Observing and modelling the impact of arctic and tropical cirrus clouds on far-infrared radiance spectra

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The work described in this thesis concerns the effect of cirrus clouds on far-infrared (FIR) radiance spectra. Though the importance of both FIR radiation and cirrus clouds to the Earth’s energy budget is well recognised, few high spectral resolution measurements have been made at FIR wavelengths to date. Observations taken during two diverse field campaigns, along with spectra simulated using a radiative transfer model, are used here to investigate the FIR signature of cirrus. The FIR observations presented are made using the TAFTS spectrometer, which measures spectral radiances from either an aircraft or the ground.

The deployment of TAFTS during the RHUBC campaign based in Barrow, Alaska is described. TAFTS was used to make ground-based FIR observations of the arctic atmosphere, both with and without cirrus. Comparing these with modelled spectra, which assume a parameterised particle size distribution (PSD) when describing the cirrus microphysics, suggested that the PSD parameterisation underestimates the fraction of ice water content contributed by small ice crystals. This conclusion is corroborated by AERI-ER observations made simultaneously at the Barrow site during RHUBC.

TAFTS observations of convective tropical cirrus made during EMERALD-II near Darwin, Australia are also presented here. During EMERALD-II TAFTS was deployed on an aircraft, enabling spectral measurements of cirrus at wavenumbers between 100 and 200cm$^{-1}$ to be made for the first time. Comparisons with LBLDIS spectra calculated using PSDs measured using cloud probes indicate that the number of small crystals measured may be too high by a factor of three. This result is in agreement with previous studies suggesting that small crystal populations are over-counted by in-situ
cloud probes, due to shattering of larger crystals on the probe inlets. The results from both campaigns illustrate the sensitivity of FIR radiances to cirrus properties, with particular emphasis on the effect of small ice crystals.
ACKNOWLEDGEMENTS

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LIST OF ACRONYMS

ACRF-NSA  ARM Climate Research Facility - North Slope of Alaska

AERI-ER  Atmospherically Emitted Radiation Interferometer - Extended Range

ARA     Airborne Research Australia

ARM     Atmospheric Radiation Measurement program

BT      Brightness Temperature

CPI     Cloud Particle Imager

DISORT  DIScrete Ordinates method for Radiative Transfer

DFT     Differential Fringe Timing

EMERALD Egrett Microphysics Experiment with RADiation Lidar and Dynamics

ERB     Earth Radiation Budget

FFT     Fast Fourier Transform

FIR     Far InfraRed

FPH     Frost Point Hygrometer

FSSP    Forward Scattering Spectrometer Probe

FTS     Fourier Transform Spectroscopy
<table>
<thead>
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<tr>
<td>GCM</td>
<td>General Circulation Model</td>
</tr>
<tr>
<td>GSR</td>
<td>Ground-based Scanning Radiometer</td>
</tr>
<tr>
<td>GVR</td>
<td>G-band Vapour Radiometer</td>
</tr>
<tr>
<td>HITRAN</td>
<td>High-resolution TRANsmission molecular absorption database</td>
</tr>
<tr>
<td>IR</td>
<td>InfraRed</td>
</tr>
<tr>
<td>ISCCP</td>
<td>International Satellite Cloud Climatology Project</td>
</tr>
<tr>
<td>ITCZ</td>
<td>Inter-Tropical Convergence Zone</td>
</tr>
<tr>
<td>IWC</td>
<td>Ice Water Content</td>
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<td>LBLDIS</td>
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<tr>
<td>LBLRTM</td>
<td>Line By Line Radiative Transfer Model</td>
</tr>
<tr>
<td>LW</td>
<td>LongWave</td>
</tr>
<tr>
<td>MMCR</td>
<td>MilliMetre Cloud Radar</td>
</tr>
<tr>
<td>MPL</td>
<td>MicroPulse Lidar</td>
</tr>
<tr>
<td>MWR</td>
<td>MicroWave Radiometer</td>
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<tr>
<td>NOAA</td>
<td>National Oceanic and Atmospheric Administration</td>
</tr>
<tr>
<td>OLR</td>
<td>Outgoing Longwave Radiation</td>
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<tr>
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<tr>
<td>POB</td>
<td>Pointing Optics Box</td>
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<tr>
<td>PWV</td>
<td>Precipitable Water Vapour</td>
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<td>Relative Humidity</td>
</tr>
<tr>
<td>RHUBC</td>
<td>Radiative Heating in Under-explored Bands Campaign</td>
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RMS  Root Mean Square
SW   ShortWave
TAFTS Tropospheric Airborne Fourier Transform Spectrometer
TDL  Tunable Diode Laser
UTC  Co-ordinated Universal Time
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1. INTRODUCTION

This introductory chapter provides some of the essential background necessary to provide some context to the work covered by this study. Section 1.1 illustrates the concept of the Earth’s Radiation Budget (ERB) through simple calculations, leading to a demonstration of how absorption and re-emission of radiation by the atmosphere results in a greenhouse effect which enhances the terrestrial surface temperature. The processes by which radiation is absorbed and re-emitted by molecules in the atmosphere are outlined in Section 1.2. Section 1.3 then highlights the significance of far-infrared (FIR) radiation in the ERB, and briefly reviews the observational and theoretical studies carried out on FIR radiation in the atmosphere to date. Finally Section 1.4 reviews the current state of knowledge regarding cirrus clouds, focusing on their effect on the ERB. Cirrus in-situ and remote sensing studies are also reviewed in this section.

1.1 The Earth’s radiative energy budget

The primary driving force for the Earth’s atmosphere is the absorption of solar energy at the surface (Salby, 1996). Over long timescales the surface-atmosphere system is in thermal equilibrium, so this input of radiative energy must somehow be balanced. This balance is achieved by the emission to space from the Earth’s surface and atmosphere of thermal radiation, which is concentrated in the infrared part of the electromagnetic spectrum. By balancing the incoming shortwave (SW) radiation (concentrated on the visible and ultraviolet parts of the spectrum) with the outgoing longwave (LW) radiation, it is possible to perform a simplistic estimate of the Earth’s mean
temperature.

A beam of SW radiation of cross section $\pi a^2$, where $a$ is the Earth’s radius, and flux $F_s$ is incident on the Earth. A fraction $A$ (the albedo) of this incident radiation is reflected back to space, either by the surface or by components of the atmosphere. The rest of the incident SW flux, $(1-A)F_s$, is absorbed by the Earth-atmosphere system and distributed across the globe.

The thermal equilibrium of the earth-atmosphere system is maintained by emitting LW radiation at the same rate at which SW radiation is absorbed. Assuming that the earth-atmosphere system radiates like a blackbody, the emission to space of thermal radiation is given by the Stefan-Boltzmann law

$$\pi B = \sigma T^4,$$

(1.1)

where $\pi B$ is the total outgoing flux over all wavelengths emitted by a blackbody at temperature $T$, and $\sigma$ is the Stefan-Boltzmann constant. The equation for the energy balance is obtained by integrating the emitted LW flux over the Earth’s surface, and equating this result to the total SW energy absorbed by the earth-atmosphere system, which gives

$$(1-A)F_s \pi a^2 = 4\pi a^2 \sigma T^4.$$ (1.2)

$T_e$ is the equivalent blackbody temperature of the Earth. By rearranging Equation 1.2, the following expression is found for $T_e$:

$$T_e = \left[ \frac{(1-A)F_s}{4\sigma} \right]^{\frac{1}{4}}$$ (1.3)

Typical values for $F_s$ and $A$ are 1372 Wm$^{-2}$ and 0.30, respectively. Substituting these values into Equation 1.3 gives an equivalent blackbody temperature for the Earth of $T_e = 255$ K, which is about 30 K less than the global mean surface temperature, $T_s = 288$ K.

The simple calculation presented here underestimates the Earth’s mean surface temperature, since SW and LW radiation interacts with the atmosphere in different ways. The atmosphere transmits a significant fraction of the incoming SW radiation, but is almost opaque to the LW radiation
emitted by the Earth’s surface. Consequently, this LW radiation is largely absorbed by the atmosphere. Each atmospheric layer re-emits the energy it absorbs, half upwards and half downwards. The upwelling radiation is re-absorbed by higher layers, and which then re-emit in a similar manner. This is repeated until the remaining LW energy is radiated beyond all absorbing material in the atmosphere, and is transmitted to space. This inhibition of the radiation of LW energy from the Earth’s surface to space traps energy in the earth-atmosphere system, and elevates the surface temperature.

This enhancement of surface temperature owing to the different ways in which the atmosphere interacts with SW and LW radiation is known as the *greenhouse effect*. The effect is controlled by the presence of certain atmospheric constituent species which are opaque in the infrared region, and therefore radiatively insulate the planet. The most important absorbers are water vapour, carbon dioxide and clouds, whilst other contributors include aerosols, ozone, methane and nitrous oxide.

Figure 1.1 is a schematic representation illustrating an estimate of the Earth’s globally averaged energy budget (Kiehl & Trenberth, 1997). The incoming SW radiation is equal to \( F_s/4 \), owing to the ratio of the cross sectional area intercepted by the Earth to the Earth’s surface area. The amount of SW radiation reaching the surface is governed by three processes; absorption by the atmosphere (19.6 %), reflection by clouds, aerosols and the atmosphere (22.5 %), and reflection by the surface (8.8 %), where the percentages in brackets are the fractions of incoming SW radiation removed by each process. The remaining 49.1 % is absorbed by the Earth’s surface.

To maintain thermal equilibrium at the Earth’s surface as described above, the 168 Wm\(^{-2}\) absorbed must be re-emitted. Given that the global mean surface temperature \( T_s = 288 \) K, by Equation 1.1 the surface emits 390 Wm\(^{-2}\) of LW radiation, which is much more energy than the surface absorbs. This excess is balanced by the transfer of energy from other sources. Through the greenhouse effect, the surface receives LW radiation emitted downwards by the atmosphere, a total of 324 Wm\(^{-2}\). Adding these three
1. Introduction

Fig. 1.1: Estimate of the global mean energy budget from Kiehl & Trenberth (1997). All quantities are given in units of Wm\(^{-2}\).

contributions results in a net gain in radiative energy of 102 Wm\(^{-2}\) at the surface. Equilibrium is maintained through the transfer of sensible and latent heat to the atmosphere, by means of convection and evapotranspiration respectively.

The energy budget of the atmosphere must also balance to zero to remain in thermal equilibrium. 67 Wm\(^{-2}\) of SW radiation is absorbed by the atmosphere. Of the 390 Wm\(^{-2}\) emitted by the surface, 40 Wm\(^{-2}\) passes straight though the atmosphere to space. The remainder is absorbed by a combination of clouds, aerosols and absorbing gases such as water vapour and carbon dioxide. The atmosphere emits 324 Wm\(^{-2}\) downwards towards the surface, and 195 Wm\(^{-2}\) upwards to space. The net radiative forcing of the atmosphere, given by the sum of these components, is \(-102\) Wm\(^{-2}\). This radiative cooling of the atmosphere is balanced by the latent and sensible heat transfer from the surface described above.

This section has given a brief outline of the different contributions to the Earth’s radiative energy budget. Section 1.2 now describes the physical
processes behind the spectral absorption and emission of radiation by the atmosphere’s constituent particles.

1.2 Spectroscopy in the atmosphere

The absorption and emission of thermal photons by atmospheric molecules is dependent on the energy differences between their vibrationally or rotationally excited states (Andrews, 2000). These energy differences are quantized (Rae, 2002), meaning that they may only take particular discrete values determined by the quantum mechanical selection rules. It follows from this that atmospheric radiation is only absorbed or emitted at certain discrete frequencies, since a photon’s energy $E$ is proportional to its frequency $f$ ($E = hf$, where $h$ is Planck’s constant). Throughout this study, photon energy is quantified either by its wavenumber ($\nu$, $\nu = f/c$ where $c$ is the speed of light), which is expressed in inverse centimetres, or by its wavelength ($\lambda$, $\lambda = c/f = 1/\nu$).

In this idealised scenario, the atmosphere’s extinction coefficient $k_\nu$ (the constant of proportionality determining the decrease in the spectral radiance of a beam as it passes through a medium, see Section 2.1) would be zero at all wavenumbers except for certain discrete values $\nu_n$ at which transitions occur. At these wavenumbers $k_\nu$ is large – these spikes in the extinction coefficient are known as spectral lines. The extinction coefficient in this form may be expressed as a sum of these spectral lines as follows:

$$k_\nu = \sum_n S_n \delta(\nu - \nu_n) \quad (1.4)$$

Here $S_n$ are constants representing the line strengths, and $\delta(\ldots)$ is the Dirac delta function. In practice spectral lines may not be described by Dirac delta functions, since there are physical effects which lead to broadening of the lines in wavenumber space. The Dirac delta functions in Equation 1.4 are replaced by line-shape functions $f_n$ such that:
\[ k_\nu = \sum_n S_n f_n(\nu - \nu_n) \] (1.5)

Each line-shape function \( f_n \) is normalised, so that the integral of \( f_n \) over all wavenumbers is equal to one. When considering the line-shapes of atmospheric spectral lines, there are two main physical processes contributing to their overall shape: collisional broadening and Doppler broadening.

Collisional broadening occurs when the emission process is interrupted through collisions with other molecules. Since the transition between states now has a finite collisional lifetime \( \tau_c \), the energy levels associated with the transition are not precisely defined but instead (according to Heisenberg’s Uncertainty Principle) have a spread of values. In wavenumber space this spread is described by the Lorentz line-shape:

\[ f_L(\nu - \nu_n) = \frac{1}{\pi c} \frac{\gamma_L}{(\nu - \nu_n)^2 + \gamma_L^2} \] (1.6)

The line half-width at half maximum is given by \( \gamma_L = (2\pi \tau_c c)^{-1} \). Through kinetic theory (Goody & Yung, 1989), the pressure and temperature dependence of \( \gamma_L \) is found to be:

\[ \gamma_L \propto pT^{-1/2} \] (1.7)

The Doppler effect describes the shift in frequency of electromagnetic radiation emitted by a moving particle. If the emitter is moving with velocity \( u \) away from the observer, the Doppler shift \( \nu - \nu_0 \) of a spectral line at wavenumber \( \nu_0 \) is given by \( \nu - \nu_0 = (u/c)\nu_0 \) in the limit \( u << c \). The velocities of particles in a given sample follow the Maxwell-Boltzmann distribution, whereby the probability that a particle has a velocity between \( u \) and \( u + du \) is given by \( P(u)du \), where

\[ P(u) = \left( \frac{m}{2\pi kT} \right)^{1/2} \exp \left( -\frac{mu^2}{2kT} \right) \] (1.8)

Here \( k \) is the Boltzmann constant and \( m \) is the particle mass. The variation in spectral shifts owing to the range of particle velocities as described
by the Maxwell-Boltzmann distribution is what leads to Doppler broadening of a spectral line. The spectral line-shape for a Doppler broadened line is obtained by convolving $P(u)$ with a Dirac delta function, giving

$$f_D(\nu - \nu_n) = \frac{1}{\gamma_D \sqrt{\pi c}} \exp \left( -\frac{(\nu - \nu_n)^2}{\gamma_D^2} \right) \quad (1.9)$$

where

$$\gamma_D = \frac{\nu_n c}{2kT m^{1/2}}. \quad (1.10)$$

This is a Gaussian distribution with a half-width at half-maximum of $\gamma_D (\ln 2)^{1/2}$. It can be seen from Equation 1.10 that $\gamma_D$ is proportional to $T^{1/2}$, but is independent of $p$. Since $\gamma_L$ is proportional to $p$, collisional broadening dominates at low altitudes (higher pressures) whereas Doppler broadening becomes more important at higher altitudes. In addition, the wavenumber dependence of $\gamma_D$ means that Doppler broadening has a greater effect on spectral lines in the visible and ultra-violet than it does in the infra-red.

Collisional and Doppler effects are both taken into account by the *Voigt line-shape*, which is derived by convolving the Lorentz and Doppler line-shape functions shown above. It is used throughout this study for the simulation of spectral radiances as described in Section 2.1.1.

### 1.3 The significance of far-infrared radiation in the Earth’s atmosphere

Having discussed the physics behind the spectral lines observed as a result of absorption and emission of radiation by atmospheric particles in Section 1.2, this section will now focus on the significance of far-infrared (FIR) radiation in the ERB. A thorough review illustrating the current understanding of the far-infrared properties of the Earth’s atmosphere may be found in Harries *et al.* (2008).

---

1 The far-infrared is defined here as all wavelengths longer than 15 $\mu$m, or all wavenumbers lower than 667 cm$^{-1}$. 

The Earth emits longwave radiation to space with an effective temperature of \( \sim 255 \text{ K} \), as shown in Section 1.1. At this temperature the Planck curve peaks at around \( 500 \text{ cm}^{-1} \), thus making the Earth a far-infrared object when viewed from space (Harries et al., 2008). Calculations by Sinha & Harries (1995) indicate that between 27 and 35% of the absolute clear-sky greenhouse effect \( G \) (defined as the difference between the surface emission \( E \) and the outgoing longwave radiation \( F \)) is at far-infrared wavelengths, with the fraction dependent on latitude and humidity. It should be noted here that the authors defined the FIR as all wavenumbers less than \( 500 \text{ cm}^{-1} \), so these fractions would be higher if they had used the FIR definition adopted in this work. The authors also show that the fraction of water vapour forcing (the change in OLR owing to an increase in column water vapour) in the FIR can be as high as 53% in the case of a sub-arctic winter atmosphere (the equivalent value for a tropical atmosphere was found to be 17%).

These results were examined in further detail by Sinha & Harries (1997), quantifying the effects of the FIR on a global scale (using observational data on a \( 10^\circ \times 10^\circ \) grid) by examining the difference in the calculated clear-sky OLR between January and July. They also investigated the effect of changes in the vertical structure of water vapour on the OLR, concluding that the FIR is important in the determination of the ERB and local atmospheric cooling rates, whilst suggesting that there may also be a systematic seasonal, geographical and altitude dependent variation to FIR effects. This implies that the FIR may have a significant influence on the dynamical behaviour of the atmosphere in addition to its impact on the ERB. Brindley & Harries (1998) investigate further the altitude dependence by simulating the spectral response of the clear-sky greenhouse effect to perturbations in carbon dioxide and water vapour at different altitudes. They demonstrate that the main response of the Earth to predicted increases in both \( \text{CO}_2 \) and \( \text{H}_2\text{O} \) may be expected to occur in the IR window (800 to 1250 cm\(^{-1}\)) within the lower troposphere, and in the water vapour pure rotation band in the FIR (between 250 and 450 cm\(^{-1}\)) within the upper troposphere.
The study of Brindley & Harries (1998) also looks at both the vertical and spectral dependence of the atmospheric heating rate, which measures the rate of energy gain at a particular wavenumber (typically in units of Kday$^{-1}$(cm$^{-1}$)$^{-1}$), following the work of Clough et al. (1992). The results of the calculations for standard tropical and sub-arctic winter atmospheres are shown in Figure 1.2.

Fig. 1.2: Cooling rates as a function of wavenumber and pressure calculated for: (a) a tropical atmosphere, and (b) a sub-arctic winter atmosphere, calculated by Brindley & Harries (1998).

These plots illustrate the dominant effect that the pure rotational bands of water vapour at wavelengths longer than 12 µm have on the ERB. The red and orange regions in the plots indicate cooling of the atmosphere at that
altitude and wavenumber. In both cases shown here, the majority of cooling to space occurs in the mid- to upper troposphere at FIR wavenumbers. There is also cooling of the lower troposphere in the tropical case via the mid-infrared window between 800 and 1200 cm$^{-1}$, owing to weaker water vapour lines and to continuum absorption and emission$^2$. The sub-arctic winter atmosphere exhibits much less cooling via the mid-infrared window since it contains less water vapour by a factor of $\sim 10$, and because it is colder than the tropical case in the lower troposphere by up to 40 K. In the far-infrared, the altitude at which the cooling rate peaks for a given wavenumber depends on the opacity of the atmosphere at that wavenumber – the less opaque the atmosphere, the lower the altitude at which the cooling rate peaks, since emitted radiation is able to pass further through the atmosphere without being absorbed.

In spite of the growing appreciation of the importance of the role played by the FIR in the ERB (as highlighted by the work described above), to date there have been relatively few observations of the Earth’s atmosphere at these wavelengths. Though the spectral OLR peaks at FIR wavelengths as discussed previously in this section, signals measured in the FIR region remain intrinsically weak since the Planck function at typical atmospheric temperatures falls to low values at longer wavelengths. The majority of experiments making FIR observations use cryogenic detectors to work around this issue. Some of the more recent attempts to make FIR observations of the Earth’s atmosphere are listed here.

**TAFTS** The Tropospheric Airborne Fourier Transform Spectrometer (Canas \textit{et al.}, 1997) was developed at Imperial College London specifically for the observation of FIR spectral radiances from an aircraft platform. It is able to observe both zenith and nadir radiances at high spectral resolution. Most of the observational results presented in Chapters 4 and 5 of this study were obtained using the TAFTS; a more detailed description of the instrument may be found in Chapter 3. Clear-sky ra-

$^2$ See Section 2.1.2 for a definition of the water vapour continuum
diances measured by TAFTS during the EAQUATE (European AQUA Thermodynamic Experiment, Taylor et al. (2008)) field campaign are presented by Cox et al. (2007), who demonstrate that spectra simulated using the techniques described in Chapter 2 reproduce those measured using TAFTS within instrument uncertainties. A preliminary overview of the results taken during the Radiative Heating in Underexplored Bands Campaign (RHUBC), described in detail in Chapter 4, is given by Humpage et al. (2008).

**AERI-ER** The Atmospherically Emitted Radiation Interferometer (Knuteson et al., 2004a,b) is a ground-based zenith viewing FTS which operates continuously and autonomously at each of the ARM (Atmospheric Radiation Measurement program, Ackerman & Stokes (2003)) sites situated across the globe. The majority of the AERIs currently in operation observe spectral radiances between 520 and 3300 cm$^{-1}$. The AERI-ER (AERI-Extended Range) located at the ACRF-NSA (ARM Climate Research Facility - North Slope of Alaska) site near Barrow, Alaska is different from other AERIs in that its lower wavenumber limit is 400 cm$^{-1}$, enabling it to make observations of the far-infrared ‘sub-arctic window’ (see Section 4.1). Data from an AERI-ER deployed during the Surface Heating Budget of the Arctic Ocean (SHEBA) campaign have been used to investigate the foreign-broadened water vapour continuum, resulting in a significant revision of the CKD continuum model at far-infrared wavelengths (Tobin et al., 1999).

**REFIR** The Radiation Explorer in the Far-Infrared is a space mission concept developed by a European consortium to address important issues related to the water cycle and climate by exploiting the far-infrared. Two prototypes have been built to demonstrate this concept: REFIR-Breadboard (BB, Palchetti et al. (2005)), a ground-based instrument which makes zenith view observations, and REFIR-Prototype for Applications and Development (PAD, Palchetti et al. (2006)), designed to
be flown on board a balloon-lifted gondola making nadir view observations from stratospheric altitudes. Both instruments observe a spectral range of $100 - 1400\,\text{cm}^{-1}$ at a spectral resolution of $0.25\,\text{cm}^{-1}$. A key feature of their design is that they employ uncooled detectors, yet are still able to provide accurate measurements of radiances throughout the far-infrared, even at the long wavelength end of the FIR spectrum. The successful demonstration of these instruments during a number of field campaigns (Bhawar et al., 2008; Serio et al., 2008a,b; Bianchini et al., 2008; Palchetti et al., 2008), using uncooled components, makes the development of a far-infrared satellite-based Earth observation instrument in the near future highly feasible (Harries et al., 2008).

**FIRST** The Far-Infrared Spectroscopy of the Troposphere instrument, developed by the NASA Langley Research Center, is another satellite instrument prototype designed to take nadir measurements from a gondola on a high-altitude balloon (Mlynczak et al., 2006). The design is based on a compact plane mirror Michelson interferometer, and is able to measure spectral radiances at frequencies between $50$ and $2000\,\text{cm}^{-1}$ with a spectral resolution of $0.643\,\text{cm}^{-1}$. The FTS and detector optics are cooled to $\sim 180\,\text{K}$ by liquid nitrogen to reduce background radiation and simulate possible spacecraft conditions, whilst the detectors themselves are cooled to $4.2\,\text{K}$. Mlynczak et al. (2006) present results from the first balloon flight carrying the FIRST instrument. The measured spectra compare well with both radiative transfer model output at far-infrared wavelengths, and with measurements between $800$ and $950\,\text{cm}^{-1}$ from the AIRS (Atmospheric Infrared Sounder) instrument on board the NASA AQUA satellite taken as it passed over the FIRST measurement site.

During this section, an overview of why it is important to study the far-infrared in the context of the Earth’s energy budget has been presented, along with some of the recent attempts to measure far-infrared spectral radiances in the Earth’s atmosphere from a variety of platforms. This introductory
chapter now concludes by considering the diverse properties of cirrus clouds and the techniques used to measure them.

1.4 A review of cirrus cloud properties and techniques for their observation

Cirrus clouds are high clouds composed principally of ice crystals. They are ubiquitous in nature, covering an average of \( \sim 30\% \) of the globe (compared with \( \sim 75\% \) coverage due to all cloud types), and are most frequently found at tropical latitudes (Wylie et al., 2005). The macrophysical properties of cirrus (here the term ‘macrophysics’ is used to describe the bulk physical properties of a whole cirrus cloud layer, whilst ‘microphysics’ refers to the properties of individual ice crystals) are highly variable and as such defy characterization by a single set of values (Dowling & Radke, 1990). The Dowling & Radke (1990) review concludes with a list of typical values and previously measured ranges, which is repeated here in Table 1.1. The ice water content (IWC) is defined as the mass of cloud ice per unit volume of atmospheric air. The values that would actually be observed in a given cirrus cloud are dependent on local conditions such as temperature and relative humidity.

<table>
<thead>
<tr>
<th>Property</th>
<th>Typical value</th>
<th>Measured range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thickness</td>
<td>1.5 km</td>
<td>0.1 to 8 km</td>
</tr>
<tr>
<td>Cloud centre altitude</td>
<td>9 km</td>
<td>4 to 20 km</td>
</tr>
<tr>
<td>Crystal concentration</td>
<td>30 L(^{-1})</td>
<td>(10^{-4}) to (10^{4}) L(^{-1})</td>
</tr>
<tr>
<td>Ice water content</td>
<td>0.025 gm(^{-3})</td>
<td>(10^{-4}) to 1.2 gm(^{-3})</td>
</tr>
<tr>
<td>Crystal length</td>
<td>250 (\mu)m</td>
<td>1 to 8000 (\mu)m</td>
</tr>
</tbody>
</table>

**Tab. 1.1:** Typical values and measured ranges of the physical properties of cirrus as reported by Dowling & Radke (1990).

The large variation in cirrus physical properties makes it very difficult to accurately represent cirrus in the General Circulation Models (GCMs) used
to predict future climate change (Waliser et al., 2009). This is particularly important when considering the radiative impact of cirrus on climate (Liou, 1986) since the infrared radiative properties of cirrus are intrinsically linked to their physical properties (Slingo & Slingo, 1988; Stephens et al., 1990; Stackhouse & Stephens, 1991), such that the sign of the radiative heating of the upper troposphere due to cirrus clouds may be either positive or negative depending upon height, thickness and crystal size (Maestri & Rizzi, 2003). Maestri et al. (2005) investigated the heating rates due to different cloud layer geometries (whilst holding the cloud microphysics constant), finding that thinner and higher cirrus induces greater heating rates within the cloud layer compared with thicker, lower cirrus. The spectral contributions to the heating rates are divided between the mid-infrared window region, where the cloud layer experiences net absorption, and the far-infrared where it experiences a net emission.

Fig. 1.3: Spectral flux convergence difference (cirrus minus clear-sky) calculated by Maestri & Rizzi (2003) as a function of wavenumber and altitude, assuming a tropical atmosphere. Colour scale is in mW/(m²•km•cm⁻¹).

This is illustrated by Figure 1.3, taken from Maestri & Rizzi (2003), which shows the difference in calculated spectral flux convergences as a function of
altitude due to the presence of a cirrus layer (the two horizontal lines in the
figure indicate the altitudes of the cirrus layer top and base). The figure
also shows how, at far-infrared wavenumbers, the presence of a cirrus layer
results in heating of the atmosphere below the cloud relative to a cloud free
atmosphere. This happens because radiative energy emitted and scattered
by the cirrus layer enhances the downwards flux below its base. This extra
energy is absorbed by tropospheric water vapour via the pure rotational
band, resulting in net heating of the atmosphere below the cloud.

The microphysical properties of cirrus are also important, and like the
macrophysical properties are also difficult to characterize. As mentioned in
Table 1.1 the size of ice crystals in cirrus can vary by over three orders of
magnitude, whilst a wide range of different crystal shapes (or ‘habits’) have
also been observed in cirrus clouds (Lawson et al., 2001, 2006; Gallagher
et al., 2005; Korolev et al., 1999). The main habit classifications used in
radiative transfer calculations are shown in Figure 1.4.

Fig. 1.4: Types of ice crystal habit used to calculate cirrus optical properties,
taken from Yang et al. (2005).

Previous work has demonstrated that the assumed crystal habit has a
non-trivial effect when calculating the infrared radiative properties of cir-
rus (Wendisch et al., 2007; Yang et al., 2001, 2003; Baran & Francis, 2004;
Baran, 2005). The infrared scattering properties used in this study (see Baran
& Francis (2004) for details) are calculated for both hexagonal columns and for aggregates, so one must assume one shape or the other when simulating infrared spectral radiances (see Section 2.2.3). More recent work by Baum et al. (2005) models cirrus scattering properties by assuming a mixture of crystal habits, a more realistic assumption which improves agreement with in-situ observations of bulk cirrus microphysics.

The range and spread of ice crystal sizes within a cirrus cloud is described by the particle size distribution (PSD). PSDs for cirrus formed under many different meteorological conditions have been observed in-situ using aircraft mounted cloud probes during the past three decades (Heymsfield & Platt, 1984; Francis et al., 1994; McFarquhar & Heymsfield, 1997; Whiteway et al., 2004). There has been some debate regarding the validity of cloud probe measurements of small crystals (maximum dimension less than 100 µm), owing to shattering of larger crystals on probe inlets resulting in an artificial enhancement of the measured small crystal concentrations (Field et al., 2003, 2006; McFarquhar et al., 2007; Heymsfield, 2007). Heymsfield (2007) even goes as far as suggesting that conclusions from previous studies of the radiative effects of small ice particles may need to be re-evaluated.

Though ice crystal PSDs are typically bimodal (Mitchell et al., 1996), in climate models they are generally assumed to adopt an exponential distribution (Kristjansson et al., 1999; Wilson & Ballard, 1999). A number of parameterizations have been developed which enable the PSD to be estimated from cloud temperature and IWC (McFarquhar & Heymsfield, 1997; Ivanova et al., 2001; Field et al., 2005, 2007). Examples of some of these are shown in Figure 1.5. A quantity known as the effective diameter is frequently used to characterize PSDs. Its definition (Foot, 1988) is given in Equation 4.8.

Since the parameterizations are typically derived from aircraft observations they are not universally applicable to any cirrus cloud, but tend to be best suited to clouds formed in similar meteorological conditions to those

---

3 The Field et al. (2005) parameterization is described in detail in Section 4.4.1.
1. Introduction

Fig. 1.5: PSDs calculated from the parameterizations of McFarquhar & Heymsfield (1997), Ivanova et al. (2001) and Field et al. (2005) assuming IWC = \(1.19 \times 10^{-2} \text{ gm}^{-3}\) and cloud temperature \(T_c = 236.5 \text{ K}\). Effective diameters for each PSD are given in microns.

encountered during the flight campaign providing the in-situ data. This explains the differences in the parameterization results shown in Figure 1.5, despite the same input values of IWC and cloud temperature being used in each case.

Much progress has recently been made in the development of techniques for the remote sensing of cirrus properties from both satellite and ground-based observations (Stephens & Kummerow (2007), Comstock et al. (2007) and references therein). Knowledge of cirrus optical properties (which are covered in detail in Section 2.2.3), how they vary with wavenumber and how they are related to the physical properties of cirrus enables the retrieval of these properties from radiometric observations. It is the influence of the refractive index of ice which causes the variation of cirrus optical properties with wavenumber (see Section 2.2.3). Figure 1.6 shows the real and imaginary components of ice refractive index for wavenumbers less than \(1000 \text{ cm}^{-1}\), as measured by Warren (1984).

Remote sensing techniques may be divided into two categories: passive remote sensing and active remote sensing. Passive remote sensing involves the retrieval of atmospheric properties from observations of electromagnetic radiance which has been emitted or scattered by the atmosphere. A number
of radiometers, operating in a range of different spectral bands, are currently used for retrieving cirrus cloud properties. The ISCCP cloud product (International Satellite Cloud Climatology Project, Rossow & Schiffer (1999)) uses data collated from a suite of meteorological satellites, each of which have imaging radiometer channels measuring emitted infrared and reflected visible radiances. Global coverage is provided by four satellites in geostationary orbits (Meteosat, GMS, GOES-East and GOES-West), along with the NOAA series of polar-orbiters. A simple cloud detection algorithm identifies the cloudy pixels (Rossow & Garder, 1993). The cloudy pixels are then classified in terms of cloud top pressure (obtained from the infrared channel) and optical depth (obtained from the visible channel, defined by Equation 2.5), properties which are retrieved by comparing the measured radiances with
expected radiances calculated using a radiative transfer model (Rossow & Schiffer, 1991). Clouds occurring at pressures lower than 440 mbar in the ISCCP product are automatically classified as ice clouds.

More advanced retrievals of ice cloud properties based on passive remote sensing techniques utilise specifically chosen radiance bands to simultaneously infer cirrus optical depth, ice water content and effective radius. MODIS (the Moderate Resolution Imaging Spectroradiometer), which flies on board the NASA Terra satellite, measures reflected radiance in three different visible wavelength channels. These observations are used in conjunction with look-up tables of cirrus and surface reflectances, calculated from sets of in-situ measured PSDs and theoretical ice scattering properties, to retrieve cirrus properties on a global scale (Rolland et al., 2000; Meyer et al., 2004). The sensitivity of mid-infrared spectral radiances to ice properties (Bantges et al., 1999; Yang et al., 2001) has also been exploited for cirrus retrievals. The slope of the brightness temperature observed from space (nadir geometry) as a function of wavenumber between 750 and 1000 cm\(^{-1}\) is sensitive to the ice crystal effective particle size, whilst the 1100 – 1250 cm\(^{-1}\) spectral region is more sensitive to optical depth (Huang et al., 2004), enabling the simultaneous retrieval of both properties from high spectral resolution infrared spectrometers such as AIRS (Atmospheric Infrared Sounder, Aumann et al. (2003)) and IASI (Infrared Atmospheric Sounding Interferometer, Chalon et al. (2001)).

Active remote sensing, which involves firing pulses of electromagnetic radiation towards the target sample and then measuring the backscatter, has also proved to be very useful for observing cirrus properties from both space- and ground-based platforms. The principal active remote sensing techniques are radar and lidar, which utilise microwave and visible radiation respectively. The Cloud Profiling Radar on board the NASA CloudSat platform operates in the W-band (94 GHz), and has provided authoritative estimates of global ice water paths since its launch in 2006 (Stephens et al., 2008). The Cloud/Aerosol Lidar and Infrared Pathfinder Satellite Observa-
tion (CALIPSO) satellite, which flies just behind CloudSat as part of the A-Train satellite constellation (see Figure 1.7 for a schematic illustration of the A-Train concept), carries a three channel elastic backscatter lidar as part of its payload (Hunt et al., 2009). The A-Train (Stephens & Vane, 2007) also includes the MODIS and AIRS instruments on board the Aqua satellite.

![Artist impression of the A-Train satellite constellation](image)

**Fig. 1.7:** Artist impression of the A-Train satellite constellation from Stephens et al. (2002).

Whilst the lidar is useful in its own right for remotely observing cloud top height and optical extinction (Young & Vaughan, 2009), it is the synergy of observations from instruments utilising different optical methods which can potentially provide the greatest amount of information on cirrus physical properties (Haladay & Stephens, 2009; Grenier et al., 2009). Multi-instrument retrievals of ice cloud properties which include combinations of radar, lidar and infrared observations (Comstock et al. (2007) and references
therein) have also been developed for ground-based sites such as the ARM (Atmospheric Radiation Measurement) Climate Research Facilities (Acker- man & Stokes, 2003).

**Fig. 1.8:** Left: Simulated FIR brightness temperature spectra for a nadir view from altitude 20 km for cirrus clouds of different visible optical thicknesses. The clouds are 1 km thick and at 10 km altitude with effective particle size $D_e = 50 \mu m$. Right: Spectra obtained with different $D_e$, visible optical thickness $\tau = 10$. Taken from Yang et al. (2003).

At present, there are no operational far-infrared observations of cirrus clouds, though there have been some modelling studies investigating the sensitivity of far-infrared radiance spectra to cirrus cloud properties. Yang *et al.* (2003) found that the differences between brightness temperatures at certain wavenumbers, corresponding to regions of strong and weak water vapour absorption, are sensitive to optical depth and effective radius (see Figure 1.8), with the sensitivity being dependent on cloud temperature and altitude as well as the atmospheric temperature and water vapour profiles. However, Baran (2007) showed in addition that far-infrared spectral radiances also demonstrate sensitivity to the assumed ice crystal geometry and the shape of the particle size distribution. Altogether this means that although there is great potential in using the far-infrared for cirrus observations, a lot of *a priori* information would be needed to perform operational retrievals of cirrus properties from far-infrared radiance spectra.
1.5 Summary

This chapter has introduced some of the basic concepts regarding the Earth’s radiative energy budget and the influence of cirrus clouds. Understanding the effect of cirrus on the propagation of both thermal and visible radiation through the atmosphere remains a difficult challenge in spite of the recent advances in the field reviewed in this chapter. However, the ongoing development of new measurement techniques – particularly in satellite based remote sensing – will continue to inform and guide efforts to model and understand the relationships between cirrus microphysics and the Earth’s radiation budget. The importance of the far-infrared band of the electromagnetic spectrum has also been highlighted here, focussing on the effects of both water vapour (discussed in more detail in Section 2.1.2) and cirrus clouds at these wavelengths.

The core theme of this work is to demonstrate the potential for using far-infrared spectral radiances to remotely observe cirrus properties, through comparisons between observations and radiative transfer modelling. Chapter 2 describes how a line-by-line radiative transfer model is used in conjunction with a scattering code to simulate the effect of cirrus on far-infrared radiance spectra. The instrument used to measure these spectra, the Tropospheric Airborne Fourier Transform Spectrometer (TAFTS), is described in Chapter 3 along with the method used to perform the radiometric calibration of the data. Chapters 4 and 5 report on the deployment of TAFTS during two very distinct field campaigns. Chapter 4 describes ground-based measurements of far-infrared radiance spectra emitted by the Arctic winter atmosphere at the North Slope of Alaska ARM Climate Research Facility in Barrow during RHUBC (Radiative Heating of Underexplored Bands Campaign), whilst Chapter 5 shows results from EMERALD-II (Egrett Microphysics Experiment with Radiation Lidar and Dynamics), which investigated the properties of tropical anvil convective outflow cirrus by performing in-situ measurements near Darwin, Australia. Both of these chapters describe the other instrumentation available during each campaign, and how their data
were used to create the radiative transfer modelled spectra which are compared with the TAFTS observations. This work concludes with a discussion of the findings (Chapter 6), and their implications for the future study of cirrus at far-infrared wavelengths.
2. MODELLING THE IMPACT OF CIRRUS ON
FAR-INFRARED RADIANCE SPECTRA

This chapter describes the modelling of infrared radiance spectra using line-
by-line radiative transfer models. The theory of radiative transfer in a non-
scattering atmosphere is outlined in Section 2.1, with particular focus on the
effect of water vapour given that it is by far the most significant absorbing
gas at far-infrared (FIR) wavelengths (Section 2.1.2). The scattering effect
of ice crystals on infrared radiative transfer is then considered in Section 2.2.

2.1 Infrared radiative transfer in a non-scattering atmosphere

Consider a beam of radiation of spectral radiance $I_\nu$ at wavenumber $\nu$ as it
enters an absorbing material. The beam will undergo a fractional decrease
in radiance as it passes through the material. This decrease along a path of
length $ds$ is proportional to the mass of absorbing and scattering material
encountered along the path (the Beer-Lambert law, Andrews (2000)), so for a
beam of unit cross-sectional area passing through a density $\rho_a$ of a radiatively
active gas the change in spectral radiance at wavenumber $\nu$ is given by:

$$dI_\nu = -k_{e,\nu} \rho_a I_\nu ds$$  \hspace{1cm} (2.1)

The (wavenumber dependent) constant of proportionality $k_{e,\nu}$ is known
as the mass extinction coefficient, and is expressed in units of $m^2 kg^{-1}$. Ex-
tinction of radiation from a beam as it passes through a radiatively active
medium is by one of two processes: absorption from the beam, and scat-
tering out of the beam. The extinction coefficient is therefore expressed as
the sum of a mass absorption coefficient $k_{a,\nu}$, and a mass scattering coefficient $k_{s,\nu}$. These quantities are often converted into extinction, absorption and scattering cross sections $\kappa_{e/a/s,\nu}$ with units m$^2$, through multiplication by the particle mass $m_a$:

$$\kappa_{e,\nu} = m_a k_{e,\nu}, \kappa_{a,\nu} = m_a k_{a,\nu}, \kappa_{s,\nu} = m_a k_{s,\nu}$$ (2.2)

The scattering component will now be neglected for the remainder of this section, and then re-introduced to the discussion in Section 2.2.

The final process contributing to the change in $I_\nu$ as it passes through the material is the emission of radiation into the beam by particles within the material. This is incorporated into Equation 2.1 by adding a source function $S_\nu$, defined such that the increase in spectral radiance along the path $ds$ is given by $k_{e,\nu} \rho_a ds S_\nu$. Including this source function gives the basic radiative transfer equation (Equation 2.3), which is also known as Schwarzschild’s equation (Andrews, 2000).

$$\frac{dI_\nu}{ds} = -k_{e,\nu} \rho_a (I_\nu - S_\nu)$$ (2.3)

Solving this differential equation gives radiance as a function of wavenumber, location and direction. The directional dependence of the equation may be greatly simplified by applying the plane parallel approximation. This assumes that the intensity of the spectral radiance varies only with the vertical coordinate $z$ and the zenith angle $\theta$, such that the radiation field displays axial symmetry about the $z$ axis. The assumption of a plane parallel atmosphere enables $ds$ to be replaced by $dz/\mu$, where $\mu$ is the cosine of $\theta$.

A further simplifying assumption is that of local thermodynamic equilibrium (LTE). For a medium in LTE, the radiation is isotropic. LTE is a valid approximation for the lowest $\sim 70$ km of the atmosphere (Houghton, 2002), where collisions between molecules are sufficiently frequent that radiative energy absorbed by molecules is transferred through collisions to other molecules in the form of kinetic energy, rather than being re-emitted as radiative energy through a molecular transition.
Under LTE, the emissivity of a medium is equal to the absorptivity (Kirchhoff’s Law, Andrews (2000)), enabling the Planck function $B_\nu$ to be used to describe the radiation emitted by the medium. The source function $S_\nu$ may therefore be replaced by $B_\nu$ in Equation 2.3. The equation is re-written here, applying the two approximations described above.

$$
\mu \frac{dI_\nu(z, \mu)}{dz} = -k_{e, \nu} \rho_a (I_\nu(z, \mu) - B_\nu(z))
$$

(2.4)

Here $B_\nu(z)$ is the Planck function at altitude $z$ and temperature $T(z)$. The vertical coordinate $z$ is finally converted into optical depth $\tau$, a dimensionless quantity defined by:

$$
\tau = \int_0^\tau d\tau' = \int_z^z k_{e, \nu}(z') \rho_a(z')dz' = -\int_z^{z_\infty} k_{e, \nu}(z') \rho_a(z')dz' = -\int_z^{z_\infty} k_{e, \nu}(z') \rho_a(z')dz' (2.5)
$$

In terms of $dz$, a small increment in optical depth is therefore given by $d\tau = -k_{e, \nu} \rho_a dz$. Optical depth has the opposite sign to altitude, since it is defined from the top of the atmosphere ($z_\infty$) whereas altitude is defined from the bottom. Replacing the $z$ dependence in Equation 2.4 gives the final, simplified version of Schwarzschild’s equation, Equation 2.6:

$$
\mu \frac{dI_\nu(\tau, \mu)}{d\tau} = I_\nu(\tau, \mu) - B_\nu(\tau)
$$

(2.6)

Solution of this equation requires knowledge of the boundary conditions, in addition to temperature, pressure and atmospheric composition profiles. The lower boundary is assumed to consist of an isotropically emitted flux described by the Planck function at the surface temperature $T_{\text{surface}}$. The upper boundary at the top of the atmosphere may be approximated to zero at infrared wavelengths.

This section provides only a brief outline of the radiative transfer theory used to calculate the propagation of thermal radiation through the atmosphere. For further detail, the reader is advised to consult one of the many textbooks on the subject, such as those by Goody & Yung (1989) and Liou (1992).
2.1.1 Line-by-line modelling of radiative transfer

The line-by-line method of performing radiative transfer calculations involves calculating the contribution to the spectral radiance of every spectral line within the desired wavenumber range. These are performed on a very high resolution wavenumber grid to account for contributions from the thinnest of spectral lines. The LBLRTM (Line By Line Radiative Transfer Model) code used here (Clough \textit{et al.}, 2005) has been regularly updated since its original release in the early 1990s (Clough \textit{et al.}, 1992), and has compared well with both observations and other radiative transfer codes when applied to mid-infrared (Tjemkes \textit{et al.}, 2003) and far-infrared wavelengths (Kratz \textit{et al.}, 2005). The Voigt line shape (see Section 1.2) is used to model the absorption of radiation due to the various vibrational and rotational transitions encountered by molecules in the atmosphere. The line parameters, including the line absorption strength and half-width temperature dependence, are all taken from the 2004 edition of HITRAN (High-resolution Transmission molecular absorption database, Rothman \textit{et al.} (2005)) with the exception of the water vapour line parameters, which have been updated more recently (Gordon \textit{et al.}, 2007). An outline of the code showing the required inputs and produced outputs is shown in Figure 2.1.

The user has the option to either specify one of six standard atmospheres (Anderson \textit{et al.}, 1986) as input for the temperature and absorbing gas vertical profiles, or to define their own profiles e.g. from radiosonde observations. The code also includes the capacity to convolve the calculated spectral radiances with an instrument function. This feature is essential for comparing modelled spectra with those observed using all types of spectrometers, including the Fourier Transform Spectrometers (FTS) used during this study (see Chapter 3 for a description of how a non-ideal FTS instrument affects the observed radiance spectrum). The final key element used as input by the LBLRTM code is the formulation used to describe the continuum absorption. Version 2.1 of the MT_CKD (the initials are taken from the names of those who have worked on the model: Mlawer, Tobin, Clough, Kneizys and
2. Modelling the impact of cirrus on far-infrared radiance spectra

Fig. 2.1: Flow diagram illustrating the calculation of clear sky radiance spectra using LBLRTM.

Davies) continuum model (Clough et al., 2005) is used here, and its effect at far-infrared wavelengths is described further (with particular emphasis on the water vapour continuum absorption) in Section 2.1.2.

2.1.2 The influence of water vapour at far-infrared wavelengths

Water vapour is the dominant radiatively active species in the far-infrared, and is so effective at absorbing FIR radiation that the Earth’s surface when viewed from space is completely obscured at these wavelengths (Harries et al., 2008). The only other gas influencing the shape of FIR radiance spectra is carbon dioxide, through the $\nu_2$ vibrational-rotational band centered at 667 cm$^{-1}$ which has an impact at the high frequency end of the FIR. The contribution of all other absorbing molecules at far-infrared wavelengths is negligible in comparison.

The absorption of FIR radiation by water vapour is due to the pure rotational band, a broad spectral feature centered around 150 cm$^{-1}$ which includes contributions from hundreds of strongly absorbing pure rotational
transition lines. Its influence on radiance spectra covers the whole range of far-infrared energies. The transition intensities peak at $\sim 150 \text{ cm}^{-1}$, and decrease with increasing wavenumber such that the atmosphere becomes partially transparent between 400 and 600 cm$^{-1}$. This spectral region is exploited during the ground-based study of arctic cirrus described in Chapter 4. The transition intensities, the Gordon et al. (2007) values for which are shown in Figure 2.2, also decrease with decreasing wavenumber for wavenumbers smaller than the band centre. The positions of the transition lines on the wavenumber scale are known to within 0.0001 cm$^{-1}$.

![Fig. 2.2](image)

**Fig. 2.2:** Intensities (in cm$^{-1}$/molecule cm$^{-2}$) of all water vapour transition lines between 100 and 600 cm$^{-1}$ according to the Gordon et al. (2007) dataset. The colours correspond to the uncertainty index given for each transition, as defined in Table 2.1.

Figure 2.2 also shows the uncertainty given in the updated HITRAN database for each transition line intensity (see Table 2.1 for definitions of the uncertainty indices used in HITRAN). There are 2050 lines shown in the figure, most of which have intensities with uncertainties less than 10%. All of the lines have self-broadened half-widths between 0.1 and 0.5 cm$^{-1}$/atm with uncertainties between 10 and 20%, whilst the air-broadened half-widths of most lines are between 0.01 and 0.1 cm$^{-1}$/atm. The air-broadened half-widths are derived from a variety of different sources comprising experimental measurements where available, and semi-empirical model calculations otherwise. The sources of measurements included in the dataset are chosen for each transition on the basis of the quality of the experimental data available, as described by Gordon et al. (2007). The accuracy of the air-broadened half-
Tab. 2.1: Definition of the uncertainty indices used in the HITRAN 2004 database.

<table>
<thead>
<tr>
<th>Code</th>
<th>Uncertainty range</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>Unreported or unavailable</td>
</tr>
<tr>
<td>1</td>
<td>Default or constant</td>
</tr>
<tr>
<td>2</td>
<td>Average or estimate</td>
</tr>
<tr>
<td>3</td>
<td>$\geq 20%$</td>
</tr>
<tr>
<td>4</td>
<td>$\geq 10%$ and $&lt; 20%$</td>
</tr>
<tr>
<td>5</td>
<td>$\geq 5%$ and $&lt; 10%$</td>
</tr>
<tr>
<td>6</td>
<td>$\geq 2%$ and $&lt; 5%$</td>
</tr>
<tr>
<td>7</td>
<td>$\geq 1%$ and $&lt; 2%$</td>
</tr>
<tr>
<td>8</td>
<td>$&lt; 1%$</td>
</tr>
</tbody>
</table>

widths in the dataset should not be expected to be better than 5\% (Gordon et al., 2007). Both the original HITRAN 2004 database and the Gordon et al. (2007) updated version were tested against infrared spectrometer observations at wavenumbers between 520 and 1360 cm$^{-1}$ by Esposito et al. (2007), who found that the updated HITRAN yielded the highest consistency with observations.

Given that the Earth is a far-infrared object when viewed from space (see Section 1.3), water vapour performs an important role through FIR absorption as a greenhouse gas. As discussed above, it is very effective at absorbing energy emitted by the Earth’s relatively warm surface before re-emitting it to space from a higher altitude (typically the upper troposphere, see Clough et al. (1992)), and therefore at a colder temperature. Water vapour therefore exhibits a strong positive feedback (Held & Soden, 2000) – greenhouse warming due to anthropogenic increases in carbon dioxide levels is amplified by water vapour, if more water vapour enters the atmosphere as a result of the original warming.

Whilst absorption close to the centre of water vapour pure rotational lines is theoretically well understood and represented adequately by the Voigt
line-shape, the properties of absorption lines further from the line centre are not well understood. The total contribution of the far-wings of collisionally broadened spectral lines to radiative absorption in the atmosphere is known as the \textit{continuum absorption}. Spectral observations of water vapour show that the assumption of the Lorentz line-shape for the far-wings of the absorption lines is inadequate (Burch, 1981). The continuum absorption derived assuming Lorentzian behaviour is too high at some infrared wavelengths, and too low at others (Clough \textit{et al.}, 1989). The source of these deviations in the continuum absorption from the Lorentz line-shape is usually explained through errors in the line-shape assumed for molecule-molecule interactions (Ma & Tipping, 2002), though some researchers believe that the effect comes from absorption by water vapour dimers created as a result of collisions (Ptashnik \textit{et al.}, 2004; Vaida \textit{et al.}, 2001).

An empirical approach to the continuum absorption problem has been developed by Clough \textit{et al.} (1989), resulting in the CKD continuum model commonly used for radiative transfer calculations. The CKD model describes the continuum absorption coefficient \( k_c(\nu) \) as follows:

\[
k_c(\nu) = R(\nu, T) \sum_i \left\{ S_i(T) \left[ f_c(\nu - \nu_i) \chi'(\nu - \nu_i) + f_c(\nu + \nu_i) \chi'(\nu + \nu_i) \right] \right\}
\]

(2.7)

Here \( R(\nu, T) \) describes the radiation field (accounting for stimulated emission), \( S_i(T) \) is the line strength of absorption line \( i \) (see Section 1.2), and \( f_c(\nu \pm \nu_i) \) is the contribution of the line shape to the continuum absorption. The CKD model takes the Lorentz line-shape for wavenumbers more than 25 cm\(^{-1}\) from the line centre, and assumes a flat shape within these limits as follows (\( \gamma_i \) is the Lorentzian half-width of absorption line \( i \)):

\[
f_c(\nu \pm \nu_i) = \begin{cases} 
\frac{1}{\pi} \frac{\gamma_i}{25^2 + \gamma_i^2} & \text{if } |\nu \pm \nu_i| \leq 25 \text{ cm}^{-1} \\
\frac{1}{\pi} \frac{\gamma_i}{(\nu \pm \nu_i)^2 + \gamma_i^2} & \text{if } |\nu \pm \nu_i| > 25 \text{ cm}^{-1}
\end{cases}
\]

(2.8)
Figure 2.3 shows \( f_c(\nu \pm \nu_i) \) as defined in the CKD model, along with the Lorentz line-shape. The function \( \chi'(\nu \pm \nu_i) \) in Equation 2.7 is a semi-empirical function used to correct for duration of collision effects and to ensure agreement between calculated and measured spectra (Clough et al., 1989).

\[ \chi'(\nu \pm \nu_i) = \frac{1}{1 + (\nu \pm \nu_i)^2/\gamma_i^2} \]

The latest release of this continuum model, MT\textunderscore CKD v2.1 (which includes revisions following ground-based observations of the arctic winter atmosphere between 400 and 600 cm\(^{-1}\) by Tobin et al. (1999)), is used throughout this study. The continuum absorption is comprised of two terms: a contribution from the self-broadening effect, caused by collisions between water molecules; and a contribution from the foreign-broadening effect caused by collisions between water molecules and molecules of other gases (the most abundant of these being N\textsubscript{2} and O\textsubscript{2}). Unlike the original CKD model, the self and foreign components are each based on contributions from two components: a collision induced component introduced for the MT\textunderscore CKD model, and a new line-shape component (Clough et al., 2005). The line-shape component differs from the one used in the CKD model in that super-Lorentzian
absorption is no longer allowed, since this cannot be explained physically. The collision induced component, which accounts for any super-Lorentzian absorption previously attributed to the line-shape in the CKD model, is based on dipole-allowed transitions with widths on the order of $50 \text{ cm}^{-1}$.

**Fig. 2.4:** Plots of simulated FIR radiance and residual spectra (upper panel: 100 to 350 cm$^{-1}$; lower panel: 350 to 600 cm$^{-1}$) showing contribution of water vapour continuum absorption – downwelling radiances for an arctic clear sky atmosphere observed from sea level (residual in each case shows ‘with continuum’ minus ‘without continuum’). Black: MT$_{CKD}$ v2.1; Blue: No continuum

The effect of the water vapour continuum absorption on simulated far-infrared radiance spectra is shown in Figures 2.4 and 2.5. The spectra simulated here are similar to those observed during the field campaigns described in Chapters 4 and 5 respectively. In the first example (Figure 2.4), which shows the spectrum observed from sea level looking upwards at an arctic
winter atmosphere, the continuum has very little effect at wavenumbers less than 350 cm\(^{-1}\) since all of the absorption is accounted for by the multitude of strong pure rotational transition lines which dominate at these wavelengths. At wavenumbers between 350 and 600 cm\(^{-1}\) the continuum contribution becomes much more significant as there are fewer transition lines within this spectral range, such that a greater fraction of the absorption is due to the far reaches of the transition lines.

![Fig. 2.5: Plots of simulated FIR radiance and residual spectra (upper panel: 100 to 350 cm\(^{-1}\); lower panel: 350 to 600 cm\(^{-1}\)) showing contribution of water vapour continuum absorption – downwelling radiances for a tropical clear sky atmosphere observed from 10.4 km above sea level (residual in each case shows ‘with continuum’ minus ‘without continuum’). Black: MT\_CKD v2.1 Blue: No continuum](image)

The second example (Figure 2.5) shows spectra that would be observed from an aircraft at 10.4 km above sea level looking upwards at a clear sky atmosphere.
tropical atmosphere. In this situation there is very little contribution from the continuum at the high wavenumber end of the far-infrared, since the amount of water vapour above the observation height is so small that there is hardly any absorption other than that occurring close to the centres of the strongest transition lines. At smaller wavenumbers, the cumulative effect of the greater number of transition lines means that even with so little water vapour there is sufficient absorption far enough from the line centres that the continuum makes some contribution to the simulated radiance spectrum. These two examples illustrate the importance of correctly modelling the continuum absorption at far-infrared wavelengths to the calculation of accurate spectral radiances.

In this section, the components required to simulate the FIR radiance spectrum viewed when observing a clear sky, non-scattering atmosphere have been described and discussed. In order to model the effect of cirrus on FIR spectra, the theory of radiative transfer in a scattering atmosphere must be incorporated into the calculations. The theory of scattering with particular emphasis on infrared scattering due to ice crystals is covered in the next section.

2.2 Infrared radiative transfer in a scattering atmosphere: accounting for the effect of cirrus on far-infrared radiance spectra

When cloud particles are present in the atmosphere, they introduce electromagnetic scattering to the radiative transfer problem (for a more detailed description of the scattering of electromagnetic radiation by small particles, the reader is advised to consult the textbooks by van de Hulst (1981) and Salby (1996)). As well as including a scattering component in the extinction coefficient to describe scattering of radiation out of the beam ($k_{\nu,s}$, as described in Section 2.1), an additional source term must also be considered to account for scattering of radiation into the beam. The source function
$S_\nu$ is therefore now written as the sum of contributions from emission and scattering, as shown in Equation 2.9:

$$S_\nu = \frac{\beta_{a,\nu}B_\nu + \beta_{s,\nu}J_\nu}{\beta_{e,\nu}}$$  \hspace{1cm} (2.9)

Here $J_\nu$ is the scattering source function. $\beta_{a,\nu}$ and $\beta_{s,\nu}$ are the absorption and scattering coefficients, and their sum is equal to the extinction coefficient $\beta_{e,\nu}$. These quantities are related to the mass extinction coefficient $k_{e,\nu}$ by the particle density $\rho_a$, and are expressed in units of $m^{-1}$:

$$\beta_{e,\nu} = \rho_a k_{e,\nu}, \beta_{a,\nu} = \rho_a k_{a,\nu}, \beta_{s,\nu} = \rho_a k_{s,\nu}$$  \hspace{1cm} (2.10)

The expression for the source function $S_\nu$ (Equation 2.9) may be simplified further by introducing the single scattering albedo, $\omega_{0,\nu}$, which is the ratio of the scattering coefficient to the extinction coefficient as shown in Equation 2.11:

$$\omega_{0,\nu} = \frac{\beta_{s,\nu}}{\beta_{e,\nu}}$$  \hspace{1cm} (2.11)

$S_\nu$ is then re-written as shown below:

$$S_\nu = (1 - \omega_{0,\nu})B_\nu + \omega_{0,\nu}J_\nu$$  \hspace{1cm} (2.12)

The scattering of radiation by particles occurs in all directions. The angular distribution of scattered radiation depends strongly on particle shape, orientation and size, and is described by the phase function $P(\cos \theta)$. It is usually normalised to unity (Baran, 2009), as in Equation 2.13:

$$\int_0^{2\pi} \int_0^\pi \frac{P(\cos \theta)}{4\pi} \sin \theta d\theta d\phi = 1$$  \hspace{1cm} (2.13)

Making the plane-parallel approximation means that the azimuthal dependence may be dropped. The scattering source function $J_\nu$ is then written in terms of the azimuthal-independent phase function $P(\mu, \mu')$, as shown in Equation 2.14:
Here, $\mu = \cos \theta$ where $\theta$ is the zenith angle of the beam being considered, whilst $\mu'$ is the cosine of the zenith angle from which radiation between $\mu'$ and $\mu' + d\mu'$ is being scattered into the beam. In this form, the phase function acts as a directional weighting function for radiation scattered into the beam from all zenith angles. The final radiative transfer equation for a scattering atmosphere may now be obtained by substituting the scattering source function as given by Equation 2.14 into Equation 2.12 to give the general source function $S_\nu$:

$$S_\nu = (1 - \omega_{0,\nu})B_\nu + \frac{\omega_{0,\nu}}{2} \int_{-1}^{1} P(\mu, \mu')I_\nu(\tau, \mu')d\mu' \quad (2.15)$$

This is then used in Equation 2.6 in place of $B_\nu$ (which is the source function in the absence of scattering), to obtain the end result given by Equation 2.16:

$$\mu \frac{dI_\nu(\tau, \mu)}{d\tau} = I_\nu(\tau, \mu) - (1 - \omega_{0,\nu})B_\nu(\tau) - \frac{\omega_{0,\nu}}{2} \int_{-1}^{1} P(\mu, \mu')I_\nu(\tau, \mu')d\mu' \quad (2.16)$$

During this study, the discrete ordinate method is used to solve the radiative transfer equation in a scattering atmosphere. This method is briefly introduced in Section 2.2.1. Its implementation in the LBLDIS (Line-by-line Discrete Ordinate Method for Radiative Transfer) scattering code is then described in Section 2.2.2.

### 2.2.1 The discrete ordinate method for calculating radiative transfer in a scattering atmosphere

The discrete ordinate method (Stamnes et al., 1988) involves the transformation of the radiative transfer equation (Equation 2.6) into a set of simultaneous first order differential equations. It is a generalisation of the two stream
approximation, which reduces the radiative transfer equation to a single pair of simultaneous equations.

The method begins by making the plane parallel approximation, a method of dividing the atmosphere up into discrete layers. The approximation is based on two key assumptions:

- The curvature associated with the sphericity of the Earth may be ignored.
- The atmosphere is horizontally homogeneous, and the radiation field is horizontally isotropic.

The slant path $ds$ of a beam of radiation inclined at zenith angle $\theta$ through a parallel slab of atmosphere of thickness $dz$ is given by

$$ds = \frac{dz}{\mu}$$  \hspace{1cm} (2.17)

where $\mu = \cos \theta$. It is also useful to define the optical depth, $\tau_\nu$, which is simply the optical path (Equation 2.5) from the top of the atmosphere down to altitude $z$:

$$\tau_\nu(z) = \int_z^\infty \rho(z') a k_\nu(z') dz'$$  \hspace{1cm} (2.18)

When considering scattering, the solution of the radiative transfer equation requires an integration over all zenith angles, and the full range of optical depths. The discrete ordinate method simplifies this calculation by replacing $\mu$ with $\mu_i$, where $i = -n, \ldots, n$ and $n = 1, 2, \ldots$, and by replacing the integral over zenith angle (i.e. over $\mu$) with a weighted sum,

$$\int_{-1}^{1} f(\mu) d\mu = \sum_{j=-n}^{n} f(\mu_j) a_j,$$  \hspace{1cm} (2.19)

where $f(\mu)$ is a continuous function of $\mu$, such as the source function, and $a_j$ is the weighting. The general solution to the system of $2n$ coupled ordinary differential equations derived by applying the discrete ordinate method is given by

$$I(\tau, \mu_i) = \sum_{j=-n}^{n} L_j \phi_j(\mu_i) e^{-k_j \tau} + I_p(\tau, \mu_i),$$  \hspace{1cm} (2.20)
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where $k_j$ and $\phi_j$ are the eigenvalues and eigenvectors, and $L_j$ are constants to be determined from appropriate boundary conditions. $I_p(\tau, \mu_i)$ is the particular solution. A detailed description of the discrete ordinate method and its solution may be found in Liou (1992) or Goody & Yung (1989).

2.2.2 The LBLDIS radiative scattering code

LBLDIS (Line-by-line Discrete Ordinate Method for Radiative Transfer) was initially developed to provide forward models for the retrieval of cloud phase from infrared spectral measurements during the SHEBA (Surface Heat Budget of the Arctic ocean) campaign (Turner et al., 2003). It is used in conjunction with LBLRTM, as shown in Figure 2.6, to calculate the modelled FIR cirrus radiance spectra shown throughout this study.

**Fig. 2.6:** Flow diagram illustrating the calculation of cirrus scene radiance spectra using LBLDIS and LBLRTM.

Clear sky optical depths are calculated for each horizontal layer using LBLRTM and then used as input by LBLDIS, along with the mean temperature for each layer. The discrete ordinate calculation is performed using 16
streams \((n = 16\) in Equation 2.20) to account for angular variability. The bulk scattering properties for the layers containing cloud are calculated as described in Section 2.2.3 and read into the code, such that it is now able to account for emission, absorption and scattering throughout the model atmosphere. The assumption made here is that within each layer the atmosphere is homogeneous. The surface temperature and emissivity, providing the boundary condition for the bottom of the atmosphere, completes the list of inputs required by LBLDIS.

### 2.2.3 Infrared optical properties of ice crystals

This section discusses how the infrared optical properties of ice crystals are calculated and then used in the LBLDIS code. Initially, the method used to create the single scattering property database using theoretical optical scattering models for small particles is outlined. The section then concludes by demonstrating how these properties are applied to a population of ice crystals to obtain the bulk scattering properties of a cirrus layer.

**Theoretical calculation of the single scattering properties of ice crystals**

The single scattering properties of ice crystals, which include the scattering phase function (the dependence of the scattered radiation intensity on scattering angle) and the scattering and extinction cross sections (see Equation 2.2), vary significantly with crystal shape, size and incident wavelength. A number of theories have been developed for calculating these properties (see Baran (2009) for a review of the theories currently in use), none of which are universally applicable to all ice crystal size parameters (the size parameter is defined as the ratio of the crystal maximum dimension to incident wavelength, Equation 2.21). For example, at far-infrared wavelengths the size parameter may have values between 0.1 and 600, covering a range greater than two orders of magnitude. It is therefore common practice to apply different methods to different regions of size parameter space, depending on which is most appropriate (Baran, 2009).
In this section, two different single scattering theories will be discussed. Mie theory describes the interactions of an electromagnetic wave with a sphere - although ice crystals are generally non-spherical, it is a useful approximation when the incident wavelength is much greater than the crystal dimension (i.e. when the size parameter is very small). The geometric optics method on the other hand is applicable to intermediate/large size parameter space, since it explicitly considers the crystal shape.

Mie theory formally describes the interaction of a plane electromagnetic wave with a dielectric sphere. A more thorough description may be found in either Liou (1992) or Goody & Yung (1989); this brief summary of the key results of the theory follows that of Salby (1996).

Mie theory solves Maxwell’s equations for the electric and magnetic field vectors incident on a dielectric sphere of radius $a$. Through the separation of variables, this solution may be expressed in terms of an expansion in spherical harmonics and Bessel functions. Properties of the scattered wave field are expressed in terms of the size parameter, $x$, given by

$$ x = 2\pi \frac{a}{\lambda} \tag{2.21} $$

where $\lambda$ is the wavelength of the incident radiation. From the scattered wave field obtained through Mie theory, it is possible to derive useful quantities such as the scattering and extinction efficiencies, which are defined as the fractional area of the incident beam removed by interaction with the sphere through scattering and extinction, respectively. These quantities are defined by:

$$ Q_{s,\nu} = \frac{\kappa_{s,\nu}}{\pi a^2} \tag{2.22} $$

and

$$ Q_{e,\nu} = \frac{\kappa_{e,\nu}}{\pi a^2} \tag{2.23} $$

where $\kappa_{s,\nu}$ is the scattering cross section at wavenumber $\nu$ and $\kappa_{e,\nu}$ is the extinction cross section (defined in Equation 2.2). The scattering and absorption efficiencies (the absorption efficiency is given by extinction minus scattering) are strongly wavelength dependent, and are related to the real
and imaginary components of the sphere’s refractive index respectively (see Figure 1.6 for the wavenumber dependence of ice refractive index).

Consider a non-absorbing sphere (imaginary component of refractive index \( n_i = 0 \) and \( Q_{s,\nu} = Q_{c,\nu} \)). For \( x \ll 1 \), \( Q_{s,\nu} \propto \lambda^4 \) (Salby, 1996), corresponding to the Rayleigh scattering regime which describes scattering from point-like particles. \( Q_{s,\nu} \) reaches a maximum where \( x \approx 2\pi \) (\( \lambda \approx a \)), and then passes through a series of oscillations with increasing \( x \). These oscillations are due to interference of light diffracted and transmitted by the particle (Hansen & Travis, 1974). As \( x \to \infty \), these oscillations in scattering efficiency damp out until \( Q_{s,\nu} \sim 2 \), as shown in Figure 2.7. This result is in agreement with geometric optics calculations, in which the amount of light diffracted by the particle into the beam is equal to the amount striking the particle, independently of particle shape and refractive index (Hansen & Travis, 1974). The scattered radiation therefore includes equal contributions from energy diffracted about the sphere, and from energy redirected by reflection and refraction within the sphere.

Figure 2.7 also illustrates the effect on the scattering efficiency of including absorption. As the imaginary component of the refractive index is increased from zero (thus introducing absorption) \( Q_{s,\nu} \) continues to exhibit similar behaviour, except that the oscillations caused by interference are reduced by increasing absorption (since the amount of radiation transmitted by the particle is reduced). Once \( n_i = 0.1 \), only a single peak near \( x \approx 2\pi \) remains, and the limiting value of \( Q_{s,\nu} \) at large \( x \) is also reduced, owing to absorption of radiation by the sphere that was transmitted in the non-absorbing case. Compared with Rayleigh scattering, Mie scattering also possesses much stronger directionality, with strong forward scattering exhibited at all wavelengths (Salby, 1996). The directionality of the scattered radiance field is often quantified using the asymmetry parameter, \( g \), which is defined as the average cosine of the scattering angle (Baran, 2009):

\[
g = \langle \cos \theta \rangle = \int_{-1}^{1} P(\cos \theta) \cos \theta \, d(\cos \theta)
\]
Fig. 2.7: Scattering efficiency $Q_s$ calculated using Mie scattering theory by Hansen & Travis (1974) for a dielectric sphere with real refractive index $n_r = 1.33$ and four different values of imaginary refractive index $n_i$. $Q_s$ is plotted as a function of the size parameter $x$.

Here $\theta$ is the scattering angle and $P(\cos \theta)$ is the scattering phase function (see Equation 2.13). $g$ can take on values between 1 and $-1$ depending on the crystal size, shape and refractive index. When scattering is completely in the forward/backward direction, $g = 1/−1$ respectively. When scattering is equally likely to occur in either the forwards or backwards direction, $g = 0$.

**Geometric optics:** Cirrus clouds consist largely of non-spherical ice crystals, such as bullet rosettes, columns, and plates (see Figure 1.4 for diagrams of these different shapes). The scattering and absorption properties of hexagonal ice crystals like these are much more difficult to calculate, com-
pared with the relatively straightforward Mie theory used for spherical particles. The method described here (following that of Liou (1992)) invokes the laws of geometric optics, where it is assumed that the incident light beam consists of separate rays which pursue independent paths. Each ray in the beam hitting the crystal then undergoes reflection and refraction, depending on the angle of incidence at each surface it encounters. The rays finally emerge from the crystal in various directions, and with different amplitudes and phases. The method is only an approximation of the wave equation approach, and requires that the width of the incident beam is much greater than the wavelength, but smaller than the particle size.

The first problem to be overcome when applying the geometric ray tracing method to a non-spherical ice crystal is defining the geometry of the orientation of the crystal with respect to the incident ray. The electric field components perpendicular and parallel to the scattering plane (defined by the crystal face on which the beam is incident) are required for the ray tracing calculation. The electric fields reflected or refracted by a given surface are calculated using Snell’s law and the Fresnel reflection coefficients. Snell’s law, given by Equation 2.25, determines the angle by which a ray is refracted when it is incident on a surface dividing two materials with different refractive indices.

\[
\frac{\sin \theta_i}{\sin \theta_t} = \frac{v_1}{v_2} = n
\]  

(2.25)

Here, \( \theta_i \) and \( \theta_t \) are the angles of the incident (i) and refracted (t) rays with respect to the normal, \( v_1 \) and \( v_2 \) are the wave velocities in the two media, and \( n = n_r + i n_i \) is the refractive index of the second medium relative to the first (\( n_r \) and \( n_i \) are the real and imaginary components, respectively). The Fresnel reflection coefficients, which govern the fraction of incident power that is reflected at the boundary between two dielectric media, are derived from the requirement of continuity at a boundary for the tangential components of the magnetic and electric field vectors; see Born & Wolf (1999) for details of their derivation. The coefficients depend on the polarisation of the incident
ray – if the light is polarised such that the electric field is perpendicular to the plane of incidence (the plane formed by the incident ray and the normal to the boundary), or \(s\)-polarised, then the reflected power is \(R_s\) as given by Equation 2.26. On the other hand, if the light is polarised such that the electric field is parallel to the plane of incidence, or \(p\)-polarised, then the reflected power is \(R_p\) (Equation 2.27).

\[
R_s = \left(\frac{\sin(\theta_t - \theta_i)}{\sin(\theta_t + \theta_i)}\right)^2
\]

\[
R_p = \left(\frac{\tan(\theta_t - \theta_i)}{\tan(\theta_t + \theta_i)}\right)^2
\]

The transmission coefficient in each case (\(T_{s,p}\), the fraction of incident power transmitted by the boundary between two dielectric media) may then be deduced from the principle of conservation of energy, since all power incident on the boundary must be either transmitted or reflected:

\[
T_{s,p} = 1 - R_{s,p}
\]

In the case of unpolarised light consisting of an equal mix of \(s\)- and \(p\)-polarisations, the reflection coefficient is given by \(R = (R_s + R_p)/2\).

Figure 2.8 illustrates some of the possible paths which can be taken by a scattered light ray, and classifies them in terms of the number of times \(l\) that the ray encounters the particle surface. Under the geometric optics approximation, light rays (labelled \(l = 0\) in Figure 2.8) which miss the particle are partially diffracted into its geometrical cross-section (Hansen & Travis, 1974). The amount of light diffracted into the beam in this way is equal to the amount striking the particle. As mentioned previously, this is independent of particle shape and refractive index. The derived scattering phase function for randomly orientated hexagonal crystals, obtained by summing the results from reflection (\(l = 1\) and \(l \geq 3\) in Figure 2.8), refraction (\(l = 2\)) and diffraction (\(l = 0\)), consists of a large peak for scattering angles less than \(\sim 10^\circ\) owing to the diffraction component. Another component con-
tributing to this peak is the $\delta$-function transmission, which occurs only at the $0^\circ$ scattering angle, and is the result of the double refraction of a ray through parallel crystal faces. Overall, the majority of light in the scattered beam comes from rays which are either diffracted or double-refracted, with only a small contribution coming from rays undergoing reflections (Hansen & Travis, 1974).

Fig. 2.8: Illustration from Hansen & Travis (1974) showing possible paths of light rays scattered by a spherical particle under the geometrical optics method. $\gamma$ is the angle of incidence for rays striking the particle.

The extinction cross section, $\kappa_e$, is equal to two times the geometric cross section (in the limit of the geometric optics approximation – particle dimension much greater than incident wavelength) since the amount of energy diffracted into the beam is equal to that incident on the particle cross section presented to the incoming beam. These two equal contributions therefore combine to give an extinction cross section which is double the geometric cross section. The average geometric cross section, $A$, for any randomly orientated crystal, is equal to one quarter of the total surface area (van de Hulst, 1981). For hexagonal prisms of length $L$, radius $r$, this is given by

$$A = \frac{3r^2}{4} \left( \sqrt{3} + \frac{2L}{r} \right). \quad (2.29)$$

From this, the extinction cross section $k$ is

$$\kappa_e = 2A = \frac{3r^2}{2} \left( \sqrt{3} + \frac{2L}{r} \right). \quad (2.30)$$
The relative contributions of scattering and absorption to the extinction cross section are then determined by the imaginary and real refractive indices of the crystal.

The method used by Baran & Francis (2004) to compute the infrared single scattering property database used in this study is the T-matrix method of Mishchenko et al. (1996). The T-matrix method is applied to simulated aggregate crystal shapes, which are defined as ensembles of circular ice cylinders of varying aspect ratios whilst conserving the aspect ratio of the whole crystal (Yang & Liou, 1998). The phase function is calculated from the asymmetry parameter using the modified Henyey-Greenstein approach described by Baran et al. (2001). This approach of representing a complex crystal geometry by an ensemble of simpler shapes is described by Baran (2003), who shows that the single scattering properties calculated using this approach are generally well within 4% of the exact calculations for an ice aggregate at mid-infrared wavelengths. The refractive indices of ice used here are from the measurements collated by Warren (1984), though it should be noted that this dataset has recently been updated by the same author (Warren & Brandt, 2008). In the updated version, there is no change in the refractive index at frequencies between 350 and 600 cm$^{-1}$. The refractive indices at lower frequencies have been altered following the availability of new laboratory measurements (Curtis et al., 2005).

Estimating the bulk scattering properties for a cirrus layer

The bulk scattering properties of the cirrus layer (defined in this section) required as input by the LBLDIS code are the cloud layer optical depth, the bulk scattering albedo and the bulk asymmetry parameter. These are estimated using the single scattering properties for ice aggregates (calculated theoretically as described in Section 2.2.3) in conjunction with a particle size distribution. The Baran database of single scattering properties (Baran & Francis, 2004) is used in this study, and is in the form of look-up tables consisting of values of extinction cross section (see Equation 2.10), single
scattering albedo (Equation 2.11) and asymmetry parameter (Equation 2.24) calculated for 88 different wavenumbers and 23 different crystal dimensions.

![Contour plot (log scale) of extinction cross section $\kappa_{e,\nu}$ (µm$^2$) for ice aggregates at infrared wavenumbers (Baran & Francis, 2004).](image)

**Fig. 2.9:** Contour plot (log scale) of extinction cross section $\kappa_{e,\nu}$ (µm$^2$) for ice aggregates at infrared wavenumbers (Baran & Francis, 2004).

Before calculating the bulk properties, the single scattering properties are each interpolated onto the same crystal dimension scale as the size distribution (in this study the scale used is 1 µm through to 4400 µm in 1 µm intervals). The interpolated values are plotted in Figures 2.9 (extinction cross section), 2.10 (single scattering albedo) and 2.11 (asymmetry parameter). Note the strong dependence on wavenumber: this is due to the sensitivity of the refractive index of ice to changes in wavenumber (Warren (1984), Warren & Brandt (2008) – see Figure 1.6).

The cloud optical depth is determined by integrating the bulk extinction coefficient over the geometrical depth of the cloud. The bulk extinction coefficient $\tilde{\beta}_{e,\nu}$ is defined in terms of the extinction cross section $\kappa_{e,\nu}$ in Equation 2.31, where $N(D)$ is the particle size distribution (PSD) and $D$ is the maximum crystal dimension. The PSD may come from in-situ observations if they are available (using instruments such as the Cloud Particle Imager, Lawson et al. (2001)) or from parameterizations, which typically re-
Fig. 2.10: Contour plot of single scattering albedo $\omega_{0,\nu}$ (dimensionless) for ice aggregates at infrared wavenumbers (Baran & Francis, 2004).

turn gamma-type distributions given cloud macroscopic properties such as ice water content and temperature as inputs (Field et al., 2005, 2007; Ivanova et al., 2001; McFarquhar & Heymsfield, 1997).

$$\bar{\beta}_{e,\nu} = \int_{0}^{\infty} \kappa_{e,\nu}(D) N(D) \, dD \approx \sum_{D=1\mu m}^{D=4400\mu m} \kappa_{e,\nu}(D) N_D$$ (2.31)

Here, $N_D$ is the number of particles per cubic metre with maximum dimension within the limits $D \pm 0.5\mu m$. The bulk extinction coefficient is calculated for each of the reference wavenumbers and then interpolated onto a high resolution ($0.01 \text{ cm}^{-1}$) wavenumber scale for use in the LBLDIS scattering code. LBLDIS requires the optical depth of the layer as input, which is obtained by multiplying the extinction coefficient by the cloud geometrical thickness. This step uses the simplifying assumption that the cloud layer is vertically homogeneous. In addition to the bulk extinction coefficient, the bulk scattering coefficient $\bar{\beta}_{s,\nu}$ is also required to calculate the bulk scattering albedo $\bar{\omega}$. This is determined by first calculating the scattering cross section from the single scattering albedo $\omega_{\nu}$ and the extinction cross section $\kappa_{e,\nu}$.
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Fig. 2.11: Contour plot of asymmetry parameter $g$ (dimensionless) for ice aggregates at infrared wavenumbers (Baran & Francis, 2004).

\[ \kappa_{s,\nu}(D) = \omega_\nu(D) \kappa_{e,\nu}(D) \]  
\[ (2.32) \]

Then, the scattering cross section is integrated over the PSD to obtain the bulk scattering coefficient $\bar{\beta}_{s,\nu}$:

\[ \bar{\beta}_{s,\nu} = \int_0^\infty \kappa_{s,\nu}(D) N(D) \, dD \approx \sum_{D=1,\mu m}^{D=4400,\mu m} \kappa_{s,\nu}(D) N_D \]  
\[ (2.33) \]

The bulk scattering albedo, given by Equation 2.34, is then calculated from the ratio of the bulk scattering and bulk extinction coefficients:

\[ \bar{\omega}_\nu = \frac{\bar{\beta}_{s,\nu}}{\bar{\beta}_{e,\nu}} \]  
\[ (2.34) \]

The bulk asymmetry parameter, $\bar{g}$, is then defined by Equation 2.35, where $g(\nu, D)$ is the asymmetry parameter for a given ice crystal dimension and wavenumber (see Equation 2.24).
\[
\bar{g}(\nu) = \frac{\int_0^\infty \bar{\beta}_{s,\nu} g(\nu, D) N(D) \, dD}{\int_0^\infty \bar{\beta}_{s,\nu} N(D) \, dD} \approx \frac{\sum_{D=1 \mu m}^{D=4400 \mu m} \bar{\beta}_{s,\nu} g(\nu, D) N_D}{\sum_{D=1 \mu m}^{D=4400 \mu m} \bar{\beta}_{s,\nu} N_D} \tag{2.35}
\]

This section has described how the bulk scattering properties of a cirrus layer required to calculate radiative transfer through the layer are estimated, given a database of single scattering properties for a particular crystal habit (in this case aggregates) and an assumed PSD. Recent work has demonstrated that the bulk scattering properties are better represented by assuming different crystal shapes for different sections of the PSD function. For example, Baum et al. (2005) assume that all crystals with maximum dimension \( D < 60 \mu m \) are droxtals, whilst larger crystals are represented by differing mixtures of bullet rosettes, hexagonal columns, hexagonal plates and aggregates depending on the value of \( D \). The assumption of a single crystal habit therefore introduces an error in the simulated cirrus scene radiance spectra shown later in this study.

### 2.3 Summary

This chapter has briefly described the radiative transfer theory used to simulate the atmospheric radiances observed by spectrometers such as the Tropospheric Airborne Fourier Transform Spectrometer used in this study (TAFTS, see Chapter 3). Starting with the specific case of a cloud-free atmosphere, the effects of absorption and emission on radiative transfer are quantified in the radiative transfer equation, with particular emphasis on the effect of the water vapour continuum absorption at far-infrared wavelengths. The radiative transfer theory is then expanded though the addition of scattering processes, as would be observed when viewing a cirrus scene. An overview of ice crystal single scattering properties has been presented, along with a description of how these properties are used to calculate the bulk properties for a whole cirrus layer required by the LBLDIS scattering code.

When compared with line-by-line radiative transfer calculations of the sort described in this chapter, high spectral resolution observations of the
2. Modelling the impact of cirrus on far-infrared radiance spectra

atmosphere such as those presented in Chapters 4 and 5 are very useful for testing some of the theory behind radiative transfer. Of particular interest for this study is whether the theoretical ice crystal scattering properties used in the calculations are successful in their current form at replicating observed cirrus radiance spectra. The next chapter describes in detail the theory and operation of the TAFTS spectrometer used to make the far-infrared spectral radiance observations presented later in this study, whilst Chapters 4 and 5 show and discuss the results obtained during two distinct field campaigns in which TAFTS was involved.
In this chapter the technique for measuring radiance spectra known as Fourier Transform Spectroscopy (FTS) is described, with emphasis on how the theory is applied in practice using TAFTS (Tropospheric Airborne Fourier Transform Spectrometer). FTS is based on the mathematical theory of Fourier analysis, which states that any mathematical function may be rewritten as a superposition of sinusoids (Davis et al., 2001). Fourier analysis may be applied in one of two symmetrical methods: Fourier decomposition, where a function is broken down into its constituent sinusoids, and Fourier synthesis, the construction of a function from appropriately weighted sinusoids. Fourier analysis is best described by the following pair of equations. Here, $F(\nu)$ is the Fourier transform of $f(x)$:

$$f(x) = \int_{-\infty}^{+\infty} F(\nu) e^{i2\pi\nu x} d\nu$$  \hspace{1cm} (3.1)

$$F(\nu) = \int_{-\infty}^{+\infty} f(x) e^{-i2\pi\nu x} dx$$  \hspace{1cm} (3.2)

An interferometer such as TAFTS works by taking an input signal $B(\nu)$, where $\nu$ is wavenumber, and effectively performing a physical Fourier decomposition. The resulting output is an interferogram, $I(x)$, where $x$ is the optical path difference. The original signal $B(\nu)$ is retrieved by computing the Fourier transform of the output signal. For an ideal instrument, $B(\nu)$ and $I(x)$ are related by Equations 3.4. A discussion of the limitations of a real instrument compared with the ideal case described here may be found in Section 3.1.
\[ I(x) = \int_{-\infty}^{+\infty} B(\nu) e^{i2\pi \nu x} d\nu \quad (3.3) \]

\[ B(\nu) = \int_{-\infty}^{+\infty} I(x) e^{-i2\pi \nu x} dx \quad (3.4) \]

FTS offers many advantages over other types of spectrometer, such as diffraction grating spectrometers (Davis et al., 2001; Thorne, 1988). The spectra obtained are high resolution with accurate photometry (the measured intensity is representative of the incident radiation), and the instrument response function is usually well known, since only a single detector is required. An FTS also provides high efficiency with minimal losses owing to diffraction, high throughput (the Jacquinot advantage), the simultaneous observation of all frequencies (the multiplex, or Fellgett, advantage), and a wide spectral range. However, since an FTS requires spectra to be computed from the raw interferograms produced by the instrument, it is difficult for the instrument operator to quickly judge whether a scan has been successful without a lot of experience.

The theory behind FTS, specifically covering how practical limitations affect the classical Fourier transform theory as expressed by Equations 3.4, will be covered in Section 3.1. A brief description of the TAFTS design, and how this is optimised for the observation of far infrared radiance from an aircraft platform, is given in Section 3.2, whilst the methods used to convert the interferograms output by TAFTS into radiance spectra are outlined in Section 3.3.

### 3.1 FTS theory

The theory of FTS presented here follows closely the work of Thorne (1988) and Davis et al. (2001).

The ideal interferogram, given a steady, uniform radiation source, an infinite scanning path length, and a ‘perfect’ instrument (one in which no information is lost as it passes through) would be a real, symmetric function.
However, this is generally not the case, owing to instrumental imperfections and limitations. This non-ideal nature may be represented by an instrument function, which is defined as the intensity distribution of the output recovered from a monochromatic input. For the ideal instrument, this is simply a delta function.

### 3.1.1 Finite path difference and apodisation

In practice, it is not possible to measure an interferogram over an infinite path length. The scan length must be stopped at some maximum path difference $L$, such that the interferogram is only measured for path differences given by $-L \leq x \leq +L$. This is equivalent to multiplying the whole interferogram by a rectangular function equal to 1 between $\pm L$, and 0 elsewhere.

Consider a monochromatic input at wavenumber $\nu_0$. The interferogram is then

$$I(x) = B(\nu_0) \cos 2\pi\nu_0 x. \quad (3.5)$$

On performing the Fourier transform, the spread of intensity due to the finite path difference is given by

$$S(\nu) = B(\nu_0) \int_{-L}^{+L} \cos(2\pi\nu_0 x) \cos(2\pi\nu x) dx \quad (3.6)$$

which, when evaluated, becomes

$$S(\nu) = B(\nu_0)L \left[ \frac{\sin 2\pi(\nu_0 + \nu)L}{2\pi(\nu_0 + \nu)L} + \frac{\sin 2\pi(\nu_0 - \nu)L}{2\pi(\nu_0 - \nu)L} \right]. \quad (3.7)$$

The first of these two terms is negligibly small compared with the second for all positive $\nu$, so will be ignored here. The $L$ outside the square brackets is a normalisation factor, so the instrument function owing to the finite optical path length is a sinc function,

$$\text{sinc} 2\pi\nu L \equiv \frac{\sin 2\pi\nu L}{2\pi\nu L}. \quad (3.8)$$

The first zeroes of the sinc function are at $\pm 1/2L$. This is normally used to define the resolving power, $R = \nu/\delta\nu$, of the spectrometer, where $\delta\nu$ is
known as the resolution limit:

\[ \delta \nu = \frac{1}{2L} \Rightarrow R = 2L \nu \quad (3.9) \]

TAFTS has a total scan length \( 2L = 100 \text{ mm} \), which gives the instrument a resolution limit \( \delta \nu = 0.1 \text{ cm}^{-1} \).

The truncation of the interferogram at \( x = \pm L \) introduces side-lobes on either side of the central peak, through the sinc function form taken by the instrument function. The first of these side-lobes are 20% of the height of the central peak, and are clearly undesirable features. They can be corrected for through a process called apodisation, which involves multiplying the interferogram by a smoothing function. The smoothing function is chosen to make the discontinuity at \( x = \pm L \) less sharp, and therefore reduce the size of the side-lobes. However, apodisation broadens the central peak as well as smoothing out the side-lobes, so therefore reduces the resolution. The selection of an appropriate apodisation function becomes a balance between achieving good resolution and the need to reduce the size of the side-lobes introduced by the finite maximum path difference.

### 3.1.2 Sampling the interferogram

Up to this point, it has been assumed that the output interferogram is smooth and continuous. However, in reality the interferogram is sampled at regular intervals of length \( \Delta x \). This is equivalent to multiplying the interferogram by a Dirac comb with spacing \( \Delta x \). The result of the Fourier transform is, by the convolution theorem, equal to the transform of the original interferogram, multiplied by the transform of the Dirac comb, which is another Dirac comb, but with spacing \( \Delta \nu = 1/\Delta x \). This means that the transformed spectrum is replicated at intervals \( \Delta \nu = 1/\Delta x \) in the wavenumber domain, as a result of the discrete sampling of the interferogram. This replication is known as aliasing.

Aliasing can result in the desired spectrum overlapping with its replica spectra, unless the sampling interval is smaller than a maximum value \( \Delta x_m \).
3. The TAFTS instrument

If the original spectrum extends from $\nu = 0$ to some maximum value $\nu_m$, then the computed spectrum occupies a range $2\nu_m$, owing to the mirror image. To avoid overlap, the interval between replicas $\Delta \nu = 1/\Delta x$ must therefore be greater than $2\nu_m$. This condition gives an expression for the maximum sampling interval:

$$\Delta x_m = \frac{1}{2\nu_m} = \frac{\lambda_{\text{min}}}{2} \quad (3.10)$$

To avoid aliasing, it is therefore necessary to sample the interferogram at least every half wavelength of the shortest wavelength sampled. If this condition is not satisfied, it becomes impossible to extract the spectrum unambiguously from the interferogram. The shortest wavelength sampled by TAFTS is 12.5 $\mu$m, which gives a maximum sampling interval $\Delta x_m = 6.25 \mu$m. The sampling interval used by TAFTS is 2.5 $\mu$m, a factor of 2.5 smaller than the maximum sampling interval required to prevent aliasing.

3.1.3 Phase corrections

A further problem associated with real Fourier transform spectrometers arises when the interferogram sampled is asymmetric. Asymmetry may be introduced as a result of experimental, instrumental or computational limitations. An asymmetric function may be expressed as the sum of a symmetric and an antisymmetric component, so it follows that the cosine transforms used previously in this section are insufficient, since the integral of a cosine multiplied by an antisymmetric function is zero. Therefore, the antisymmetric component of the signal would be lost, reducing the signal-to-noise ratio.

The reconstruction of the full spectrum from an asymmetric interferogram requires a complex Fourier transform, like that given by Equations 3.4, which produces a complex spectrum. The real component cannot be used alone, since without the information contained in the imaginary component it would contain errors in the spectral line shapes, intensities and locations. The true spectrum is recovered by calculating the phase shift, and correcting for it.
A general, sampled interferogram is described by

\[ I(x_n) = \sum_{j=-N}^{N} B(\nu_j) e^{i2\pi\nu_j x_n} \quad (3.11) \]

where \( I(x_n) \) is one of the samples of the interferogram, taken when the optical path difference is \( x_n \). The introduction of a phase shift means that the exponent in Equation 3.11 does not go to zero when \( x \) is zero, so the equation is rewritten as

\[ I(x_n) = \sum_{j=-N}^{N} B(\nu_j) e^{i(2\pi\nu_j x_n + \phi_j)} \quad (3.12) \]

where \( \phi_j \) is the phase shift corresponding to wavenumber \( \nu_j \). Two of the more common causes of phase shifts are listed here:

1. The sampling grid has no point coinciding with \( x = 0 \). Then \( x_n = n\Delta x + \alpha \), and the exponent in Equation 3.11 is

\[ 2\pi\nu_j (n\Delta x + \alpha) = 2\pi\nu_j n\Delta x + 2\pi\nu_j \alpha \Rightarrow \phi_j = 2\pi\nu_j \alpha, \quad (3.13) \]

resulting in a phase shift proportional to wavenumber.

2. There is some unbalanced dispersion in one arm of the interferometer caused, for example, by the compensator plate not having the correct thickness. This produces an exponent given by

\[ +2\pi\nu_j (x_n + r_j d), \quad (3.14) \]

where \( r_j \) is the refractive index of the plate material at wavenumber \( \nu_j \), and \( d \) is the plate thickness. This introduces the \( \nu \) dependence of the refractive index into the phase shift.

3.1.4 Systematic errors

Here, two FTS problems specifically relevant to TAFTS are discussed. The first concerns the error introduced by an unwanted systematic signal, whilst the second addresses the effect of a periodic error in interferogram sampling.
Unwanted systematic signal} Electrical pickup from a stray source is a common cause of systematic noise, and may be represented by a sinusoidal signal $\epsilon_1$. The interferogram of a monochromatic source sampled between $-L < x < +L$ with systematic noise is given by

$$I(x) = \cos(2\pi\nu_0 x) + \epsilon_1. \quad (3.15)$$

By the principle of superposition, a complex waveform may be formed from the linear superposition of its harmonic components in either domain (Davis et al., 2001). The output spectrum may therefore be written as $B'(\nu) = B(\nu) + B_1(\nu)$, where $B(\nu)$ is the desired spectrum

$$B(\nu) = \int_{-L}^{+L} \cos(2\pi\nu_0 x)e^{-i2\pi\nu x}dx, \quad (3.16)$$

and $B_1(\nu)$ is the spectral feature introduced by the systematic noise:

$$B_1(\nu) = \int_{-L}^{+L} \epsilon_1 e^{-i2\pi\nu x}dx \quad (3.17)$$

The effect on the spectrum depends on the rate at which the interferogram is sampled.

**Periodic sampling error** A periodic error in the interferogram sampling intervals produces replicas of the spectrum shifted by a wavenumber interval which depends on the frequency of the error. A periodic error in the path difference may be expressed by

$$\epsilon_2 = \epsilon_{2,0}\sin(2\pi x\delta\nu + \phi), \quad (3.18)$$

where $\epsilon_{2,0}$ is the amplitude, and $\delta\nu$ the spatial frequency of the error. For a monochromatic source, the resulting interferogram is

$$I(x) = B(\nu_0)\cos(2\pi\nu(x + \epsilon_{2,0}\sin(2\pi x\delta\nu + \phi))). \quad (3.19)$$

In the limit $2\pi\nu\epsilon_{2,0} \ll \frac{\pi}{2}$:

$$I(x) = B(\nu)(\cos(2\pi\nu x) - \pi\nu\epsilon_{2,0}(\sin(2\pi x(\nu \pm \delta\nu)))) \quad (3.20)$$
for $\phi = 0$, and

$$I(x) = B(\nu)(\cos(2\pi \nu x) - \pi \nu x \cos(2\pi x(\nu \pm \delta \nu)))$$  \hspace{1cm} (3.21)

for $\phi = \frac{\pi}{2}$. Once the interferogram is transformed, replica lines appear at wavenumbers $\nu \pm \delta \nu$, with amplitudes $B(\nu)\pi \nu x \cos(2\pi x(\nu \pm \delta \nu))$. The relative phase of the interferogram sampling frequency and the error frequency rotates the replica lines through the real and imaginary phases.

Figure 3.1 shows the two perpendicular extremes described by Equations 3.20 and 3.21, where the effect of the replica lines is maximised. In the real component, the replicas have opposite signs, whereas in the imaginary component both are positive. Therefore, given the intensity and position of the replica lines it is possible to determine the size of the periodic sampling error, its phase and its period.

**Fig. 3.1**: Schematic diagram of the replica lines created by a periodic sampling error.
In the case of broadband spectra, the replica lines are instead copies of the entire spectrum, centered at \( \pm \delta \nu \) and added to the original spectrum. Given a large shift, the replica shifted in the negative wavenumber direction is mirrored at 0 cm\(^{-1}\), and folded back into the spectrum.

### 3.2 TAFTS instrument description

This section gives a brief description of the TAFTS instrument; a more detailed breakdown of all the components may be found in Green (2003). Figure 3.2 is a photograph of TAFTS with the three main modules labelled, which are described in each of the next three sections.

![Figure 3.2: A photograph of the TAFTS instrument (electronics box not included). A: Pointing optics B: Interferometer C: Cryostat](image)

Fig. 3.2: A photograph of the TAFTS instrument (electronics box not included). A: Pointing optics B: Interferometer C: Cryostat
3. Pointing optics

The pointing optics box (POB) steers inputs from the nadir and zenith angles into the main interferometer, with an acceptance solid angle of $6.28 \times 10^{-4}$ steradians in each of the two directions. The Martin-Puplett interferometer design (see Martin & Puplett (1970) and Section 3.2.2) was chosen for its easy use of both input ports simultaneously. The incoming radiation is directed into the interferometer via three gold-plated mirrors in each arm. The first mirror in each arm is steerable, by means of a stepper motor, to allow the input to alternate between either the incoming radiation or one of the internal blackbodies (also contained within the pointing optics box) used for calibration. The blackbodies, of which there are two for each arm, are short cavities coated in carbon epoxy. The ‘hot’ blackbody in each arm is kept well above the ambient temperature during operation, whilst the ‘cold’ blackbody is heated to a few degrees above ambient, to prevent condensation.

The number of mirrors in each arm are equal, in order to balance the inputs with respect to the reflectivities of the mirrors. In the downwelling path, one of the mirrors is positioned inside the evacuated main interferometer unit, whilst in the upwelling arm all three are housed in the pointing optics box. If the reflectivities of the mirrors within the pointing optics box degrade over time, owing to exposure to the outside environment, then the unequal number of mirrors in each arm within the POB will cause an imbalance in the total reflectance of each arm. The radiation is passed from the POB through to the interferometer via two polypropylene windows (one for each input arm), which transmit very effectively at FIR wavelengths.

3.2.2 Interferometer

The Martin-Puplett interferometer design (Martin & Puplett, 1970) is used in TAFTS, and is illustrated schematically in Figure 3.3.

The inputs from each of the two arms, $I_1$ and $I_2$, are combined at beam combiner $BC$. Both the beam combiner and the beam splitter, $BD$, are polarizers, consisting of 1.5 $\mu$m thick mylar membranes upon which grids of
Fig. 3.3: A schematic representation of the Martin-Puplett interferometer design used in TAFTS. Note that the angle of the apex in the real TAFTS roof mirror is 90°.

1 μm wide copper wires with a 2 μm spacing are mounted. The divided beam is then reflected by a pair of rooftop mirrors, which in TAFTS are mounted back-to-back. The rooftop mirror, though it rotates the polarization angle by 90°, greatly reduces the sensitivity of the scan to misalignment in one vertical plane relative to a plane mirror. The back-to-back design enables a greater maximum path length to be obtained from the physical distance available. Combining this reduction with the return path to the beam splitter gives a four times folding of the optical path. The required maximum path difference of 100 mm is achieved with a carriage movement range of only 25 mm, which is ideal given the spatial constraints imposed on the design by the size of the aircraft mount.
Smooth, accurate motion of the back-to-back rooftop mirror is maintained by positioning the mirror mount upon three steel balls, mounted within three v-grooves cut in the direction of travel. The mount is driven by a micro-stepping motor located below the assembly, geared to the mirror mount via a pulley system.

The interferometer is contained within a vacuum tight aluminium tank. A vacuum tight tank is required for two reasons; to provide isolation of the system from acoustic noise, and to prevent path dependent absorption discrepancies between the beams passing along the two arms during scans.

3.2.3 Cryostat

Once the two beams are recombined at the beam splitter, the single resultant beam is passed via another polypropylene window through to the cryostat, which contains the instrument detector optics. The detector optics are kept within a shielded vacuum enclosure below a liquid helium tank, which keeps the detectors at 4.2 K during operation to negate the effect of thermal radiation from the optics container. A liquid nitrogen cooled shield and a layer of super insulation surround the liquid helium tank.

The beam from the main interferometer module initially passes through a pair of polypropylene filters, which prevent all radiation with wavelengths greater than 12 \( \mu m \) from entering the detector optics. The beam is then split by the analyser (see Figure 3.3), reflecting and transmitting the complementary differential spectra. Since the spectral range cannot be covered by a single detector, each output beam is passed through a dichroic filter, which splits the spectral range at 330 cm\(^{-1}\). TAFTS therefore has four detector channels; two ‘longwave’ channels (0 and 1, spectral range 80 – 330 cm\(^{-1}\)) using Ge:Ga photoconductor detector crystals, and two ‘shortwave’ channels (2 and 3, spectral range 330 – 800 cm\(^{-1}\)) using Si:Sb crystals. The two channels in each spectral range correspond to the two complementary outputs available as a result of using the Martin-Puplett interferometer design. The current output by each detector is then amplified by a pre-amplifier mounted
on the cryostat before being passed on to the TAFTS electronics box.

3.2.4 Electronics box

The TAFTS electronics are contained separately from the instrument in a box which is located inside the aircraft during science flights. An analogue to digital converter digitizes the detector output, which is then stored on two hard disks along with the laser fringe timings and housekeeping data for later processing. The data is divided between two disks to ensure that only half of the data is lost in the event of a disk failure. A single on-board PC within the electronics box controls the sampling, the orientation of the pointing mirrors and the internal blackbody temperatures.

3.3 Conversion of TAFTS interferograms into radiance spectra

This section describes how a radiance spectrum is derived from the raw interferogram recorded by TAFTS. The process is broadly divided into two stages, which will be referred to as the processing and calibration stages. The first stage (Section 3.3.1) involves performing a Fourier transform on the measured interferogram to obtain a raw differential spectrum. The second stage (Section 3.3.2) then derives the instrument response function from internal blackbody spectra, which is then applied to external view spectra to obtain a full radiometric calibration of the sky view radiance spectrum.

3.3.1 Processing TAFTS interferograms

The interferograms recorded by TAFTS are sampled at regular time intervals. The Fourier transform performed to obtain the spectrum requires that the interferogram is sampled at regular spatial intervals, so the interferogram must be interpolated onto a uniform spatial grid. This is achieved through the Brault method, which is described briefly below.
3. The TAFTS instrument

The Brault sampling scheme

In a continuous scanning FTS like TAFTS, the central issues in retrieving radiance spectra are when and how to sample the interferogram. The reliable recognition of equal path difference is needed to reduce errors in sampling position accuracy, and therefore reduce the systematic periodic sampling errors introduced in Section 3.1.4.

The sampling scheme devised by Brault (1996) achieves this by sampling the interferogram constantly in time, and using a co-aligned He:Ne laser to monitor the fluctuations in scan velocity. The velocity data is then used to interpolate the equal-time sampled interferogram into one with equal spatial sampling. The entire sampling system is locked to a high speed master clock signal at 40 MHz. This signal is divided to obtain the sampling clock used by the analogue to digital converters (ADCs), and is also fed to a counter. The counter accumulates 25 ns clock ticks until a laser fringe is detected, at which point the counter value is stored. The result is an array of time values (the differential fringe timings) indicating the times at which the optical path difference through the interferometer was an integral number of wavelengths. These timings, recorded for every four fringes detected, are used to establish a uniform spatial sampling grid $t(x_n)$, which is used in conjunction with the interferogram recorded at equal-time intervals, $I(t_j)$, to determine $I(x)$.

As mentioned above, in order to apply the Brault method it is necessary to also know the position of the scanning mirror as a function of time. This information is contained within the differential fringe timings (DFTs). Figure 3.4 shows DFTs for a single scan along with the Fourier transform of the DFTs, which shows that the timings are essentially constant apart from a clear sinusoidal oscillation at a frequency of $\sim 200$ Hz. This is due to vibrations of the motor driving the scanning mechanism.

Before the DFTs are used to interpolate the interferogram, a couple of corrections are applied. Firstly, the laser fringes are occasionally over-counted (for example, the time is recorded after only three fringes are detected, rather than the required four fringes) resulting in lower DFTs than would be ex-
Fig. 3.4: Upper panel: differential fringe timings for a single scan. Lower panel: Fourier transform of the fringe timings.

pected. Examples of this phenomenon are highlighted by the red crosses in Figure 3.4, and are detected automatically using a threshold minimum DFT which allows for some natural variation in the scan mirror velocity. It is corrected for by replacing the anomalous fringe timing with the mean value of the DFTs recorded just before and just afterwards. Secondly, the opposite problem occurs when some fringes are missed completely, resulting in higher than expected DFTs (see Figure 3.5 for an example). This is particularly common when TAFTS is mounted on an aircraft, since the vibrations encountered in flight cause the fringe contrast to be reduced (Green, 2003).

These ‘spikes’ in the DFTs are corrected for by dividing them by the regular fringe spacing to determine the number of fringes missed between successfully detected fringes, and then inserting the missing fringes into the DFT array. The lower panel of Figure 3.5 demonstrates how this method
Fig. 3.5: Upper panel: differential fringe timings for a single scan affected by missed fringes. The corrected DFTs are shown in red. Lower panel: Close up of the upper panel focussing on the DFT ‘spike’.

Interpolation of the interferogram

The kernels used to interpolate the interferogram from a uniform temporal grid to a uniform spatial grid are 75 point long skewed sinc functions, stored in a three dimensional array of 75 by 1024 by 4 elements. The extent of the skew and the kernel amplitude are the corrections for the variations in scan velocity and sample timing (Brault, 1996). The effect of this interpolation on the interferogram is shown in Figure 3.6. The lower panel, which is a close-up view about the point of zero path difference, clearly illustrates how the sampling error jitter due to scan velocity variations is removed by the interpolation process.
Fig. 3.6: An example channel 0 interferogram before (black) and after (red) interpolation. The lower panel shows a close up view of the interferogram, centered on the position of zero path difference.

The next stage of processing before performing the Fourier transform is to apodise the interferogram (see Section 3.1.1). A Kaiser function is chosen for the apodisation, since it provides the best compromise between achieving good resolution and reducing the impact of side-lobes appearing in the final spectrum due to the instrument’s finite maximum path difference. Apodisation of the interferogram requires knowledge of the centre burst position. This is determined automatically by calculating the correlation between the interferogram and a theoretical centre burst shape, as shown in Figure 3.7. The position for which the correlation is highest is taken to be the centre burst position when performing the apodisation.
3. The TAFTS instrument

Fourier transform of the interferogram and spectral phase correction

Once processing of the interferogram is complete, it is then Fourier transformed using a Fast Fourier Transform (FFT) routine to obtain the observed differential spectrum. An example interferogram and its Fourier transform are shown in the top two panels of Figure 3.8. Note the non-zero imaginary component of the spectrum, plotted in red in the second panel. This is due to the presence of an asymmetric component in the interferogram, as described in Section 3.1.3, which must be corrected for to recover the true observed spectrum and increase the signal-to-noise ratio.

Before calculating the phase, the spectral resolution is reduced to save processing time (see the third panel of Figure 3.8). The phase ($\phi$) is given
Fig. 3.8: T AFTS interferogram processing routine output. From top to bottom: Interferogram set up for FFT; FFT of interferogram (real component in black, imaginary component in red); Low resolution FFT used to calculate phase; Phase; Final phase corrected spectrum.
by the following expression, where \( S_{\text{orig}} \) is the original spectrum:

\[
\phi(\nu) = \tan^{-1}\left( \frac{\text{Im}(S_{\text{orig}}(\nu))}{\text{Re}(S_{\text{orig}}(\nu))} \right)
\]  

(3.22)

The phase calculated for the example used in this section is shown in the fourth panel of Figure 3.8. This is used to shift the imaginary component into the real domain via the transformation shown below (the wavenumber dependences of \( S \) and \( \phi \) are omitted for clarity):

\[
\text{Re}(S_{\text{corr}}) = \text{Re}(S_{\text{orig}}) \cos \phi + \text{Im}(S_{\text{orig}}) \sin \phi
\]

(3.23)

\[
\text{Im}(S_{\text{corr}}) = \text{Re}(S_{\text{orig}}) \cos \left( \frac{\pi}{2} - \phi \right) - \text{Im}(S_{\text{orig}}) \sin \left( \frac{\pi}{2} - \phi \right)
\]

(3.24)

Applying this transformation gives the final corrected spectrum, \( S_{\text{corr}} \), shown for the example interferogram used during this section in the bottom panel of Figure 3.8. The spectrum is now ready for use in the radiometric calibration procedure described in Section 3.3.2.

### 3.3.2 Radiometric calibration of TAFTS spectra

This section presents the procedure for obtaining radiometrically calibrated radiance spectra from raw TAFTS spectra, which have been processed from measured interferograms as described in Section 3.3.1. The procedure is illustrated throughout this section using laboratory measurements of an external blackbody as an example. The procedure described also applies only to calibration of down-welling external view spectral radiances (the procedure for up-welling radiances is very similar, but left out here for clarity). Firstly the instrument response functions are calculated from internal view spectra, which involve using different combinations of the TAFTS internal blackbodies as input for each of the two interferometer arms. These response functions are then used to determine the instrument self emission term, known as the residue. Finally, these components are brought together and applied to external view spectra to calculate the external spectral radiance.
Calculating the instrument response function

The instrument response functions are determined from internal view spectra, which comprise measurements of the TAFTS internal blackbodies only. Within the pointing optics box there are four internal blackbodies; two for the up-welling input, and two for the down-welling. Each input has one hot blackbody and one cooler blackbody associated with it, and the corresponding blackbodies for each input are maintained at approximately the same temperature. During each cycle of measurements four different internal view combinations are observed in turn, which are labelled as $hc$, $ch$, $cc$ and $hh$, along with observations of the external source consisting of two view combinations: $hd$ and $cd$. In each case, the first letter refers to the internal blackbody being viewed via the up-welling interferometer arm (hot or cold), whilst the second letter indicates the down-welling target (hot, cold or down-welling external source).

\[\begin{array}{c}
\text{HC: 334.2 K, 307.1 K} \\
\text{CH: 307.8 K, 333.2 K} \\
\text{HH: 334.2 K, 333.2 K} \\
\text{CC: 307.9 K, 307.2 K}
\end{array}\]

**Fig. 3.9:** Raw TAFTS channel 0 internal view spectra used to determine the instrument response functions. Note that response functions must be derived separately for each of the four TAFTS channels. The internal blackbody temperatures for each combination of views (see text for a description of how the different view combinations are labelled) shown here are given in the top right corner of the figure.

Examples of the four different internal view spectra are shown in Fig-
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Figure 3.9, along with the internal blackbody temperatures. Each spectrum is taken from the mean of eight individual scans, to reduce the effect of random detector noise. As mentioned previously, the spectra measured by TAFTS are difference spectra – the measured quantity is the difference between the contributions from the up-welling and down-welling inputs, plus a term owing to the instrument self-emission. The measured internal view difference spectrum \( S_{xy} \) may be represented by the following equation:

\[
S_{xy} = S_{up} - S_{down} + X
\] (3.25)

Here \( S_{up} \) and \( S_{down} \) are the contributions from the up-welling and down-welling arms, whilst \( X \) is the residue term. The subscripts \( x \) and \( y \) refer to the up-welling and down-welling blackbodies respectively, and may take the value \( h \) or \( c \). The up- and down-welling contributions are sums of the thermal radiance emitted by the up- or down-welling blackbody, the radiance reflected by that blackbody into the interferometer, and a contribution from the interferometer optics:

\[
S_{up} = R[r_u\epsilon_{BB}B(T_{ux}) + r_u(1 - \epsilon_{bb})B(T_{amb}) + (1 - r_u)B(T_{amb})]
\] (3.26)

\[
S_{down} = R[r_d\epsilon_{BB}B(T_{dy}) + r_d(1 - \epsilon_{bb})B(T_{amb}) + (1 - r_d)B(T_{amb})]
\] (3.27)

In these expressions \( R \) is the instrument response function, \( \epsilon_{BB} \) is the internal blackbody emissivity, \( r_u \) and \( r_d \) are products of the reflectivities and transmittances of the optical components along the up- and down-welling arms respectively. \( B(T_{ux}) \) and \( B(T_{dy}) \) are the Planck functions of the internal blackbodies, whilst \( B(T_{amb}) \) is the Planck function of the ambient instrument temperature. \( T_{amb} \) is assumed to be uniform across the whole interferometer, and therefore independent of input arm. Substituting these into Equation 3.25 gives an expression for \( S_{xy} \) in terms of the instrument properties described above:
\[ S_{xy} = R_{ru}[\epsilon_{BB}B(T_{uh}) - \epsilon_{BB}B(T_{amb})] - R_{rd}[\epsilon_{BB}B(T_{dy}) - \epsilon_{BB}B(T_{amb})] + X \]  

(3.28)

Rather than obtaining \( R \) for the whole instrument, it is sufficient to calculate arm-specific response functions equal to \( R_{ru} \) and \( R_{rd} \). This removes the need to determine \( r_u \) and \( r_d \) explicitly. \( R_{ru} \) and \( R_{rd} \) are calculated by taking the difference between two calibration spectra. For example, by subtracting \( S_{cc} \) from \( S_{hc} \) the contributions from the down-welling arm and the residue are cancelled out, enabling the up-welling response function to be calculated from the remaining terms, as follows:

\[ S_{hc} - S_{cc} = R_{ru}[\epsilon_{BB}B(T_{uh}) - \epsilon_{BB}B(T_{uc})] \]  

(3.29)

\[ R_{ru} = \frac{S_{hc} - S_{cc}}{\epsilon_{BB}B(T_{uh}) - \epsilon_{BB}B(T_{uc})} \]  

(3.30)

Subtracting \( S_{ch} \) from \( S_{hh} \) provides an alternative expression for \( R_{ru} \). The same method is applied to the internal view spectra to derive \( R_{rd} \), using \( S_{ch} - S_{cc} \) and \( S_{hh} - S_{hc} \) to cancel out the up-welling contribution. The arm specific response functions calculated from the internal view spectra in Figure 3.9 are shown in Figure 3.10.

Calculating the residue

Though the instrument self-emission term (or residue) cancels out when determining the response functions, it is still necessary to calculate it since it contributes to the external view spectrum, and will therefore need to be taken into account when deriving the observed spectral radiance. It is estimated by re-arranging the expression for the internal view spectrum (given by Equation 3.28):

\[ X = S_{xy} - R_{ru}[\epsilon_{BB}B(T_{ux}) - \epsilon_{BB}B(T_{amb})] + R_{rd}[\epsilon_{BB}B(T_{dy}) - \epsilon_{BB}B(T_{amb})] \]  

(3.31)
Fig. 3.10: $R_{ru}$ and $R_{rd}$ calculated from the internal view spectra shown in Figure 3.9. In each case, the dotted lines are the functions calculated from each combination of internal view spectra (see text) and the solid line is the mean function.

The residue $X$ is calculated for each of the four internal blackbody combinations ($hc$, $ch$, $cc$ and $hh$). The mean residue is then used in the final part of the calibration procedure as described in the following section. The residue should be identical for all four combinations (see Figure 3.11), though may be different for each detector channel. A separate residue is therefore calculated for each channel.

Calculating the external view spectral radiance

The final stage of the calibration involves applying the response functions and residue calculated in the previous two sections to the external view spectra to obtain the desired quantity, which is the external view spectral radiance. Ex-
Fig. 3.11: Spectral residue calculated for channel 0 from the internal view spectra in Figure 3.9 and the response functions in Figure 3.10. The coloured lines correspond to residues calculated from the four different blackbody combinations whilst the black line is the mean spectral residue.

amples of external view spectra are shown in the upper panel of Figure 3.12. They differ from the internal view spectra in that the down-welling input is the external view spectral radiance, such that \( S_{\text{down}} \) is given by:

\[
S_{\text{down}} = R[r_d S_{\text{ext}} + (1 - r_d) B(T_{\text{amb}})]
\]  

(3.32)

The measured differential external view spectrum, after subtracting \( S_{\text{down}} \) from \( S_{\text{up}} \) and adding the residue \( X \), is therefore:

\[
S_{\text{xd}} = R[r_u \epsilon_{\text{BB}} B(T_{\text{ux}}) - \epsilon_{\text{BB}} B(T_{\text{amb}})] - R[r_d S_{\text{ext}} - B(T_{\text{amb}})] + X
\]  

(3.33)

In each of these expressions \( S_{\text{ext}} \) is the external view spectral radiance, which is the desired quantity. It is obtained for each of the two external view spectra combinations (\( h_d \) and \( c_d \)) by re-arranging Equation 3.33, giving:

\[
S_{\text{ext}} = \left( \frac{r_u}{r_d} \right) \epsilon_{\text{BB}} B(T_{\text{ux}}) - \frac{S_{\text{xd}}}{R[r_d]} + \left[ 1 - \left( \frac{r_u}{r_d} \right) \epsilon_{\text{BB}} \right] B(T_{\text{amb}}) + \frac{X}{R[r_d]}
\]  

(3.34)
The lower panel of Figure 3.12 shows the external view spectral radiances calibrated using the example data shown throughout this section, as determined from $hd$ spectra (drawn in red) and $cd$ spectra (drawn in blue). Note that, as mentioned previously, it is not necessary to explicitly know $R$, $r_u$ and $r_d$ to perform the calibration using Equation 3.34 – it is sufficient to know the products $Rr_u$ and $Rr_d$ only.

**Fig. 3.12:** The upper panel shows the raw external view spectra obtained using the two different view combinations available ($hd$ and $cd$). The internal blackbody temperature is given for each case, along with the external target blackbody temperature (263.6 K). The lower panel shows external view spectral radiances calibrated using each of the two view combinations shown in the upper panel. The external blackbody curve is overplotted (black dashed line).
3.3.3 Uncertainties in TAFTS radiance spectra

This section summarises the uncertainties associated with each component of the calibration procedure, and how much they contribute to the error in the final radiance spectrum. Descriptions of the error sources and more detailed error analyses of TAFTS radiance spectra may be found in Section 4.9 of Green (2003) and Section 3.4 of Cox (2007). The contribution of the detector noise to spectral radiance error is discussed here, since TAFTS is a detector noise (rather than photon noise) limited instrument. Other contributing factors are then briefly tabulated at the end of this section.

For a function \( f(x, y, z) \) of quantities \( x, y \) and \( z \), the error \( \sigma_f \) is given by:

\[
\sigma_f^2 = \left( \frac{df}{dx} \right)^2 \sigma_x^2 + \left( \frac{df}{dy} \right)^2 \sigma_y^2 + \left( \frac{df}{dz} \right)^2 \sigma_z^2
\]  \hspace{1cm} (3.35)

Here, \( \sigma_{x,y,z} \) is the uncertainty in \( x, y \) and \( z \). This expression assumes that these uncertainties are independent, random and uncorrelated (Barlow, 1989). Errors may also be systematic as well as random, and these should be treated separately from the random errors. Individual systematic errors may however be combined to find the total systematic error by using Equation 3.35. If the random and systematic errors are independent of each other, they may then be added in quadrature to determine the total error.

The main sources of random error in TAFTS radiance spectra are from detector noise and the temperature measurement of the internal blackbodies. The detector noise may be determined using the imaginary component of the raw uncalibrated spectra. The uncertainty \( \sigma_{\text{noise}} \) owing to detector noise of \( N \) co-added spectra is given by:

\[
\sigma_{\text{noise}} = \frac{SD_{\text{Im}}}{\sqrt{N}}
\]  \hspace{1cm} (3.36)

In this equation \( SD_{\text{Im}} \) is the standard deviation of the imaginary component of the processed spectra. The imaginary component is used since it consists purely of noise, assuming that the phase correction applied during the interferogram processing stage (Section 3.3.1) is correct. Green (2003)
demonstrated that the detector noise levels in the real and imaginary components of the processed spectra are comparable, enabling the use of the imaginary component to determine the detector noise if the real component is varying rapidly (e.g. the external view scene is changing).

\[ \sigma_S^2 = \left( \frac{\delta S_{\text{ext}}}{\delta S_{\text{xd}}} \right)^2 \sigma_{\text{noise}}^2 = \left( \frac{1}{Rr_d} \right)^2 \left( SD_{\text{Im}} \right)^2 N \] (3.37)

Figures 3.13 and 3.14 show the noise equivalent radiances and signal to noise ratio (SNR) for the RHUBC cirrus scene (see Section 4.4), for the long-
wave (channel 0) and shortwave (channel 2) channels respectively. The SNR is simply the ratio obtained through dividing the final calibrated spectrum \( S_{\text{ext}} \) by the noise equivalent radiance \( \sigma_S \).

![Graph showing radiance and SNR](image)

**Fig. 3.14:** Channel 2 noise equivalent radiance and SNR for the RHUBC cirrus scene described in Section 4.4.

The radiance uncertainty due to detector noise is generally between 0.03 and 1.0 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\) for the longwave detector, compared with 0.4 - 1.0 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\) for the shortwave detector. In both cases, the detector noise has been reduced by co-adding 30 scans. The equivalent values for a single scan (estimated by multiplying the previous values by \( \sqrt{30} \)) are approximately 0.15 - 5.5 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\) and 2.2 - 5.5 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\) for the longwave and shortwave detectors respectively.

The further sources of error in TAFTS spectra are listed in Tables 3.1 and 3.2. The errors for each set of channels (longwave (Green, 2003) and
shortwave (Cox, 2007)) are divided into response function and calibration errors. Within each of these divisions, errors are listed as either random or systematic. The response function errors contribute to the total error for the calibrated radiance spectrum, since the response function is used as part of the calibration process. The primary contributors to systematic errors are the internal blackbody temperature measurements, and the assumed blackbody emissivity (which varies between 0.982 at 80 cm\(^{-1}\) and 0.991 at 1000 cm\(^{-1}\)). These systematic errors in the accuracy may lead to biases in the final calibrated spectra as they propagate through the calibration procedure, as demonstrated in Green (2003) and Cox (2007).
3.3.4 Summary

This chapter has described the TAFTS instrument and some of the theory behind its operation and design. The routines for processing the interferograms and calibrating the radiance spectra based on this theory have been outlined, using a laboratory measurement of an external blackbody source as an example. Finally, the sources of error contributing to uncertainty in TAFTS radiance spectra have been listed with particular emphasis on the effect of detector noise. The methodology behind the production of TAFTS spectra laid out here is used to obtain the spectral radiance observations presented in Chapters 4 and 5.
4. GROUND-BASED OBSERVATIONS OF ARCTIC CIRRUS DURING RHUBC

In this chapter, the deployment of TAFTS (Tropospheric Airborne Fourier Transform Spectrometer) during the Radiative Heating in Under-explored Bands Campaign (RHUBC) is described. The author, along with P.D. Green of Imperial College London, was directly involved in maintaining and operating TAFTS during RHUBC, and has been responsible for processing, calibrating and analysing the TAFTS RHUBC data.

The chapter begins with an overview of the campaign, including descriptions of the other instrumentation involved. The focus here is on a single case study, concerning measurements made in the presence of a thin cirrus layer at 0420 UTC on 9th March 2007. The estimation of the cirrus layer properties and atmospheric profile for input into the radiative transfer model is outlined, and the resulting calculated spectrum is compared with both TAFTS and AERI-ER (Atmospherically Emitted Radiation Interferometer - Extended Range, see Section 4.2.1) observations.

4.1 The motivation for RHUBC

RHUBC took place in February and March 2007 at the Atmospheric Radiation Measurement programme (ARM) Climate Research Facility - North Slope of Alaska (ACRF-NSA) site near Barrow, Alaska. There were three main objectives to be addressed during RHUBC:

1. The performance of clear-sky radiative closure studies in order to reduce important uncertainties in the water vapour spectroscopy, particularly
the foreign-broadened water vapour continuum and water vapour absorption line parameters (see Section 2.1.2 for a description of the water vapour continuum).

2. The cross-calibration and validation of two state of the art far-infrared spectrometers; These are the AERI-ER, which is permanently based at the ACRF-NSA site, and the TAFTS (see Chapter 3), which was based at the site for the duration of the campaign only. This inter-comparison allows a higher confidence in the results obtained from both instruments.

3. The investigation of the radiative properties of sub-arctic cirrus. The combination of the AERI-ER and TAFTS allows simultaneous high-resolution measurements of arctic cirrus emission in both the far-IR and mid-IR. The additional instrumentation available at the ARM site provides an extensive array of auxiliary data for estimating cirrus properties.

![Fig. 4.1: LBLRTM calculations of down-welling spectral radiance performed using a US standard atmosphere and a sub-arctic winter atmosphere. The precipitable water vapour (PWV) for each case is given in centimetres.](image)

The campaign exploited the very dry atmospheric conditions prevalent during winter at high latitudes, which enable observations of the $17 \text{–} 100 \mu m$ wavelength region by ground based Fourier transform spectrometers such as
the TAFTS since the atmosphere becomes semi-transparent at these wave-
lengt hs during the arctic winter, as shown by the LBLRTM (Line-By-Line
Radiative Transfer Model) calculations in Figure 4.1. In most parts of the
globe, this wavelength band is almost opaque when observed from the ground
(as shown by the radiance spectrum calculated for a US standard atmosphere
in Figure 4.1), owing to the strong absorption by water vapour of far-infrared
radiation through the pure rotation band. However, its importance to radia-
tive cooling in the mid- to upper-troposphere is well known (Sinha & Harries,
1995), so it is necessary that detailed spectral measurements are made of this
wavelength range for the validation and improvement of radiative transfer
models. The relative transparency of the atmosphere in the far-infrared dur-
ding the dry Arctic winter makes the ACRF-NSA site ideal for such a study,
whilst the wide range of other instruments available helps to characterize
the atmospheric state used as input by radiative transfer models. Some of
these instruments are described in Section 4.2.2. As described previously
(Section 1.4), radiative transfer modelling efforts have also highlighted the
sensitivity of far-infrared radiation to cirrus microphysical properties (Baran,
2007; Yang et al., 2003), though to date very few spectral far-infrared obser-
vations of cirrus have been made.

4.2 Instrumentation involved in RHUBC

As mentioned previously, RHUBC took place at the ACRF-NSA climate
research facility near Barrow, Alaska. Since July 1997 a large number of
instruments at the facility have continuously provided data describing the
cloud and radiative processes characteristic of the high latitude climate. The
facility was set up to address the need for long term monitoring of the arctic
atmosphere, since the Arctic is particularly sensitive to changes in the
global climate. Most of the instruments are divided between the two trailer
platforms shown in Figure 4.2.

For the duration of RHUBC, TAFTS was deployed inside the trailer plat-
Fig. 4.2: Left and middle: The two main instrument platforms at the ACRF-NSA site. Right: Location of the ACRF-NSA (yellow circle, image © 2008 TerraMetrics)

Fig. 4.3: The steel cylinder connecting the TAFTS down-welling window to the outside atmosphere. The pump, visible in the right hand photograph, maintains a constant and uniform airflow throughout the cylinder and POB.

form shown on the left of Figure 4.2. A steel cylinder was installed to connect the TAFTS down-welling window to a hole in the roof of the trailer, in order to prevent contamination of the field of view from within the trailer (see Figure 4.3). A pump, visible in the right hand photograph of Figure 4.3, was attached to the cylinder to move air from the outside atmosphere through the TAFTS pointing optics box, and then up through the cylinder back to the outside atmosphere, ensuring that the air throughout the cylinder and POB was as similar as possible in temperature and water content to the air immediately outside the trailer platform. The TAFTS instrument and its op-
4. Ground-based observations of arctic cirrus during RHUBC

The following sections summarise the other instrumentation at the ACRF-NSA used during RHUBC.

4.2.1 The AERI-ER infrared spectrometer

In addition to the far-infrared radiances measured by TAFTS, the infrared radiance spectrum was also sampled by the AERI-ER (Knuteson et al., 2004a). This instrument measures a spectral range of $400 - 3300 \text{ cm}^{-1}$ at $1.0 \text{ cm}^{-1}$ spectral resolution with an accuracy better than $\pm 1\%$, and is located in the main trailer platform at the ACRF-NSA (middle image in Figure 4.2). The radiometric calibration methodology used is described in detail by Knuteson et al. (2004b). Similarly to the method used for TAFTS data, it involves alternating the input between views of the scene being calibrated and the two onboard reference blackbodies, which are kept at ambient temperature and $\sim 333 \text{ K}$ respectively. The final stage of AERI-ER data processing involves applying a noise filter, which utilises principal component analysis to identify the noise in the radiance spectrum and correct for it (Turner et al., 2006).

![Infrared radiance spectra measured by the TAFTS channels 0 and 2 and the AERI-ER channel 1 at 0420 UTC on 9th March 2007. Blackbody curves corresponding to the TAFTS instrument temperature, the outside air temperature and the estimated stratospheric temperature are overplotted for comparison (see text for full description).](image-url)

**Fig. 4.4:** Infrared radiance spectra measured by the TAFTS channels 0 and 2 and the AERI-ER channel 1 at 0420 UTC on 9th March 2007. Blackbody curves corresponding to the TAFTS instrument temperature, the outside air temperature and the estimated stratospheric temperature are overplotted for comparison (see text for full description).
The AERI-ER measurements of the infrared ‘window’ region (between 800 and 1200 cm\(^{-1}\)), which is particularly sensitive to cirrus properties (Yang et al., 2001; Bantges et al., 1999), are useful for testing the consistency of radiative transfer model results across the entire infrared spectrum from the far-infrared through to the mid-infrared. The overlap of the AERI-ER spectral range with that measured by TAFTS also provides a valuable opportunity for instrument intercomparison between 400 and 600 cm\(^{-1}\). Figure 4.4, which shows coincident TAFTS and AERI-ER spectra for a thin cirrus scene during RHUBC, illustrates both the wavenumber region in which the TAFTS and AERI-ER overlap and the mid-IR window region measured solely by the AERI-ER. The three Planck curves at 275.3 K, 245.9 K and 228.0 K represent the TAFTS instrument temperature, the outside air temperature at ground level and the stratospheric temperature (as inferred from the ozone band centered at 9.6 \(\mu\)m) respectively. Spectral radiances measured by the TAFTS and AERI-ER in the wavenumber region where the spectral ranges of the two spectrometers overlap are compared in Section 4.5.

4.2.2 Other instrumentation available at the ACRF-NSA

A number of other instruments were used during RHUBC to characterize the atmospheric state coincident with the infrared spectral radiance measurements. This is a very important input for the radiative transfer modelling procedure as described in Chapter 2, since the uncertainty in the inferred FIR cirrus signature increases if the assumed atmospheric profile is incorrect. The temperature and humidity profiles such as those shown in Figures 4.6 and 4.18 were measured using Vaisala RS-92 radiosondes, which were launched twice a day as standard, and as often as once every two hours when the atmosphere was particularly dry.

The accuracy and performance of the RS-92 radiosonde have been assessed using a cryogenic frost point hygrometer (Miloshevich et al., 2006). This study concludes that the RS-92 has a mean percentage accuracy relative to the hygrometer of < 5% for most conditions in the lower troposphere, and
< 10% in the middle and upper troposphere. After corrections are applied for time lag and calibration errors (Miloshevich et al., 2004), these may be reduced to < 1%, < 2% and < 3% respectively. However, the time lag correction is not currently applied as standard to the ARM data archive, so the first set of mean percentage accuracies should be assumed when using the RS-92 data.

Since radiosondes only provide information on the atmospheric water content intermittently, it is useful to remotely sense the precipitable water vapour (PWV) between sonde launches. This can be achieved using a microwave radiometer (MWR, Liljegren (2000)), which measures microwave radiation at 23.8 and 31.4 GHz. The two channels are sensitive to water vapour and liquid water respectively. The PWV is obtained from the brightness temperature (BT) measurements through a statistical retrieval method (Westwater, 1993). It is common practice at the ACRF-SGP (Southern Great Plains), a mid-latitude ARM site, to scale the radiosonde relative humidity profile such that the PWV obtained from the profile matches that retrieved from the MWR measurements. However, at low microwave brightness temperatures such as those measured in the very dry atmospheres encountered during the arctic winter, the MWR BT measurements show a positive bias (Westwater, personal communication). This leads to a systematic overestimate of the PWV (of approximately 0.3 mm) by the MWR compared with values derived from RS-92 RH profiles during RHUBC (see Figure 4.5). This overestimate is also evident in recent results comparing MWR retrievals with retrievals from two new microwave radiometers specially designed for ground-based monitoring of PWV in very dry atmospheres: the G-Band Vapour Radiometer (Pazmany, 2007) and the Ground-based Scanning Radiometer (Cimini et al., 2007b). The results from both of these instruments showed good agreement with radiosonde PWV data (Cadeddu et al., 2007; Cimini et al., 2007a) during previous observation campaigns at the ACRF-NSA. The RS-92 relative humidity profiles used here have therefore not been scaled to match the MWR retrievals of PWV.
Cloud properties over the ACRF-NSA may be inferred from lidar and radar measurements. The micropulse lidar, or MPL (Campbell et al., 2002), consists of a 532 nm Nd:YAG laser which emits pulses of light at a frequency of 2.5 kHz. These pulses are backscattered by small atmospheric particles such as aerosols and cloud droplets. The time between the light pulse being emitted and the backscattered light being detected by a Geiger mode avalanche photodiode indicates the distance to the scattering particles, whilst the amount of backscattered light depends on the bulk backscatter cross section. The MPL is particularly sensitive to smaller atmospheric particles because of their enhanced scattering at visible wavelengths. The cloud boundaries and the vertical profile of optical scattering cross section are routinely derived from MPL measurements.

The millimetre cloud radar, or MMCR (Moran et al., 1998), is the other main instrument used for active remote sensing of cloud properties at the ARM climate research facilities. The MMCR operates at a wavelength of 8.7 mm (frequency 35 GHz), which provides very good sensitivity to hydrometeors due to the inverse fourth-power dependence of echo intensity on wavelength $\lambda$ for particles with diameters small enough that the Rayleigh scattering approximation applies (diameter $D << \lambda$, Born & Wolf (1999)). Radar reflectivity, doppler velocity and spectral width are reported for altitudes up to $\sim 20$ km, at a vertical resolution of 45 m and a temporal
4. Ground-based observations of arctic cirrus during RHUBC

4.3 Measurements of the arctic atmosphere under clear sky conditions on 10\textsuperscript{th} March 2007

On 10\textsuperscript{th} March 2007, TAFTS was successfully operated for several hours under clear sky conditions. Two radiosondes were launched during this period at 1333 and 1547 UTC respectively. The atmosphere remained dry throughout, with the PWV increasing from 1.4 mm up to 1.64 mm as measured by the RS-92’s (see Figure 4.5), owing to an increase in humidity between 4.5 and 8.5 km above sea level observed in the sonde profiles shown in Figure 4.6.

![Fig. 4.6: Temperature (left panel) and relative humidity with respect to water (right panel) profiles measured by RS-92 radiosondes launched at 1333 UTC (solid curves) and 1537 UTC (dotted curves) on 10\textsuperscript{th} March 2007.](image-url)
Though the focus of this work is on cirrus, it is important to test the radiative transfer model under clear sky conditions. This helps to identify whether discrepancies between the modelled and observed radiance spectra are due to either the scattering model employed to represent the cirrus layer in the calculations, or the clear sky transmittances calculated with LBLRTM which are used as input by the Line-By-Line Discrete Ordinate Method scattering code (LBLDIS). The extent of the discrepancy between TAFTS observations and LBLRTM output is quantified by calculating the root-mean-square (RMS) of the residual between the observed and calculated spectra across the wavenumber range of the TAFTS shortwave channel. When making the comparisons, the sensitivity of the simulated clear sky spectrum to both PWV and the foreign broadened water vapour continuum strength is considered. The aim of this section is to determine by how much each of these quantities need to be altered to provide the best agreement with TAFTS and AERI-ER observations. The optimal scaling factors derived for each quantity and for each spectrometer are then applied when calculating the clear sky transmittances for use in the cirrus comparison described in Section 4.4.

Although the PWV and the foreign broadened water vapour continuum strength are considered to be the principle sources of uncertainty when attempting to simulate the observed clear sky spectral radiances, there are other possible sources of error. The LBLRTM utilises the HITRAN line parameter database (Rothman et al., 2005), which itself introduces uncertainty to the calculations since it is based on experimental measurements. The uncertainties in the far-infrared water vapour line parameters listed in HITRAN are discussed in Section 2.1.2. Another possibility is contamination of the observed scene by a thin cirrus layer, which would result in an increase of the spectral radiances within the absorption microwindows compared with an uncontaminated scene. The lidar measurements coincident with the clear sky observations do not indicate the presence of cloud. Lidar systems similar to the one used here have been used previously to detect subvisual cirrus with optical depths as low as $\sim 1 \times 10^{-3}$ (Sassen & Cho, 1992), so this
may be considered to be the upper limit in optical depth of any thin cirrus which may have contaminated the clear sky scenes presented throughout this chapter. The radiative effect of such a cloud would be well within the uncertainty levels of the two spectrometers used during the RHUBC campaign, so the possibility of the contamination of clear sky scenes by thin cirrus is not considered further during this work.

4.3.1 Sensitivity of LBLRTM output to PWV

The first quantity considered in this sensitivity study is the PWV. As mentioned in Section 4.2.2, the mean uncertainty of the RS-92 humidity measurements compared with frost-point hygrometer data has been estimated to be as high as 10% (Miloshevich et al., 2006) under certain conditions. This uncertainty in the water vapour profile, and consequently the PWV, contributes a related uncertainty to simulated clear sky transmittances and radiance spectra which requires quantification. The original LBLRTM spectrum calculated with the water vapour profile taken directly from the radiosonde launched at 1333 UTC on 10th March 2007 and the MT_CKD v2.1 water vapour continuum formulation is shown in Figure 4.7, along with the TAFTS observation recorded between 1401 and 1407 UTC.

The residual (defined as the simulated minus the observed spectral radiance) is positive across almost the whole wavenumber range of the TAFTS shortwave channel, with an RMS of 3.94 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$. The largest differences (residual greater than 5 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$) are observed in the microwindows between the water vapour absorption lines. This suggests that the atmosphere represented in the LBLRTM calculation is less transparent at these wavenumbers than that observed by TAFTS. Since water vapour is the primary absorbing medium here, a reduction in the PWV assumed in the radiative transfer model would result in a more transparent atmosphere in the FIR and improved agreement with the TAFTS measured spectrum. The sensitivity of the simulated radiance spectrum to PWV is tested by applying a scaling factor to the whole RS-92 measured relative hu-
Fig. 4.7: Observed (black) and simulated (blue) TAFTS clear sky spectra from 10\textsuperscript{th} March 2007. The simulation uses the water vapour profile measured by the 1333 UTC radiosonde. The observed spectrum is an average of down-welling sky view spectra recorded between 1401 and 1407 UTC. The residual (simulation minus observation) is offset by -20 mW m\textsuperscript{−2} sr\textsuperscript{−1} (cm\textsuperscript{−1})\textsuperscript{−1} for clarity.

These results demonstrate the strong dependence of the FIR radiance spectrum of an arctic winter atmosphere viewed from the ground looking directly upwards on the amount of water vapour present. When water vapour is removed, the atmosphere becomes more transparent between the absorption lines. The majority of the radiance observed is therefore emitted by the atmosphere at higher altitudes, where the atmosphere is colder. Consequently, the radiances observed in the absorption microwindows are smaller for drier atmospheres. The effect is greatest in the microwindows centred at 365 cm\textsuperscript{−1}, 391 cm\textsuperscript{−1} and 409 cm\textsuperscript{−1}, where a 40\% reduction in PWV results in decreases in spectral radiance of up to 15 mW m\textsuperscript{−2} sr\textsuperscript{−1} (cm\textsuperscript{−1})\textsuperscript{−1}. The impact of removing water vapour from the atmosphere on microwindows at shorter wavelengths is smaller, but still significant. The RMS of the residual
Fig. 4.8: Observed (black) TAFTS clear sky spectrum from 10th March 2007, along with LBLRTM - TAFTS radiance residuals (each offset by -20 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$ from the previous residual) calculated for a selection of water vapour scaling factors (as labelled on the right hand side of the figure). The LBLRTM spectrum calculated with the PWV scaled by a factor of 0.88 (the scaling factor giving the smallest RMS residual) is overplotted in blue.

was calculated for a range of scaling factors, and it was found that the smallest RMS value was obtained by using 0.88 as the scaling factor. Reducing the column amount of water vapour by 12% in this manner improves the RMS residual from 3.94 down to 3.09 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$. The RMS values obtained from the selection of PWV scaling factors used in Figure 4.8 are listed in Table 4.1.
4.3.2 Sensitivity of LBLRTM output to the foreign broadened water vapor continuum

Previous observational studies have highlighted the uncertainties in the water vapor continuum model used in radiative transfer calculations at far-infrared wavelengths (Tobin et al., 1999; Serio et al., 2008b). The continuum absorption in the Earth’s lower atmosphere at wavelengths near to the centres of water vapor absorption bands is dominated by the foreign broadened component, whereas the self broadened component dominates in extended regions of low absorption (Clough et al., 1989). We therefore look solely at the sensitivity of the far-infrared radiance spectrum to the foreign broadened continuum, since there is a high abundance of water vapor rotational absorption lines across this part of the spectrum.

Figure 4.9 illustrates the effect of reducing the foreign broadened continuum strength on the radiance calculations. In the absorption microwindows, the calculated spectrum does not demonstrate the same level of sensitivity to this parameter as it does to the column water vapor amount, though the impact of adjusting the continuum strength is not insignificant. The difference introduced by adjusting the continuum strength is also more wavelength dependent than the effect of adjusting the PWV. The greatest difference is ob-

<table>
<thead>
<tr>
<th>PWV scaling factor</th>
<th>RMS residual (mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.00</td>
<td>3.94</td>
</tr>
<tr>
<td>0.90</td>
<td>3.13</td>
</tr>
<tr>
<td><strong>0.88</strong></td>
<td><strong>3.09</strong></td>
</tr>
<tr>
<td>0.80</td>
<td>3.53</td>
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<tr>
<td>0.70</td>
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</tr>
<tr>
<td>0.60</td>
<td>7.41</td>
</tr>
<tr>
<td>0.50</td>
<td>10.23</td>
</tr>
</tbody>
</table>

**Tab. 4.1:** RMS LBLRTM - TAFTS residuals for a selection of PWV scaling factors. The scaling factor giving the smallest RMS value is highlighted in bold.
Fig. 4.9: Observed (black) TAFTS clear sky spectrum from 10\textsuperscript{th} March 2007, along with LBLRTM - TAFTS radiance residuals calculated for a range of foreign broadened continuum scaling factors (as labelled on the right hand side of the figure). The LBLRTM spectrum calculated with the continuum scaled by a factor of 0.64 (the scaling factor giving the smallest RMS residual) is overplotted in blue.

erved in microwindows at smaller wavenumbers (less than 420 cm\textsuperscript{-1}), since the foreign broadened continuum increases in strength (and consequently its contribution to the observed spectral radiances is greater) from high to low wavenumbers by approximately an order of magnitude across the TAFTS shortwave channel.

In a similar way to the study of the effect of PWV in Section 4.3.1, the RMS radiance residuals are calculated for a number of different continuum scaling factors (a selection of these are listed in Table 4.2). The smallest RMS residual is obtained using a scaling factor of 0.64, which reduces the RMS value from 3.94 to 3.01 mW m\textsuperscript{-2} sr\textsuperscript{-1} (cm\textsuperscript{-1})\textsuperscript{-1}. This suggests that if the
4. Ground-based observations of arctic cirrus during RHUBC

A scaling factor giving the smallest RMS value is highlighted in bold.

<table>
<thead>
<tr>
<th>$C_{fgn}$ scaling factor</th>
<th>RMS residual (mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.00</td>
<td>3.94</td>
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<td>3.19</td>
</tr>
</tbody>
</table>

**Tab. 4.2:** RMS LBLRTM - TAFTS residuals for a selection of foreign broadened water vapour continuum scaling factors. The scaling factor giving the smallest RMS value is highlighted in bold.

 radiosonde measured PWV is assumed to be correct, a significant reduction in the water vapour foreign broadened continuum strength is required to provide the best agreement with the TAFTS observation. However, it is more likely that the differences between the observed and calculated radiances are due to a combination of errors in both the assumed PWV and the continuum strength. This possibility is investigated in Section 4.3.3.

4.3.3 Sensitivity of LBLRTM - TAFTS residual to variations in both PWV and $C_{fgn}$

In this section the combined effect of the PWV and the foreign broadened water vapour continuum strength on the far-infrared radiance spectrum is investigated. 2601 different radiance spectra are calculated for combinations of 51 PWV scaling factors and 51 $C_{fgn}$ scaling factors, each spanning the range 0.50 to 1.00 with an interval of 0.01. Each of these calculated spectra is compared with the TAFTS observation, and the RMS radiance residual is obtained.

Unlike the RMS residuals calculated in Sections 4.3.1 and 4.3.2 which use all wavenumber data points between 350 and 600 cm$^{-1}$, these RMS values are obtained using only data points lying within the absorption microwindows
4. Ground-based observations of arctic cirrus during RHUBC

Fig. 4.10: Positions of absorption microwindows (blue crosses) used to optimise PWV and foreign broadened continuum scaling factors for the comparison of TAFTS with simulated radiances. The microwindows labelled with green crosses are not used in the analysis.

(see Figure 4.10). The microwindows were chosen for this analysis since these spectral regions are the most sensitive to changes in PWV and $C_{fgn}$.

For this particular case using the 1401 UTC TAFTS observation, only microwindows at wavenumbers less than 525 cm$^{-1}$ are considered. This is because the signal-to-noise ratio in the microwindows at wavenumbers greater than 525 cm$^{-1}$ is very low ($< 20$, see Figure 3.14), so they are omitted from the analysis to provide a more reliable fit. By plotting the RMS values as a function of the two scaling factors, it is possible to estimate the optimal combination of PWV and $C_{fgn}$ scaling factors which minimise the differences between the observed and calculated spectra. The resulting contour plot is shown in Figure 4.11.

The minimum RMS residual (1.92 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$) is indicated on the contour plot by the cross, and is obtained using scaling factors of 0.95 for PWV and 0.74 for the foreign broadened water vapour continuum strength. The lowest contour (the area of which comprises all RMS values between 1.92 and 2 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$) covers a wide range of both scaling factors, though its narrow shape suggests that the optimal pair of scaling factors
Fig. 4.11: Contour plot of RMS LBLRTM - TAFTS residuals (calculated in the microwindow regions only) as a function of PWV and \( C_{fgn} \) scaling factors. The cross indicates the location of the minimum RMS value. The blue contour shows the region where the difference from the minimum RMS value is less than the estimated TAFTS uncertainty.

can at least be constrained to a single line in scaling factor space. The radiance spectrum calculated using the two scaling factors which minimise the RMS value is plotted in Figure 4.12, alongside the 1401 UTC TAFTS observed spectrum. It should be borne in mind that these scaling factors merely represent a ‘best estimate’ of the conditions observed by the TAFTS at the time of the measurements. Given that the uncertainty in the TAFTS observations during RHUBC was approximately 1 mW m\(^{-2}\) sr\(^{-1}\) (cm\(^{-1}\))\(^{-1}\)
(see Section 3.3.3), it is possible that the ‘true’ scaling factors representative of the observed atmosphere could in fact lie anywhere within the blue contour shown in Figure 4.11. The scaling applied to the PWV is within the TAFTS uncertainty, though the scaling factor used for $C_{fgn}$ reduces it further than can be explained by uncertainty in the TAFTS spectral radiances alone.

4.3.4 Sensitivity of LBLRTM - AERI-ER residual to variations in both PWV and $C_{fgn}$

Here a similar analysis is performed comparing LBLRTM output calculated using the 1333 UTC radiosonde profiles with the 1401 UTC AERI-ER measurement. The initial LBLRTM spectrum calculated with unscaled PWV and $C_{fgn}$, and convolved with the AERI-ER response function, is plotted in Figure 4.13 along with the AERI-ER observation.

Unlike the TAFTS measurement taken at the same time (see Figure 4.7), the AERI-ER spectrum has higher radiances in the absorption microwindows than the initial unperturbed LBLRTM spectrum, with the exception of the window centred at 409 cm$^{-1}$. It is therefore anticipated that an increase in PWV and/or $C_{fgn}$ is necessary to provide the best agreement.
between the calculated and measured spectra, since greater absorption by water vapour (either through greater column amount or greater continuum absorption strength) means that the radiation observed in the microwindows reaching the surface will have been emitted by a lower (warmer) atmospheric layer, and will consequently have higher energy resulting in greater radiances observed in the absorption microwindows.

The positive residual in the microwindow at 409 cm\(^{-1}\) poses a problem, since an adjustment to the PWV or the continuum strength to correct for it results in an increase in the negative residual observed in the other microwindows. Conversely, adjusting to correct for the negative residuals in the other microwindows leads to an increase in the positive residual at 409 cm\(^{-1}\). The AERI-ER signal-to-noise ratio decreases rapidly with decreasing wavenumber below approximately 420 cm\(^{-1}\), close to the limit of the detector response (Knuteson et al., 2004b), so for the purposes of this analysis the 409 cm\(^{-1}\) microwindow is discarded.

The contour plot of the RMS values obtained using the AERI-ER measured radiance spectrum is shown in Figure 4.15. Note the different horizontal scale compared with that used in Figure 4.11, since an increase in PWV
is required here as discussed previously in this section. The minimum RMS residual in the absorption microwindows is obtained using scaling factors of 1.22 for PWV and 0.85 for $C_{fgn}$. The radiance spectrum obtained using the two scaling factors minimising the RMS value in the microwindows is plotted in Figure 4.16, alongside the 1401 UTC AERI-ER observed spectrum. As was the case with the TAFTS results shown in Section 4.3.3, these scaling factors only represent an estimate of the conditions which best represent the spectra observed by the AERI-ER. The uncertainty in the noise filtered AERI-ER observations is approximately $0.16 \text{ mW m}^{-2} \text{ sr}^{-1} (\text{cm}^{-1})^{-1}$ (Turner et al., 2006), and is illustrated by the blue contour in Figure 4.15. The changes in both PWV and $C_{fgn}$ applied here are both greater than can be explained by uncertainty in the AERI-ER spectral radiances alone.

The reduction of the foreign broadened continuum strength required in both the TAFTS and AERI-ER cases gives some confidence in the results obtained using this method. However, the perturbations in PWV required to optimise the modelled spectrum to observations is, in both cases, greater than the uncertainty in the PWV measurement obtained using the RS-92 ra-
This section has addressed the sensitivity of clear sky spectral radiances
diosonde (see Section 4.2.2). This raises the possibility of calibration errors
in both spectrometers, with the TAFTS and AERI-ER measured spectral
radiances in the absorption microwindows being too low and too high re-
respectively. See Section 4.5 for a more detailed inter-comparison between the
two instruments.

**Fig. 4.15:** Contour plot of RMS LBLRTM - AERI-ER residuals (calculated in
the microwindow regions only) as a function of PWV and $C_{fgn}$ scaling
factors. The cross indicates the location of the minimum RMS value. The blue contour shows the region where the difference from
the minimum RMS value is less than the estimated AERI-ER uncertain-

This section has addressed the sensitivity of clear sky spectral radiances
calculated using the LBLRTM code to water vapour amount and continuum absorption strength. It is likely that differences between clear sky spectra observed by TAFTS and calculated with LBLRTM are due to a combination of clear sky model and measurement errors. By attempting to determine an optimal pair of scaling factors for the PWV and the foreign broadened continuum strength, the contribution of clear sky model errors to these differences should be minimised. A more sophisticated treatment of the continuum strength would take into account its variation with wavenumber, and attempt to introduce a wavenumber dependent correction to $C_{fgn}$. However, given the purpose of this study is to provide a basis for the investigation of cirrus observations, the approach described in this section is considered to be sufficiently adequate. The minimisation of clear sky model errors in this way enables a simpler treatment of the cirrus case study presented in Section 4.4, since it becomes possible to then attribute further increases in the RMS residuals to problems with the assumptions made in formulating the cirrus model.
4.4 Measurements and modelling of arctic cirrus observed over the NSA-ACRF on 9th March 2007

This section describes the observation of a thin cirrus layer on 9th March 2007 by a number of instruments (including both TAFTS and AERI-ER) located at the NSA-ACRF during RHUBC. The method for remotely estimating the cirrus microphysics from the observations available is also outlined here. The downwelling spectral radiances measured during this cirrus event by TAFTS and AERI-ER are then compared with the radiance spectra simulated using the LBLDIS code.

![Graph](image)

**Fig. 4.17:** MWR retrieval and RS-92 measurements of PWV above the ACRF-NSA on 9th March 2007. Note that the y-axis begins at 1.0 mm.

The TAFTS and AERI-ER spectrometers were both successfully operational between 0420 and 0430 UTC on 9th March 2007. The nearest radiosonde launch to this time was at 0519 UTC (see Figure 4.18). Unfortunately the nearest launch before the radiance observations was at 1726 UTC on the previous day, so the 0519 UTC measured profiles are the closest approximation available to the atmospheric state at the time of the radiance measurements. The MWR retrieved PWV however was constant to within 0.1 mm between 0400 and 0600 GMT (see Figure 4.17), suggesting that the total amount of water vapour in the atmosphere remained roughly constant between the times of the radiance observations and the radiosonde flight.

The radiosonde launched at 0757 GMT measured temperature and rela-
4. Ground-based observations of arctic cirrus during RHUBC

Fig. 4.18: Temperature (left panel) and relative humidity with respect to water (right panel) profiles measured by RS-92 radiosondes launched at 0519 UTC (solid curves) and 0757 UTC (dotted curves) on 9th March 2007.

Relative humidity profiles (shown by the dotted curves in Figure 4.18) with very similar vertical structures to those observed by the 0519 UTC radiosonde, suggesting that the vertical distributions of water vapour and temperature were also consistent during this period. This justifies the use of the 0519 UTC radiosonde temperature and relative humidity profiles as input for LBLRTM when simulating the radiances observed by TAFTS and AERI-ER at 0420 UTC.

The MPL and MMCR observations (Figure 4.19) indicate the presence of a cirrus layer with a cloud base height of 4 km. The cloud is approximately 2 km thick until just before 0400 UTC, after which it gradually becomes thinner before dissipating completely by 0500 UTC. The radar reflectivities measured between 0420 and 0430 UTC are averaged over the 10 minute
4. Ground-based observations of arctic cirrus during RHUBC

Fig. 4.19: MMCR (left panel) and MPL (right panel) observations of cirrus over the Barrow ARM site on 9th March 2007.

period, during which the peak radar reflectivity observed is -32.9 dBZ at altitude 4.88 km.

The algorithm of Field et al. (2005), outlined in Section 4.4.1, is used to produce an estimate of the ice particle size distribution (PSD). The size distribution is then used to calculate the bulk scattering properties of the cirrus layer needed by the LBLDIS code to compute the spectral radiance. These bulk properties are obtained by integrating the single scattering properties described in Section 2.2.3 over the estimated PSD, as shown in Section 4.4.2. Mid-infrared radiance spectra computed using a range of different IWC values are then compared with the AERI-ER observation of the ice cloud to estimate the cloud IWC for this case, and the estimated PSD is scaled to match this IWC before spectra are calculated to compare with observations in Section 4.4.3.

4.4.1 Estimation of the ice crystal size distribution from ice water content and cloud temperature

This section describes how a particle size distribution for the cirrus layer is estimated. Owing to the absence of in-situ measurements of the cirrus microphysics during RHUBC, this must be derived from a parameterization.
For this work, the method of Field et al. (2005) is used to obtain size distributions. In their work, they show that a wide range of measured ice PSD data taken in-situ during flights in cloud associated with mid-latitude frontal systems can be rescaled to a single underlying ‘universal’ size distribution. This universal distribution may then be used to recover any measured PSD, given the knowledge of two moments of the distribution. They then reduce the number of moments required to predict the PSD from two to one by developing temperature-dependent power laws from the in-situ measurements relating pairs of moments. The definition used for the PSD moment of order $n$ is given in Equation 4.1.

$$\mathcal{M}_n = \int_0^\infty D^n N(D) \, dD \approx \sum_{D=100 \mu m}^{D=4400 \mu m} D^n N_D$$  \hspace{1cm} (4.1)

Here, $D$ is crystal size parallel to direction of flight, $N(D) \, dD$ is the concentration of crystals with sizes between $D$ and $D + dD$, and $N_D$ is the crystal concentration in the size-bin centred on size $D$. Note the upper and lower limits of the sum used to calculate the moments: a lower limit of 100 $\mu$m is used since these small particles are poorly sampled by the 2D-C probe used to measure the PSDs, whilst the upper limit of 4400 $\mu$m avoids the problem of spurious counts in large size-bins caused by low sample rates. The scaling function relating the universal distribution $\phi_{i,j}$ to the observed PSD for any pair of moments $i$ and $j$ of the PSD is given by Equations 4.2 and 4.3.

$$N(D) = \mathcal{M}_i^{(j+1)/(j-i)} \mathcal{M}_j^{(i+1)/(i-j)} \phi_{i,j}(x)$$  \hspace{1cm} (4.2)

$$x = D \left( \frac{\mathcal{M}_i}{\mathcal{M}_j} \right)^{1/(j-i)}$$  \hspace{1cm} (4.3)

The second moment $\mathcal{M}_2$ is used as the reference moment, since previous studies have demonstrated that the aggregate ice crystal mass is proportional to the square of the particle size (Heymsfield et al., 2004; Brown & Francis, 1995). It may therefore be assumed that $\mathcal{M}_2$ is proportional to the IWC. Field et al. derive a power law relationship (Equation 4.4) between $\mathcal{M}_2$ and
\( \mathcal{M}_3 \) by calculating the two moments for each of \( \sim 9000 \) measured PSDs and fitting a two-dimensional polynomial to the power law exponents and coefficients as a function of cloud temperature (Equations 4.5 and 4.6, \( T_c \) in °C).

\[
\mathcal{M}_3 = a(T_c)\mathcal{M}_2^{b(T_c)} \tag{4.4}
\]

\[
\log_{10}(a(T_c)) = -1.650 + 0.0545T_c + 0.000327T_c^2 \tag{4.5}
\]

\[
b(T_c) = 1.421 + 0.0119T_c + 0.0000960T_c^2 \tag{4.6}
\]

The final task is to derive the universal distribution for this pair of PSD moments, \( \phi_{2,3}(x) \). This is achieved by rescaling all \( \sim 9000 \) measured PSDs, by using moments \( \mathcal{M}_2 \) and \( \mathcal{M}_3 \) for each size distribution in Equation 4.2 to obtain a universal distribution for each measured PSD. A combination of exponential and gamma distributions are then fitted to provide the single universal distribution given in Equation 4.7.

\[
\phi_{2,3}(x) = \kappa_0 e^{-\Lambda_0 x} + \kappa_1 x^\nu e^{-\Lambda_1 x} \tag{4.7}
\]

\( \kappa_0 = 490.6, \kappa_1 = 17.46, \Lambda_0 = 20.78, \Lambda_1 = 3.290, \nu = 0.6357 \)

This universal function is now used in Equation 4.2 along with the second and third PSD moments to estimate the size distribution of the ice crystals in the cloud observed at 0420 UTC on 9th March 2007. The mass-dimension relation \( m = 0.0185D^2 \) (Brown & Francis, 1995), where mass is in kilograms and dimension in metres, is used to obtain \( \mathcal{M}_2 \) from the IWC. The cloud temperature is then used along with \( \mathcal{M}_2 \) in Equation 4.4 to obtain \( \mathcal{M}_3 \). Finally, these values of \( \mathcal{M}_2 \) and \( \mathcal{M}_3 \) are used in Equations 4.2 and 4.3 to rescale the universal distribution to the estimated ice PSD for the cloud. For example, a typical IWC of \( 1.0 \times 10^{-2} \) gm\(^{-3} \) and cloud temperature -36.6 °C
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Fig. 4.20: Universal distribution (upper panel) and predicted ice crystal size distribution (lower panel) for 0420 UTC derived using the method described by Field et al. The dashed curves indicate the contributions of the small and large crystal modes.

gives the second and third PSD moments $5.41 \times 10^{-4}$ m$^{-1}$ and $1.40 \times 10^{-7}$ respectively. The universal distribution and predicted PSD for these values are shown in Figure 4.20.

Size distributions are often characterized by the effective dimension, $D_e$. This quantity is determined by taking the ratio of the volume-integrated and cross section-integrated size distributions, as shown in Equation 4.8. $V(D)$ is the crystal volume as a function of maximum dimension $D$, and $A(D)$ is the cross-sectional area.

$$D_e = \frac{3}{2} \frac{\int_0^\infty V(D) N(D) \, dD}{\int_0^\infty A(D) N(D) \, dD} \approx \frac{3}{2} \frac{\sum_{D=4400 \mu m} V(D) N_D}{\sum_{D=1 \mu m} A(D) N_D}$$ (4.8)

For the size distribution derived in this section, the effective dimension is
105.1 µm. The predicted PSD is now used in Section 4.4.2 to calculate the bulk scattering properties of the ice crystal layer.

4.4.2 Calculation of the cirrus bulk scattering properties

The bulk scattering properties calculated using the predicted PSD derived in Section 4.4.1 in conjunction with the ice crystal single scattering property database described in Section 2.2.3 are plotted in Figure 4.21. The mid-infrared wavenumber range shown in the lower panel is observed by the AERI-ER, and is particularly useful for studying the infrared radiative impact of cirrus since there is very little absorption or emission at these energies by water vapour or any other gaseous species.

Sensitivity of model radiance spectra to cirrus microphysics

Previous studies using radiative transfer calculations (Baran, 2007; Yang et al., 2003) have demonstrated that far-infrared spectral radiances in the presence of an ice cloud are sensitive to the shape of the PSD (as characterized by the effective dimension). The effect of varying the PSD effective dimension is tested here by adjusting the fraction of IWC accounted for by the ‘small mode’ of the PSD, whilst keeping the total IWC fixed. In this study the small mode is defined as the contribution of the exponential term to the PSD (the first term of Equation 4.7), whilst the ‘large mode’ is the contribution of the gamma distribution. Figure 4.22 shows the 11 different PSDs tested, including the original unadjusted PSD given here by the black curve.

The effect of adjusting the PSD in this manner on the FIR bulk scattering properties is shown in the upper panel of Figure 4.23. Increasing the small crystal IWC fraction reduces the optical depth for wavenumbers less than 500 cm$^{-1}$, owing to the reduced number of large crystals which provide a stronger contribution to the optical extinction. The scattering albedo is increased uniformly across the FIR due to the reduction of the extinction coefficient caused by decreasing the number of large crystals. The asymmetry parameter is insensitive to increases in the small IWC fraction, though it
increases significantly when this fraction is decreased since the mean particle size becomes much greater than the incident wavelength, such that the particle shape has a significant effect on the scattering phase function. The net effect of these changes to the bulk scattering properties is that the simulated radiances in the microwindows are reduced by increasing the small crystal IWC fraction, and vice versa (see lower panel of Figure 4.22).

The differences to the mid-IR bulk scattering properties and simulated radiances are also shown in Figure 4.24. The magnitude of the shift in these radiances is never greater than 1.5 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$, compared with
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Fig. 4.22: PSDs obtained by varying the fraction of IWC accounted for by the small crystal mode. The curves are coloured green when the small crystal IWC fraction is reduced, and blue when it is increased. The bold curves indicate the two extremes (100% large crystals and 100% small crystals). The effective diameter in microns (as defined in Equation 4.8) for each PSD is listed on the right hand side.

A maximum change of 5 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$ in the far-infrared. This suggests that far-infrared radiances are more sensitive to PSD shape than those observed in the mid-infrared window. However, the lack of atmospheric gases which absorb or emit radiation at mid-infrared window wavelengths means that mid-IR radiances are still very useful for remotely estimating cirrus properties, whilst the uncertainty in the water vapour foreign broadened continuum strength at far-infrared wavelengths makes the study of ice clouds with FIR radiances more problematic. A further interesting observation to be made from the lower panel of Figure 4.24 is the insensitivity of radiances between 915 and 920 cm$^{-1}$ to changes in the PSD shape.

In addition to estimating the shape of the size distribution, it is also necessary to infer the ice water content from the measurements available. A number of empirical parameterizations relating radar reflectivity to IWC via power law relationships have been derived from in-situ measurements (Matrosov et al., 2003; Shupe et al., 2005; Sassen et al., 2002), whilst more sophisticated algorithms include cloud temperature (Hogan et al., 2006; Boudala
Fig. 4.23: Upper three panels: Difference to the FIR bulk scattering properties achieved by varying the small crystal IWC fraction. Lowest panel: Difference to the FIR radiance spectrum achieved by varying the small crystal IWC fraction (Same colour scheme as Figure 4.22).

et al., 2006) or infrared radiance (Mace et al., 1998) as additional inputs. A comprehensive review of a variety of IWC retrieval techniques using different
Fig. 4.24: Upper three panels: Difference to the mid-IR bulk scattering properties achieved by varying the small crystal IWC fraction. Lowest panel: Difference to the mid-IR radiance spectrum achieved by varying the small crystal IWC fraction (same colour scheme as Figure 4.22).

ground based sensors is given by Comstock et al. (2007). However, the radar based methods are deemed unsuitable for this particular case, owing to the
low peak reflectivity observed falling outside the range of reflectivities used to empirically derive the Z-IWC relations. The dependence of radar reflectivity on the sixth moment of the PSD also means that radar is insensitive to the contribution of smaller crystals to the IWC (which is itself proportional to the second moment (Sassen et al., 2002)).

For this case, the AERI-ER measurements of the mid-infrared window are used to estimate the IWC. Across most of this window, there is very little absorption or emission by the atmosphere, such that the observed spectral radiances are solely due to the presence of ice crystals in the interferometer field of view. The LBLDIS model is used to calculate the spectral radiance for a small wavenumber range within the mid-infrared window, assuming the Field et al. size distribution shape described in Section 4.4.1, for a range of different IWC values. Wavenumbers between 915 and 920 cm\(^{-1}\) are used since radiances at these energies are insensitive to perturbations in the shape of the size distribution, as demonstrated by the sensitivity study (see lower panel of Figure 4.24). The IWC which gave the best agreement with the AERI-ER observation of the mid-IR window at 0420 – 0430 UTC was 1.19 \(\times\) 10\(^{-2}\) gm\(^{-3}\). This value is used for all LBLDIS calculated spectra shown in Section 4.4.3.

4.4.3 Comparison of LBLDIS output with spectra measured at 0420 – 0430 UTC on 9\(^{th}\) March 2007

This section shows comparisons of LBLDIS output with radiance spectra recorded by both TAFTS and AERI-ER whilst the ice cloud shown in Figure 4.19 was directly overhead. Figure 4.25 shows radiances observed by AERI-ER across the whole infrared range (from 6.67 to 25.00 \(\mu\)m).

The blue curve overplotted is the radiance spectrum calculated from the clear-sky transmittances output by LBLRTM, which shows the spectrum that would be observed if the cirrus layer was absent. Throughout this section, the clear-sky transmittances used as input by LBLDIS are calculated using the PWV and \(C_{fgn}\) scaling factors determined in Sections 4.3.3 and 4.3.4.

The cirrus layer has the effect of increasing the observed spectral radiance
Fig. 4.25: Comparison of AERI-ER (black) and clear-sky LBLRTM (blue) IR radiance spectra using PWV and $C_{fgn}$ scaling factors obtained in Section 4.3.4. The RMS value is calculated using only wavenumbers greater than 420 cm$^{-1}$. The lower panel shows the FIR (region ‘A’) in more detail.

across the mid-infrared window labelled ‘B’ in the figure (with the exception of the ozone band centred at 9.6 $\mu$m), and also in the far-infrared absorption microwindows between 16 and 25 $\mu$m (the spectral region labelled ‘A’). This difference between the observed cloudy and calculated clear-sky spectral radiances is known as the surface radiative forcing. Figure 4.26 and the lower panel of Figure 4.25 show the far-infrared (region ‘A’) in more detail, as observed by TAFTS and AERI-ER. Both spectrometers observe an increase in spectral radiance in the absorption microwindows which has a peak of around
Fig. 4.26: Comparison of TAFTS (black) and clear-sky LBLRTM (blue) FIR (region ‘A’ in Figure 4.25) radiance spectra using PWV and $C_{fgn}$ scaling factors obtained in Section 4.3.3. The RMS value is calculated using the whole spectral range shown.

10 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$ at $\sim$ 560 cm$^{-1}$, which becomes smaller in magnitude with decreasing wavenumber. The mid-IR window (labelled region ‘B’ in Figure 4.25) is observed by AERI-ER only. The surface radiative forcing in this region has a similar peak magnitude to that observed in the FIR. The gradient of this mid-IR forcing is sensitive to the PSD effective dimension (see lower panel of Figure 4.24), making it useful for retrievals of ice cloud properties (Huang et al., 2004).

Spectra calculated using LBLDIS are now compared with observations across the full infrared spectral range. The PSD derived in Section 4.4.1 is scaled up to give the estimated IWC of $1.19 \times 10^{-2}$ gm$^{-3}$, and used to calculate the bulk scattering properties as described in Section 4.4.2. The resulting spectra are compared with TAFTS and AERI-ER observations in Figures 4.27 and 4.28. The agreement with the AERI-ER observation of the mid-infrared window in Figure 4.27 is very good, with observed radiances agreeing with the calculation to within 1 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$ across the whole of this spectral range. However, both the TAFTS and AERI-ER observations of the FIR (lower panel of Figure 4.27 and Figure 4.28)
indicate that the simulated radiances are too high in the microwindows by approximately 3 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$.

The sensitivity study (Section 4.4.2) found the magnitude of reductions in the simulated radiances to be as high as 5 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$ at certain wavenumbers when all of the IWC is in the small mode. The differences obtained by adjusting the PSD in this way are therefore of the right order of magnitude to be able to account for the residuals observed when comparing the TAFTS and AERI-ER measurements with the model output. The microphysical and optical properties of the 11 PSDs tested are listed in Table 4.3,
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Fig. 4.28: Comparison of TAFTS (black) and LBLDIS (blue) FIR radiance spectra using the bulk scattering properties calculated in Section 4.4.2. The RMS value is calculated using the whole spectral range shown.

Fig. 4.29: Comparison of TAFTS (black) and LBLDIS (blue) FIR radiance spectra using the PSD giving best agreement in the FIR absorption windows. The RMS value is calculated using the whole spectral range shown.

where IWC$_{small}$ is the small crystal IWC fraction. The visible optical depth $\tau_{vis}$ is estimated for each PSD using the geometric optics asymptotic value of 2 for the extinction efficiency (Yang et al., 2001). The visible optical depth may therefore be obtained from the IR optical depth $\tau_{IR}$:
\[ \tau_{\text{vis}} = \left( \frac{2}{Q_e} \right) \tau_{\text{IR}} \] (4.9)

\( Q_e \) is the mean extinction efficiency for the PSD (see Section 2.2.3 for a definition). The visible extinction coefficient \( \beta_{e,\text{vis}} \) is then given by the visible optical depth divided by the cloud geometrical thickness in metres, which in this case is taken from the MPL observations (see Figure 4.19) to be 620 m. The visible optical depth is fairly insensitive to the effective dimension \( D_e \), demonstrating that it is primarily dependent on the total IWC rather than the size distribution shape.

The two columns on the right of Table 4.3 contain the RMS residuals calculated in the absorption microwindows when comparing the LBLDIS simulations to observations from each spectrometer. The AERI-ER RMS residuals are much more sensitive to the PSD shape than those calculated for TAFTS, since the changes in the FIR radiances shown in the lower panel

<table>
<thead>
<tr>
<th>IWC_{small}</th>
<th>( D_e ) (( \mu \text{m} ))</th>
<th>( \tau_{\text{vis}} )</th>
<th>( \beta_{e,\text{vis}} ) (m(^{-1}))</th>
<th>AERI-ER RMS (mW m(^{-2}) sr(^{-1}) (cm(^{-1}))(^{-1}))</th>
<th>TAFTS RMS</th>
</tr>
</thead>
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<tr>
<td>0.000</td>
<td>118.7</td>
<td>0.405</td>
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<td>3.28</td>
<td>3.07</td>
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<td>113.1</td>
<td>0.405</td>
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<td>3.08</td>
<td>3.01</td>
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<td>107.4</td>
<td>0.406</td>
<td>6.55 \times 10^{-4}</td>
<td>2.88</td>
<td>2.96</td>
</tr>
<tr>
<td>0.219</td>
<td>96.0</td>
<td>0.407</td>
<td>6.56 \times 10^{-4}</td>
<td>2.48</td>
<td>2.86</td>
</tr>
<tr>
<td>0.328</td>
<td>84.7</td>
<td>0.408</td>
<td>6.58 \times 10^{-4}</td>
<td>2.08</td>
<td>2.80</td>
</tr>
<tr>
<td>0.438</td>
<td>75.4</td>
<td>0.409</td>
<td>6.60 \times 10^{-4}</td>
<td>1.69</td>
<td>2.77</td>
</tr>
<tr>
<td>0.547</td>
<td>62.1</td>
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<td>6.62 \times 10^{-4}</td>
<td>1.30</td>
<td>2.78</td>
</tr>
<tr>
<td>0.656</td>
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<td>6.72 \times 10^{-4}</td>
<td>0.89</td>
<td>3.19</td>
</tr>
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**Tab. 4.3:** Microphysical and optical properties of the PSDs shown in Figure 4.22. RMS residuals are calculated from the absorption microwindows shown in Figures 4.10 and 4.14.
4. Ground-based observations of arctic cirrus during RHUBC

Fig. 4.30: Comparison of AERI-ER (black) and LBLDIS (blue) whole IR (upper panel) and FIR (lower panel) radiance spectra using the PSD giving best agreement in the FIR absorption microwindows. The RMS value is calculated using only wavenumbers greater than 420 cm\(^{-1}\).

of Figure 4.23 are mostly within the TAFTS instrument noise level for this campaign (see Section 3.3.3 for a discussion of the TAFTS instrument errors), whereas the noise-filtered AERI-ER data product does not have this problem. However, results from both spectrometers indicate that an increase in the small crystal IWC fraction is required to provide the best agreement between modelled and observed FIR radiances. The LBLDIS spectrum calculated using \(D_e = 75.4 \mu m\) is plotted alongside the TAFTS observation in Figure 4.29, though the range of RMS residuals obtained using all eleven different PSDs falls within the uncertainty in the TAFTS spectra. Figure 4.30
shows the spectrum calculated with $D_e = 28.5 \mu m$ plotted alongside the AERI-ER measurement. Only one of the other ten PSDs ($D_e = 39.7 \mu m$) produces an RMS residual within the estimated AERI-ER uncertainty. The comparison in the lower panel of Figure 4.30 shows that when using the $D_e = 28.5 \mu m$ PSD, LBLDIS is able to accurately simulate the AERI-ER observed radiances across the entire infrared spectral range. Note that the RMS residuals are still greater than the AERI-ER uncertainty even when the ‘best’ PSD shape is selected. This is likely to be due to errors in the spectroscopic line parameter database assumed during the radiative transfer calculations, as discussed in Section 2.1.2.

4.5 Comparison of TAFTS and AERI-ER observed spectral radiances

The deployment of TAFTS alongside the AERI-ER during RHUBC enabled observations taken by the two spectrometers to be compared with one another in the 400 – 600 cm$^{-1}$ region where the spectral ranges overlap. Before the measured radiances can be compared, the AERI-ER instrument response must be convolved with the TAFTS spectrum so that the spectral resolutions are matched. This is done by following these steps:

1. Perform a Fourier transform on the TAFTS spectrum to obtain the TAFTS interferogram in optical path difference space.

2. Multiply the interferogram by the Fourier transform of the AERI-ER instrument response function. This is a sinc function, so the Fourier transform of the response function is simply a rectangular function of width $2 \times 1.037$ cm, where 1.037 cm is the AERI-ER maximum optical path length (Knuteson et al., 2004a,b).

3. Perform the reverse Fourier transform on the apodised interferogram to return to wavenumber space, thus giving the TAFTS spectral radiances at AERI-ER spectral resolution.
Fig. 4.31: Comparison of TAFTS (black) and AERI-ER FIR spectral radiance observations of three scenes (from top panel to bottom panel); 0420 UTC 9th March 2007 (cirrus, AERI-ER radiances in red), 1401 UTC 10th March 2007 (clear sky, green), and 1605 UTC 10th March 2007 (clear sky, blue).

Figure 4.31 shows TAFTS and AERI-ER spectra (with the TAFTS spectra re-interpolated to the AERI-ER spectral resolution as described above) plotted for three different scenes observed during RHUBC. It can be seen from the residuals that the TAFTS measured radiances are consistently lower than the AERI-ER measured radiances by up to 5 mW m$^{-2}$ sr$^{-1}$ (cm$^{-1}$)$^{-1}$. The residuals are generally greater for the two clear sky scenes (which have lower radiances in the absorption microwindows) than they are for the cirrus scene. In Figure 4.32, the results from the three scenes are compared by plotting each wavenumber interval on the AERI-ER wavenumber grid on a scatter diagram as a function of the TAFTS spectral radiance (x-axis) and
the AERI-ER spectral radiance (y-axis) at that wavenumber.

**Fig. 4.32:** Scatter plot of TAFTS and AERI-ER observed radiances for the three scenes shown in Figure 4.31. The dotted line indicates where the two instruments observe equal radiances. The solid black line is the least-squares linear fit (coefficients listed in Table 4.4) to all of the data points.

A least-squares linear fit (solid black line in Figure 4.32) has been applied to all of the data to provide an estimate of the baseline offset $a_0$ and scaling factor $a_1$. The linear fit coefficients to the equation $S_{AERI-ER} = a_0 + a_1 S_{TAFTS}$ are listed in Table 4.4. $S_{AERI-ER}$ and $S_{TAFTS}$ are the spectral radiances observed by AERI-ER and TAFTS respectively.
In this case of two idealised identical instruments observing the same scenes, the values for the coefficients would be $a_0 = 0.00$ and $a_1 = 1.00$ and the linear fit would look like the dotted black line in Figure 4.32. However, when comparing TAFTS and AERI-ER there is a baseline offset of $4.48 \text{ mW m}^{-2} \text{ sr}^{-1} \text{ (cm}^{-1})^{-1}$ (i.e. when $S_{\text{TAFTS}}$ is zero, $S_{\text{AERI-ER}} = 4.48$). Given that the initial comparisons of clear-sky spectra with LBLRTM output indicated that the TAFTS measured radiances were too low and the AERI-ER radiances were too high (see Sections 4.3.3 and 4.3.4), it is likely that there are contributions from errors in the calibrations of both instruments to this offset. Also, the responses of the instruments to the input radiance are not quite equal, as indicated by the linear fit gradient $a_1 = 0.975 \neq 1.00$. For the scenes considered here the agreement between the two spectrometers is therefore best for higher radiances, such as those observed in the water vapour absorption line centres, since it is at these radiances that the linear fit and the idealised case are closest together.

<table>
<thead>
<tr>
<th>Coefficient</th>
<th>Result of least-squares fit</th>
<th>Error estimate</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a_0$</td>
<td>4.48</td>
<td>0.185</td>
</tr>
<tr>
<td>$a_1$</td>
<td>0.975</td>
<td>0.00391</td>
</tr>
</tbody>
</table>

**Tab. 4.4:** Coefficients from least-squares fit of the scatter-plot data in Figure 4.32 to the linear equation $S_{\text{AERI-ER}} = a_0 + a_1S_{\text{TAFTS}}$ with $S$ in units of $\text{mW m}^{-2} \text{ sr}^{-1} \text{ (cm}^{-1})^{-1}$.

### 4.6 Summary of RHUBC results

This chapter has described the deployment of TAFTS during RHUBC in February and March 2007. During RHUBC, high resolution far-infrared spectra covering wavenumbers between 400 and $600 \text{ cm}^{-1}$ of the Arctic atmosphere were measured simultaneously and independently by two different spectrometers (the TAFTS and the AERI-ER) for the first time, under both clear-sky and cloudy conditions. Clear sky observations were used along with
Ground-based observations of arctic cirrus during RHUBC

Radiosonde measurements to characterise the response of each instrument to the PWV and foreign broadened water vapour continuum absorption. In both cases a small reduction of the continuum strength was required to reconcile model output with observations, though the adjustments in PWV needed (which were opposite in sign for the two cases) were beyond the uncertainty in the radiosonde measurements. The uncertainties in the spectral observations made by each instrument were also such that the obtained scaling factors for PWV and $C_{fgn}$ may only be considered as estimates of the atmospheric conditions which best fit the observed spectra. The inter-comparison between the two instruments also revealed an offset with the AERI-ER typically measuring higher spectral radiances than the TAFTS. Considering both of these results, it appears that there are calibration errors in both spectrometers which contribute to this offset.

The optimal scaling factors for the PWV and continuum absorption estimated during the clear-sky sensitivity study in Sections 4.3.3 and 4.3.4 were applied to the modelled layer optical depths used by the scattering model, so that any errors in the modelled cirrus spectra could then be attributed with confidence to the cirrus model employed. In the absence of in-situ microphysical observations, the PSD parameterization of Field et al. (2005) is used along with a single scattering property database (Baran & Francis, 2004) to model the distribution of crystal sizes within the observed cloud layer. The sensitivity of far- and mid-infrared spectral radiances to the PSD shape is demonstrated, and utilised to show that in order for the model output to be consistent with the observations across the whole infrared spectrum, the fraction of small crystals contributing to the PSD needed to be increased. This was particularly evident from the AERI-ER results, which suggest that the effective diameter of the PSD required to best fit the observations was less than half of that predicted by the PSD parameterization. Whilst no conclusions are made about the quality of the single scattering property database or the assumption of aggregate shaped crystals, the results do indicate that observations of the far-infrared region of the electromagnetic spectrum (in
addition to observations of the mid-infrared) are potentially very useful for the testing of cirrus scattering models.
5. AIRCRAFT-BASED OBSERVATIONS OF TROPICAL CIRRUS DURING EMERALD-II

This chapter describes far-infrared spectral radiance observations of tropical anvil outflow cirrus taken during the Second Egrett Microphysics Experiment with Radiation, Lidar and Dynamics (EMERALD-II), which took place off the north coast of Australia during November and December 2002. J.E. Murray, P.D. Green and G.K. Straine of Imperial College London operated and maintained TAFTS during EMERALD-II, whilst J.E. Murray was primarily responsible for processing and calibrating the data. The author has taken a lead role in analysing the TAFTS cirrus data acquired during the campaign.

The motivation for the campaign and the instrumentation involved are outlined in Section 5.1, whilst Section 5.2 presents results from the 2nd December 2002 flight. The results are divided into two subsections: the first concerning a TAFTS clear sky observation, and the construction of water vapour and temperature profiles from in-situ measurements which successfully represent the observed atmosphere in radiative transfer calculations (Sections 5.2.2 and 5.2.3), and the second comparing a TAFTS cirrus observation with radiative transfer calculations incorporating a scattering layer with bulk scattering properties derived from cloud probe measurements (Sections 5.2.4 and 5.2.5). The results of this chapter are discussed in Section 5.3.

5.1 The EMERALD-II campaign

EMERALD was an airborne field campaign, based in Australia, designed to investigate the dynamical, microphysical and infrared radiative properties of cirrus clouds, using a combination of in-situ and remote observations.
EMERALD consisted of two phases: EMERALD-I, based in Adelaide during September 2001, which focused on mid-latitude frontal cirrus (Gallagher et al., 2005); and EMERALD-II, based in Darwin during November and December 2002, which looked at tropical convective outflow cirrus. The results discussed in this chapter are from the second phase only. Both phases utilised the same set of instruments, which are described briefly in Section 5.1.2, and in more detail by Cook (2004). Whiteway et al. (2004) provide a brief snapshot of results from both phases of the campaign.

5.1.1 The motivation for EMERALD-II

The EMERALD campaigns were designed to improve the realistic representation of cirrus ice crystal formation and evolution in climate prediction models through the study of the interactions between microphysics, dynamics and radiation (Whiteway et al., 2004), as it is understood that cirrus microphysics and radiative effects can play an important role in climate feedback mechanisms (Stephens et al., 1990). The impact of tropical convective cirrus on climate is of particular interest, since cirrus coverage above the inter-tropical convergence zone (ITCZ) due to convection can be as high as 90% (Wylie & Menzel, 1999).

Fig. 5.1: Photograph of the ‘Hector’ storm system taken from a distance of approximately 100 km.

The Darwin base chosen for EMERALD-II was ideal for studying convective cirrus, since an isolated system of thunderstorms named ‘Hector’
Aircraft-based observations of tropical cirrus during EMERALD-II (see Figure 5.1 for a photograph) occurs on an almost daily basis during the transition between the dry and wet seasons over the Tiwi Islands, located approximately 100 km to the north of Darwin. During the day, convective cells formed over the islands grow in height through convection, up to heights of around 17 km. These regularly produce outflow cirrus by late afternoon each day (see photograph in Figure 5.2) as moisture transported upwards through the middle of the storm cells freezes and aggregates at the low temperatures encountered at such high altitudes (colder than \(-60\)°C).

Fig. 5.2: Photograph of tropical convective outflow cirrus produced by the ‘Hector’ storm system.

5.1.2 EMERALD-II instrumentation

The EMERALD experimental set-up incorporates a range of instruments mounted on two aircraft which fly in tandem (see schematic in Figure 5.3); the Egrett (Grob G520T Egrett), which provided in-situ measurements of the cirrus, and the Kingair (Beech B200T Super King Air), which remotely mapped the cloud structure. The TAFTS was flown on-board the Egrett,
5. Aircraft-based observations of tropical cirrus during EMERALD-II

Fig. 5.3: Schematic representation of the EMERALD campaign concept. The Egrett flies at high altitude and samples the cirrus in-situ, whilst the Kingair flies directly below the Egrett and provides lidar observations of the same cirrus.

and operated using both zenith and nadir viewing geometries. Both aircraft were operated by Airborne Research Australia (ARA). The instrumentation on-board each aircraft is described here in turn.

Instrumentation on board the Egrett

Cloud Particle Imager

The Cloud Particle Imager (CPI, Lawson et al. (2001)) is an instrument which records high-definition images of cloud particles, enabling measurements of particle shape as well as particle size and number concentration. It works by casting an image of a particle onto a solid state CCD camera using a high-power 25 ns pulsed laser diode. A particle detection system using a pair of upstream lasers ensures that there is always at least one particle in the image in focus, by precisely defining the position of the focal plane. The CCD array consists of 2.3 μm pixels, enabling the imaging of particles between 10 μm and 2 mm in
diameter. The camera is operated at rates of up to 40 Hz, whilst it is possible to process more than a hundred particles per frame. Consequently it is possible to obtain data rates of at least 1000 imaged particles per second. When the data are processed, they are corrected for particle over-sizing owing to poor focus, probe triggering inefficiency for small particles and errors due to collections of particles within the depth-of-field (Connolly et al., 2007).

**Forward Scattering Spectrometer Probe**

The Forward Scattering Spectrometer Probe (FSSP, Knollenberg (1981); Baumgardner (1983)) determines cloud particle diameter by measuring the intensity of light forward-scattered by a particle passing through a focused He:Ne laser beam. It is able to detect and count particles with diameters between 3 and 47 µm. The measured intensity is related to particle dimension through Mie theory, which provides theoretical intensities as a function of particle dimension and refractive index.

Data obtained by the FSSP should be treated with caution, since the Mie theory used to obtain the particle dimension assumes spherical particles. The non-spherical nature of ice crystals therefore means that they may not be sized correctly. There has also been some debate regarding whether the measured small particle number concentrations are artificially enhanced through the shattering of larger ice crystals on the probe inlet (Field et al., 2003, 2006; Heymsfield, 2007), resulting in the total particle concentration being overestimated by up to a factor of 5.

**Frost Point Hygrometer**

The Frost Point Hygrometer (FPH, Busen & Buck (1995)) measures the frost point temperature through the ‘chilled mirror’ technique. A mirror is cooled cryogenically, and raised to the frost point temperature by a heating element. The heating element is driven by a feedback loop which responds to changes in the mirror reflectance, and holds
the mirror temperature at the point which maintains a constant layer of condensate on the mirror surface (and therefore a constant mirror reflectance). The frost point temperature is then measured by a temperature sensor embedded within the mirror.

The FPH used during EMERALD-II is sensitive to a frost point temperature range of $-90^\circ C$ to $+30^\circ C$, with response times typically between 6 and 30 seconds. The relatively long response times, which increase with decreasing humidity, introduce uncertainties in the frost point temperature measurements during ascents and descents during which the humidity may change rapidly. Sudden increases in humidity during descents can also result in the mirror being flooded with condensation, producing substantial errors until the excess condensation evaporates.

**Tunable Diode Laser**

The Tunable Diode Laser (TDL, May (1998)) system obtains the atmospheric humidity by measuring the attenuation by water vapour of laser light emitted at a wavelength of $1.3 \, \mu m$ over a known absorption path. The attenuation is quantitatively related to the concentration of water vapour by the Beer-Lambert Law, which states that the reduction in intensity is proportional to the optical path length, the incident intensity and the absorption cross section (which is a function of pressure and incident wavelength). Since the TDL calibration is pressure dependent a laboratory calibration was not possible. The calibration was therefore achieved during EMERALD-II by using the FPH measurements, as described by Cook (2004).

**Best Aircraft Turbulence probe**

The Best Aircraft Turbulence probe (BAT probe\(^1\)) was primarily designed by NOAA (National Oceanic and Atmospheric Administration) and ARA to provide measurements of atmospheric turbulence. This however requires accurate knowledge of air temperature, pressure and

\(^1\) see http://www.noaa.inel.gov/capabilities/bat/ for a detailed description
the local 3D wind vector, all of which are sampled by the BAT probe. Temperature is sampled at a rate of 14 Hz with accuracy $\pm 0.1 \text{ K}$, whilst local pressure is observed at 1 kHz with accuracy $\pm 0.05 \text{ hPa}$.

*Instrumentation on board the Kingair*

**Lidar**

The upward viewing lidar system on board the Kingair provided real time mapping of the cloud structure sampled by the Egrett, enabling the mission scientists to guide the flight path of the Egrett relative to the cloud layer. The system employed a Nd:YAG laser which transmitted pulses of light at a wavelength of 532 nm through an aperture in the ceiling of the Kingair aircraft. The measured backscatter was recorded as a function of time (which translates to vertical distance above the lidar) giving a vertical resolution of 30 m with a range of up to 13 km above the aircraft.

### 5.2 Case study: 2\textsuperscript{nd} December 2002 flight

The remainder of this chapter describes results from an EMERALD-II flight which took place on 2\textsuperscript{nd} December 2002. The flight paths of the two aircraft and the conditions encountered on the day are outlined first. Section 5.2.2 then describes how water vapour and temperature measurements are used to construct the profiles used as input by the LBLRTM (Line-By-Line Radiative Transfer Model, see Chapter 2), and the radiative transfer model output is compared with a TAFTS clear sky observation in Section 5.2.3. CPI and FSSP data from the flight is used to estimate the cirrus optical properties used by LBLDIS (Line-By-Line DIScrete ordinate radiative transfer model, see Chapter 2) in Section 5.2.4. Finally, Section 5.2.5 compares TAFTS cirrus observations with far-infrared radiance spectra calculated using LBLDIS.
5.2.1 Flight summary

The flight paths taken by the two aircraft on 2nd December 2002 are shown in Figure 5.4. After approaching the vicinity of the outflow cirrus from the south, both aircraft made a number of horizontal runs back and forth along the 12°S line of latitude as shown in the lower panel. The Kingair flew at a constant altitude of 5.13 km, providing a real time map of the cloud structure above it using the lidar observations. These were used to guide the Egrett aircraft to the required altitude for sampling the cirrus.

The Egrett initially climbed to 10.4 km before performing the first horizontal run. It is during this run that the clear sky and cirrus TAFTS observations discussed in Sections 5.2.3 and 5.2.5 respectively are made. According to CPI measurements, the Egrett passes all the way through the outflow cirrus during this run. The Kingair lidar observations taken from below the Egrett during this run are shown in Figure 5.5, with the Egrett flight path plotted in black. This shows the limitations of the lidar system in the presence of thick, lower level cloud, since no optical backscatter is observed at the Egrett flight level for the majority of this horizontal run. However, the CPI recorded the presence of ice crystals for almost the whole duration of the run, as illustrated by the crosses plotted in Figure 5.4. The lidar signal is completely attenuated in this instance by the presence of thick lower level cloud, effectively restricting the lidar vertical range.

Following this run, the Egrett then climbed to 13.7 km whilst turning around to approach the cirrus from the west. The cirrus profile is then sampled by flying a zig-zag pattern, with descents flown in the eastwards direction and horizontal runs in the westwards direction. This continues until the Egrett is below the cirrus at 6.95 km. One final ascent up to 10.8 km through the cirrus completes the mission. The zig-zag profiling of the cloud reveals the vertical structure of the cirrus composition through the CPI and FSSP measurements, which are discussed further in Section 5.2.4.

The temperature and humidity measurements recorded during the flight are shown in Figure 5.6. The sections of the profile are coloured to indi-
Fig. 5.4: Flight paths of the Egrett (solid lines) and Kingair (dashed lines) on 2nd December 2002. Crosses indicate where the Egrett was flying through cirrus according to CPI measurements, whilst the colour code (used in the temperature and water vapour profiles in Figure 5.6) corresponds to the different flight legs. The black and red diamonds show the position of the Egrett during the TAFTS clear sky (black) and cirrus (red) zenith view observations discussed in this chapter.
cate measurements taken during the corresponding coloured flight legs in Figure 5.4. From the temperature profile shown in the left panel, it is clear that the temperature varies very little with horizontal location, and is almost solely dependent on altitude. The humidity profile does however show horizontal dependence, particularly in the presence of cloud. Both hygrometers used during EMERALD-II suffered problems with ice clogging up their inlets whilst flying through cloud (Vaughan, personal communication). Humidity data taken whilst within cloud cannot therefore be trusted and is not considered when constructing the humidity profile used in the LBLRTM calculations, as described in the following section.

5.2.2 Modelling the 2nd December 2002 atmosphere from observations

The profiles used in the LBLRTM calculations are mostly derived from temperature and humidity measurements taken in-situ by instruments on board the Egrett. However, the in-situ measurements are obviously only available
Fig. 5.6: Temperature (left panel) and water vapour (right panel) measurements taken by the Egrett during the 2nd December 2002 flight. The colours correspond to the same coloured flight legs in Figure 5.4.

for altitudes up to the maximum height at which the Egrett flew during the mission, which in this case is 13.7 km. Above 13.7 km, the standard tropical atmosphere (Anderson et al., 1986) is assumed. The standard tropical atmosphere temperature and relative humidity profiles are plotted alongside the Egrett measurements in Figure 5.7. The mixing ratio measurements have been converted into relative humidity (RH) with respect to water using the Goff-Gratch equation for saturation vapour pressure (Goff, 1957).

The observed temperature profile closely follows the standard tropical atmosphere throughout the altitude range covered by the Egrett, with the two profiles converging at the maximum height flown. This gives some confidence in the assumption that the standard tropical atmosphere provides a suitable representation of the temperature profile above 13.7 km. The measured wa-
Fig. 5.7: Final temperature (left panel) and water vapour (right panel) profiles (in blue) used in LBLRTM to calculate clear sky layer transmittances, alongside the Egrett measured (in black) and standard tropical (in red) profiles. The green water vapour profile includes a moist layer (100% RH with respect to ice – the x-axis of the plot is RH with respect to water) coincident with the cirrus measured by TAFTS. The dotted line indicates the Egrett flight level at the time of the TAFTS observations (10.4 km).

The water vapour profile is however very different from the standard tropical atmosphere, which is not surprising given the spatial variability of water vapour concentration throughout the Earth’s atmosphere. The hygrometer observations taken during the first two ascents of the Egrett flight (coloured in blue and green in Figure 5.4) are used for the bottom 13.7 km of the atmosphere, since during these ascents the Egrett does not pass through cloud and the problems experienced with ice build-up in the hygrometer inlets (see Sec-
tion 5.2.1) are therefore avoided. Above 13.7 km, the water vapour mixing ratio is linearly interpolated between the Egrett 13.7 km humidity measurement and the standard tropical atmosphere water vapour mixing ratio value at the tropopause, located at 17.0 km above sea level. The standard tropical atmosphere humidity profile is then assumed for the stratosphere upwards. The suitability of these assumptions regarding the temperature and water vapour profiles is tested by comparing LBLRTM output with TAFTS clear sky observations, as described in Section 5.2.3.

Since the Egrett hygrometer observations used for the water vapour profile were all taken in cloud free conditions, it is likely that this profile is incorrect where there is cloud present. To account for this, it is assumed when making the LBLDIS calculations described in Section 5.2.5 that the relative humidity with respect to ice within the cirrus layer is 100%. The impact of enhancing the water vapour concentration within the cloud in this way on the water vapour profile is shown by the green curve in Figure 5.7.

5.2.3 Comparison of modelled clear sky spectral radiances with TAFTS observations

In this section LBLRTM calculations of clear sky spectral radiances are performed using the temperature and water vapour profiles described in Section 5.2.2. These are then compared with a TAFTS zenith view clear sky observation. Although the instrument was operated using both zenith and nadir viewing geometries, only zenith spectra are investigated in this chapter. This is because the atmosphere below the Egrett flight level was extremely moist, with precipitable water vapour (PWV) in the atmosphere below 10.4 km equal to 21.98 mm. The PWV above 10.4 km (assuming the standard tropical atmosphere above the tropopause) is only 0.1 mm, so when looking at downwelling spectral radiances the microwindows between the water vapour absorption lines are much more transparent than when looking in the nadir direction, and are therefore more sensitive to cirrus properties.

The TAFTS observation used here is an average of 12 zenith view spectra
recorded between 06:20:04 and 06:20:47 UTC (15:50:04 and 15:50:47 local time). The location of the Egrett at these times is indicated by the black diamond in Figures 5.4 and 5.5. The TAFTS observed spectrum is plotted in Figure 5.8 along with the initial LBLRTM calculated spectral radiances.

There are two ways in which the radiative transfer model output spectrum differs from the TAFTS observation. Firstly, there are a number of spikes in the residual coincident with the strongest water vapour absorption lines. These are due to absorption by water vapour within the TAFTS pointing optics box along the path imbalance that exists between the zenith and nadir input arms (see Chapter 3). This is therefore an instrumental effect, and not a problem with the atmosphere assumed by the radiative transfer model.

Secondly, the calculated spectral radiances at wavenumbers between the absorption lines are generally greater than the observed values across the wavenumber range considered here. There are a number of possible issues which can explain this residual within the absorption microwindows. If the assumed humidity profile contains too much water vapour, this would result in the calculations introducing more absorption at far-infrared wavenumbers than occurs in reality. The incident radiances would therefore, on average, be

Fig. 5.8: Comparison of TAFTS zenith clear sky observed (black) with LBLRTM calculated (blue) spectral radiances, calculated using the blue temperature and water vapour profiles shown in Figure 5.7.
emitted by a lower (warmer) region of the atmosphere, and would be greater in magnitude than those actually observed using TAFTS. A second issue is the foreign broadened water vapour continuum absorption strength assumed by the model – if this is too high at these wavenumbers, then it will have a similar effect on the calculated spectrum to that of assuming too much water vapour. These are considered to be the two main issues, so are investigated further in this section. It is also possible that errors in the absorption line intensities and half-widths obtained from the HITRAN line parameter database (Rothman et al., 2005) could contribute to the discrepancy. Section 2.1.2 discusses the uncertainties in the far-infrared water vapour line parameters listed in the HITRAN database.

The presence of a thin cirrus layer undetected by the Kingair lidar would be another possible source of any residual between the observed and simulated clear sky spectra. Lidar systems similar to the one used here (Nd:YAG laser at a wavelength of 532 nm) have been demonstrated to have a cirrus optical depth detection limit of $\sim 1 \times 10^{-3}$ (Sassen & Cho, 1992). The far-infrared radiative effect of a cirrus layer optically thinner than this detection limit would be within the uncertainty level of the TAFTS spectral radiance observations, so the possibility of contamination of the TAFTS clear sky scene by thin cirrus is not pursued further here.

The sensitivity of the radiances observed in the absorption microwindows to these two parameters is investigated here. The microwindows used in the study are shown in the upper panel of Figure 5.9. LBLRTM spectra were calculated using a range of scaling factors between 0.5 and 1.0 for both PWV and foreign broadened continuum absorption strength. For each combination of scaling factors, the root mean square (RMS) of the calculation minus observation residual is obtained, using only spectral radiances within the absorption microwindows. These RMS residuals are plotted as a function of the two scaling factors in the lower panel of Figure 5.9.

The contour plot shows that reducing both the PWV and the foreign broadened continuum absorption strength ($C_{fgm}$) improves the agreement be-
Fig. 5.9: Upper panel: Positions of absorption microwindows (blue crosses) used to optimise PWV and foreign broadened continuum scaling factors for the comparison of TAFTS with simulated radiances. Lower panel: Contour plot of LBLRTM - TAFTS RMS residuals as a function of PWV and $C_{f_{gn}}$ scaling factors. The cross indicates the location of the minimum RMS value. The blue contour shows the region where the difference from the minimum RMS value is less than the estimated TAFTS uncertainty.
between the LBLRTM calculation and the TAFTS observation. The RMS spectral radiance residual in the microwindows is minimised by reducing the PWV by 29% and $C_{fgn}$ by 7%. Using these scaling factors has the effect of reducing the microwindow RMS residual from 3.16 down to $0.84 \text{ mW m}^{-2} \text{ sr}^{-1} (\text{cm}^{-1})^{-1}$. The LBLRTM spectrum calculated using these optimal scaling factors is plotted alongside the TAFTS observed radiance spectrum in Figure 5.10.

![Figure 5.10](image)

**Fig. 5.10:** Comparison of TAFTS clear sky observed (black) with LBLRTM calculated (blue) spectral radiances. The PWV and $C_{fgn}$ used in the LBLRTM calculation have been scaled by the optimal scaling factors shown in Figure 5.9.

The scaling factors obtained here should be regarded as an estimate of the conditions which best reproduce the spectra observed by the TAFTS when applied to the radiative transfer model input. The uncertainty in the longwave channel TAFTS observations is $\sim 0.15 \text{ mW m}^{-2} \text{ sr}^{-1} (\text{cm}^{-1})^{-1}$ (see Section 3.3.3), and is illustrated by the blue contour in Figure 5.9. The two scaling factors obtained here (0.71 for PWV and 0.93 for $C_{fgn}$) are assumed in all subsequent radiative transfer calculations in this chapter. This enables any microwindow residuals between TAFTS cirrus observation and LBLDIS output greater than $\sim 0.84 \text{ mW m}^{-2} \text{ sr}^{-1} (\text{cm}^{-1})^{-1}$ to be attributed to errors in the cirrus model, rather than issues with the effect of water vapour absorption on the spectral radiances within the absorption microwindows. The
change in PWV applied here is greater than can be explained by uncertainty in the TAFTS spectral radiances alone (most likely due to the assumption of a standard tropical atmospheric profile for water vapour higher than the tropopause), though the scaling applied to $C_{fgn}$ does fall within the TAFTS longwave channel uncertainty.

5.2.4 Estimation of cirrus layer scattering properties from CPI, FSSP and lidar observations

This section concerns the calculation of the cirrus layer scattering properties to be used in the LBLDIS code. The availability of cloud probe measurements removes the need to assume a particle size distribution (PSD) shape, though care must still be taken to ensure that the subset of CPI and FSSP data taken from the flight to construct the PSD is representative of the section of cirrus being observed by the TAFTS instrument. Figure 5.11 shows, in black, the mean PSD measured by each instrument for all in-cloud measurements taken during the flight. The error bars show the fractional standard deviation for each crystal dimension bin – for example, if the standard deviation is 50% of the mean value, then the error bars indicate the mean value plus/minus 50%.

The coloured PSDs are calculated by grouping the cloud probe data into temperature ranges, as indicated by the legend on the figure. Since temperature decreases with altitude, this effectively shows the dependence of the PSD on cloud height. Whilst the PSD shapes are similar, the cloud water content (CWC) within each temperature range varies, with the smallest concentrations of particles measured by both the FSSP and CPI occurring in the coldest temperature range (-50 to -60 °C). However, the PSDs calculated for each of the four temperature ranges all fall within the natural variability of the mean PSD for the whole flight. This firstly demonstrates the inhomogeneous nature of cirrus composition, and secondly suggests that the use of a single mean PSD to model the whole cirrus layer in the LBLDIS calculation is justified.
**Fig. 5.11:** Particle size distributions measured by the FSSP (left panel) and CPI (right panel). In each case, the thick black line is the mean PSD from the whole flight, whilst the coloured thin lines are the mean PSDs for certain temperature ranges.

The initial PSD assumed for the radiative transfer calculation is derived from the mean of all PSDs measured by the FSSP and CPI between altitudes...
of 11.5 and 13.0 km, as shown in Figure 5.12. These altitudes are taken as the cloud boundaries, using the lidar observations at the time of the TAFTS cirrus measurement labelled in Figure 5.5. It is difficult to tell from the lidar observation whether the upper boundary inferred from the optical backscatter data is the real upper boundary, or the point at which the lidar signal is saturated. The cloud thickness of 1.5 km should therefore be taken to be a minimum possible value in the following discussions.

**Fig. 5.12:** Mean PSDs measured by the CPI (black) and FSSP (red) between 11.5 and 13.0 km. The blue PSD is the composite used in the LBLDIS calculation (see Section 5.2.4).

The FSSP and CPI are combined by taking the FSSP size distribution for crystals below 20 $\mu$m, and then scaling the CPI PSD such that it meets the FSSP PSD at this point. The composite PSD, plotted in blue in Figure 5.12, is then interpolated onto a higher resolution grid to be used in calculating the bulk scattering properties. It has an effective diameter ($D_e$, see Equation 4.8) of 58.5 $\mu$m. The bulk scattering properties obtained by integrating the single crystal scattering properties for ice aggregates over this PSD (as described in Section 4.4.2) are plotted in Figure 5.13.
Fig. 5.13: Bulk scattering properties obtained by integrating single scattering properties over the composite PSD shown in Figure 5.12. Top panel: optical depth; middle panel: scattering albedo; bottom panel: asymmetry parameter.

The visible optical depth (as defined by Equation 4.9) is equal to 3.87, and the ice water content (IWC) estimated using the mass-dimension relation $m = 0.0185D^2$ (Brown & Francis, 1995) is $4.66 \times 10^{-2} \text{ gm}^{-3}$. The LBLDIS radiance spectrum calculated using bulk scattering properties obtained using the composite PSD is compared with the TAFTS cirrus observation in Section 5.2.5.
5.2.5 **Comparison of modelled cirrus spectral radiances with TAFTS observations**

Here a comparison is made between the TAFTS cirrus observation (the position of the Egrett at the time of this observation is denoted by the red diamond in Figures 5.4 and 5.5) and the radiance spectrum calculated with LBLDIS assuming the composite PSD described in Section 5.2.4. The TAFTS observation is an average of 12 zenith view spectra measured between 06:53:04 and 06:54:06 UTC (16:23:04 and 16:24:06 local time). The two spectra are plotted together in Figure 5.14, along with the TAFTS clear sky observation discussed in Section 5.2.3.

![Fig. 5.14: Comparison of TAFTS observed (red) with LBLDIS calculated (blue) spectral radiances in the presence of a cirrus layer. The clear sky TAFTS observation discussed in Section 5.2.3 is shown in black.](image)

The calculated radiance spectrum is similar to the TAFTS observation, though the radiances in the absorption microwindows are greater in the modelled spectrum (plotted in blue) than they are in the observed spectrum (in red). The RMS residual in the microwindows (as defined in the upper panel of Figure 5.9) is \(1.47 \text{ mW m}^{-2} \text{ sr}^{-1} (\text{cm}^{-1})^{-1}\). Given that the difference is positive for all absorption microwindows, it appears that the modelled cirrus layer used in the LBLDIS calculation is too opaque. This possibility is investigated by re-calculating the radiances for a number of different PSDs.
Fig. 5.15: Upper panel: RMS residual between TAFTS and LBLDIS in the microwindows as a function of the PSD scaling factor. Lower panel: Same as Figure 5.12, but also including (in green) the scaled PSD giving best agreement with the TAFTS observation.

which have been scaled down from the original composite PSD used initially. The effect of this is to reduce the cirrus optical depth whilst keeping the PSD shape (and therefore $D_e$) constant. The microwindows RMS residual, plotted as a function of the scaling factor in the upper panel of Figure 5.15, is minimised to $0.392 \text{ mW m}^{-2} \text{ sr}^{-1} (\text{cm}^{-1})^{-1}$ when the PSD is scaled by a factor of 0.63. The scaled PSD is plotted in green in the lower panel of Figure 5.15, along with the original composite PSD and the cloud probe observations.
The uncertainty in the TAFTS observation ($\sim 0.15 \text{ mW m}^{-2} \text{ sr}^{-1} \text{ (cm}^{-1})^{-1}$) implies that the actual scaling factor providing best reconciliation between the modelled and observed spectra could take any value between 0.56 and 0.7.

The scaled PSD falls within the lower bounds of the cloud probe observation error bars, which demonstrates that the method used to construct the composite PSD in Section 5.2.4 is justified. However, it should be noted at this point that the scaled PSD still represents an upper limit for the true size distribution, since the cloud geometrical thickness could be greater than that assumed here. If the cloud was thicker (greater than 1.5 km), then the ice crystal concentrations required to match the TAFTS observation would have to be reduced to compensate. Therefore, if the cirrus layer is thicker, then the PSD would have to be scaled down even further for radiative closure between the TAFTS observation and the LBLDIS output.

The bulk scattering properties calculated by integrating over the scaled PSD are shown in Figure 5.16. The scattering albedo and asymmetry parameter are unchanged, since they are each defined as the ratio of two integrals, both of which include the size distribution $N(D)$ in the integrand. If the whole PSD is multiplied by the same scaling factor as is done here, then the scaling factor cancels out when calculating the ratio (see Equations 2.34 and 2.35). They are therefore dependent on the PSD shape rather than the total ice crystal concentration. The optical depth is reduced at all wavenumbers by the same scaling factor as the PSD. The visible optical depth for this re-scaled PSD is 2.44 and the estimated IWC is $2.93 \times 10^{-2} \text{ gm}^{-3}$.

Figure 5.17 shows the effect of scaling down the assumed PSD on the LBLDIS - TAFTS residual. The residual shows no obvious spectral dependence within the absorption microwindows, which indicates that both the PSD effective diameter and the assumed single scattering properties are accurate enough to reproduce the observed radiance spectrum between 100 and 200 cm$^{-1}$. Unfortunately, the unavailability of the TAFTS shortwave channels during this flight prevents testing of the effectiveness of the cirrus model.
Having demonstrated that the composite PSD used here results in a modeled cloud layer which is too opaque, the contribution of small crystals to this discrepancy is now investigated. Recent work suggests that small ice crystals are over-sampled by cloud probes due to the shattering of larger crystals on probe inlets (Field et al., 2003, 2006; Heymsfield, 2007). The sensitivity of far-infrared spectra to small crystals (defined here as all crystals with maximum dimension $D < 100\,\mu m$) can be determined by comparing the bulk scattering properties calculated for the composite PSD with those obtained when small crystals are removed, as shown in Figure 5.18.

Removing small crystals in this way not only has the effect of reducing the optical depth (as the quantity of ice in the cloud is reduced, less radiance
is absorbed by the cloud layer), but also increasing the asymmetry parameter (since scattering in larger crystals is concentrated in the forwards direction, whereas small crystals tend to scatter isotropically). The scattering albedo is increased by a small amount at wavenumbers between 120 and 200 cm$^{-1}$. The small crystals in the composite PSD account for 59.0% of the total ice water content. The effect on the calculated spectral radiances, shown in the lower panel of Figure 5.18, is a reduction in spectral radiances within the absorption microwindows. This is primarily caused by the decrease in optical depth, which results in a greater contribution to the observed radiance from transmitted, rather than emitted radiation. Since in the zenith view the radiances transmitted by the cloud are very low, a decrease in total radiance is observed when the optical depth is reduced.

The effect of scaling down the small crystal component of the PSD is shown in Figure 5.19. The upper panel shows the RMS residual between the TAFTS observation and the LBLDIS output as a function of the scaling factor applied to the composite PSD for $D < 100 \mu m$. Applying a scaling factor of 0.34 to the small crystals, which gives the PSD plotted in green in the lower panel of Figure 5.19, minimises the RMS residual. There is
Fig. 5.18: Upper three panels: Effect on bulk scattering properties of removing crystals with $D < 100 \mu m$ from the original measured PSD. Blue curves are the properties calculated with the original composite PSD, green curves are the properties for PSDs with small crystal concentrations scaled by factors 0.9 to 0.0 in 0.1 intervals. Lowest panel: Effect on LBLDIS radiance spectrum.
some uncertainty in the scaling factor obtained here, due to uncertainty in
the TAFTS observation ($\sim 0.15 \text{ mW m}^{-2} \text{ sr}^{-1} (\text{cm}^{-1})^{-1}$). Applying this
uncertainty to these results suggests that the scaling factor which provides
the best representation of the scene observed by TAFTS has some value
between 0.22 and 0.46.

The cloud properties obtained assuming this PSD are similar to those
obtained for the scaled PSD shown in Figure 5.15, with visible optical depth $\tau_{\text{vis}} = 2.35$ and an ice water content of $2.85 \times 10^{-2} \text{ gm}^{-3}$ (compared with values of 2.44 and $2.93 \times 10^{-2} \text{ gm}^{-3}$ respectively obtained when scaling the whole PSD). The effective diameter has however increased from 58.5 to 83.3 $\mu$m. Figure 5.20 shows the LBLDIS spectrum calculated using the PSD in Figure 5.19 alongside the TAFTS observation. The comparison shows that the scattering properties calculated using this PSD result in LBLDIS output which matches observations very well across this spectral range.

![Figure 5.20: Same as Figure 5.14, but also including (in green) the LBLDIS radiance spectrum calculated using the scattering properties obtained from the PSD (with small crystal concentrations scaled by a factor of 0.34) in Figure 5.19.](image)

The model spectrum shown here is very similar to the one shown in Figure 5.17, despite the different PSD shape (and therefore effective diameter) used in each case. This implies that spectral radiances in the 100 to 200 $\text{cm}^{-1}$ range alone cannot be used to distinguish remotely between PSDs which have similar ice water content but different effective diameter. This study of the effect of reducing the small crystal concentration does however support the conclusions of previous work regarding the artificial enhancement of small crystal concentrations measured by cloud probes such as the FSSP due to the shattering of large crystals on the probe inlet (Field et al., 2003, 2006;
McFarquhar et al., 2007; Heymsfield, 2007). The reduction in concentrations of particles with $D < 100 \mu m$ required here for the radiative transfer model output to agree with observations was 66%, suggesting that these particles may have been over-counted by a factor of $\sim 3$. This is in good agreement with the results of Field et al. (2003) who found in their data that the FSSP typically over-counted small particles by a factor of $\sim 2$, and in the worst case over-counted by a factor of 5.

### 5.3 Summary of EMERALD-II results

During this chapter, TAFTS observations taken during the 2nd December 2002 EMERALD-II flight have been compared with line by line radiative transfer calculations. Using the TAFTS instrument’s zenith viewing geometry at aircraft altitude has enabled calculations of spectral radiances at wavenumbers less than 200 cm$^{-1}$ to be tested against observations. When viewed in the nadir from aircraft altitude (or in the zenith from ground level) the atmosphere is far too opaque to investigate the spectral radiance signature of cirrus, due to strong absorption by the water vapour pure rotation band.

Comparison of a TAFTS clear sky spectrum with LBLRTM output showed that reductions of the assumed column water vapour amount by 29% and the foreign broadened water vapour continuum absorption strength by 7% were required to give the best agreement between the calculated and observed radiances in the absorption microwindows. Some of the required change to the PWV may be explained by the assumption of a standard tropical atmospheric profile above the tropopause, though similarly to the RHUBC results presented in Chapter 4 it is also possible that there is a contribution from errors in the TAFTS calibration (as discussed in Section 4.5). The water vapour continuum at these wavenumbers is based on a semi-empirical function fitted to the experimental data obtained by Burch (1981) taken at wavenumbers between 350 and 600 cm$^{-1}$ at 296 K. Previous studies involving
5. Aircraft-based observations of tropical cirrus during EMERALD-II

atmospheric spectral observations at wavenumbers down to 240 cm$^{-1}$ have noted less absorption by the foreign broadened continuum than predicted by Burch’s laboratory observations (Tobin et al., 1999; Green, 2003; Serio et al., 2008b), consistent with the findings from the clear sky comparison made here. The scalings of PWV and water vapour continuum found during this clear sky comparison are applied when using LBLDIS to simulate the TAFTS cirrus observation, so that differences between the observed and calculated cirrus spectra may be attributed with greater confidence to errors in the cirrus model.

The comparison of a TAFTS cirrus view spectrum with LBLDIS output demonstrated that, for this particular case, the assumed crystal single scattering properties were able to reproduce the observed spectral radiances with some success when used as input for the scattering code. However, the uncertainty in the particle size distribution used (owing to the inhomogeneity of the cirrus sampled by the cloud probes) makes it difficult to determine from this single case how representative they are of the real scattering properties of the cirrus layer measured. The contribution of small crystals to the spectral signature was also investigated, and it was found that reducing the small crystal concentrations in the measured PSD by a factor of $\sim 3$ resulted in the best agreement between the radiative transfer code and the TAFTS observation. This result is consistent with previous work by Field et al. (2003) which suggested that the FSSP in particular over-counts small particles by up to a factor of 5.

The results presented in this chapter demonstrate for the first time that measurements of the effect of tropical cirrus on far-infrared radiance spectra are able to provide an insight into the microphysical properties of the cirrus layer. The observations show that, given knowledge of the crystal size distribution, it is possible to successfully calculate far-infrared spectral radiances using a database of theoretical scattering properties.
6. CONCLUSIONS AND FUTURE WORK

This thesis has presented an original body of work investigating the influence of cirrus on far-infrared radiance spectra, through a combination of radiative transfer model calculations and spectrometer observations. Though the significance of far-infrared radiative transfer to the Earth’s radiative energy budget is now recognised (Harries et al., 2008), it is only recently that observations have begun to test this theoretical understanding. The comparisons made in this thesis between radiance spectra observed using TAFTS and simulated using a state-of-the-art radiative transfer model, under both clear sky and cirrus conditions, represent a step forward in the interpretation of far-infrared radiative transfer through the Earth’s atmosphere using spectral radiance observations.

The most significant achievements of this work are summarised in Section 6.1. Highlights include the first observations of tropical cirrus at far-infrared wavelengths, and the first simultaneous observation of an arctic ice cloud by two far-infrared spectrometers. Section 6.2 discusses possibilities for future research that would utilise and extend upon the results presented in this thesis.

6.1 Summary of main results

6.1.1 Radiative Heating in Underexplored Bands Campaign (RHUBC)

The deployment of TAFTS in Barrow, Alaska during RHUBC was groundbreaking for a number of reasons. The campaign was the first time that TAFTS had performed observations of the atmosphere from the ground via its zenith view geometry. The extremely dry conditions prevalent during the
Arctic winter allowed ground based observations of the troposphere which wouldn’t be possible at lower latitudes, owing to strong absorption by the water vapour pure rotation band. The spectral observations taken during clear sky conditions showed potential for remotely sensing both the amount of tropospheric water vapour and the strength of the water vapour continuum absorption from the ground under these conditions. Successful characterisation of these properties is vital for the study of cirrus using far-infrared radiances, since the wavenumbers most sensitive to cirrus cloud (those within the absorption microwindows) are also the most sensitive to perturbations in water vapour amount and continuum absorption. Clear sky observations were used here to determine by how much the water vapour parameters in the model needed adjusting to provide the best agreement (see Sections 4.3.3 and 4.3.4). These adjustments were then applied when producing simulated cirrus spectra, in order to minimise the discrepancies caused by errors in the model water vapour profile and continuum absorption strength.

The co-location of the TAFTS and AERI-ER spectrometers at the NSA-ACRF site provided a unique opportunity to make simultaneous observations of the Arctic atmosphere. This was the first time that such an inter-comparison had been performed between two spectrometers observing far-infrared spectral radiances. TAFTS and AERI-ER showed similar responses to the scenes observed during RHUBC, though differences were noted in the baseline calibrations of the two instruments (see Section 4.5). The inter-comparison between the two spectrometers and the radiative transfer model output instilled greater confidence in the measured spectra than would have been possible if only one spectrometer was used during the campaign.

The sensitivity of the LBLDIS simulated far-infrared radiances in the presence of cirrus to the particle size distribution (PSD) shape was tested using the parameterization of Field et al. (2005). Increasing the proportion of small crystals in the PSD whilst holding the ice water content (estimated using the AERI-ER mid-infrared window observations) constant resulted in a reduction in spectral radiances in the absorption microwindows
In the mid-infrared window, the effect altering the PSD shape in this way is to increase the gradient of the radiance vs. wavenumber slope (see Figure 4.24). This mid-infrared window effect was found to be smaller in magnitude than the reduction in spectral radiances calculated for the far-infrared absorption microwindows.

On comparison with observations, it was found that the best agreement between the observed and simulated spectra was obtained by increasing the small crystal fraction. The fraction of ice water content attributed to small crystals needed to be increased from $\sim 11\%$ (the fraction predicted using the original unperturbed PSD generated by the parameterization) to $\sim 88\%$ to match the AERI-ER far-infrared observations, whilst for TAFTS the best fit was achieved when it was increased to $\sim 44\%$. It should be noted that TAFTS noise levels during this campaign were much higher than those for the AERI-ER spectra analysed, since a principal component analysis noise filter is applied operationally to the AERI-ER data. This means that the TAFTS observations are less sensitive to small changes in the PSD shape, since the resulting changes in spectral radiance fall beneath the TAFTS noise level (note how the RMS residuals listed in Table 4.3 do not vary significantly with PSD shape compared with those obtained using the AERI-ER). Nonetheless, it is encouraging that measurements from both spectrometers suggest that an increase in small crystal fraction is necessary to reconcile the cirrus model with observations. This is consistent with in-situ observations made by Lawson et al. (2001), who found from Cloud Particle Imager (CPI) measurements of arctic cirrus that the PSDs were typically narrow in shape with large concentrations of small ice crystals.

### 6.1.2 Egrett Microphysics Experiment with Radiation Lidar and Dynamics II (EMERALD-II)

The EMERALD-II campaign of science flights targeted the tropical anvil outflow cirrus produced by a highly predictable, isolated system of thunderstorms known as ‘Hector’. ‘Hector’ occurs on an almost daily basis during
the transition between the dry and wet seasons over the Tiwi Islands, located off the north coast of Australia approximately 100 km north of Darwin. During the campaign convective cirrus was sampled in-situ by the high-altitude Egrett aircraft. The Egrett carried instrumentation designed to measure the cirrus microphysical properties, in addition to the TAFTS spectrometer which was operated in both its zenith and nadir viewing geometries. A second aircraft, the Kingair, flew directly beneath the Egrett and carried a lidar to remotely determine the altitude and optical thickness of the target cirrus. The lidar data was used in real time during the flights by the mission scientist to guide the Egrett to the correct altitude for sampling the cloud.

The results presented in Chapter 5 are from the flight which took place on 2nd December 2002. During this flight TAFTS recorded a number of clear sky and cloudy spectra using both viewing geometries, though only the longwave channel was operating successfully (80 - 330 cm\(^{-1}\)). Radiance spectra are presented with a wavenumber range of 100 - 200 cm\(^{-1}\), since it is at these wavenumbers that the signal-to-noise ratio is greatest (see Section 3.3.3). Only the zenith view spectra are used to investigate the sensitivity of the far-infrared radiances to water vapour and cirrus clouds, as the atmosphere below the Egrett was so moist that the absorption microwindows in the nadir view were no longer sensitive to small changes in water vapour and cirrus properties.

The clear sky spectra were used to test the water vapour profile assumed as input for the LBLRTM model. Relative humidity measurements made by the frost point hygrometer were only available at the altitudes flown by the Egrett, so a standard tropical atmosphere model was used to extrapolate upwards beyond the upper troposphere. On comparison with the TAFTS observations, it was found that modelled atmosphere in the absorption microwindows was too opaque. Reducing the PWV by 29% and the foreign broadened water vapour continuum absorption by 7% in the model gave the best agreement with the observed spectrum (see Section 5.2.3). These scalings (which were then applied when simulating the cirrus spectra) are similar
to those used to match clear sky simulated spectra to the TAFTS observations made during RHUBC, even though a different wavenumber range is used during EMERALD-II.

The CPI and FSSP measurements from the 2nd December 2002 flight were used to create a composite PSD, which in turn was then used to calculate the cirrus single scattering properties needed to run the LBLDIS radiative transfer code. The LBLDIS output was then compared with a TAFTS spectrum recorded when the Egrett was flying approximately 1 km beneath the cloud base. The initial comparison showed that the modelled cirrus layer was too opaque, even when allowing for scaling of the water vapour profile used in the calculations as mentioned above. The largest source of uncertainty in the cirrus model is in the measured PSD. A number of spectra were simulated, with the entire PSD multiplied by a range of different scaling factors, and compared with the TAFTS observation. Reducing the whole PSD by 37% in this way minimised the RMS residual between simulation and observation. This re-scaled PSD remained just within the lower limit of the cloud probe measurement uncertainties, as shown in Figure 5.15.

An alternative approach was also considered which tests the hypothesis that small crystals are over-counted by the cloud probes due to larger crystals shattering on the probe inlet, resulting in an artificial enhancement of the concentrations of small crystals measured by the probes. In this case, the PSD obtained from the CPI and FSSP measurements is only reduced for crystal dimensions $D < 100 \mu$m, and is unchanged for all other $D$. The agreement between simulated and observed spectra was best when the small crystal component of the PSD was reduced by 66%. This is in broad agreement with the results of Field et al. (2003), who estimated that the FSSP over-counts small crystals typically by a factor of $\sim 2$, and in the worst cases by factors as high as $\sim 5$. The results presented in Section 5.2.5 demonstrate how far-infrared spectral radiance observations may be used in conjunction with radiative transfer and scattering codes to assess the performance of microphysical measurements, though this implies the assumption that the
scattering property database theoretically calculated for ice crystals is correct at the wavenumber range considered.

6.2 Possibilities for future work

The results presented in this thesis indicate the potential for more extensive use of far-infrared spectral radiance measurements for the investigation of cirrus properties in the future. This will come with the continuing development and testing of far-infrared Fourier transform spectrometer technology, which is already being demonstrated not only with TAFTS but also with other FIR spectrometers, which have recently been deployed for their first measurement campaigns (REFIR and FIRST, see Section 1.3). Some possibilities for further far-infrared investigations and developments (not necessarily specifically related to TAFTS) are discussed here.

The RHUBC results in Chapter 4 would benefit enormously from the development of a principle component analysis (PCA) noise filter for application to the TAFTS data, similar to that used with the AERI-ER (described by Turner et al. (2006)). The difference this would make is noticeable in the sensitivity of the two spectrometers to perturbations in the PSD shape. The changes in spectral radiance expected as a result of such perturbations (illustrated in Figure 4.23) tended to fall within the TAFTS noise levels, whereas the PCA noise filtered AERI-ER data could be used to distinguish between the radiative effects of these perturbations. The TAFTS vs. AERI-ER inter-comparison would also be more definitive if the respective instrument noise levels were roughly similar in magnitude.

A further improvement upon these results would come from a proper treatment of the model water vapour parameters, particularly the foreign broadened continuum absorption. Whilst this has been crudely accounted for in this study with the aim of minimising the effect of errors in the water vapour model on simulated cirrus spectra, a thorough investigation of the wavenumber dependency of the far-infrared water vapour continuum absorp-
tion under atmospheric conditions is deemed necessary. Initial results from the REFIR spectrometer have begun to address this problem (Serio et al., 2008b), and the involvement of all three FIR spectrometers mentioned above in future observational campaigns will expand greatly on the clear sky data currently available to resolve this issue. The success of future studies of cirrus at FIR wavenumbers depends strongly on a realistic characterisation of the effect of water vapour on radiances within this spectral range.

There is also a need to rigorously test the scattering properties used under atmospheric conditions. In the case of the two campaigns described here, there have been missing elements resulting in too many uncertainties to be able to draw conclusions regarding the accuracy of the scattering calculations used to derive the property databases. During RHUBC the absence of in-situ microphysical measurements meant that a PSD shape had to be assumed, and the cloud ice water content had to be inferred remotely. This meant that errors in the simulated spectra were more likely to be caused by problems with these assumptions rather than errors in the scattering properties. Whilst microphysical observations were available during EMERALD-II, the lack of radiosondes close to the measurement site meant that the water vapour profile above the highest flight level has to be assumed. In addition, only a limited spectral range was available for the radiance measurements. The results from RHUBC showed the usefulness of having spectral radiance measurements in the mid-infrared window region between 800 and 1200 cm\(^{-1}\), in addition to the far-infrared observations. In summary, future campaigns dedicated to the studying the influence of cirrus far-infrared spectral radiances should include spectral observations from the far-infrared through to the mid-infrared, in addition to microphysics probes and adequate radiosonde coverage.

Though field campaigns such as RHUBC and EMERALD-II have provided significant insight into far-infrared radiative transfer in the Earth’s atmosphere, the next logical step would be to observe far-infrared spectral radiances from space. Both REFIR and FIRST have been designed as test-beds for possible future space missions, whilst the proposed NASA CLARREO
(Climate Absolute Radiance and Refractivity Observatory, Revercomb et al. (2007)) mission envisages a polar-orbiting nadir view interferometer which can observe radiances at wavenumbers as low as 200 cm$^{-1}$ at 0.5 cm$^{-1}$ spectral resolution. A satellite based instrument such as CLARREO would provide global coverage, and thus enable investigation of the impact of cirrus on FIR radiances (and therefore on the FIR radiative energy budget) for the whole range of possible atmospheric conditions, rather than the very localised conditions encountered during aircraft or ground-based field campaigns. Future investigations should be geared towards developing the ability to exploit such a dataset once it becomes available.


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