 183.31 ± 1.00 GHz). Hence spectroscopic effects of ozone is relatively small compared...
<table>
<thead>
<tr>
<th></th>
<th>NICAM</th>
<th>SAAD</th>
</tr>
</thead>
<tbody>
<tr>
<td>Between 327 GHz and 181.5 GHz</td>
<td>Corr.=0.9999</td>
<td>Corr.=1.0000</td>
</tr>
<tr>
<td>a = 0.9676</td>
<td>a = 0.9598</td>
<td></td>
</tr>
<tr>
<td>b = 7.7147 K</td>
<td>b = 9.6755 K</td>
<td></td>
</tr>
<tr>
<td>Between 456 GHz and 181.5 GHz</td>
<td>Corr.=0.9997</td>
<td>Corr.=0.9999</td>
</tr>
<tr>
<td>a = 0.9403</td>
<td>a = 0.9218</td>
<td></td>
</tr>
<tr>
<td>b = 14.6834 K</td>
<td>b = 19.3405 K</td>
<td></td>
</tr>
<tr>
<td>Between 494.5 GHz and 181.5 GHz</td>
<td>Corr.=0.9998</td>
<td>Corr.=0.9998</td>
</tr>
<tr>
<td>a = 0.9389</td>
<td>a = 0.9228</td>
<td></td>
</tr>
<tr>
<td>b = 14.8899 K</td>
<td>b = 18.9389 K</td>
<td></td>
</tr>
<tr>
<td>Between 663.5 GHz and 181.5 GHz</td>
<td>Corr.=0.9996</td>
<td>Corr.=0.9998</td>
</tr>
<tr>
<td>a = 0.9597</td>
<td>a = 0.9404</td>
<td></td>
</tr>
<tr>
<td>b = 10.2730 K</td>
<td>b = 15.1163 K</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.2: Shows the correlation values of BT obtained at other frequency channels with that of 181.5 GHz on both data sets with best fit line values.

to that of nitrogen, oxygen and water vapor. Although including ozone lines in the RT calculation will improve the result, there is balance between the little effect they will have on the signal obtained and computational time.

This simulation was set up to make the gas absorption calculations by pre-calculating the look-up table (LUT), which is later used in the RT simulations. LUT approach stores pre-calculated absorption cross-sections as a function of pressure, temperature, and water vapor concentrations. The look-up table is used by interpolating the table to the current atmospheric grid parameters and frequencies (Buehler et al., 2011).

Finally, this simulation was carried out by the so-called batch calculations in ARTS. In ARTS batch agenda is basically a loop function, that uses different input atmospheric parameters to perform the same RT simulations. Batch calculations enable the processing of substantial amounts of input data in an efficient way, and used to prevent performing RT calculations by calling ARTS repeatedly.

It should be noted that most of the clear-sky set up are also used in the cloudy case simulation. So clear-sky RT simulation can be viewed as a subset of the cloudy RT simulation.

### 4.4 Cloudy RT simulation ARTS set up

As mentioned earlier the cloudy case simulation was carried out with ART-MC reverse approach. Like in the clear-sky simulations, model atmosphere is set up to consist of nitrogen, oxygen, and water vapour as absorbing species, and gas absorption was calculated in the same manner. All cloudy case RT calculations are carried out with
the same selected frequencies.

Here we make some modifications in ARTS since we are doing the cloudy RT simulations. In addition to the absorption species in the model atmosphere, particles such as liquid water content (LWC), and ice water content (IWC) are included in it. The particle specie used for the LWC is ”LWC-H98_STCO”, and the of IWC is ”IWC-H13”.

The tag on the LWC implies that the particles size distributions of the LWC is according to modified gamma function in Hess et al. (1998), and that the water cloud type is continental stratus (STCO).

\[
\frac{dN}{dr} = N \alpha r^\beta e^{\frac{-\beta}{\gamma} \left( \frac{r}{r_{\text{mod}}} \right)} = N \alpha r^\beta e^{(-B r^\gamma)} \quad (4.3)
\]

where

\[
B = \frac{\beta}{\gamma r_{\text{mod}}} \quad (4.4)
\]

where \( \alpha, \beta, \gamma, r_{\text{mod}} \) are positive constants. The parameters, \( \beta, \gamma, \) and \( r_{\text{mod}} \), completely determine the shape of the distribution curve while the constant \( \alpha \) is (a normalization constant ensuring that the integral of the size distribution over all radii yields \( N \)) only related to the total number \( N \) of droplets per unit volume [cm\(^{-3}\)]. It should be noted that \( r_{\text{mod}} \) is the radius of maximum frequency and that \( \beta \) takes only positive integer. The effective radius of this size distribution is given as

\[
\text{r}_{\text{eff}} = \frac{\int r^3 \frac{dN}{dr} dr}{\int r^2 \frac{dN}{dr} dr} \quad (4.5)
\]

The values of these constants which are applied to continental stratus cloud include: \( r_{\text{mod}} = 4.7 \mu m \), \( \beta = 5 \), \( \gamma = 1.05 \), \( \alpha = 9.792 \times 10^{-3} \), \( B = 0.938 \), \( r_{\text{eff}} = 7.33 \mu m \), \( N = 250 \text{cm}^{-3} \), and liquid water content \( L = 0.28 \text{gm}^{-3} \). The value of the liquid water content (L) depends on the particle number density (N).

The tag on the IWC means that the particle size distributions (PSD) are parameterized in the form of gamma distributions in (Heymsfield et al., 2013).

\[
n(D_{\text{max}}) = N_0 D_{\text{max}}^\mu e^{-\lambda D_{\text{max}}} \quad (4.6)
\]

where \( D_{\text{max}} \) is the particle maximum dimension (is the particle diameter since they are assumed to be spherical in our study), \( n(D_{\text{max}}) \) is the particle per unit volume, \( N_0 \) is the intercept, \( \lambda \) is the slope, and \( \mu \) is the dispersion. PSD dispersion as a function of temperature is given as \( \mu = -0.84 - 0.0915T - 2.93 \times 10^{-3}T^2 - 3.653 \times 10^{-5}T^3 - 2.157 \times 10^{-8}T^4 \), for all temperatures.

The ice cloud is assumed to consists of different particle sizes but with spherical ice shapes, so the orientation of the particles is not a concern.
As cloudy RT simulation is complicated as a results of particles’ scattering, this calculation is restricted to a special atmospheric domain called cloud box. The cloud box is defined to be rectangular in the used coordinate system, with limits exactly at points of the involved grids. This implies, for example, that the vertical limits of the cloud box are two pressure levels (Eriksson et al., 2003). The cloud box is defined to cover the region: lat(20°N,20°S), lon(20°W,20°E), and pressure point between 20000 hPa and 0 hPa. This region of the cloud box is defined internally in ARTS and has nothing to do with the regions of the NICAM selected for this study since ARTS-MC is a statistical approach, the more statistical samples, which in this case is called photons, are used the more accurate it is. To have a very accurate result means that the simulation could be allowed to run forever. So in order to find a balance between accuracy and running time a threshold has to be defined. Since ART-MC simulation is run with radiance, the radiances will change a lot over the frequency range (160 - 664 GHz) that correspond to 1 K in BT. As a result we have to specify the threshold in terms of radiances over the range of frequencies we have. At each of these frequencies, the BT corresponds to largely different radiances. So the threshold should be sufficient enough for a certain BT (in our case 1 K), which implies what radiance change does BT change in the different frequencies correspond to. The threshold of the radiance change over 1 K change in BT was set to 1e-17 [Wsr⁻¹m⁻²Hz⁻¹]. This threshold value is the least value of change in radiance for 1 K change in BT over the frequency range of 160 - 664 GHz.

Another step taken to get more realistic results of the simulation is to define a threshold for water amount mass density. As the NICAM is solved in a numerical way, there is basically no lower threshold of water amount defined in it. The way it is set up and solved introduces some kind of mathematical noise, which have some values that we have to cut out. Although they will probably not make much difference in the ARTS forward calculation because of its small amount, it will take more computational time. Also we have cases where the mass density is negative, but obviously in reality we cannot have negative mass density. Another problem is NICAM predicts some liquid clouds at altitudes where it is very cold, where no liquid cloud could have existed physically. Again we could run ARTS with this but it would include unphysical data in our simulation. So a mass density threshold was set for each batch of the cloudy case simulations. Some batch calculations were made with made with mass density threshold of 1e-15 kg/m³, and some were made with 1e-8 kg/m³, and 1e-7 kg/m³. It can be seen the values are very small, we do not know if it affects our result but we hope it does not.

It should be noted here that both clear-sky, and cloudy RT simulations were carried with the sensor in nadir direction, and with the same sea surface reflectivity.
Chapter 5

Results and discussion

In the chapter, the results of the clear-sky and cloud inhomogeneity effects in the FOV of a passive microwave sensor on up-welling signals (BT) will be discussed. First we will discuss the results of clear-sky case.

5.1 Clear-sky inhomogeneity results

As explained in the Section 3.2.1, Figure 5.1 shows the pencil beam calculation result, $BT_{IPA}$ obtained from the 181.5 GHz channel in the clear-sky RT simulation. It can be seen that the $BT_{IPA}$ varies from 247 K to 261 K over the atmospheric region used for the clear-sky simulations.

![Figure 5.1: Pencil beam simulation results ($BT_{IPA}$ [K]) obtained at 181.5 GHz channel in the clear-sky RT simulation](image)

Figure 5.1: Pencil beam simulation results ($BT_{IPA}$ [K]) obtained at 181.5 GHz channel in the clear-sky RT simulation
This result is quite similar to the results obtained in the other remaining frequency channels. The similarity in the results, of course, is because the temperature Jacobians obtained at these frequencies are quite similar in the clear-sky case. In order to show how well the BT_{IPA} results made at these frequencies agree with one another, we subtract the BT_{IPA} results obtained at the other frequency channels from the BT_{IPA} result obtained at 181.5 GHz. That is,

\[
BT_{IPA}^{181.5\text{GHz}} - BT_{IPA}^{X\text{GHz}}
\]

where X in this expression represents 327, 456, 494.5, and 663.5 GHz channels. The differences are shown in Figure 5.2.

![Figure 5.2: Differences between BTs measured at the other frequency channels and that measured at 181.5 GHz channels](image)

In the Figure 5.2, the upper left panel shows the difference between the BT_{IPA} obtained at 181.5 GHz and that of 327 GHz, the upper right panel shows that of 456 GHz, and the lower left and right panel show that of 494.5 GHz and 663.5 GHz respectively. The differences in BT_{IPA} are quite different in each of the frequency channels. This is due to the fact that though the temperature Jacobians at these selected frequencies are quite similar, they are not completely exact. This can be seen
5.1. Clear-sky inhomogeneity results

From the Figure 4.3 where there are slight differences in the temperature Jacobians, but these differences are so small that it will not affect the goal of this study. From Figure 5.2, the differences in BT_{IPA} varies between -0.3 K and 1 K which are relatively small to the absolute BT_{IPA} obtained at these frequency channels.

However, in order to determine the clear-sky inhomogeneity, \( \Delta BT \), we have to compare BT_{IPAFOV} and BT_{ATMFOV} as explained in Section 3.2.1.

Figure 5.3: Pencil beam simulation results BT [K], averaged over the sensor’s FOV, obtained at 181.5 GHz channel in the clear-sky case. Left panel: Inhomogeneous atmospheric field (BT_{IPAFOV}). Right panel: homogenized atmospheric scene (BT_{ATMFOV}).

In the Figure 5.3, the left and right panel show BT_{IPAFOV} and BT_{ATMFOV} results obtained at the 181.5 GHz channel respectively. In order to determine the clear-sky inhomogeneity effect in this passive microwave sensor channel, we have to, according to Equation (3.11), subtract the results obtained in the right panel from that of the left panel in the Figure 5.3. This difference \( \Delta BT \), which is as a result of clear-sky inhomogeneity, is known as beam filling effect and it is shown in Figure 5.4.

Figure 5.4 shows the result of the clear-sky \( \Delta BT \) in the simulated BT at the 181.5 GHz channel. This result was found out to be the same with the other remaining frequency channels, this similarity is evident from Figure 5.5 where the correlations of the \( \Delta BT \) obtained at the other frequencies are approximately equal to 1 with that obtained at 181.5 GHz channel. From Figure 5.4, it can be seen that the values of the \( \Delta BT \) are always positive and that it varies from 0 to 1.5 K. This implies that in the actual inhomogeneous atmospheric field, over the regions considered in this study, the sensor sees warmer body in all the selected frequency channels. This is because BT_{IPAFOV} results are closer in reality as the atmospheric parameters in the sensor’s FOV are not averaged contrary to the averaged atmospheric parameters in BT_{ATMFOV}. However the clear-sky \( \Delta BT \) results, obtained here, could change both in sign and magnitude if the temperature and water vapor profiles are different from the atmospheric profiles used in this simulation. This can also be verified, but it was
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not reviewed in this study.

Figure 5.4: Clear-sky $\Delta BT$ [K] obtained at 181.5 GHz channels over the clear-sky atmospheric region.

Figure 5.5: Plots of clear-sky $\Delta BT$ [K] obtained at the other selected frequency channels with that obtained at 181.5 GHz channel.

In the Figure 5.4, it can be seen that the clear-sky $\Delta BT$ is well pronounced at the regions where there are steep gradient in the BT at both panels in Figure 5.3. This implies that this $\Delta BT$ effect at these regions could be as a result of non-uniformity of the local temperature and/or water vapor present in these atmospheric regions.
In order to examine the cause of this clear-sky inhomogeneity, we calculated the weighted standard deviations ($\sigma$) of the following parameters: pencil beam simulation (BT) results $\sigma_{BT}$, the local temperature $\sigma_T$ at the altitude of the peak Jacobians (8.349 km) and first two atmospheric levels above the peak Jacobians (which correspond to 9.154 km and 10.029 km respectively), and the water vapor $\sigma_{H20}$ of the atmosphere at the same altitude levels with that of local temperature. These altitudes were selected as the sensor is most sensitive to the signals obtained there. The $\sigma$ is calculated as

$$\sigma = \sqrt{\frac{\sum_{i=1}^{N} w_i (x_i - \bar{x}_w)^2}{(N' - 1) \sum_{i=1}^{N} w_i}}$$

(5.2)

where $\bar{x}_w$ is the weighted mean of the observation, $N'$ is the number of non-zero weights, and $w_i$ is the weight of the $i$th observation. Here the weighting function is the 2D Gaussian function of the sensor. It is the same Gaussian function used to average the pencil beam calculation results (as in BT_{IPAFOV}), and the atmospheric parameters (as in BT_{ATMFOV}) in the sensor’s FOV. This implies that each footprint new value, of the atmospheric parameters, is set to a weighted average of the footprint’s neighborhood. The original footprint’s value receives the highest weight (having the highest Gaussian value) and the neighboring footprints receive smaller weights as their distance to the original footprint increases.

Ideally by definition, regions with small $\sigma_{BT}$, $\sigma_T$, and/or $\sigma_{H20}$ are homogeneous regions and should result to small $\Delta BT$, as $\Delta BT$ is the inhomogeneity effect. Also regions with large $\sigma$, of these parameters, are inhomogeneous regions should result to large $\Delta BT$.

Figures 5.6, 5.8, and 5.10 show the variations of the $\sigma_{BT}$, $\sigma_T$, and $\sigma_{H20}$ over the atmospheric region considered in the clear-sky case respectively. It can be seen, from these figures, that the regions where their respective $\sigma$ are high correspond to the regions where the inhomogeneity effects $\Delta BT$ are high as in Figure 5.4, though this is not always the case as some regions in these Figures have large $\sigma$ where $\Delta BT$ is small. These inconsistencies raise a question on what characteristics of the atmospheric parameters cause the inhomogeneity effect.

Figure 5.6 shows the results of the weighted standard deviation ($\sigma_{BT}$) on BT_{IPA} obtained at the selected frequency channels. The upper left panel shows $\sigma_{BT}$ at 181.5 GHz channel, the upper right panel shows $\sigma_{BT}$ at 327 GHz channel, the middle left panel shows $\sigma_{BT}$ at 456 GHz channel, the middle right panel shows $\sigma_{BT}$ at 494.5 GHz channel, and the lower panel shows $\sigma_{BT}$ at 663.5 GHz channel. There is a systematic pattern with the results in all these frequency channels, which looks quite similar to
that in Figure 5.4. The region where $\sigma_{BT}$ is large at each of the frequency channel is where $\Delta BT$ is large, and the variation in $\sigma_{BT}$ looks quite similar in all the frequency channels.

In order to understand how $\sigma_{BT}$ corresponds to $\Delta BT$, plots of $\sigma_{BT}$ against $\Delta BT$ at each of the frequency channel are shown in Figure 5.7. In Figure 5.7 each point represents atmospheric footprint in the sensor’s FOV. The colors of these points are just to differentiate each footprint point. The positions of the panels in this Figure is the same with that of Figure 5.6. It can be seen from Figure 5.7 that there is a systematic pattern at which $\sigma_{BT}$ varies over $\Delta BT$ at these frequency channels. Some footprint points with small $\sigma_{BT}$ are in line where $\Delta BT = 0$, indicating that these regions are homogeneous as expected. But some footprint points with small $\sigma_{BT}$ result in larger $\Delta BT$, in contrary to what is expected from homogeneous regions.

As explained earlier, that the clear-sky inhomogeneity is as a result of non-uniform trace gases and local temperature of the atmospheric profiles. It is therefore important to see how $\sigma_T$, and $\sigma_{H2O}$ vary over $\Delta BT$. The calculations of $\sigma_T$, and $\sigma_{H2O}$ were carried at the same atmospheric altitudes, explained above, where the sensor is most sensitive to the radiation signals. Figure 5.8 shows the variation in $\sigma_T$ obtained at these different altitudes. It can be seen that the variation of $\sigma_T$ at these altitudes are quite small, this implies that the regions of inhomogeneity are quite similar at these altitudes. However, the results are much different from the results of $\sigma_{BT}$ in Figure 5.6. There are three distinct areas in Figure 5.8 with large $\sigma_T$, indicating inhomogeneous regions, unlike that of $\sigma_{BT}$ which only have two of such distinct areas. However, the variations in $\sigma_{H2O}$ in Figure 5.10, are quite similar to that of $\sigma_{BT}$ where both have 2 distinct area with large $\sigma$ values at the same regions. This implies that H2O vapor contributes more to the clear-sky $\Delta BT$ than local atmospheric temperature of the region of the atmosphere selected for these simulations.

Figures 5.9, and 5.11 show the plots of $\sigma_T$, and $\sigma_{H2O}$ of the atmospheric profiles, at the specified altitudes against $\Delta BT$ obtained in the 181.5 GHz channel respectively. In Figure 5.9, the upper left panel shows $\sigma_T$ at 8.349 km, the upper right panel shows $\sigma_T$ at 9.154 km, and the lower panel shows $\sigma_T$ at 10.029 km. Correspondingly, the panels in Figure 5.10 shows $\sigma_{H2O}$ at these altitudes. From Figures 5.9, and 5.11, just as in Figure 5.7, some footprints points with small $\sigma_T$ and $\sigma_{H2O}$ are in line where $\Delta BT = 0$, indicating that these regions are homogeneous as expected. However some footprint points with small $\sigma_T$ and $\sigma_{H2O}$ result in larger $\Delta BT$, in contrary to what is expected from homogeneous regions. Knowing where these deviation points are arising from will make us understand the actual characteristics of the atmosphere that cause the clear-sky inhomogeneity and to what degree. The analysis of these regions resulting in clear-sky inhomogeneity will be an extension of this study.
Figure 5.6: $\sigma_{BT}$ within the sensor’s FOV obtained at the selected frequency channels over the region in the clear-sky case.
Figure 5.7: Shows the plots of $\sigma_{BT}$ [K] obtained at each of the selected frequency channels against the clear-sky $\Delta BT$ [K] at each respective frequency channels.
5.1. Clear-sky inhomogeneity results

Figure 5.8: $\sigma_T$ within the sensor’s FOV obtained at different atmospheric altitudes over the region in the clear-sky case. Upper left panel: $\sigma_T$ at 8.349km. Upper right panel: $\sigma_T$ at 9.154km. Lower panel: $\sigma_T$ at 10.029km

Figure 5.9: Shows the plots of $\sigma_T$ [K] against clear-sky $\Delta BT$ [K] obtained in 181.5 GHz frequency channel. $\sigma_T$ at 8.349km. Upper right panel: $\sigma_T$ at 9.154km. Lower panel: $\sigma_T$ at 10.029km
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Figure 5.10: $\sigma_{H20}$ [kg/kg] within the sensor’s FOV obtained at different atmospheric altitudes over the region in the clear-sky case. Upper left panel: $\sigma_{H20}$ at 8.349km. Upper right panel: $\sigma_{H20}$ at 9.154km. Lower panel: $\sigma_{H20}$ at 10.029km.

Figure 5.11: Shows the plots of $\sigma_{H20}$ [kg/kg] against the clear-sky $\Delta$BT [K] obtained in 181.5 GHz frequency channel. Upper left panel: $\sigma_{H20}$ at 8.349km. Upper right panel: $\sigma_{H20}$ at 9.154km. Lower panel: $\sigma_{H20}$ at 10.029km.
5.2 Cloud inhomogeneity results

This section presents the results of cloud ΔBT on down looking space-borne mm/sub-mm observations.

Figure 5.12: Pencil beam simulation results BT_{IPA} [K] obtained at the selected frequency channels in the cloudy case. Upper left panel: BT_{IPA} at 181.5 GHz. Upper right panel: BT_{IPA} at 327 GHz. Middle left panel: BT_{IPA} at 456 GHz. Middle right panel: BT_{IPA} at 494.5 GHz. Lower panel: BT_{IPA} at 663.5 GHz.
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As the different frequency channels are sensitive to different optical properties (particle sizes) of the ice cloud, it is expected that cloud ΔBT to be different for each of the frequency channels under investigation. It is important to remind us here that the essence of the clear-sky case is basically to exclude the possibility that there is large ΔBT arising already from the clear-sky, and to correctly interpret the retrievals in the cloudy case.

Figure 5.12 shows the BT_{IPA} obtained at each of the selected frequency channels. The differences in the obtained BT_{IPA} is due to the fact that each channel is sensitive to different particles’ sizes in the ice cloud. This difference in sensitivity affects the amount of radiation (in terms of BT) which is obtained at each channel as scattering of radiance depends on the particle’s size in respect to the wave length of the incident radiation (χ).

Figure 5.13: Pencil beam simulation results BT [K] obtained at 181.5 GHz channel in the cloudy case. Left panel: Inhomogeneous atmospheric field (BT_{IPAFOV}). Right panel: homogenized atmospheric scene (BT_{ATMFOV}).

In Figure 5.13, the left panel shows BT_{IPAFOV}, and the right panel shows BT_{ATMFOV} results obtained at 181.5 GHz in the cloudy RT simulations. From this Figure, it is evident that there are differences in the two simulations results, as the right panel (BT_{ATMFOV}) shows more significant changes within the sensor’s FOV than that of the left panel. This difference is as a result of averaging the atmospheric parameters within the sensor’s FOV, which affects the particles’ sizes which the frequency channel are sensitive to.

Correspondingly, Figures 5.14-5.17 present the BT_{IPAFOV} and BT_{ATMFOV} results at 327, 456, 494.5, and 663.5 GHz frequency channels respectively. It can be seen from these figures, that the variation of the results (BT_{IPAFOV} and BT_{ATMFOV}) within the considered atmospheric region is in the magnitude 100 K. And that BT_{IPAFOV} and BT_{ATMFOV} are quite different within the same frequency channel, and also when
5.2. Cloud inhomogeneity results

compared with the results in the other frequency channels.

Figure 5.18 shows the cloudy ∆BT at each of the selected frequency channels. The upper left panel shows ∆BT at 181.5 GHz channel, the upper right panel shows ∆BT at 327 GHz channel, the middle left panel shows ∆BT at 456 GHz channel, the middle right panel shows ∆BT at 494.5 GHz channel, and the lower panel shows ∆BT at 663.5 GHz channel. Unlike the results we obtained in the clear-sky case, the magnitude of the cloudy ∆BT are different at each frequency channel and are not always positive in all the frequency channels. These imply that the channels are sensitive to different optical properties of the cloud.

Figure 5.14: Pencil beam simulation results BT [K] obtained at 327 GHz channel in the cloudy case. Left panel: Inhomogeneous atmospheric field (BT_{IPAFOV}). Right panel: homogenized atmospheric scene (BT_{ATMFOV}).

Figure 5.15: Pencil beam simulation results BT [K] obtained at 456 GHz channel in the cloudy case. Left panel: Inhomogeneous atmospheric field (BT_{IPAFOV}). Right panel: homogenized atmospheric scene (BT_{ATMFOV}).
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Figure 5.16: Pencil beam simulation results BT [K] obtained at 494.5 GHz channel in the cloudy case. Left panel: Inhomogeneous atmospheric field (BT_{IPAFOV}). Right panel: homogenized atmospheric scene (BT_{ATMFOV}).

Figure 5.17: Pencil beam simulation results BT [K] obtained at 663.5 GHz channel in the cloudy case. Left panel: Inhomogeneous atmospheric field (BT_{IPAFOV}). Right panel: homogenized atmospheric scene (BT_{ATMFOV}).
5.2. Cloud inhomogeneity results

Figure 5.18: $\Delta BT$ [K] at each of the selected frequency channel over the region considered in the cloudy case. Upper left panel: $\Delta BT$ at 181.5 GHz. Upper right panel: $\Delta BT$ at 327 GHz. Middle left panel: $\Delta BT$ at 456 GHz. Middle right panel: $\Delta BT$ at 494.5 GHz. Lower panel: $\Delta BT$ at 663.5 GHz.

The Table 5.1 shows the absolute magnitudes of cloudy $\Delta BT$ obtained at the selected frequency channels.
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<table>
<thead>
<tr>
<th>Frequency channels [GHz]</th>
<th>Cloud inhomogeneity effects (ΔBT [K])</th>
</tr>
</thead>
<tbody>
<tr>
<td>181.5</td>
<td>-3 - 6</td>
</tr>
<tr>
<td>327.0</td>
<td>-2 - 12</td>
</tr>
<tr>
<td>456.0</td>
<td>-2 - 16</td>
</tr>
<tr>
<td>494.5</td>
<td>-2 - 18</td>
</tr>
<tr>
<td>663.5</td>
<td>0 - 22</td>
</tr>
</tbody>
</table>

Table 5.1: The magnitude of cloudy ΔBT at the frequency channels under investigation

After determining the magnitude of cloudy ΔBT on the passive microwave measurement retrievals, we decided to characterize the cause(s) of these effects. In order to determine the cause of the cloudy ΔBT, we calculated the weighted standard (σ_BT) of the pencil beam simulation results (BT_IPA) in the cloudy case. We chose to investigate the cause cloudy ΔBT using σ_BT (instead of atmospheric parameters like IWC, LWC, local temperature, and H2O vapor) because BT_IPA is a simulation result which has taken into account all these atmospheric parameters over the altitudes of the whole atmospheric profiles considered in this study.

As described in the clear-sky results, regions with small σ_BT are homogeneous regions and should result to small ΔBT, as ΔBT is the inhomogeneity effect. Also regions with large σ_BT, are inhomogeneous regions and should result to large ΔBT. Figure 5.19 shows the variation of the σ_BT over the atmospheric region considered in the cloudy case. In this Figure, the upper left panel shows σ_BT at 181.5 GHz channel, the upper right panel shows σ_BT at 327 GHz channel, the middle left panel shows σ_BT at 456 GHz channel, the middle right panel shows σ_BT at 494.5 GHz channel, and the lower panel shows σ_BT at 663.5 GHz channel. It can be seen that there is a wide range in the σ_BT over the atmospheric region at each of the frequency channel.

In order to understand how σ_BT corresponds to cloudy ΔBT, plots of calculated σ_BT against the cloudy ΔBT at each of the frequency channel are shown in Figure 5.20. Each point in Figure 5.20 represents atmospheric footprint in the sensor’s FOV. As can be seen that there is a pattern in 327 GHz, 456 GHz, 494.5 GHz and 663.5 GHz channels, where some footprints with high values of σ_BT are in line with ΔBT = 0. This indicates that at these points where the atmospheric parameters are actually inhomogeneous, gives ΔBT = 0 which contradicts the definition of cloudy ΔBT. However, some footprint points with higher σ_BT gives higher values of ΔBT as expected. The features of the σ_BT in 181.5 GHz are quite different from that of the other channels. They tend not to vary so much from the region where ΔBT = 0. These differences in results are due to different sensitivity of these channels to different cloud particles. To characterize the atmospheric regions where these cloudy ΔBT are coming from will be an extension of this thesis.
5.2. Cloud inhomogeneity results

Figure 5.19: Calculated \( \sigma_{BT} \) at the selected frequency channels over the region considered in the cloudy case. Upper left panel: \( \sigma_{BT} \) at 181.5 GHz. Upper right panel: \( \sigma_{BT} \) at 327 GHz. Middle left panel: \( \sigma_{BT} \) at 456 GHz. Middle right panel: \( \sigma_{BT} \) at 494.5 GHz. Lower panel: \( \sigma_{BT} \) at 663.5 GHz.
Figure 5.20: Shows the plots of $\sigma_{BT}$ obtained at each of the selected frequencies against cloudy $\Delta BT$ [K] at each respective frequencies.
This thesis presents detailed 1-D pencil beam simulations, of microwave and sub-millimeter cloud observations in ARTS model, to study cloud inhomogeneity effects. The simulation results show that clear-sky inhomogeneity effects ($\Delta BT$) on up-welling signal are the same on the selected frequency channels that have similar temperature Jacobians. The clear-sky $\Delta BT$ was found to vary from 0 to 1.5 K, where $\Delta BT = 0$ is a complete homogeneous region, and $\Delta BT > 0$ is an inhomogeneous region.

The cloudy RT simulation results show that cloud inhomogeneity effects depend on the frequency of the sensor’s channel. As sensitivity of a sensor’s channel to cloud particles’ sizes depends on the frequency of the channel. This thesis has demonstrated that the cloud inhomogeneity effect, which is as a result of the beamfilling effect, is a direct consequence of non-uniform particles, trace gases and atmospheric temperature in the sensor’s line of sight. The results obtained in this study, of course, depend on the sensor’s ground resolutions and the frequencies under investigation.

This thesis tried to understand the nature of the atmospheric region that causes the cloud inhomogeneity effects. The analysis of the atmospheric regions was carried out by plotting weighted standard deviations ($\sigma$) of the $BT_{IPA}$, local atmospheric temperature and water vapor against the obtained cloud $\Delta BT$. By definition any atmospheric region with small $\sigma$ is homogeneous and should correspond to a point in the plot where $\Delta BT = 0$. These plots showed some deviations from the above definition, that is, some regions with small $\sigma$ correspond to a point where $\Delta BT = 0$ (as expected) while some regions with small $\sigma$ correspond to points where $\Delta BT > 0$. Also some of the atmospheric regions with high $\sigma$ correspond to a point where $\Delta BT = 0$, indicating that these regions where the atmospheric parameters are highly inhomogeneous do not result in cloud $\Delta BT$. Analyzing these regions to understand the reason for these deviations will be an extension of this thesis. Another future work of this study will be the investigation of cloud inhomogeneity effects as a result of 3D radiative effects.
References


