LINKING REMOTE SENSING AND IN-SITU DATA OF CLOUD PROPERTIES

A thesis submitted to the University of Manchester for the degree of Doctor of Philosophy in the Faculty of Science and Engineering, School of Earth and Environmental Sciences

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ABSTRACT

Clouds play a major role in many earth-atmosphere system processes, such as precipitation processes and radiative transfer affecting the weather and the climate. Although these processes are included in atmospheric numerical models, they still remain uncertain. This thesis focuses on the comprehensive microphysical analysis of a warm front of a wide low pressure system over the north Atlantic ocean, which affected the United Kingdom on 21 January 2009. For this analysis, airborne and ground based measurements are compared in order to investigate and understand microphysical properties and structures of the clouds.

During the investigation of the warm front, a warm conveyor belt was always activated transporting humid air aloft and causing the formation of massive stratiform clouds with embedded convective elements, which are called “generating cells”. Such cells consist of a high reflectivity factor ($Z_H$) core with rimed crystals and high differential reflectivity ($Z_{DR}$) boundaries with pristine planar crystals. Expanding on the effect and the evolution of these convective features, which were entering the warm conveyor belt dominated by strong and sheared westerlies, they were transformed into two slanted fall streaks of different polarimetric and microphysical characteristics. The microphysical processes within these structures together with the presence of strong winds along the profile of the atmosphere can affect the time and the place of the occurrence of surface precipitation.

Further analysis of the warm front focused on the processes occurring at the upper regions of the clouds and within the warm conveyor belt. For first time, we observed high concentrations of an ice particle with distinctive shape referred to as “ice-lolly”, the formation of which is explained. These particles can potentially play an important role in the lifetime of the clouds and the surface precipitation, being produced due to the interaction between the warm conveyor belt and the ice multiplication mechanism.

The latter part of the present thesis focuses on the evaluation of the NCAR’s Hydrometeor Classification Algorithm (HCA) for S-band radar suggesting a
validation methodology based on in-situ aircraft measurements. Alongside 2DS and CIP-100 probe images, we also used calculated parameters such as mean particle size measurements, concentration of particles of sizes <1000µm and >1000µm, Liquid Water Content (LWC), weighted shape factor, weighted area ratio and weighted transparency retrievals. Although they provided useful information about the ice particles in the clouds, further work could improve the efficiency of using such parameters. The HCA results seem to generally agree with the in-situ observations. However, the presence of liquid water close to the melting layer may be problematic for hydrometeor classification.
DECLARATION

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PhD by published work Candidate Declaration

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Faculty: Science and Engineering
Thesis Title: Linking Remote Sensing and In-Situ Data of Cloud Properties

Declaration to be completed by the candidate:

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CHAPTER ONE

INTRODUCTION

1.1 Motivation

Clouds play a major role in many earth-atmosphere system processes, such as precipitation formation and radiative transfer. In this study, microphysical processes occurring in mixed-phase clouds are investigated. The concentration and the size of ice particles in clouds can significantly alter their lifetime, component, structure and radiative forcing [Lohmann and Feichter, 2005] through Bergeron-Findeisen process [Bergeron, 1935]. In mixed-phase clouds, when ice particles are found under specific conditions (see section 2.1.3), ice multiplication mechanisms are activated producing significant amounts of ice particles in clouds affecting their properties [Hallett and Mossop, 1974; Phillips, 2003]. Although cloud microphysical processes have been subject of extensive research [e.g. Beard and Pruppacher, 1969; Hobbs et al., 1973; Hallett and Mossop, 1974; Choularton, 1980; Matejka et al., 1980; Lord et al., 1984; Jameson and Johnson, 1990; Biggerstaff and Houze, 1993; Mitchell, 1996; Khain et al., 2004; Pruppacher and Klett, 2012] and have been parametrised to be used in atmospheric numerical models [e.g. Walko et al., 1995; Lohmann and Roeckner, 1996; Meyers et al., 1997; Hong and Lim, 2006; Seifert and Beheng, 2006], they still remain uncertain [e.g. Baumgardner et al. 2012, Theriault et al. 2006]. As a result, from a short-term prediction of the evolution of a hazardous weather system to the estimation of the long-term impact of the Earth climate and energy balance change, improving the understanding of cloud microphysics is still challenging and critically important.

This study focuses on comparing and linking polarimetric radar data with in-situ cloud properties data. One of the main purposes is to further investigate and understand the evolution of the cloud microphysical processes associated with frontal systems. In addition, we aim to enhance the confidence of interpreting radar data, which can lead to the development of robust hydrometeor classification
algorithms (HCAs). The existence of accurate HCAs is of high importance, as they can assist in improving our understanding of atmospheric dynamics and provide higher quality of assimilation and verification in atmospheric numerical models. Finally, HCAs can make a significant contribution to operational nowcasting applications associated with civil safety. As an illustration, identifying the location of dendritical ice particles can assist in avoiding aircraft icing hazards [Bringi et al., 2007; Nygaard et al. 2011; Smith et al. 2012]. In addition, improved HCAs can provide more accurate information about the hydrometeor composition of the clouds. This can improve the estimation of rain rate, which depends on the hydrometeor type and the density of ice crystals, which differs for different ice habits [Vivekanandan et al. 1994, Mitchell et al. 1996, Wolfe and Snider 2012].

1.2 The APPRAISE Project

The Aerosol Properties, Processes And InfluenceS on the Earth’s climate (APPRAISE) project took place in south UK between 2007-2011 [Bower and Burgess, 2012]. The overall purpose of the project was to investigate the role and the effectiveness of a wide range of aerosol types (such as dust, organic material, black carbon etc), which can act as cloud condensation (CCN) or ice nuclei (IN), in ice and mixed phase clouds. For this project, a fully equipped (see description of some instrumentation in chapter 3) FAAM (Facility for Airborne Atmospheric Measurements) aircraft and ground based instrumentation (the Chilbolton lidar and radar) were adopted for obtaining measurements of chemical composition and size distribution of aerosol particles, but also of cloud properties. The main focus of the present study is to a.) obtain insight into the microphysics of clouds of frontal systems analysing and comparing both airborne and ground measurements, b.) identify regions of clouds with specific characteristics and properties using polarimetric radar data and c.) make use of in-situ measurements to evaluate a HCA.

Based on the APPRAISE measurements, Crosier et al. [2011] examined thin mixed-phase stratiform clouds with large planar crystals in small concentrations
(0.2 L⁻¹) being nucleated at ~ -11°C and precipitated as virga. They found that ice formed through slow activation process of droplets containing INs. In addition, they observed embedded convective clouds, with (at least a magnitude) higher than predicted concentrations of ice, which formed due to Hallett-Mossop process and caused light surface precipitation. Analysing a different case, Crawford et al. [2012] investigated the importance of the Hallett-Mossop process in the rapid glaciation of slightly supercooled wintertime cumulus clouds under the presence of dust, biological particles and small numbers of primarily produced ice crystals at temperatures ~ -7.5°C. Studying the role of supercooled water, Westbrook et al. [2011] observed that supercooled droplets are always present at temperatures > -20°C, which implies that ice can mostly originate from freezing supercooled water drops. Finally, Cui et al. [2012] examined the formation and characteristics of wave clouds, which formed between the maximum updraft and downdraft, where vertical velocity is opposite, while temperature reaches the lowest values in the middle of the cloud.

1.4 Thesis overview

The main theme throughout this thesis is to investigate mixed-phase clouds, as there is significant uncertainty in different aspects of microphysics, especially when ice phase is involved (e.g. ice nucleation, secondary ice production etc). In particular, we try a.) to analyse detailed microphysical measurements of cloud properties, which were collected in-situ by Optical Array and Mie Scattering probes, including ice/droplet number concentrations, ice/liquid water content and particle images, b.) to investigate, and link polarimetric radar features of frontal system clouds with the observed in-situ properties of clouds (examples of such radar features are the warm conveyor belt signature in Doppler velocity parameter and the ice fall streak signatures in reflectivity and differential reflectivity parameters). This analysis can reveal useful details for the evolution of the in-cloud microphysical processes and c.) to use the in-situ data to evaluate the NCAR HCA.
The present thesis is structured as follows: chapter 2 describes all the necessary background knowledge based on peer-reviewed literature, which is essential for the understanding of cloud microphysical processes. The way that a weather radar operates and the various polarimetric radar variables are interpreted is also explained in this chapter. Chapter 3 describes the instrumentation, which was used for the data collection, and demonstrates the methodology, which was followed for the data analysis. Chapter 4 presents a paper that analyses the cloud microphysical properties and radar polarimetric features during the passage of a warm front across the UK. Chapter 5 is a paper focusing on the formation of particular ice particles (ice-lollies), which were observed during the warm front of chapter 4. Chapter 6 includes an analysis of the in-situ data, which is conducted in order to evaluate the NCAR HCA. We intend to submit this work to the Atmospheric Measurement Techniques (AMT) journal. Chapter 7 is the conclusion chapter providing a summary of the key conclusions of the overall work presented in this thesis.
CHAPTER TWO

LITERATURE REVIEW

In this chapter, we describe and explain some fundamental theories on the physics of the atmosphere, the microphysics of the clouds and the radar technology. This is important and essential in order to fully understand the content of the current thesis.

2.1 Cloud Microphysics

2.1.1 Types of Clouds and General Properties

There are twelve basic types of clouds, which form in the atmosphere, falling into four main categories: a.) the low-level (fog, cumulus, stratus, stratocumulus, nimbostratus), a.) the mid-level (altostratus, altocumulus), c.) the high-level (cirrus, cirrocumulus, cirrostratus) and d.) the vertically developed clouds (cumulus congestus, cumulonimbus). The present research focuses mainly on the investigation of mixed-phase clouds, which contain water in both ice and liquid phase. Thus, we mostly focus on low-level, mid-level and vertically developed clouds.

In order to understand the composition of the different types of clouds that this work focuses on, some of their characteristics are described below. **Stratus clouds** are observed between 0-2km height and their lifetime is usually long. Such clouds play an important role in the global radiation balance due to their large global area coverage (24%) and the high albedo that their tops demonstrate [e.g. Zuidema and Hartmann, 1995; Warren et al., 1998]. Fairly low vertical air motions (few cm/s) and modest Liquid Water Content (LWC) (0.4g m\(^{-3}\)) and ice number concentrations (up to 0.2L\(^{-1}\)) [Hobbs and Rangno, 1985] within stratus clouds cannot trigger significant precipitation. However, according to the measurements of Hobbs and Rangno [1985], **stratocumulus/nimbostratus** can be associated with surface precipitation as they present much larger ice number concentration (up to 80/40L\(^{-1}\).
and LWC (up to 1.6/2g m⁻³). Stratus clouds are capped by a temperature inversion, which is the reason that they cannot demonstrate further vertical development. Light downdrafts can occur due to the evaporation and cooling of the air at the top of the cloud, which is caused by the mixing of the dry air over the cloud top and the moist air of the upper regions of the cloud [Rogers and Yau, 1989; Crosier 2011; Houze, 2014]. **Cumulus clouds** are usually less than 1km thick, but they present much stronger downdrafts (-5m s⁻¹) and updrafts (2.5-7.5m s⁻¹), which can cause larger LWC (up to 2g m⁻³) [Gerber, 2000]. The sizes of the droplets, which are usually measured, are between 2-30μm and their concentration decreases with height due to the collision process that takes place along the updraft zones. Downdrafts can be triggered, as in stratus clouds, due to entrainment at cloud tops. The content of cumulus clouds can differ for different cloud origins. Maritime cumulus demonstrate smaller concentration of larger droplets, while continental cumulus usually present larger concentrations of smaller droplets [Rogers and Yau, 1989]. Under specific circumstances, cumulus can be further developed to **cumulonimbus clouds**, which are associated with thunderstorms. Cumulonimbus clouds can demonstrate high LWC (>2g m⁻³), ice number concentrations (>50L⁻¹) and updraft velocities (>10m s⁻¹) [Barker and Davies, 1992]. Regarding the mid-level clouds, which can be observed between 2-6km, **altostratus** and **altocumulus** clouds can demonstrate ice concentrations of up to 50L⁻¹ and LWC from <0.1 to 1.3g m⁻³ [Hobbs and Rangno, 1985]. The percentage of the earth surface, which is covered by these two cloud types, is estimated at 22% [Gedzelman, 1988; Heymsfield et al., 1991].
All of the cloud types, which were mentioned above, can fall into the category of mixed-phase clouds, as they can be observed at temperatures between 0 and -38°C, where water can remain in liquid phase as supercooled droplets. The ratio between solid and liquid particles depends on the ambient temperature, the ambient aerosol concentration/size and the age of the cloud [Rogers and Yau, 1989; Moss and Johnson, 1992; Cober, 2001; Korolev et al., 2003; Morrison et al., 2005; Wallace and Hobbs, 2006; Houze, 2014]. Figure 1 shows the dependence between the temperature and the probability of observing ice, liquid or both phases within a cloud.

2.1.2 Cloud Droplet Formation and Growth Processes

The formation of liquid cloud droplets starts when water molecules condense on a cloud condensation nucleus (CCN), which is usually around tenths of a micron in size. When CCNs are involved in the formation of cloud droplets, the process is called heterogeneous nucleation. Condensation nuclei, cloud drops and raindrops present different average sizes, concentrations and terminal velocities in the clouds (Figure 2). Cloud droplets can potentially form homogeneously.
During the **homogeneous nucleation**, a droplet forms from random vapor collisions in high supersaturation environments (300-400%), where no CCNs are present. As supersaturation in the atmosphere rarely exceeds 1%, this process seems to be theoretical and plays no role in real clouds [e.g. Rogers and Yau, 1989, Pruppacher and Klett, 2012, Houze, 2014].

![Figure 2](image_url)

**Figure 2.** Comparative sizes, concentrations, and terminal velocities of some of the particles included in cloud and precipitation properties [Adapted from McDonald, 1958].

The **Köhler curve** explains the variation of the relative humidity over a solution droplet as a function of its radius, which plays an important role in the droplet growth. The Köhler curve is different for droplets containing different solutes and is calculated by the following equation (1):

\[
\frac{e}{e_s} = \left[ \exp \frac{-2\sigma}{n k_B T R} \left[ 1 + \frac{i m M_w}{M_s \left( \frac{4}{3} \pi R^3 \rho - m \right)} \right] \right]^{-1}
\]

where \(e\) is the vapor pressure over a droplet surface (in Pa), \(e_s\) is the saturation vapor pressure over a flat water surface (in Pa), \(\sigma\) is the surface tension (in N m\(^{-1}\)), \(n\) is the number concentration of water molecules (mol m\(^{-3}\)), \(k_B\) is the Boltzmann constant (J K\(^{-1}\) molecule\(^{-1}\)), \(T\) is the temperature of the system droplet-ambient air (in K), \(R\) is the droplet radius (in m), \(i\) is the degree of ionic dissociation, \(m\) is the
mass of the solute (in kg), $M_w$ is the molecular weight of water (kg kmol$^{-1}$), $M_s$ is the molecular weight of the solute (kg kmol$^{-1}$), and $\rho$ is the density of the water (kg m$^{-3}$).

Figure 3. Saturation ratio as a function of the radius of a droplet formed on a CCN of ammonium sulphate (mass=10$^{-6}$g). The upper dashed line presents the Kelvin equation and the bottom dashed line depicts the Raoult law. The solid line represents the Köhler curve. The term is related to the molar weight of the water, the term $b$ refers to the characteristics of the solute (CCN), $S^*$ is the critical saturation ratio and $r^*$ is the critical radius ($R_c$) [Adapted from Rogers and Yau, 1989].

There is a certain value of the ratio $e/e_s$, which is called critical saturation ratio ($S_c$), at which an embryonic droplet attain a certain radius (critical radius, $R_c$) due to the chance collision of water molecules. After attaining this size, the droplet can continue grow spontaneously by diffusion. The Köhler curve of ammonium sulphate, which shows the change of supersaturation over the solution droplet ($S_{c, adj}$) against to its radius, is illustrated in figure 3. Explaining further the Köhler equation, if a particle, which can act as CCN (e.g. NaCl, (NH$_4$)$_2$SO$_4$ etc.), is placed in a sample of air with supersaturation ratio $S_{amb}$ larger than its $S_{c, adj}$ (over the particle), a solution droplet will form due to condensation. As $S_{c, adj}$ will always be
lower than $S_{\text{amb}}$, the droplet will continue growing passing over $R_c$, which corresponds to $S_{c,\text{adj}}$. As a result, the droplet is assumed as **activated** and can grow further via diffusion. In contrast, in cases that $S_{\text{amb}} < S_{c,\text{adj}}$, the particle will grow as a solution droplet until $S_{\text{amb}} = S_{c,\text{adj}}$. Then, if the droplet should either grow or evaporate slightly, $S_{c,\text{adj}}$ would increase above or decrease below $S_{\text{amb}}$ preserving, eventually, its size [Wallace and Hobbs, 2006].

The **equation 1** is a combination of two different equations, the **Kelvin equation** and the **Raoult’s law**. The critical radius ($R_c$) can be calculated by the **Kelvin equation** (relation 2a), which is the first part of the **equation 1**. The Kelvin equation dominates the diffusional growth process after the activation of the droplet, as the droplet size increases significantly and the droplet curvature decreases:

$$\ln\left(\frac{e}{e_s}\right) = \frac{2\sigma V_m}{R_c RT} \quad (2a) \quad \Leftrightarrow \quad R_c = \frac{2\sigma}{n k_B T \ln(e/e_s)} \quad (2b)$$

where $V_m$ is the molar volume, $R=k_B \cdot N_A$ is the universal gas constant, $N_A=V_m \cdot n$ is the Avogadro number. According to **equation (2b)**, a cloud droplet cannot form if $e/e_s \to 1$, as $R_c \to \infty$, and $R_c$ becomes positive, if air is supersaturated ($e/e_s > 1$). Furthermore, for higher supersaturations, the size that a drop needs to exceed ($R_c$) in order to be activated becomes smaller. Although the temperature factor ($T$) appears in the denominator of the **equation (2b)**, when it takes values that correspond to the real atmosphere, it does not significantly affect $R_c$ [e.g. Pruppacher and Klett, 2012; Houze, 2014].

The second part of **equation 1** comes from the **Raoult’s law**, which describes the way that the chemical composition of the solute (thus the CCN) can affect the growth of a droplet. According to this law, the diffusional growth depends on the volume of the droplet, the volatility and the mass of the solute [e.g. Rogers and Yau, 1989; Wallace and Hobbs, 2006; Pruppacher and Klett, 2012].

After the formation of a small cloud droplet, an unstable cloud environment can lead this droplet to grow further. An unstable environment can be achieved when particles in the cloud demonstrate various sizes. In this case, large drops grow against smaller ones by vapor diffusion, as the vapor pressure is lower over larger
drops and higher over smaller droplets. Subsequently, as a water drop becomes larger, the probability of colliding (collision efficiency) with smaller water droplets, and growing further, increases. However, it is not necessary that all of the collided water droplets will be merged. The droplets can temporarily merge and, then, separate maintaining their characteristics, or they can merge maintaining their characteristics, or, even, they can merge and, then, break into several smaller droplets. The coalescence efficiency increases for larger size differences between the collector and the collected droplet. When collector and collected drops present similar sizes, the possibility of some air to be trapped between them is higher, which can prevent them from coalescence [Wallace and Hobbs, 2006]. In addition, similar sized droplets exhibit smaller terminal velocity difference. In this case, the collision cannot overcome the surface tension. In case that the two drops coalesce, the growth can be calculated by the continuous model:

\[
\frac{dm_1}{dt} = E_c \pi (\alpha_1 + \alpha_2)^2 (U_{\infty,1} - U_{\infty,2}) w_L
\]

where \(m_1\) is mass of the collector drop, \(E_c\) is the collision efficiency between particles with radius \(\alpha_1\) and \(\alpha_2\) and critical distance \(y\) between the centre fall line of the droplet that makes just a touching with the collector drop and the centre of the collector drop (\(E_c = y^2/(r_1 + r_2)^2\)), \(U_{\infty,1}\) and \(U_{\infty,2}\) the fall speed of the two droplets and \(w_L\) is the ambient liquid water content [Pruppacher and Klett, 2012].

**Terminal velocity (\(V_{t,\text{drop}}\))** of the cloud particles is the result of the effect of drag and gravity force and is, generally, calculated by the following equation:

\[
V_{t,\text{drop}} = \left( \frac{2 \cdot m \cdot g}{\rho_a \cdot A \cdot C_D} \right)^{1/2}
\]

where \(m\) is particle mass, \(g\) is gravitational acceleration, \(\rho_a\) is the air density, \(A\) is the particle’s area projected normal to the flow and \(C_D\) is the drag coefficient [Mitchell, 1996]. It should be noted that the terminal fall speed of a droplet <100\(\mu\)m is <1m s\(^{-1}\), but it can increase to \(~12\) m s\(^{-1}\) for drops with diameter >4mm [Beard and Pruppacher, 1969; Beard, 1976]. Collision and coalescence process explains the precipitation formation in warm clouds, where there is no ice.
2.1.3 Ice Crystal Formation and Growth Processes

Ice nucleation mechanisms fall into two main categories, the **homogeneous** and **heterogeneous** processes.

The **homogeneous ice nucleation** is divided into **homogeneous deposition** and **homogeneous freezing**. During the **homogeneous deposition**, an ice particle can form after random collisions of water vapors. However, molecules in the vapor phase can aggregate forming an ice embryo only at temperatures $<-65^\circ\text{C}$ and supersaturations $\sim 1000\%$. As such high supersaturation regimes cannot be observed in the atmosphere, this type of homogeneous process cannot be involved in the formation of natural ice clouds. **Homogeneous freezing** occurs when molecules of a pure liquid supercooled droplet obtain an ice-like structure at temperatures $<-38^\circ\text{C}$ and freezes [e.g. Rogers and Yau, 1989, Pruppacher and Klett, 2012; Houze, 2014; Liou and Yang, 2016].

![Figure 4](image.png)

**Figure 4.** Schematic representation of **a.** heterogeneous ice nucleation and **b.** secondary ice production mechanisms [Adapted from Houze, 2014].
In **heterogeneous ice nucleation**, water freezes at temperatures >-38°C due to the existence of **ice nuclei**, which are particles that demonstrate an ice-like structure. Such particles can be mineral dust and biological particles [DeMott et al., 2010]. There are four different modes of heterogeneous nucleation (figure 4a). The first is **deposition** and occurs when ice forms directly from water vapors deposited on an ice nucleus. The second is **condensation followed by freezing**, during which a nucleus acts as CCN (hence, firstly a liquid droplet forms, which then freezes). The third is **immersion** occurring when a nucleus penetrates a supercooled droplet, which then freezes. The last process is **nucleation by contact**, which occurs when a liquid drop freezes being contacted by an ice nucleus.

Although ice nuclei play an important role in ice formation in clouds, they are often observed in much smaller concentrations (up to an order of 10⁴) than these of ice particles [Hobbs, 2010]. This fact triggered the theory of **ice multiplication/secondary ice production** mechanisms in the clouds (figure 4b). The four main processes that are considered to cause ice multiplication are a. **collisional fracture**, b. **evaporative fracture**, c. **freezing and shattering of water drops** and d. **rime splintering** (known as Hallett-Mossop process). The **collisional fracture** can produce ice fragments due to the collisions between ice crystals. Larger ice crystals fall with larger terminal velocities increasing the collision efficiency and, thus, the potential for fragmentation [Hobbs and Farber, 1972; Vardiman, 1978]. **Evaporative fracture** occurs at relative humidity <90%, when ice crystals break up due to evaporation [Dong et al., 1979]. **Freezing and shattering of water drops** occurs when a supercooled water drop freezes and breaks into several fragments [Johnson and Hallett, 1968; Takahashi and Yamashita, 1969; Pruppacher and Schlamp, 1975]. Although the theories mentioned above describe the multiplication of ice, they cannot explain the large observed difference between ice nuclei and ice crystal concentrations. The **rime splintering** or **Hallett-Mossop (H-M)** process is considered as the main mechanism that can explain large ice particle concentrations in the clouds [Mossop and Hallett, 1974; Mossop, 1976; Mossop, 1978; Choularton 1978; Choularton, 1980]. The leading theory describing the H-M process [Choularton, 1980] suggests
that when supercooled droplets, with diameter >24μm, come into contact with pre-rimed ice particles (Mossop [1978] had suggested that droplets of diameter <13μm are also necessary to be present), they freeze in an explosive way. This explosive freezing occurs after the formation of an ice crust on the surface of a supercooled water droplet and the subsequent freezing of the subsurface water. The latter stresses and, eventually, fractures the crust. The fracturing of the surface crust is thought to result in the generation of multiple small ice splinters (Mossop [1985] suggests that ~300 ice splinters per milligram of accreted rime can be produced).

As the H-M process operates at temperatures between -3°C and -8°C (with a peak at -5°C), the ejected ice splinters typically grow into columnar ice particles. Generally, secondary ice production can potentially result in the glaciation of a cloud, as instant freezing occurs whenever a supercooled droplet comes into contact with ice at temperatures <0°C.

At first stages, ice grows via diffusion. This is explained by the following equation, which gives the ratio between the saturation vapor pressure over liquid water $e_s$ and ice $e_i$ at a given temperature ($T$):

$$\frac{e_s(T)}{e_i(T)} \approx \exp\left\{\frac{L_f}{R_v 273} \left(\frac{273}{T} - 1\right)\right\}$$

(5)

where $L_f$ is the latent heat of fusion of water ($3.34 \times 10^5$ J kg$^{-1}$) and $R_v$ is the gas constant for water vapour (461 J K$^{-1}$ kg$^{-1}$). According to equation 5, the vapor pressure over ice is always smaller than the vapor pressure over water drop surface (figure 5a). Moreover, as figure 5b clearly shows, the difference between the vapor pressure over ice and water surface presents a maximum (~0.25mb) between -10°C and -15°C. Because of this difference, ice can grow against water droplets in mixed-phase clouds (clouds where different water phases co-exist) via diffusion and deposition exhibiting a maximum growth rate at such temperatures [Lamb and Hobbs, 1971; Bohren and Albrecht, 2000]. The process described above, where ice particles grow faster than droplets, is called Bergeron-Findeisen process [Bergeron, 1935].

The shape, which ice can grow into, depends on the temperature and supersaturation regime. Thus, columnar crystals form between -3°C and -10°C and
planar crystals between -10°C and -22°C. It should be highlighted that dendrites can form in the planar temperature region, but only in high supersaturation regime [Kobayashi, 1961; Magono and Lee, 1966; Libbrecht, 2005; Bailey and Hallett, 2009]. Bailey and Hallett [2009] suggest that in the temperature region below -20°C, the shape of ice crystals becomes polycrystalline with assemblages of columns, plates and rosettes.

Figure 5. a. Saturation vapor pressure (mb) as a function of temperature for liquid water (solid line) and ice (dashed line). b. The difference between the saturation vapor pressures (mb) for water and ice (e_s - e_a) as a function of temperature. [Adapted by Bohren and Albrecht, 2000].

Ice particles in clouds can grow further via aggregation and/or riming. Aggregation is the process that two or more ice crystals collide and stick together. During the riming (accretion) process, supercooled cloud droplets are contacted by a falling ice particle and freeze instantaneously sticking to the ice particle. Both processes depend on the ice habit, size, terminal velocity and the collision efficiency [Wang, 1983; Pruppacher and Klett, 2012]. Terminal falling velocities (V_t,ice) of ice have been estimated by Mitchel [1996]:

\[
V_{t,\text{ice}} = a_{Re} v \left( \frac{2ag}{\rho_a v^2 \gamma} \right)^b D_{b(\beta+2-\sigma)-1} \quad (6)
\]
where \( \text{Re} \) and \( b \) are constants depending on the flow around the ice particle (Reynolds number), \( \nu \) is kinematic viscosity (in kg m\(^{-1}\) s\(^{-1}\)), \( g \) is the gravitational acceleration (m s\(^{-2}\)), \( D \) is the maximum dimension of the ice particle (in m), \( \rho_a \) is the air density (in kg m\(^{-3}\)), \( \alpha, \beta \) and \( \gamma, \sigma \) are constants depending on the mass and the projected area of the ice particle. Typically, large ice columns (\( D > 2 \text{mm} \)) present the highest fall speed (\( > 1 \text{m s}^{-1} \)) from the unrimed pristine ice crystals, while large lump graupels (\( > 2 \text{mm} \)) can exceed \( 2 \text{m s}^{-1} \) and large hailstones (\( > 20 \text{mm} \)) can present fallspeeds \( > 15 \text{m s}^{-1} \). Ice dendrites usually demonstrate the lowest fall speeds (\( < 1 \text{m s}^{-1} \)), as they face stronger drag force [Mitchel, 1996].

![Figure 6](image_url)

**Figure 6.** Habit diagram for atmospheric ice crystals depending on temperature and ambient supersaturation [Adapted from Libbrecht, 2005].

As previously mentioned, the probability of two particles to collide is called **collision efficiency** (see equation 3). The collision efficiency between a pristine ice crystal of a certain size and a liquid droplet becomes larger for lager droplets due to the difference between their terminal velocities (an example for ice plates is shown in figure 7). For very small droplets or droplets that exhibit similar terminal velocities to ice crystals, the collision efficiency is close to zero. In the first case,
small droplets demonstrate small inertia relative to the strength of the hydrodynamic drag force and they move around the ice crystal. Furthermore, very small ice (diameter <50-100μm) do not present large collision efficiencies, because they need to grow further by diffusion in order be able to collect water droplets [Bohm, 1992; Wang, 2002]. Finally, Schlamp et al. [1975] and Bohm [1992, 1994] showed that for ice columns with diameters >200μm and lengths >1000μm and droplet terminal velocities larger than ice column velocities, the collision efficiency increases (to more than unity) with increasing droplet size due to the wake effect of the droplet on the ice crystal.

![Collision efficiencies between hexagonal ice plates and supercooled droplets](Adapted from Wang and Wusheng, 2000).

**Figure 7.** Collision efficiencies between hexagonal ice plates and supercooled droplets [Adapted from Wang and Wusheng, 2000].

### 2.2 Radar and the Radiation Physics Background

The term “Radar” was introduced as an acronym by the US navy and it stands for “Radio Detecting And Ranging”. In order to understand the way that radars
operate, the fundamental radar principles and radiation physics are described in the next subsections. Radar measurements are based on the way that objects reflect or scatter the electromagnetic waves that a radar emits. Firstly, the receiving of a reflected signal by the radar is an indication of the existence of an object in the atmosphere. The wavelength of the signal is important as it determines what size of objects (e.g. aircrafts, birds, precipitation, cloud particles etc.) are visible by a radar and how penetrative the signal can be. As a general rule, smaller wavelengths are used for the detection of smaller particles, but attenuation is much more significant. An important advantage of radars compared to passive remote sensors is that they can receive signals independently of the existence of the daylight. In this chapter, the information about the radar operation principles mainly comes from Fukao et al. [2014] and Houze [2014].

2.2.1 The Radar Bands

Meteorological radars operate in the following bands: UHF (0.3-1GHz), L (1-2 GHz), S (2-4 GHz), C (4-8 GHz), X (8-12 GHz) and K (12-18 or 27-40 GHz) bands (table 1). Some radars use UHF band, which is less affected by rain and clouds, in order to obtain the wind profile of the atmosphere. L-band radars can be used for clear air turbulence studies. S-band radars usually come with a very large-diameter antenna (sometimes the dish can exceed 7 meters), but they have the significant advantage of small attenuation. Their range exceeds 100km and they are used for the detection of precipitation. C-band radars face moderate attenuation (coming with smaller antennas), while X-band radars are used for the detection of small cloud particles, but the attenuation is significant and, subsequently, their range is limited (~75km). This last band can be used for the detection of tiny particles and light precipitation, because of its small wavelength. X-band radars are usually portable (on vehicles or airplanes). Finally, K-band is mainly used by the Global Precipitation Measurements (GPM) system for taking measurements of cloud properties from the space via a system of satellites. This band splits down into Ku (12-18GHz) and Ka bands (27-40 GHz) due to strong absorption in water vapor for frequencies 18-27GHz.
Table 1. Radar bands, frequencies (f) and wavelengths (λ) that are used for atmospheric studies.

<table>
<thead>
<tr>
<th>Band</th>
<th>UHF</th>
<th>L</th>
<th>S</th>
<th>C</th>
<th>X</th>
<th>Ku</th>
<th>Ka</th>
</tr>
</thead>
<tbody>
<tr>
<td>f (Hz)</td>
<td>0.3-1</td>
<td>1-2</td>
<td>2-4</td>
<td>4-8</td>
<td>8-12</td>
<td>12-18</td>
<td>27-40</td>
</tr>
<tr>
<td>λ (cm)</td>
<td>30-100</td>
<td>15-30</td>
<td>7.5-15</td>
<td>3.75-7.5</td>
<td>2.5-3.75</td>
<td>1.11-1.67</td>
<td>0.75-1.11</td>
</tr>
</tbody>
</table>

Meteorological radars use an antenna, which is a parabolic reflector. The radar energy focuses onto the reflector, which then can direct the light waves in a concentrated beam. The diameter of the radar antenna (Dₐ) plays an important role in the beam width θ depending on the wavelength λ (Fukao et al., 2004). In general, larger Dₐ allows narrower beam width:

\[ \theta = 1.22 \frac{\lambda}{D_a} \text{ [rad]} \]  

As a matter of fact, the wavelength determines the effectiveness of the antenna as it is a factor that is involved in both the size of the antenna and the antenna aperture (see parameter A in eq. 13). Another characteristic of radars is the pulse length or width, which is the time period that each pulse lasts and should be long enough to ensure that sufficient energy will be emitted and reflected back by the scatterer.

2.2.2 Identifying the Position of Hydrometeors

For the identification of the position of a detected particle in the atmosphere, we need to retrieve its range (from the radar) and height from the earth surface. In general, radars have the ability to scan the atmosphere at various angles θₑ. The electromagnetic energy, which is emitted by the radars, travels through the air at a constant speed (which is almost the speed of light \( \sim 3 \cdot 10^8 \text{ m s}^{-1} \)). Thus, the distance (or range \( R \)) between the radar and the target is calculated for given speed \( c_0 \) and measured running time \( t \) by equation 8:

\[ \frac{R}{2} = \frac{c_0 t}{2} \]
It should be noted that the amount $C_0 \cdot t$ is divided by 2, because $t$ is the time, in which the wave travels to the target and returns back to the radar. In order to calculate height $h$, it is necessary to take into account the curvature of the earth and the refractive index of the air, which depends on the atmospheric pressure, temperature and humidity [Owens, 1976]. For given scanning angle $\theta_e$, the height of the signal beam can be calculated as a function of the radar range $R$:

$$ h = \sqrt{R^2 + (k_e r_e)^2 + 2 R k_e r_e \sin \theta_e - k_e r_e + h_a} $$  \hspace{1cm} (9)

where $h_a$ is the antenna height, $r_e$ is the earth radius and $k_e$ is a multiplier dependent on the vertical gradient of the refractive index of the air (for standard atmosphere it is $4/3$). Then, the projected distance $D$ on the earth surface of the range $R$ is calculated as a function of the the earth radius ($r_e = 6371$km) and the angle $\phi$ of the arc of the distance $D$:

$$ D = \phi \cdot r_e $$  \hspace{1cm} (10)

where $r_e$ is the earth radius (=6371km) and $\phi$ is the angle of the arc (in radians) of the distance $D$.

Figure 8. Representation of the space location of parameters used in relations (7), (8) and (9).
A radar dataset usually needs to be transformed from $\theta_e$ - R system to the Cartesian system. In this case, the range can be calculated as a function of height $h$ and angle $\varphi$, which is given by equation (11):

$$R = \sqrt{\alpha^2 + b^2} \quad (11)$$

where $\alpha = (r_e + h) \cdot \cos(\varphi) - r_e$ and $b = (r_e + h) \cdot \sin(\varphi)$. Figure 8 is a schematic representation of the parameters used in equations (9), (10) and (11). Figure 9 shows the beam position in the Cartesian system (height-projected distance on earth) for different scanning angles ($\theta_e$). According to this graph, no data can be collected close to the earth surface for distances $>\sim 30$km.

![Figure 9](image.png)

**Figure 9.** Height and distance on earth surface of radar beam in various scanning angles $\theta_e$ calculated from relations (7) and (8). The point (0,0) is considered as the radar location.

### 2.2.3 The Radar Equation

We assume that radar is a transmitter that emits a signal in all directions being, in fact, an isotropic radiator. Areas with the same amount of radiation are determined by spherical surfaces around the radar, which is the centre of all these spheres. Consequently, the signal spreads out on a larger surface, as the range increases. Hence, there is a negative correlation between the power density of the radiation $P$ and the range $R$, as $P = P_t/4\pi R^2$. For the reflected radiation by a scatterer, the power $P_{rf}$ is:

$$P_{rf} = \frac{P_t G \sigma}{4 \pi R^2} \text{[W m}^{-2}] \quad (12)$$
where $P_t$ is the power transmitted by the radar, $R$ is the distance between the radar and the target, $G$ is antenna gain (a parameter which describes the directivity and efficiency of the radar antenna) and $\sigma$ is the cross section of the scatterer (scatterer’s ability to reflect the power received in the propagation direction). The power received by the radar $P_{rc}$ can be calculated by the following equation:

$$P_{rc} = \frac{P_t A}{4 \pi R^2} \quad (13)$$

where effective antenna aperture $A = G\lambda^2/4\pi$ determines how effective is an antenna at receiving power depending on the wavelength of the signal $\lambda$. According to equations (12) and (13), the final form of equation (13), which is also called the radar equation, is:

$$P_{rc} = \frac{P_t G^2 \lambda^2}{(4\pi)^3 R^4} \sigma \quad (14)$$

The effective maximum range $R_{\text{max}}$ that a radar can operate is defined in equation (14) or for a given pulse repetition frequency (PRF) in the following equation:

$$R_{\text{max}} = \frac{C_0}{2 \text{PRF}} \quad [\text{Hz}] \quad (15)$$

where PRF is the Pulse Repetition Frequency, which is the frequency that the radar transmits a pulse. This is a very important parameter, which is used to avoid the so-called range folding error. High PRF values limit the range at which a radar can detect a target. When a radar receives the backscattered signal of the first pulse after emitting the second one, the distance calculation is based on the time difference between the transition of the second signal and the reception of the first one. As a result, there is an underestimation of the distance of a particle or a range ambiguity in case there are targets at both $R$ and $R+R_{\text{max}}$. 
2.2.4 Radiation Physics Background

The understanding of the way that different particles in the atmosphere interact with the radiation emitted by a radar is of high importance in order to interpret radar datasets.

The radar operation is based on the radiation scattering effect. Scattering is the secondary emission of electromagnetic radiation from the electrons of a particle, which are oscillated as they are affected by the incident radar radiation. It should be highlighted that the wavelength $\lambda$ of the signal remains the same after scattering. There are three different scattering regimes depending on the diameter of the size of the particles, which cause the scattering effect: a.) the Rayleigh, b.) the Mie and c.) the optical scattering. The radar operation is based mainly on the Rayleigh, but also on the Mie scattering. The Rayleigh scattering occurs when $\pi D / \lambda << 1$ (where $D$ is the particle diameter). In this scattering mode, the scattered radiation intensity $I$ reaches its maximum $I_{\text{max}}$ in the propagation and the opposite direction (figure 10). In contrast, the radiation with the minimum intensity ($I_{\text{min}} \sim I_{\text{max}}/2$) is scattered in the direction vertical to the propagation path. The Mie scattering occurs for particles that $\pi D / \lambda \sim 1$. In this regime, high intensities are observed in the forward scattering direction, while the minimum intensities are noted in the backscattering direction. The maximum intensity is even larger for larger particles. Practically, when Mie scattering occurs, two different waves are reflected back (figure 10). The one wave comes from the reflection from the near side of the particle and the other from the back side of the particle after penetrating it. Mie scattering occurs when the one radiation interferes with the other, causing augmentation or decline [Robinson, 1966; Kumjian and Ryzhkov, 2010; Fukao et al., 2004]. Figure 11 illustrates the relationship between the cross section of a particle and its size distinguishing the different scattering regions.
Generally, precipitation and cloud particles are assumed to belong to the Rayleigh region for low frequency radars (e.g. S-band). As previously stated, the radar cross section $\sigma$, which is used in the radar equation (equation 14), determines the ability of a particle to scatter radiation in the direction that the incident radiation was previously propagated. For a single particle with diameter $D$ in the Rayleigh region, $\sigma$ is given by:

$$\sigma = \frac{\pi^5}{\lambda^4} |K|^2 D^6 \ [m^2] \quad (16)$$

where $K$ is the complex dielectric factor, which depends on the density of a particle [Doviak et al, 1979]. The increased $K$ for water ($K=0.96$) implies that water has the ability to conserve the electric field in contrast to ice ($K=0.44$), in which there are losses. For constant $\lambda$, large liquid particles are able to scatter back more power than smaller ice particles. In addition, the dependence of $\sigma$ on $\lambda^4$ explains the reason that shorter wavelengths demonstrate greater sensitivity to more weakly reflecting targets.
Figure 11. Normalized radar cross section of a spherical particle as a function of particle diameter (which is presented as a function of the wavelength $\lambda$) [Adapted from Skolnik, 1990].

2.2.5 The Radar Parameters

Single polarisation radars transmit a single horizontally oscillated signal deriving only the parameters of reflectivity factor and Doppler velocity. Although these parameters are very useful for the understanding of the concentration, size and movement velocity of the particles, they cannot provide any information about the shape, the orientation and the bulk density of the hydrometeors. Such properties can be obtained only with the use of dual-polarisation radars, which are able to transmit and receive both horizontally and vertically oscillated pulses [e.g. Vivekanandan et al, 1999; Naud et al, 2004]. In this section, all the radar parameters that can be derived by a dual-polarised radar are briefly described.

2.2.5.1 The Reflectivity Factor ($Z_H$)

The first parameter, which is derived by both single and dual-polarised radars, is the reflectivity factor ($Z_H$). It is related with the power that is scattered by a target, in the atmosphere, back to the radar. Thus, $Z_H$ is proportional to the cross section $\sigma$
and, therefore, it is proportional to the sixth power of the diameter of the particle in a sample volume:

\[
Z_H = \frac{1}{\Delta V} \sum_i D_i^6 [mm^6 m^{-3}] \quad (17)
\]

where \(\Delta V\) is an elemental volume that contains \(i\) scatterers and \(D_i\) is the water droplet diameter of the particles [e.g. Doviak et al, 1979; Bringi and Chandrasekar, 2001; Fukao et al., 2004]. Equation 17 shows that larger particles contribute much more to \(Z_H\) rather than smaller particles. Hence, large snow aggregates and hail are represented by larger \(Z_H\) than single pristine ice crystals and small rain droplets. As \(Z_H\) is measured in \([mm^6 m^{-3}]\), it can vary from 0.001 for fog/clouds to 50000000 \(mm^6 m^{-3}\) for large hail. The following transformation to \([dBZ]\) offers a more efficient way of estimating and understanding this parameter, because the \(Z_H\) range becomes narrower (from -30 to 80dBZ) [e.g. Leitao and Watson, 1984; Golestani et al., 1989; Balakrishnan and Zrnic 1990; Ryzhkov and Zrnic 1998; Straka et al., 2000; Kumjian 2013a]:

\[
Z_{dBZ} = 10 \log_{10}(Z_{mm^6 m^{-3}}) \quad (18)
\]

As \(Z_H\) is proportional to the cross section of a particle, it also depends on the dielectric constant \(K\) (equation 16). This means that higher \(Z_H\) is expected for liquid (denser) rather than for ice particles of the same size. A brief presentation of \(Z_H\) thresholds for various atmospheric hydrometeors is shown in figure 12a.

![Figure 12. S-band polarimetric thresholds for various hydrometeor types according to Straka et al. [2000].](image_url)
2.2.5.2 The radial velocity ($V_{\text{rad}}$)

Another parameter that can be retrieved by both single and dual polarisation radars is the radial velocity ($V_R$). This parameter expresses the component of the particle velocity, which acts towards or away of the radar and is based on the Doppler effect firstly proposed by C.J. Doppler in 1842. According to this effect, the frequency of the backscattered radiation increases (comparing to the radiation transmitted by the radar) when the target moves towards the radar and vice versa. $V_R$ is calculated as a function of the wavelength $\lambda$:

$$V_R = -\frac{\lambda f_{\text{shift}}}{2}$$

(19)

where $f_{\text{shift}}$ is the frequency shift between two successive and consecutive pulses [e.g Fukao et al., 2004; Brown and Wood, 2007]. Hence, $V_R$ is negative when frequency increases and a particle moves towards the radar, and is positive when the particle moves away from the radar (figure 13). As a radar can only identify movements along the direction of transition, the radial velocity is not necessarily the actual wind speed. Thus, $V_R$ can be calculated for given radar scanning azimuth $\theta_R$, actual wind speed $V_{\text{act}}$ and actual wind direction $\theta_{\text{act}}$:

$$V_R = V_{\text{act}} \cos(\theta_{\text{act}} - \theta_R)$$

(20)

According to equation 20, $V_R = V_{\text{act}}$, when a particle moves along the propagation path. However, when the flow is perpendicular to the propagation path, the radar cannot detect any relative movement and $V_R = 0 \text{ m s}^{-1}$ (figure 11).

In order to obtain a correct representation of $V_R$ around the radar, it is necessary to be aware of the velocity folding error. This error occurs when a particle exceeds the maximum velocity $V_{R\text{max}}$ that a radar is able to measure. The phase difference of two successive consecutive pulses that a radar is able to detect is a number between -$\pi$ and $\pi$. As an illustration, when a particle moves faster than $V_{R\text{max}}$ towards the radar, the two consecutive pulses have a large negative phase difference exceeding -$\pi$. Thus, phase difference obtains slightly positive values, because of the limited range. As a result, the particle appears to move away from
the radar. $V_{R\text{max}}$ is proportional to the wavelength $\lambda$ and $PRF$ (see section 2.2.3) of the radar and it is also called **unambiguous velocity range**, which is given by:

$$V_{R\text{max}} = \frac{PRF \lambda}{4} \quad (21)$$

**Equations 15** and **21** highlight the so called, Doppler dilemma. According to that, a fair choice of $PRF$ to achieve large $V_{R\text{max}}$ is, simultaneously, a poor choice for a preferential large unambiguous velocity (and vice versa). Thus, according to **equation 21**, for increasing $PRF$, $V_{R\text{max}}$ increases. In this case, the effective maximum range $R_{\text{max}}$ decreases (see **equation 15**) causing range folding error (see **section 2.2.3**) [Brown and Wood, 2007]. However, for higher frequency radars (e.g. X-band), which face significant attenuation through precipitation and, hence, they are not expected to have great range in such cases, the choice of a fair PRF is an easier task. Velocity folding can be overcome with unfolding processes, which are based on the previously described principle [e.g. Bargan et al. 1980; Eilts et al., 1990; Jing et al., 1993; Liu et al., 2003; Brukx, 2015].

**Figure 13.** Representation of a wind field with constant speed of 26 m/s by vectors (left) and Doppler velocity (right) (Adapted from Brown and Wood, 2007).

### 2.2.5.2 The Differential Reflectivity ($Z_{DR}$)

As mentioned above, dual-polarisation radars can transmit two pulses, the oscillation direction of which is vertical to each other. Thus, dual polarisation radars are able to obtain reflectivity factor from both vertical ($Z_V$) and horizontal ($Z_H$) pulses. **Differential reflectivity** $Z_{DR}$ is calculated from the ratio $Z_H / Z_V$ and is
related with the size and axis ratio of hydrometeors. Seliga and Bringi [1976], firstly, introduced $Z_{DR}$, which is given by the following equation:

$$Z_{DR} = 10 \log (Z_H [dB]/Z_V [dB]) \ [dB] \quad (22)$$

where $Z_H$ and $Z_V$ are in linear units. $Z_{DR}$ is affected by size, shape, orientation, density and dielectric constant of hydrometeors, but it is independent of particle concentration and radar miscalibration. It should be highlighted that $Z_{DR}$ can be biased if any anisotropic beam obstacle interferes between the radar and the target. In the case of a large sized vertically aligned obstacle (e.g. a building), a significant amount of vertical polarization pulse is blocked and the horizontal pulse is stronger, causing strongly positively biased $Z_{DR}$ [e.g. Zrnic et al., 2006; Kumjian, 2013a].

Spherical particles, which demonstrate axis ratio ~1, are expected to scatter back similar power for both vertically and horizontally oscillated pulses. Hence, such particles are represented by $Z_{DR} \sim 0$dB. However, $Z_{DR}$ can be positive or negative for particles that their major axis is horizontally or vertically aligned respectively. Kumjian [2013a] mentions that $Z_{DR}$ is affected by the physical composition and/or density of the targets. As an illustration, a large and oblate liquid drop presents higher $Z_{DR}$ than a graupel particle of the same size and shape. Therefore, $Z_{DR}$ is used for the discrimination between different hydrometeor types (figure 12b) [e.g. Seliga and Bringi, 1978; Ryzhkov and Zrnic,1998; Straka et al., 2000; Kennedy and Rutledge, 2011; Andrić et al., 2013; Moisseev et al., 2015].

Referring to the liquid phase, $Z_{DR}$ can be a measure of the median drop size in a sample volume, because drop oblateness increases due to aerodynamic drag proportionally to the diameter (figure 14) [Pruppacher and Pitter, 1971; Wakimoto and Bringi, 1988]. Thus, for preferential reduction of small droplets (for a variety of reasons, such as evaporation), the median size within the sample volume increases and, eventually, $Z_{DR}$ also enhances [Kumjian and Ryzhkov, 2010].
Z$_{DR}$ has been a useful parameter for discriminating different types of ice crystals. Snow aggregates, rimed crystals and dry hail with spherical/quasi-spherical shape can be represented by Z$_{DR}$~0dB, while horizontally aligned anisotropic pristine ice crystals (especially planar crystals) are associated with 1dB < Z$_{DR}$ < 5.5dB. Negative Z$_{DR}$ can be observed for vertically aligned pristine ice crystals [Saunders, 1999 (figure 12b)]. The thresholds mentioned above are indicative and they may differ if various hydrometeor types coexist within the sample volume or even if the radar elevation angle is high [Wolde and Vali, 2001]. Although the discrimination between different ice habits is a fairly tough work, Z$_{DR}$ can also be used for such purpose. Hexagonal plates demonstrate Z$_{DR}$ >2dB (sometimes up to 10dB), while ice dendrites are usually represented by Z$_{DR}$~1-4dB [e.g. Seliga and Bringi, 1978; Hall et al, 1984; Ryzhkov and Zrnic,1998; Straka et al., 2000; Westbrook et al, 2010; Kennedy and Rutledge, 2011; Andrić et al., 2013; Thompson et al., 2014; Moisseev et al., 2015]. When ice crystals and hail move into a region with temperatures between 0-2°C (the so-called melting-layer), they start melting. In this case, ice is covered by a liquid water shell, which is not uniformly distributed around the ice surface (obtaining a shape similar to a large falling drop) due to drag force. As a result, Z$_{DR}$, but also Z$_{Hr}$ presents enhanced values within the melting layer and for this reason this layer is also called bright band [Leary and Houze, 1979; Stewart et al., 1984; Fabry and Zawadzki, 1995; Brandes and Ikeda, 2004; Giangrande et al., 2008; Boodoo et al., 2010; Kumjian, 2013b].
2.2.5.3 The Differential Phase Shift (Φ_{DP}) and Specific Differential Phase (K_{DP})

The **differential phase shift** Φ_{DP} expresses the phase shift difference between horizontal (Φ_{H}) and vertical polarization (Φ_{V}) [Sachidananda and Zrnic, 1986; Fuakao et al., 2014]:

\[
Φ_{DP} = Φ_{H} - Φ_{V} \quad [\text{deg}] \quad (23)
\]

For spherical particles Φ_{DP} \sim 0^\circ, while for oblate particles which cause smaller (larger) vertical than horizontal phase shift, Φ_{DP} is positive (negative). One of the differences between Φ_{DP} and Z_{DR} is that the first depends on particle concentration. As an illustration, low concentrations of large droplets affect less significantly this polarimetric variable than high concentrations of small droplets. In general, Φ_{DP} is not affected by attenuation, partial beam blockage or radar miscalibration. This is the reason why Φ_{DP} is used for attenuation correction and quantitative precipitation estimation [Kumjian, 2013a].

However, Φ_{DP} is cumulative along each radial direction and, thus, it is not used widely (figure 15) [Kumjian et al, 2013a]. For this reason, a much more effective way to use this variable is its range derivative. The range derivative of differential phase shift (Φ_{DP}) is called **specific differential phase** (K_{DP}):

\[
K_{DP} = \frac{Φ_{DP}(R_2) - Φ_{DP}(R_1)}{2(R_2 - R_1)} \quad [\text{deg km}^{-1}] \quad (24)
\]

where \(R_1\) and \(R_2\) refer to measurements at range 1 and range 2 from the radar (\(R_1 < R_2\)). Before any \(K_{DP}\) derivations, \(Φ_{DP}\) needs to be filtered as it is a fairly noisy variable [e.g. Hubbert and Bringi, 1995; Ryzhkov and Zrnic, 1996; Wang and Chandrasekar, 2009; Giangrande et al., 2013].

Similarly to Φ_{DP}, \(K_{DP}\) increases with increasing oblateness, size and concentration of hydrometeors. As \(K_{DP}\) is mainly affected by the presence of liquid water, it exhibits values > 0° km\(^{-1}\) for large liquid drops and wet graupel particles. However, hail, aggregated snow, rimed and pristine crystals demonstrate \(K_{DP} \sim 0^\circ\) km\(^{-1}\). In
cases that pristine ice crystals are observed in high concentrations, $K_{DP}$ can reach 0.6° km$^{-1}$ (Figure 10d) [e.g. Kennedy and Rutlegde, 2011; Straka et al, 2000; Kumjian, 2013a; Thompson et al., 2014].

![Figure 15. Range profile of $\Phi_{DP}$ and $K_{DP}$ showing the way that the two variables change for an area of heavy rain [Adapted from Kumjian, 2013a].](image)

2.2.5.4 The Co-polar Correlation Coefficient ($\rho_{HV}$)

The co-polar correlation coefficient $\rho_{HV}$ presents the degree that a scatterer in the sample volume contributes to the overall V- and H- polarised signals at zero lag (thus, removal of the time lag between vertical and horizontal pulses is needed) [Jameson and Mueller, 1985; Sachidananda and Zrnic, 1985; Kumjian, 2013a]. The co-polar correlation coefficient obtains values between 0 and 1:

$$\rho_{HV} = \frac{\langle |S^{VV}S^{HH}| \rangle}{\sqrt{\langle |S^{HH}|^2 \rangle \langle |S^{VV}|^2 \rangle}} \tag{25}$$

where $S^{HH}$ and $S^{VV}$ are the backscattering matrix elements (incident horizontal/scattered horizontal and incident vertical/scattered vertical...
correspondingly). Thus, in case that shapes, orientations and types of particles within the sampling volume significantly vary, $\rho_{HV}$ decreases. In addition, $\rho_{HV}$ presents decreased values for non-Rayleigh scatterers. It should be noted that various particle sizes cannot affect $\rho_{HV}$, unless their shape varies. $\rho_{HV}$ can also be very useful for the discrimination between meteorological and non-meteorological echoes (e.g. biological scatterers, tornadic debris, smoke plumes etc).

In particular, large $\rho_{HV}$ values (>0.95) are expected for liquid droplets, dry spherical hail, dry snow aggregates, rimed and pristine ice crystals. However, ice needles and plates can produce decreased $\rho_{HV}$ due to their non-spherical shape and non-preferential orientation. Large hailstones, especially when they are covered by liquid water, and wet snow also cause dramatic decrease in $\rho_{HV}$ (<0.95 or even <0.85) (figure 12e) due to their highly irregular shapes and large sizes, which may allow scattering in Mie regime. This is the reason that this polarimetric radar variable is used (in combination with others) for the identification of the melting layer [e.g. Brandes and Ikeda, 2004; Giangrande et al., 2008; Boodoo et al., 2010; Kumjian, 2013a].

### 2.2.5.5 Linear Depolarisation Ratio ($L_{DR}$)

Anisotropic and aspherical particles, which do not have any of their symmetrical axes parallel to the polarization of the emitted electromagnetic wave have the ability to depolarise the radar signal. As an example, such particles can receive a polarised wave (e.g. horizontal) and scatter back radiation in both vertical and horizontal polarisations. The power received at the same polarisation as the transmitted power (e.g. horizontal is transmitted and received) is called co-polar, while the power received at the opposite polarisation to the transmitted one (e.g. horizontal is transmitted, but both horizontal and vertical are received) is referred to as cross-polar [Tragl, 1990]. Linear Depolarization Ratio ($L_{DR}$) is the ratio of cross-polar to co-polar power and can be calculated as:

$$L_{DR} = 10 \log \frac{Z_{HV}}{Z_{HH}} \quad (26)$$
where $Z_{HV}$ and $Z_{HH}$ are the cross-polar and co-polar power respectively [Fukao et al., 2014]. According to Houze [2014] “$L_{DR}$ is a measure of how much of a horizontally transmitted signal is depolarised”. $L_{DR}$ is always negative because the cross-polar is much weaker than the co-polar power, thus the ratio $Z_{HV}/Z_{HH}$ does not exhibit values >1dB. In the ideal case of a spherical particle, which cannot depolarise a polarised signal, $L_{DR}$ obtains the value $-\infty$, because $Z_{HV}$ is zero. However, for anisotropic particles $L_{DR}$ obtains higher values. As an illustration, $L_{DR}$ can reach up to -20dB for graupel or small hail, -10dB for snow aggregates and 0 dB for hail. Although pristine ice crystals are assumed as anisotropic crystals, they are represented by relatively small $L_{DR}$ -30 to -27dB due to their small size, density (50-900kg m$^{-3}$) and dielectric constant (0.44) which generally produce lower contrasts among the polarimetric variables [Straka et al., 2000]. It should be noted that $L_{DR}$ depends on the angle of the symmetry axis of a particle. Thus, $L_{DR}$ may be increased even for a spherical particle, the symmetry axis of which is not perpendicular to either the vertical or horizontal polarisation directions of an incident beam [Matrosov et al., 1996; Meischner, 2005]. In **figure 12c**, indicative $L_{DR}$ values are presented for different hydrometeor types.

### 2.2.6 Hydrometeor Classification Algorithms (HCAs)

Various techniques have been developed and applied on weather radar data for distinguishing different types of hydrometeors in clouds. Such techniques are referred to as Hydrometeor Classification Algorithms (HCAs). The inputs used by these algorithms are usually a set of polarimetric variables, such as $Z_{H}, Z_{DR}, \rho_{HV}, \Phi_{DP}, K_{DP}$ and others.

The largest category is this of fuzzy-logic HCAs, which use theoretical/observational thresholds of radar polarimetric variables to group observations with similar characteristics into specific hydrometeor classes [Vivekanandan et al., 1999; Liu and Chandrasekar, 2000; Straka et al., 2000; Zrnic et al., 2000; Marzano et al., 2007; Park et al., 2009; Al-Sakka, 2013; Dolan et al., 2013]. For each radar polarimetric variable and class, these algorithms use
trapezoidal membership functions (MBF), which can determine the possibility of an object (which is a vector of polarimetric radar variables for a sample volume) to be classified to one of the pre-defined hydrometeor classes. Then, these MBF results are multiplied with a weighting factor, which is a degree of discriminating efficiency of each variable with respect to a particular class [Park et al., 2009]. Finally, the weighted results are summed for each hydrometeor class. The class with the largest value is considered to represent more effectively the class of the object. The fuzzy logic HCAs can be divided into these, which use temperature from radiosondes or weather model simulations [e.g. Zrnic et al., 2000] and these, which use the polarimetric signature of the melting layer (described in previous paragraphs) to estimate the height of the 0°C isotherm [e.g. Park et al., 2009]. We can also distinguish them between algorithms that use single-dimensional or dual/more-dimensional MBFs. It should be clarified that the class possibility can be calculated as a function of a single or two/more radar parameters each time for single- [e.g. Liu and Chandrasekar, 2000] or dual/more- dimensional MBFs [e.g. Zrnic et al., 2000] respectively. Although the techniques described above have been broadly used, it seems that they have some limitations due to the uncertainty of scattering properties, especially of ice crystals, and the fact that the choice of the number and the content of the hydrometeor classes is subjective [Tyynela et al., 2011; Grazioli et al., 2015].

Some of the latest HCAs make use of clustering method suggesting an alternative approach to the classification of hydrometeors using radar data [Bechini and Chandrasekar, 2015; Grazioli et al., 2015; Besic et al., 2016]. In general, cluster analysis is used to group a set of objects (here objects are multi-dimensional vectors, which contain a number of radar polarimetric variables for a determined sample volume) in such a way that objects in the same hydrometeor class (which is called cluster) demonstrate similar characteristics to each other than to those in other groups. It should be emphasized that the polarimetric radar variables are usually normalised in order to be comparable to each other during the clustering process. The scheme, which is proposed by Grazioli et al. [2005], applies the Agglomerative Hierarchical Clustering method [Ward, 1963] to objects $x=\{Z_H, Z_{DR}, K_{DP}, \rho_{HV}, \Delta z\}$, where $\Delta z$ is the difference between the altitude of i-th
resolution volume and the estimated, from ground stations, altitude of the 0°C isotherm. The optimum number of clusters is obtained applying on the dataset statistic indices (such as Accuracy Spread, Cohen’s kappa and SD index - see Grazioli et al., 2015), which determine the homogeneity within the clusters and the diversity between the different clusters. Then, each cluster is labeled with a hydrometeor class name according to its objects’ polarimetric signatures if it is a negative temperature cluster, or is used for calculation of the rain rate [Otto and Russchenberg, 2012] if it is a positive temperature class.

Clustering can be used in combination with the fuzzy-logic technique (e.g. Bechini and Chandrasekar [2015], Besic et al. [2016]). As an example, the algorithm suggested by Besic et al. (2016) uses k-medoids clustering [Park and Jun, 2009] combined with the Euclidian distance (straight-line distance) method on objects \( x = \{Z_H, Z_{DR}, K_{DP}, \rho_{HV}, \text{Ind} \} \), where \( \text{Ind} \) is a factor relevant to the 0°C isotherm altitude of i-th resolution volume, retrieved by a weather numerical model. The results from this clustering process are compared with the results from a fuzzy-logic algorithm based on Dolan et al. [2013] using Kolmogorov-Smirnoff test. A cluster is labeled with a hydrometeor class name, when it exhibits similar distribution to one of the fuzzy-logic classes. The algorithm runs multiple times for random Kolmogorov-Smirnov test parametrisation, and centroids for each class equivalent to the number of runs are produced. The dispersion of the centroids determines when a cluster is considered or not.

2.2.7 Linking Rain Rate with Radar Variables

Except for identifying the content of the clouds, dual-polarisation radars can also be useful in the estimation of the rain rate at the surface or in a cloud, which is within the range that a radar is able to transmit radiation.

Marshall and Palmer [1948] found that, broadly, smaller liquid droplets are observed in much higher concentrations than larger drops in a cloud (gamma
distribution) and suggested an exponential relationship between the number concentration $N(D)$ and the diameter of the liquid drops ($D$): 

$$N(D) = N_0 \exp(-\Lambda D) \quad (27)$$

where $N_0$ is a constant intercept parameter (usually it is $0.08 \text{ cm}^{-4}$) and $\Lambda$ is the slope of the distribution as a function of the rainfall rate $R \ (\Lambda = 41R^{-0.21})$ (Marshal and Palmer, 1948). Then, the rain rate for a given drop size distribution depends on the $3^{rd}$ power of the drop diameter:

$$R = 6 \times 10^{-3} \pi \int_{0}^{\infty} N(D) v(D) D^3 dD \quad (28)$$

where $v(D)$ is the raindrop fall speed [Atlas and Ulbrich, 1977].

Although liquid droplets of small sizes dominate the clouds, their contribution to the reflectivity factor ($Z_H$) is negligible. This occurs due to the proportionality of $Z_H$ to the $6^{th}$ power of the particle diameter (equation 17). For this reason, relationships based exclusively on $Z_H$ [Marshall et al., 1955; Sekhon and Srivastava, 1971; Jorgensen and Willis, 1982] may not provide an accurate rainfall rate estimation. On the contrary, for large droplets, $K_{DP}$ is proportional to the $4.24^{th}$ power of the particle diameter and, thus, can provide a better estimation of the rainfall rate [e.g. Ryzhkov and Zrnic, 1996; Brandes et al, 2001; Brandes et al, 2002; Teschl et al, 2008; Otto and Russchenberg, 2012]. Nevertheless, inaccurate rainfall rate estimations may arise for light rain (as $K_{DP}$ can be noisy and it is proportional to the $5.6^{th}$ power of the drop diameter), non-uniform beam filling and non-Rayleigh scatterers [Kumjian et al., 2010; Kumjian and Ryzhkov, 2010; Kumjian 2013a; Ryzhkov, 2007].
CHAPTER THREE
TOOLS AND METHODS

3.1 Tools

The work in the present thesis was based on the datasets provided from the Aerosol Properties, Processes And InfluenceS on the Earth’s climate (APPRAISE-Clouds) project, which took place in the UK. During this campaign both radar and airborne measurements of cloud properties were collected. For the collection of comparable datasets, the aircraft obtained measurements along the same azimuth relative to the Chilbolton Advanced Meteorological Radar (CAMRa). In order to obtain the accurate aircraft position relative to the CAMRa location, the haversine law was applied on the data retrieved from the GPS aircraft system. In this chapter, the instruments and techniques used for the analysis of the datasets are discussed.

3.1.1 The Bae-146 aircraft

The Facility for Airborne Atmospheric Measurements (FAAM) British Aerospace 146 (Bae-146) aircraft was used for the collection of in-situ data during the APPRAISE-Clouds project. The aircraft is able to fly for a range of ~3300km and up to altitudes of ~12km. This allows measurements within various types of clouds and meteorological conditions. The maximum aircraft speed is up to 160m s\(^{-1}\), but during measuring process is constantly ~100m s\(^{-1}\). All the instrumentation, which was used for the purposes of the present study, is attached on the aircraft wings and it will be described in the next paragraphs.
3.1.2 Optical Array Probes (OAPs)

In general, the Optical Array Probes (OAPs) are a useful tool for the investigation of cloud physics. They consist of a laser housed in one of the arms of the probe which emits radiation of defined wavelength towards a photodiode array housed in the other arm, which is an array of elements sensitive to the laser light. These elements record the degree that the laser beam intensity decreases when a particle passes through (thus the shadow of the particle). This information can be used in order to characterise the shape and the size of the particles [Knollenberg, 1981].

Since the introduction of OAPs, artefacts, which come from hydrometeors breaking and splashing when colliding with the probe tips, have been known that need to be removed from the data before any estimations/calculations are obtained. Hallett [1976] showed that ice particles of 100-200μm in diameter can shatter on probes producing dozens of fragments. Cooper [1977] proposed for 2D OAPs that the latter of two consecutively recorded particles should not be accepted if the spatial distance between them is less than 2.5cm. Field et al. [2003] noticed that inter-arrival times of particles are bimodal with modes at $10^{-2}$ and $10^{-4}$s. They explain that the physical origin of the bimodality is not clear, but if shattering is responsible for the bimodality the measured particle concentrations may be overestimated by a factor of 5 at most. Finally, Korolev and Field [2015] developed an Inter-arrival Time Algorithm (ITA) in order to exclude shattered ice from the data collected by an OAP. This algorithm is based on the assumption that the minimum inter-arrival time between two intact particles is longer than the maximum inter-arrival time of the shattered particles, which always pass through the sample volume as a group of at least two particles. The algorithm comes along with some limitations, such as the recording of a singleton, which is the only visible (for a variety of reasons) particle of a group of shattered particles, or partially viewed particles. However, they highlight that the effect of shattering in measurements is neglectable for clouds with particles less than 400μm in size.

Particles can be totally contained within the sampling area. However, another limitation of using OAPs arises from the fact that a significant amount of particles
partially appear in the sample area. Eliminating or including these particles in the measured size spectrum can either decrease the efficiency of the probe or the underestimation of the actual particle dimension and sampling volume. Two of the techniques, which are usually employed in 2D data, are the “entire-in” and the “centre-in” method. The “entire-in” method is used to reject any particles that touch one/both end diodes, while the “centre-in” keeps any particle, the centre of which is within the diode array and, hence its size is greater than x and/or y dimensions of the sampling volume [Heymsfield and Parrish, 1978]. Figure 16 shows a comparison of the sample area of the “centre-in” and “entire-in” method calculated according to Heymsfield and Parrish [1978] (equation 30).

![Graph showing sample area comparison](image)

**Figure 16.** Sample area calculated for the 2D-S probe for the “Centre-In” and “Entire-In” method.

For the processing of the OAP data, it is necessary to define the actual sampling volume (SV) of the probe using the following relation:

$$SV \text{ (in cm}^3\text{)} = SA \cdot \text{TAS} \cdot t \cdot 10^2 \quad (29)$$

where TAS is the true airspeed in m s\(^{-1}\) and \(t\) is the sampling time in seconds. \(SA\) is the sampling area, which is calculated from **equation 30:**
\[ SA = DF \cdot w \quad (30) \]

where \( w \) (in \( \mu m \)) is the effective width of the array. The term depth of field (DF; in cm) refers to the range around the object plane in which a particle can be detectable and is given by the formula:

\[ DF = \pm 7.5 \cdot 10^{-5} \cdot D^2 \cdot w/\lambda \quad (31) \]

where diameter in \( \mu m \) is \( D = I \cdot RS \) (\( I \) is the number of occulted diodes and \( RS \) the probe resolution) and \( \lambda \) (in \( \mu m \)) is the wavelength of the laser beam. The \( \pm \) indicates sign as particles can be detected in both sides beyond the object plane [Knollenberg, 1970; Heymsfield and Parrish, 1978].

3.1.2.1 Cloud Imaging Probes 15 and 100 (CIP-15 and CIP-100)

The CIP-15 and CIP-100 probes are OAPs, which consist of 64 element photodiode arrays. Both these probes image the particle shadows, which are recorded by the photodiode arrays, when a particle passes through the emitted laser beam (wavelength \( \lambda = 0.685 \mu m \)). Particles are considered as in-focus only when they pass through the object plane, which is in the middle of the distance between the two sample arms. The particles, which are detected outside the object plane are depicted as doughnut-shaped items, the outline of which decreases with increasing distance from the object plane. Although these probes operate in a similar way, there are three differences between them. The first is the element resolutions, which are 15\( \mu m \) and 100\( \mu m \) respectively. The second difference is that CIP-15 measures particles over size range 15-930\( \mu m \), while CIP-100\( \mu m \) measures larger particles of sizes 100-6200\( \mu m \). The third is that the CIP-15 is a grayscale, while the CIP-100 is a monoscale probe. For this reason CIP-15 can record a particle at a certain shadowing thresholds (usually 25, 50, 75%), while the CIP-100 produces particle images based on a single shadowing threshold (50%). It should be noted that greyscale feature allows the determination of the location of a particle in the sample.
area being a useful tool for particle size correction [Joe and List, 1986]. As ice shattering can significantly affect the measurements, a knife-sharp leading edge is placed on the arms to minimize this effect [Baumgardner et al., 2001].

3.1.2.2 The Two-Dimensional Stereoscopic (2D-S) Probe

The 2D-S is an OAP that measures particles in a size range of 10-1280μm imaging their shadows. The main feature that differentiates this OAP from CIP-15 and CIP-100 is that two laser beams are crossed (figure 17) illuminating two different 128 element photodiode arrays. Each of these elements can represent a pixel of 10μm. The two overlapped beams capture single images of particles outside the overlap region, and two images of the same particle within it, in a resolution of 10μm. The sample volume of the overlap region is 16.6 L s⁻¹ for an aircraft speed of 100m s⁻¹. As this OAP is a monoscale probe, it can record a particle event at a 50% shadowing threshold [Lawson et al., 2006].

![Figure 17. Picture showing the sampling arms of a 2D-S probe and the crossed laser beams in the sample area (Adapted from Lawson et al, 2006).](image)

3.1.3 Forward-Scattering Probes

Another category of probes, which were mounted on the observational aircraft, is the probes, which use Mie-scattering theory to detect and measure small cloud
droplets. In general, such probes emit a laser beam, which is scattered forward by liquid droplets. Then, the intensity of the light is measured by photodetectors. The size of the particles is estimated by the theoretical Mie-scattering curve, which calculates droplet sizes as a function of the intensity of the scattering light. Such probes are the Forward Scattering Spectrometer Probe (FSSP) [e.g. Baumgardner] and the Cloud Droplet Probe (CDP) [Lance et al., 2010].

General errors that can affect the measurements are the dead-time/coinidence, the airspeed and the ice shattering effects [Brenguier and Amodei, 1989; Lance et al., 2010]. Dead-time/coinidence effect is the effect that the probe cannot count two particles passing together through the sample volume. This can lead to errors of concentration estimation and mis-sizing of the particles. An uncertainty in the airspeed can lead to errors in droplet concentrations, as it defines the sample volume. Finally, small ice particles caused by ice shattering can lead to overestimations of droplet concentrations especially in FSSP by a factor of 1000 within regions of ice concentrations >5 L⁻¹ [Crosier et al., 2011; Pina et al., 2011].

3.1.3.1 Cloud Droplet Probe (CDP)

The CDP is used for the detection of liquid cloud droplets of sizes between 3-50μm. The particle data is separated into 30 size bins. Size bins are 1μm wide between 3-14μm, and 2μm wide between 15-50μm. As mentioned above, the operation of the probe is based on the Mie-scattering theory. In particular, when a cloud droplet passes through the laser beam, the photo detectors of the probe measure the intensity of the forward scattered light over the angles ~4-12°. Then, the light is equally distributed (by a beam splitter) between the qualifier, which recognise a countable particle, and the sizer, which is used for the particle size estimation (figure 18). From the data provided by this probe, Liquid Water Content (LWC), droplet size and number concentration can be estimated. It should be highlighted that CDP measurements can be affected by ice particles. However, the number concentration of ice particles in sizes, which are countable by CDP, are estimated at a few per liter [Lance et al., 2010].
3.1.5 Temperature Sensor

The attached Rosemount temperature sensor on the aircraft is a “loom”-type PRT (Platinum Resistance Thermometer), which means that the ambient air temperature measurements are based on the change of the electrical resistance to current flow of platinum when temperature changes. The sensor exhibits a response time of 1s. It is equipped with specific inlets to minimize water and particle ingress, which can cause abnormal observations. In addition, it can be placed within a de-iced housing, which includes a heater to prevent the icing effect. However, the de-iced sensor can be susceptible to wetting and, therefore, it may measure inaccurately. In general, temperature measurements should be taken under consideration only when there is a fair degree of agreement between de-iced and non-de-iced temperature (~±0.5K). Another factor that can affect temperature measurements is the degree of deceleration of the air flow in de-iced or non-de-iced housings, which is still under investigation. Finally, electrical noise has been found to cause temperature fluctuations, the sources of which have not been completely clarified yet [Price, 2016].
3.1.6 The Chilbolton Advanced Meteorological Radar

The CAMRa, [Goddard et al., 1994], is a dual-polarisation S-band radar (λ≈10cm, f=3GHz) located at Chilbolton in southern England (51.1445°N, 1.4370°W). It is equipped with a very large (diameter=25m) fully steerable dish, which allows high spatial resolution measurements due to the very narrow beam (0.28° beam width, 0.3km in range). Because of its low frequency, insignificant two way attenuation (< 1 dB km⁻¹) occurs even in cases of heavy precipitation (e.g. 16mm h⁻¹) [Curry, 2012]. The PRF is set to 610Hz, which means that the effective maximum range is estimated at ~255km (according to equation 15), while unambiguous Doppler velocity is estimated at ~15m s⁻¹ (according to equation 20). As it is a dual-polarization radar, it can provide measurements of reflectivity factor (Z_H) and Doppler velocity, but also differential reflectivity (Z_DR), differential phase shift (Φ_DP), Linear Depolarisation Ratio (L_DR) and co-polar correlation coefficient (ρ_HV).

3.1.7 The Copernicus 35GHz Cloud Radar

Copernicus, which is located within 50m of the CAMRa, is a 35GHz (λ=9mm) Ka-band dual-polarisation cloud radar, which is equipped with a zenith-pointing 2.4m diameter antenna. It can provide high resolution data as the beamwidth is 0.25° and the range resolution is 30m. As PRF is set to 5000Hz, the unambiguous Doppler velocity is 10.7m s⁻¹ and the effective maximum range is 30km, which allows observations of clouds in the entire profile of the atmosphere [Wood et al., 2009]. In contrast to the CAMRa, the Copernicus radar suffers from significant attenuation due to precipitation (exceeding two way attenuation of 5dB km⁻¹ for heavy precipitation of a rate of 16mm h⁻¹) [Curry, 2012]. Millimeter radars like this are ideal for measuring small cloud particles due to the smaller wavelength of the transmitted radiation [Godard, 1970; Kolias et al., 2007; Fukao et al., 2014].
3.2 Methodology

In this section, we discuss all the methods that were used for the in-situ and radar data processing, the results of which are used in chapters 4, 5 and 6.

3.2.1 Processing the OAP Data

There are several processes that were followed in order to obtain information about the size, the concentration and the phase of the particles measured by the OAPs. For the calculation of particle concentrations in size range 10-6200μm, 2D-S, CIP-15 and CIP-100 datasets were merged by averaging the overlapped size regions. The size of the particles was estimated as the average of the maximum size of the particle along Dy and across Dx array directions [Crosier et al., 2014]. It should be noted that concentration and size of ice particles can be affected by ice shattering effect despite the existence of anti-shattering tips. This effect can occur because of ice breakup on probe arms and ice fragmentation due to turbulence, which is produced by the probe [Korolev and Isaac, 2005]. For this reason, particles, which are measured in less than $10^{-4}$ secs inter-arrival time, are considered as artefacts and, thus, are excluded, according to Crosier et al. [2014] (see also section 3.1.2).

In order to discriminate ice from liquid particles, the total area and the perimeter around the edge of particles larger than 20μm were used for the circularity analysis (considering as highly, medium and low irregular, particles with perimeter to area ratio >1.4, 1.2-1.4 and <1.2 respectively). Thus, spherical particles were considered as liquid drops, while asymmetrical particles assumed to be ice crystals [Crosier et al., 2011]. The size threshold of 20μm was used in order to exclude particles having inefficient resolution for determining their shape. Then, Ice Water Content (IWC) was calculated exclusively for number size distributions (2DS and CIP-100) of asymmetrical particles by the general formula given by Hogan et al. [2006]:

\[
\text{IWC} = \int_{0}^{\infty} n(D) m(D) \, dD 
\]
where \( n \) is the particle size distribution, \( D \) is the particle diameter and particle mass \( m \) suggested by Heymsfield et al. [2007] is in the form of \( m = a D^b \), where \( a=0.00114 \) and \( b = 1.86 + 0.004\cdot T \) are coefficients depending on the temperature \( T \).

### 3.2.2 Pre-Processing the Radar Data

The radar data format from the CAMRa is originally in polar co-ordinate system. In the work presented here, we only used Range Height Indicator (RHI) radar scans in order to investigate entire cloud profiles. In order to produce comprehensive RHIs, we converted the radar data from polar (elevation-range) to Cartesian (height-distance) co-ordinate system according to the methodology described in the section 2.2.2.

Before any use of the radar data, a series of filtering were applied in order to improve the quality of the datasets. First, a flood-fill approach was applied on the Doppler velocity parameter in order to unfold it (see section 2.2.5.2), when it presents incorrectly values \( >15 \text{m s}^{-1} \) or \( <15 \text{m s}^{-1} \). According to this methodology, a pixel, which is assumed to be within the folding range and near the radar, is chosen. Then, regions of contiguous rainy pixels are unfolded. Sometimes, a whole region can be wrongly unfolded by a multiple of \( 30 \text{m s}^{-1} \), which can be easily spotted by eye [Rihan et al., 2008]. Second, random and unwanted noise, which arise from inside the radar circuit (detector noise) was also removed. According to the methodology suggested by Hogan [1998], data points that present signal power \( < \bar{N} + 3\Delta N \) (where \( \bar{N} \) is the mean noise power and \( \Delta N \) is the standard error of the mean noise power) should be removed. Third, ground clutter, which refers to undesirable signal coming from sea/earth surface (such as sea waves, hills, mountains, buildings etc.) was removed using a certain threshold of Doppler velocity (e.g. \(<0.15 \text{m s}^{-1} \)). Data points that present smaller Doppler velocity than this threshold are considered as motionless. Finally, speckle removal was applied for every single pixel that is surrounded by pixels which are below detection.
3.2.3 The Hydrometeor Classification Algorithm (HCA)

The NCAR’s (Nation Center for Atmospheric Research) HCA for S-band radar, which is used for the purposes of this thesis, falls into the category of fuzzy-logic algorithms, which are briefly described in section 2.2.6. The algorithm is a standard part of NCAR’s LIDAR RADAR Open Software Environment (LROSE) and it is based on the Vivekanandan et al. [1999] approach. Some recent studies, which make use of this algorithm are Plummer et al. [2010], Jung et al. [2012], Houze [2012], Friendrich et al. [2013], Kalina et al. [2016] and Kalina et al. [2017].

The algorithm primarily utilizes a fuzzy-logic scheme with two-dimensional linear piecewise membership functions (MBF). The MBFs are “two-dimensional” because they vary as both functions of the reflectivity factor $Z_H$ and one of the other variables utilized in the scheme ($Z_H$, $Z_{DR}$, $L_{DR}$, $K_{DP}$, $\rho_{HV}$, temperature $T$ and standard deviations of $Z_{DR}$, and $\Phi_{DP}$, $SD(Z_{DR})$ and $SD(\Phi_{DP})$) that the HCA uses. The term “piecewise” means that MBFs are described by a non-continuous function. It should be noted that certain thresholds (see $a$, $b$, $c$, $d$ in equation 33) for each variable are used for certain $Z_H$ steps. For every single step and each element $x$ from each one of the eight variables, a linear piecewise function is used. Although a linear piecewise function can be triangular, trapezoid or more complex, here we give the simplest form, which is a trapezoidal MBF:

$$
\mu(x) = \begin{cases} 
0 & , \ (x < a) \ or \ (x > d) \\
\frac{x-a}{b-a} & , \ a \leq x \leq b \\
1 & , \ b < x \leq c \\
\frac{d-x}{d-c} & , \ c \leq x \leq d 
\end{cases}
$$

(33)

where $a$ is the lower limit, $d$ the upper limit, $b$ the lower support limit and $c$ the upper support limit and $a < b < c < d$. A representation of the two-dimensional MBF for $Z_{DR}$ for the class of moderate rain is given in figure 19.
Figure 19. The two-dimensional membership function of $Z_H$ and $Z_{DR}$ in 2D shaded representation. The graph refers to the class of moderate rain.

Thus, the number of MBFs produced for each of the 8 input variables corresponds to the number of hydrometeor types $n$. This process of changing a real scalar value into a fuzzy one is called **fuzzification**. The fuzzification results are multiplied by weights depending on the degree that a variable is important for the process (see **table 1**). Then, during the aggregation process, the $n$ (here 8) weighted fuzzification results are summed producing a single aggregated value for each of the hydrometeor classes. The dominant hydrometeor class is defined as the class with the maximum aggregated value. Some extra limitations, which are presented in **table 2**, are applied in order to minimise errors originated in apparently erroneous designations.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>Temperature</td>
<td>0.2</td>
</tr>
<tr>
<td>$Z_H$</td>
<td>0.2</td>
</tr>
<tr>
<td>$Z_{DR}$</td>
<td>0.2</td>
</tr>
<tr>
<td>$K_{DP}$</td>
<td>0.05</td>
</tr>
<tr>
<td>$L_{DR}$</td>
<td>0.1</td>
</tr>
<tr>
<td>$\rho_{HV}$</td>
<td>0.1</td>
</tr>
<tr>
<td>$SD(Z_{DR})$</td>
<td>0.3</td>
</tr>
<tr>
<td>$SD(\Phi_{DP})$</td>
<td>0.3</td>
</tr>
</tbody>
</table>

**Table 1.** Matrix of weights W
The NCAR HCA discriminates between 15 classes of radar echo: 1) Ground Clutter, 2) Cloud, 3) Drizzle, 4) Light Rain (rain rate < 10 mm h\(^{-1}\)), 5) Moderate Rain (10 ≤ rain rate < 40 mm h\(^{-1}\)), 6) Heavy Rain (≥ 40 mm h\(^{-1}\)), 7) Supercooled Droplets, 8) Graupel and Rain mixture, 9) Rain and Hail mixture, 10) Graupe/Small Hail, 11) Hail, 12) Wet Snow, 13) Dry Snow, 14) Ice Crystals, 15) Irregular Ice crystals (smaller than snow aggregated/rimed crystals with no distinct shape that would produce significant \(Z_{\text{DR}}\) or \(K_{\text{DP}}\)).

<table>
<thead>
<tr>
<th>Hydrometeor Class</th>
<th>(Z_H) (dBZ)</th>
<th>(Z_{\text{DR}}) (dB)</th>
<th>Temperature (°C)</th>
<th>SD ((Z_{\text{DR}}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ground Clutter</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>&lt;2</td>
</tr>
<tr>
<td>Cloud</td>
<td>-</td>
<td>&gt;1</td>
<td>&lt;0</td>
<td>-</td>
</tr>
<tr>
<td>Drizzle</td>
<td>-</td>
<td>&gt;1.5</td>
<td>&lt;0</td>
<td>-</td>
</tr>
<tr>
<td>Light Rain</td>
<td>-</td>
<td>-</td>
<td>&lt;0</td>
<td>-</td>
</tr>
<tr>
<td>Moderate Rain</td>
<td>-</td>
<td>≤0.1</td>
<td>&lt;0</td>
<td>-</td>
</tr>
<tr>
<td>Heavy Rain</td>
<td>-</td>
<td>≤0.3</td>
<td>&lt;0</td>
<td>-</td>
</tr>
<tr>
<td>Supercooled Droplets</td>
<td>&lt;20</td>
<td>-</td>
<td>-40 or &gt;0</td>
<td>-</td>
</tr>
<tr>
<td>Graupel/Rain mixture</td>
<td>-</td>
<td>-</td>
<td>&lt;30</td>
<td>-</td>
</tr>
<tr>
<td>Rain/Hail mixture</td>
<td>-</td>
<td>-</td>
<td>&lt;30</td>
<td>-</td>
</tr>
<tr>
<td>Hail</td>
<td>&lt;45</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Wet Snow</td>
<td>-</td>
<td>&lt;0.3</td>
<td>&lt;8 or &gt;8</td>
<td>-</td>
</tr>
<tr>
<td>Dry Snow</td>
<td>-</td>
<td>-</td>
<td>&gt;1</td>
<td>-</td>
</tr>
<tr>
<td>Ice Crystals</td>
<td>-</td>
<td>&lt;0.5</td>
<td>&gt;0</td>
<td>-</td>
</tr>
<tr>
<td>Irregular Ice Crystals</td>
<td>-</td>
<td>-</td>
<td>&gt;0</td>
<td>-</td>
</tr>
</tbody>
</table>

**Table 2.** Empirical hard thresholds used minimise errors due to apparently erroneous designations.

This HCA has been validated by Barthazy et al. [2001] and Martini et al. [2015] showing that the algorithm is able to efficiently distinguish particles. The first used formvar slides and a ground-based optical spectrometer to observe hydrometeors and they noticed that oblate irregular ice crystals and ice needles may be misclassified due to similar fuzzy thresholds. The second compared the HCA
results with OAP imagery of cloud particles, in two different winter cases of stratiform clouds with embedded convection, highlighting the difficulty of identifying partially melted, refrozen, heavily rimed and aggregated particles in OAP images, which makes the HCA validation process difficult.
CHAPTER FOUR

MICROPHYSICAL PROPERTIES AND RADAR POLARIMETRIC FEATURES WITHIN A WARM FRONT

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Microphysical Properties and Radar Polarimetric Features within a Warm Front

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ABSTRACT

On 21 January 2009, the warm front of an extensive low pressure system affected the U.K. weather. In this work, macroscopic and microphysical characteristics of this warm front are investigated using in situ (optical array probes, temperatures sensors, and radiosondes) and S-band polarimetric radar data from the Aerosol Properties, Processes and Influences on the Earth’s Climate-Clouds project. The warm front was associated with a warm conveyor belt, a zone of wind speeds of up to 26 m s\(^{-1}\), which played a key role in the formation of extensive mixed-phase cloud mass by ascending significant liquid water (LWC ~0.22 g m\(^{-3}\)) at a level ~3 km and creating an ideal environment at temperatures ~ -5°C for ice multiplication. Then, “generating cells”, which formed in the unstable and sheared layer above the warm conveyor belt, influenced the structure of the stratiform cloud layer, dividing it into two types of elongated and slanted ice fall streaks: one depicted by large Z\(_{DR}\) values and the other by large Z\(_{H}\) values. The different polarimetric characteristics of these ice fall streaks reveal their different microphysical properties, such as the ice habit, concentration, and size. We investigate their evolution, which was affected by the warm conveyor belt, and their impact on the surface precipitation.
1. Introduction

Extratropical cyclones have been extensively studied, as such systems often demonstrate complicated cloud structures, providing an ideal environment for microphysical studies (e.g., Matejka et al. 1980; Chen and Cotton 1988; Forbes and Clark 2003; Stark et al. 2013; Crosier et al. 2014; Lloyd et al. 2014; Dearden et al. 2016). The typical structure of cyclones consists of a cold, a warm, and an occluded front, which usually affect the United Kingdom with strong winds and heavy rainfall (e.g., Browning 2004; Lavers et al. 2011). The fronts are associated with warm and cold conveyor belts, which circulate air masses with different traits. Warm conveyor belts (WCBs), which usually originate in surface maritime areas, convey humid air, making a substantial contribution to the cloud structure and surface precipitation (Harrold 1973; Browning 1986; Eckhardt et al. 2004; Pfahl et al. 2014). Although all types of fronts have particularities and mechanisms that need further investigation, studying warm fronts is important, as they account for the majority of the precipitation associated with extratropical cyclones (Keyser 1986; Wakimoto and Bosart 2001).

Dual-polarization radars have been a valuable tool for revealing cloud microphysical properties, providing details about ice crystal habits, dimension, shape, orientation, and phase of hydrometeors. Various studies have been conducted on linking polarimetric radar variable thresholds with specific hydrometeor types (e.g., Herzegh and Jameson 1992; Straka et al. 2000; Hogan et al. 2002; Kennedy and Rutledge 2011; Moisseev et al. 2015; Schrom and Kumjian 2016) and determining polarimetric signatures, which can provide insight into cloud microphysical processes (e.g., Hall et al. 1984; Illingworth et al. 1987; Hubbert et al. 1998; Smith et al. 1999; Ryzhkov et al. 2005; Kumjian and Ryzhkov 2008; Kumjian 2013). Comparing and linking in situ cloud measurements with radar data can improve and validate the interpretation of the remote sensing data and, therefore, the understanding of cloud microphysics, including the role of primary (Koop et al. 2000; Vali et al. 2015) and secondary ice production (Hobbs and Farber 1972; Hallett and Mossop 1974; Mossop and Hallett 1974; Pruppacher and Schlamp 1975; Vardiman 1978; Knight 1979; Choularton et al. 1980) and embedded convection mechanisms. Embedded convective turrets, which appear as
cells with horizontal and vertical extent of up to 6 and 2km, respectively, are
referred to as “generating cells” (GCs) and have been documented several times in
the literature (Wexler and Atlas 1959; Houze et al. 1976; Hobbs and Locatelli 1978;
Herzegh and Hobbs 1981; Kumjian et al. 2014; Plummer et al. 2014; Rosenow et
al. 2014; Plummer et al. 2015; Oue et al. 2015; Rauber et al. 2015; Keeler et al.
2016a,b, 2017; and many others). Discussing some of the latest studies, Plummer
et al. (2014, 2015) used in-situ (2D-C and 2D-P) together with a Doppler radar
(reflectivity factor and Doppler velocity) in order to investigate the microphysics
of GCs and the fall streaks originated in them. In addition, Rosenow et al. (2014)
used an airborne W-band radar to quantify the magnitude of the vertical velocities
in GCs, while Keeler et al. (2016a,b, 2017) used models to simulate GCs and
examine their origin, their forcing, and their relationship to vertical wind shear,
ambient thermal instability, and cloud-top radiative forcing. A mechanism that
commonly triggers GCs is Kelvin-Helmholtz instability, which happens when
velocity shear occurs in a single continuous fluid (Browning et al. 1973; Hogan et
al. 2002). GCs usually present updrafts of < 3m s\(^{-1}\), and they are often associated
with descending trails of ice crystals and snow, which are commonly referred to as
“ice fall streaks”. Ice fall streaks, the shape of which is affected by the vertical wind
shear, originate in and form from generating cells and play an important role in the
production of precipitation (e.g., Marshall 1953; Langleben 1956; Douglas et al.
1957; Wexler and Atlas 1959; Bader et al. 1987; Kumjian et al. 2014; Rosenow et
al. 2014; Oue et al. 2015).

In this study, we use high-resolution collocated in situ and active remote sensing
observations obtained from dualpolarization data to provide (i) a detailed picture
of the microphysical structure of a warm front and (ii) analysis of the structure, the
origin, and the effects of the ice fall streaks, which are frequently embedded within
warm fronts.

2. Methodology

The data used in this study were obtained during the NERC-funded Aerosol
Properties, Processes and Influences on the Earth’s Climate (APPRAISE)-Clouds
project, which took place in the United Kingdom during 2007-10. This cloud
project aimed to provide a comprehensive investigation of mixed-phase clouds influencing U.K. weather systems and the impact of aerosols on the microphysical properties and development of such clouds (Crosier et al. 2011; Crawford et al. 2012).

a. The in situ datasets

The U.K. Facility for Airborne Atmospheric Measurements (FAAM) Bae-146 aircraft was used to collect in situ measurements of cloud properties during APPRAISE-Clouds. In particular, on 21 January 2009 during flight B424, the aircraft extensively sampled clouds associated with a warm front over southern England, performing a series of straight and level runs, profiles, and sawtooth profiles over the period 1500-2000 UTC. The sampling occurred along a radial of 255° within ~100-km range from the Chilbolton Facility for Atmospheric and Radio Research (CFARR; located at 51.158N, 1.448W) to allow direct comparison with radar scans along the same azimuth. Hourly radiosondes were also launched from the Chilbolton ground site between 0700 and 1800 UTC.

The FAAM Bae-146 aircraft was equipped with microphysical probes, which provide measurements of a wide range of hydrometeors, as described in Crosier et al. (2014). Key instrumentation used in this work included a Mie-scattering cloud droplet probe (CDP), measuring number and size (~3 and 50mm) of cloud droplets (Lance et al. 2010); two-dimensional stereo (2D-S) and cloud imaging probe-100 (CIP-100) optical array probes (OAPs), providing shadow imagery of particles with size 10-1280 and 100-6200mm (Knollenberg 1970; Lawson et al. 2006), respectively; and a Rosemount probe to measure the ambient deiced temperature. It should be highlighted that the 2D-S probe provides better resolution (10mm) and wider size range (10-1280mm), compared to its previous version, 2D-C (25mm; 25-800mm). Particle phase was estimated based on a shape analysis of images obtained with the 2D-S and CIP-100, while ice number concentration and diameter (from the diagonal of the bounding box around the recorded particle image) was estimated according to Crosier et al. (2011, 2014). An interarrival time threshold was used to filter OAP images to remove artifacts associated with shattering of
large particles on the probes according to Field et al. (2006). To estimate bulk ice-phase parameters, including mean diameter and concentration, the 2D-S and CIP-100 datasets were merged, and measurements in overlapping sizes were averaged. Ice water content (IWC) was calculated from merged 2D-S and CIP-100 particle size distributions using estimates of particle mass from Heymsfield et al. (2007). Significant uncertainties (~50% for IWC) can arise because of the diversity in OAP image processing options. Liquid water content (LWC) was estimated from the CDP particle size distribution.

b. The remote sensing datasets

During the flight, the Chilbolton Advanced Meteorological Radar (CAMRa) performed range-height indicator (RHI) scans along the 255° (WSW) radial every ~90s to obtain near-collocated radar and in situ measurements. The CAMRa is a dual-polarization Doppler radar that operates at 3GHz with a steerable 25-m dish, resulting in a narrow 0.28° beam (Goddard et al. 1994). Additionally, the zenith-pointing, dual-polarization, 35-GHz Copernicus radar, which is located within 50m of the CAMRa, was operating throughout the day. In contrast to low-frequency radars, which are applied for precipitation observations (Kollias et al. 2007; Fukao and Hamazu 2013), high-frequency radars are significantly affected by attenuation due to precipitation (Godard 1970). The cross-sectional ratio between large and small particles at such high frequencies is smaller, compared to longer wavelengths, so it is an ideal tool for monitoring clouds and fog. All references to data from the 35-GHz vertical pointing radar will be indicated by the subscript “35”; otherwise, the data being discussed are from the 3-GHz radar.

In this study, we use radar variables, such as reflectivity factor $Z_{H}$, Doppler velocity $V_{RAD}$, and differential reflectivity $Z_{DR}$, to examine cloud dynamical and microphysical structures; $Z_{H}$ is determined by the sixth power of the diameter and concentration of the hydrometeors (Doviak et al. 1979). Doppler velocity provides an estimate of the motion of hydrometeors, which can indirectly be used to infer, depending on the mode of operation, information on wind fields or the fall speeds of hydrometeors (Browning and Wexler 1968). Finally, $Z_{DR}$ is the ratio between the radar reflectivities at horizontal and vertical polarizations, expressed in
logarithmic scale induced by hydrometeors within the radar beam (Seliga and Bringi 1976), and can be used to provide details about their shape, orientation, and phase. We highlight that $Z_{DR}$ data coming from elevations, $0.2^\circ$ have been excluded due to vertical beam blockage (Ryzhkov et al. 2002; Giangrande and Ryzhkov 2005).

3. The synoptic condition

On 21 January 2009, a deep, low pressure system in the northwest Atlantic Ocean moved eastward and developed (Fig. 1a). This system generated multiple frontal boundaries, which approached the United Kingdom from the west. The Bae-146 research flight sampled the warm front as it passed across the south of England. This warm front was observed (through fields of wind and potential temperature) to be associated with a WCB, which originated in the warm sector of the system (Fig. 1). The warm front is not located clearly in the field of relative humidity (Fig. 1b), probably because precipitation moistens the layer between the frontal zone and the surface. Figures 1a and 1b show that the WCB transported humid and warm air over the colder air found over the U.K. mainland. In Fig. 1b, the WCB becomes perceptible through the zone of enhanced relative humidity and strong south-southwest to west winds (up to 23 m s$^{-1}$), which reached a level of $\sim$450mb ($\sim$6-km height).
FIG. 1. ERA-Interim (0.1258 resolution) at 1800 UTC 21 Jan 2009. (a) Temperature (°C) on the 850-mb (1 mb = 1 hPa) pressure level (color shading and dashed contours) and mean sea level pressure (mb) (white contours). Surface fronts have been plotted in accordance with Met Office analysis charts. (b) Vertical section of relative humidity (color shading) and wind speed/direction (standard wind barbs) across the investigated warm
front at 1800 UTC [purple line in (a)]. (c) Vertical section as in (b) of potential temperature. Warm front position is highlighted by the red line according to wind and potential temperature fields.

4. A macroscopic description of the cloud features

A brief description of the entire in-situ and radar dataset is provided in order to understand the general cloud structures in the lower/mid-troposphere during the passage of the warm front shown in Fig. 1a.

The spatial structure of the frontal cloud obtained from CAMRa at 1712 UTC is depicted in Fig. 2. Several features, which can be seen in Fig. 2, were consistently observed throughout the frontal passage and will now be discussed in turn. First, we observed GCs presented as protruded cloud tops, which demonstrated high-$Z_H$/low-$Z_{DR}$ cores ($>10$dBZ/$\sim 0$ dB) and low-$Z_H$/high-$Z_{DR}$ (<10dBZ/$> 1$ dB) boundaries (Figs. 2a,d) (e.g., Bader et al. 1987; Kumjian et al. 2014; Oue et al. 2015). A $Z_H$ ice fall streak (e.g., Bader et al. 1987; Kumjian et al. 2014; Oue et al. 2015) is represented as a long (tens of km) and narrow ($\sim 1$ km) slant (angle of 38-78) zone of high $Z_H$ (enhanced by 10-15dBZ relative to the surroundings) (Fig. 2a) and $\sim 0$-dB $Z_{DR}$ (Fig. 2d) containing mostly rimed/aggregated crystals (Fig. 5; 2, altitude, 3.5 km). Then, $Z_{DR}$ fall streaks (e.g., Bader et al. 1987; Kumjian et al. 2014; Oue et al. 2015), typically located above a $Z_H$ fall streak, exhibited a zone of high $Z_{DR}$ (>1dB) (Fig. 2d) and low $Z_H$ (<10dBZ) (Fig. 2a) values consistent with the presence of pristine ice crystals (Fig. 5; 3, altitude, 5 km). Although Fig. 2 does not show an ideal example of GC, it highlights the connection between a GC and a $Z_H/Z_{DR}$ fall streak (see details in section 5b). Finally, Fig. 2f presents two banded layers of enhanced Doppler velocity associated with the so-called warm conveyor belt, an airstream that appears in RHIs between 2 and 4 km, which transports humid and warm air aloft.

An overview of the major temporal changes in the cloud structure can be seen in the vertical-pointing 35-GHz radar data (Fig. 3). It should be noted that clouds located at horizontal distances between 30- and 90-km range in the 3-GHz radar scans are estimated to overpass the 35-GHz radar 25-75 min later (based on a
typical wind speed of ~20 m s^-1 along the radar profile). Although comparisons between these two radar datasets will not be fully correlated, as the system may change during transit, it reveals useful general trends of the system over time.

In Fig. 3a, we can see a change in the height of the freezing level or “melting layer” (ML). The ML appears as a bright band in the reflectivity field, the $Z_H$ signal of which increases depending on the type and concentration of the hydrometeors that enter this region (Fabry and Zawadzki 1995). The $Z_H$ increases within the ML due to the change of the phase and shape of melting hydrometeors, which also affects $Z_{DR}$ (e.g., Stewart et al. 1984; Zrnic et al. 1993; Szyrmer and Zawadzki 1999; Giangrande et al. 2008). Before the passage of the warm front (<2215 UTC), the ML was observed at ~0.7-1 km ($Z_{H,35}$ reached 30 dBZ at 0.75 km; Fig. 3a). After the passage of the warm front, the ML height raised to ~1.6 km (2215 UTC) (Figs. 1a, 3a). The ML is also depicted as a zone of dramatic increase of Doppler velocity (from ~1 to >4 m s^-1), retrieved by the 35-GHz vertical-pointing radar (Fig. 3b), because hydrometeors may become larger in size due to coalescence and denser due to melting (Mitchell 1996; Heymsfield and Iaquinta 2000; Protat and Williams 2011).
FIG. 2. The 3-GHz radar RHI scans of (a) $Z_H$ (dBZ), (d) $Z_{DR}$ (dB), and (f) unfolded Doppler velocity (m s$^{-1}$) at 1711:00-1712:30 UTC (negative values indicate flow toward the radar). (b) Percentage of cloudy pixels ($Z_H > 40$dBZ) with height. (c),(e),(g),(h) Graphs show average (solid lines) and 95th percentiles (dots) of $Z_H$, $Z_{DR}$, Doppler velocity, and vertical wind shear produced from Doppler velocity, which were calculated from the 3-GHz radar dataset for $30 \leq \text{range} \leq 90$ km and $0 \leq \text{altitude} \leq 8$ km for the most representative scans during each time period shown in the legend. Different colors indicate different time periods for (b),(c),(e),(g),(h).
FIG. 3. (a) $Z_H$ (dBZ) and (b) Doppler velocity (m s$^{-1}$) evolution with time as captured by the 35-GHz radar. In (a), the ML height is highlighted by the blue (time, 2100 UTC), the green (2100 < time < 2215 UTC), and the red (time < 2215 UTC) dashed lines. The black dashed line shows the linear fitting of the cloud-top heights between 1500 and 2100 UTC. The colored lines on the left corner indicate the time that radiosondes of Fig. 4 were launched, compared to the time color scale of (a). In (b), negative Doppler velocity indicates flow toward the radar (for vertically pointing radars, this is usually dominated by particle fall speed). (c) Average (solid lines) and 95th percentiles (dots) of $Z_H$ as calculated from 35-GHz radar data. (d) As in Fig. 2b, but for the 35GHz radar dataset. (c),(d) Calculated from the first 10 min of data of each hour.

Data from the 35-GHz vertical-pointing radar (Fig. 3a) show that the cloud-top height decreased (from 6.5 to 5 km) at a mean rate of 0.24 km h$^{-1}$ during the frontal passage, while the cloud-top temperature increased from $\sim$35$^\circ$ to $\sim$25$^\circ$C, according to radiosondes (Fig. 4). The lowering of cloud-top height with time,
which can be seen in Fig. 4c as a decrease of RH with time by 50% at altitudes 6-7 km, is also perceptible from Fig. 2b (3-GHz radar) and Fig. 3b (35-GHz radar). The 35-GHz radar data (Fig. 3c) show that >50% of the pixels were considered as cloudy (where $Z_H > -40$ dBZ) at altitudes $\leq 5.5$ km for 1800-1900 UTC. However, this height gradually decreased to 4.5 (1900-2000 UTC) and 4 km (2100-2200 UTC). The corresponding heights for the 3-GHz radar were always observed ~0.5 km lower than the 35-GHz radar (Fig. 2b), as the 3-GHz frequency is mainly used for the detection of precipitation and not of smaller cloud particles. According to Fig. 3b, Doppler velocities between 0 and -1 m s$^{-1}$ were observed at cloud tops (height > 4 km), which is consistent with weak vertical motions and slowly precipitating low-density ice crystals (like the small ice plates and columns at altitudes > 4 km and temperatures -15°C in Fig. 5) (Jayaweera and Cottis 1969; Heymsfield 1972; Kajikawa 1972).

To gain insight into the presence of the aforementioned features, a statistical analysis of the $Z_H$ and $Z_{DR}$ datasets is shown in Figs. 2c, 2e, and 3c. In relation to $Z_H$ fall streaks, in Fig. 2c, $Z_H$ seems to increase between 1646 and 1932 UTC (95th percentiles of $Z_H \sim 20$ dBZ at 2-4 km height). At 1912-1932 UTC, although average $Z_H$ decreased significantly to ~10 dBZ, high-$Z_H$ 95th-percentile values (>20 dBZ) were observed at 2-4 km, highlighting the presence of discrete GCs. This is also demonstrated by the 35-GHz radar dataset, as 95th percentiles of $Z_{H\sim 35}$ around the ML were enhanced at 2000 UTC (from 15 to ~30 dBZ). This $Z_{H\sim 35}$ increase, as well as the enhanced 95th percentiles at 2-4 km, implies the occurrence of more frequent and more intense, but spatially restricted, GCs, which occasionally affected the ML echo (Fig. 3c). This is also demonstrated by the fact that $Z_{H\sim 35}$ increased in the ML and decreased at 2-4 km with time, while 95th-percentile signals at 2-4 km remained significantly high (~15 dBZ). In addition, the vertical pointing radar at 1900-2030 UTC observed larger Doppler velocities (-2 to -3 m s$^{-1}$), compared to the previous time period at 0.7-3-km height, which indicates the existence of denser hydrometeors (like the large ice particles at altitudes < 3.5 km in Fig. 5). Switching focus to $Z_{DR}$ fall streaks, these features can be observed in Fig. 2e as local $Z_{DR}$ maxima at altitudes of ~4 km (slightly higher than $Z_H$ fall streaks). The $Z_{DR}$ fall streaks manifest mainly as perturbations to the 95th percentile (0.5-2 dB) relative
to the rest of the profile, affecting also the mean value (0.5-1dB). The \( Z_{DR} \) fall streaks are mainly observed during the second half of the observing period, 1800-1932 UTC.

The WCB was a clearly identifiable dynamical feature that likely “fed” the cloud system with water vapor. It is presented, in the vertical-pointing radar dataset, as a zone of Doppler velocity \( \sim 0 \text{ m s}^{-1} \), where strong westerlies (typically \( >20 \text{ m s}^{-1} \)) of the WCB advected the hydrometeors almost horizontally (Fig. 3b). While the wind speed profiles suggest significant stratification at all times (Fig. 4d), it seems that the highest wind speeds were observed near cloud top, where relative humidity (Fig. 4c) and cloudiness (Fig. 3c) dramatically decrease. The main region of high wind speeds (20-30 m s\(^{-1}\)) (Fig. 4d) and positive wind shear (Fig. 2h) was first located at 5-6 km (1500-1600 UTC) but gradually lowered to 3.5-4.5 km (after 1800 UTC). However, before 1800 UTC, there were also other enhanced wind speed layers along the profile, which likely represent slight differences in air mass characteristics/origin. As an illustration, two well distinguished layers, centered at altitudes of \( \sim 2.5 \) and \( \sim 4 \) km at \( \sim 1700 \) UTC (Fig. 2f), coincided with wind speed (17-20 m s\(^{-1}\)), relative humidity (>90\%), clockwise wind veering (~60° km\(^{-1}\)), compared to the average of ~20° km\(^{-1}\) for altitudes 0-6 km), and potential temperature (~ +9 K km\(^{-1}\), compared to average rate of +4 K km\(^{-1}\) for altitudes 0-6 km) maxima (Figs. 4c-f). At the same altitudes, temperature presented the highest increase (up to 1.3°C h\(^{-1}\)) between 1500 and 1800 UTC. All the above observations reveal that there was a main warm front at \( \sim 4 \) km and a possible subfrontal zone at \( \sim 2.5 \) km, the locations of which were associated with defined structures in the wind speed and temperature data. As a general trend, it seems that as the warm front was approaching, the various wind speed layers tended to merge (only one significant wind speed peak can be observed at \( \sim 4 \) km at 1800 UTC; Fig. 4d) and were located at lower altitudes (Figs. 2g,h).
FIG. 4. Graphs demonstrating (a) temperature (°C), (b) potential temperature (K), (c) relative humidity (%), (d) wind speed (m s⁻¹), (e) wind direction, and (f) change of potential temperature with height obtained from radiosonde data. Different colors indicate different times and were selected to be comparable with the color scheme of Figs. 2b, 2c, 2e, 2g, 3b, 3c, and 3e.

FIG. 5. Example of ice particle images captured by the 2D-S probe. Ice particles are grouped by the temperature (left axis) and altitude (right axis) at which they were observed.
5. Microphysical properties of the warm front

Using in situ and remote sensing data, the microphysical properties in specific features of the warm frontal cloud, such as the WCB, ice fall streaks, and regions of secondary ice production, are investigated.

a. The warm conveyor belt

The WCB originated in the warm sector of the deep, low pressure system shown in Fig. 1a, and during the observing period, it intersected the surface in the Celtic Sea. It was responsible for the large-scale slantwise ascent of humid air, which led to widespread formation of mixed-phase clouds.

At 1600 UTC, CAMRa scans to the WSW identified three zones of enhanced Doppler velocity (up to ~20 m s\(^{-1}\)) at altitudes of 2.5, 4, and 6 km (Fig. 6a). These three zones caused significant vertical wind shear (up to -12 m s\(^{-1}\) km\(^{-1}\)), which could potentially release instability and form convective GCs. Similar regions of wind shear were also identified in the wind speed and direction data from radiosonde profiles (Fig. 6b). In regions where the radar scan azimuth (255°) is closely aligned with the wind direction (at altitudes 2.5-3 km), the radiosonde wind speed is in good agreement with Doppler velocity.
FIG. 6. Unfolded Doppler velocity (m s$^{-1}$) in RHI scans for three scanning periods: (a) 1600:50-1602:20, (c) 1800:06-1801:35, and (e) 1935:13-1936:43 UTC (the grayscale line shows the aircraft track between 1933:34 and 1940:05—black for the initial and white for the final position). Negative values indicate flow toward the radar. The orange dashed line is the $-20$ m s$^{-1}$ contour. The solid red line in (c) determines the warm front boundary location and the dashed red line the possible subfront. (b),(d) Radiosonde wind speed and direction (color scale) over the Chilbolton region (shaded curve) and Doppler velocity (blue dashed line) along the vertical black line on (a),(c), respectively. (f) IWC, LWC, and cloud droplet concentrations for the corresponding aircraft track in (e) at $T \sim -1.9^\circ$C.

At 1800 UTC, as the warm front approached the radar site, a broad slanted zone of high Doppler velocities (up to 23 m s$^{-1}$) was observed spanning altitudes of 2-5km (Fig. 6c). The high Doppler velocity zone at 3-5km highlighted the WCB location and presented enhanced wind speed (15.6-26.0 m s$^{-1}$) and wind direction shear (21°km$^{-1}$). The peak wind speed was located near the middle of the WCB (bottom ~1 km and top ~5 km in regions close to the radar), with two zones of wind speed and direction shear (calculated from radiosonde data) located below (between 3
and 4.3 km, +8.7 m s⁻¹ km⁻¹ and ~47° km⁻¹) and above it (between 4.3 and 5.5 km, -3.8 m s⁻¹ km⁻¹ and ~ +10° km⁻¹). Again, there was a good level of agreement between Doppler velocity and measured wind speed for wind directions close to the radar beam azimuth (255°) (Figs. 6a-d). Comparing and linking the Doppler velocity with radiosonde data, we estimate the main warm front location is as shown in Fig. 6c by the solid red line. Smaller fluctuations in Doppler velocity fields and radiosonde wind profiles indicate the presence of a small subfrontal zone located at ~2 km (dashed red line).

At 1935 UTC, the frontal system had transited farther to the east over the ground site. At this time, the WCB was located ~2 km closer to the surface (between 1 and 3 km), being represented by a distinct zone of Doppler velocities between 17 and 25 m s⁻¹ (Fig. 6e). Aircraft observations obtained through the WCB at this time indicate the presence of significant liquid water (Fig. 6f) (up to 0.22 g m⁻³) and cloud droplet concentrations (10-58 cm⁻³), while IWC dramatically increases outside the WCB (up to 0.044 g m⁻³). These observations support the idea that slantwise ascent from the WCB generates a large expanse of cloud containing supercooled liquid water. The lifetime and radiative properties of similar supercooled layer clouds is sensitive to the presence of ice nuclei (e.g., Pinto 1998; Jiang et al. 2000; Morrison et al. 2005; Murray et al. 2012), which can activate to form ice in the cloud, leading to poor representation in weather models.

b. The generating cells

As the warm front approached the United Kingdom, some GCs appeared in RHI, especially after 1800 UTC. Despite the lack of in situ measurements within these features, we try to investigate their microphysics using the available radar data. In general, GCs were observed above the WCB and above a braided structure (Chapman and Browning 1998) associated with sheared wind (Fig. 7a). Sheared wind is associated with Kelvin-Helmholtz instability (Browning 1971; Browning et al. 1973), which is a mechanism that can trigger convection (Hogan et al. 2002).

At altitudes where GCs appeared (4.5-6.5 km), the rate of change in potential temperature with altitude recorded by the radiosondes at 1700 and 1800 UTC was
0-5 K km\(^{-1}\) (Fig. 4f). It is important that a small region of negative rate is observed at ~5 km, which coincides with the typical location of GCs, implying that GCs formed due to the release of instability through the Kelvin-Helmholtz mechanism. In the meantime, the wind speed exhibited an increasing trend with time within the WCB (4.2 km) (Fig. 3b). This increased the wind shear from ~4 to ~8 m s\(^{-1}\) km\(^{-1}\) at 1700 and 1900 UTC, respectively, at the top layer of the WCB, which also lowered from 4.5 to 3.5km (Fig. 2h). Although a radiosonde cannot provide representative data over a wide area, the above measurements indicate the occurrence of instability and a triggering mechanism that can explain the presence of GCs. As stated earlier (beginning of section 4), GCs demonstrated a core of \(Z_h > 10\)dBZ, which increased to 20-33 dBZ as the warm front approached the Chilbolton site. GCs also demonstrated high \(Z_{DR}\) at cloud tops (> 2 dB), which indicates regions of newly formed ice, in which \(Z_{DR}\) fall streaks originated (e.g., Fig. 7e; range = 85-95 km).

In Figs. 7a-h, we present the evolution of a GC. At ~1928 UTC (Figs. 7a,e), a GC was triggered at approximately 90-km distance from CFARR, above the WCB. The Doppler velocity below the GC was ~ -23 m s\(^{-1}\), but only ~ -14 m s\(^{-1}\) in the newly formed GC, indicating wind shear at the top of the WCB (Fig. 7a). The above observations suggest that the trigger that caused the formation of GCs was the wind shear at the WCB top layer (Kelvin-Helmholtz instability). Through this mechanism, significant amounts of liquid water from the WCB are lofted via turbulence and weak updrafts. In the example presented in Figs. 7i-k, the aircraft measured high cloud droplet concentration and LWC (up to 10 cm\(^{-3}\) and 0.3 g m\(^{-3}\), respectively) at the rear region of a newly formed GC (located within the WCB at 55 < range < 65 km, 2 < altitude < 3 km). It is important to highlight that no positive values of Doppler velocity were recorded by the vertical-pointing radar (Fig. 3b). According to the CAMRa, GCs were generally forming at ranges > 30km away, passing overhead the vertical-pointing radar as dissipated cells or \(Z_{DR}/Z_h\) fall streaks. In addition, due to the short time of GC lifecycle (5-15 min) and their narrow updraft region (typically <5 km), it is not highly likely that a GC was captured overhead the radar during its genesis time.
FIG. 7. RHIs of (a)-(d),(i) $Z_H$ (dBZ) and (e)-(h),(j) $Z_{DR}$ (dB). The horizontal grayscale line shows the aircraft track between 1903:00 and 1907:48 in (i),(j) (white for the final and black for the starting position). The mean temperature recorded along the track was ~ -7°C. The black contour line in (a)-(d) is for Doppler velocity equal to -20 m s$^{-1}$ and in (e)-(h) for $Z_H = 15$ dBZ. In (a), the red contour presents wind shear $dV/dz = 7$ m s$^{-1}$ km$^{-1}$. (k) LWC and cloud droplet concentration along the aircraft track in (i),(j).
The GC was further developed in height 10-20 min later (Figs. 7b,f and 7c,g), which caused the intensification of the aggregation and riming processes and affected the $Z_H$ parameter ($Z_H$ increased to $\sim$32dBZ). At the later stage (Fig. 7g), a distinct region of very high $Z_{DR}$ and low-medium $Z_H$ was observed (up to 4 dB, <17dBZ) at the GC top. This suggests regions of newly formed (as there was not such a region earlier) pristine ice plates/dendrites at -15°C (Fig. 7f) (e.g., Kobayashi 1961; Magono and Lee 1966; Schrom et al. 2015; Bailey and Hallett 2009; Schrom and Kumjian 2016). The fact that these $Z_{DR}$ structures, which indicate the presence of unrimed ice particles, typically form at the later stages of GC formation suggests that they result from ice nucleation in regions where the initial convective feature has mostly decayed, but that significant regions of supersaturation still remain. Finally, as the GC core sediments/precipitates into the WCB (Figs. 7d,h), the associated $Z_H$ structure changes from a largely vertical orientation to being more horizontally elongated due to strong westerlies within the WCB. The evolution of this GC shows the connection between fall streaks and GCs. Another example is shown in Figs. 12c and 12f (40 < range < 60 km, 2 < altitude < 4 km).

c. The $Z_{DR}$ and $Z_H$ ice fall streaks

1) GENERAL CHARACTERISTICS

At 1800 UTC, two zones with different radar polarimetric traits were detected within sheared ice fall streaks. The $Z_{DR}$ fall streaks appeared as slanted zones of significantly high-$Z_{DR}$ (>1.5 dB) and low-$Z_H$ (<15dBZ) values, containing pristine ice crystals (hexagonal plates and dendrites) (e.g., Andrić et al. 2013; Schrom et al. 2015; Schrom and Kumjian 2016). The $Z_H$ fall streaks, which appeared as a zero-$Z_{DR}$ (-0.5 to 1 dB) and high-$Z_H$ (>15dBZ) zone, were typically observed beneath a related $Z_{DR}$ fall streak, containing mostly aggregated and rimed ice crystals. Essentially, these fall streaks were represented by a bipolar zone of high-low-$Z_{DR}$ (or low-high $Z_H$) zones [similar fall streaks have been investigated by Bader et al. (1987); Kumjian et al. (2014); Oue et al. (2015)]. These zones originate in and descend from GCs (section 5b), as ice crystals fall into sheared flow (Kumjian et
al. 2014); $Z_H$ fall streaks can be also enhanced by pristine ice crystals precipitating into it from the overlying $Z_{DR}$ fall streak.

Examining the entire in situ and polarimetric radar dataset, observations of $Z_{DR} > 1.5\text{dB}$ were mostly (> 60% of the observations) linked with low ice concentrations (< 2 L$^{-1}$) and small-sized ice particles (< 800μm) (Figs. 8a,b). Taking into account previous literature (e.g., Schrom et al. 2015; Schrom and Kumjian 2016) and the fact that the data around the ML (< 1 km) were excluded, it seems that large $Z_{DR}$ values are mainly linked with small pristine ice crystals in small concentrations within the $Z_{DR}$ fall streaks. Figure 8c also shows that regions of $Z_{DR} < 1 \text{dB}$ are partially associated (by >45%) with observations of $Z_H > 15\text{dBZ}$ due to $Z_H$ fall streaks, while regions of $Z_{DR} > 1 \text{dB}$ are associated (by 60%-70%) with smaller $Z_H$ ($Z_H < 15\text{dBZ}$) due to $Z_{DR}$ fall streaks. It should be highlighted that regions of lower-$Z_{DR}$ values (in general, < 1.5 dB, such as $Z_H$ fall streaks) were observed to have a larger variety of ice particle number concentrations (from < 4 to 10 L$^{-1}$) and mean size (from < 500 to 3600μm). Smaller crystals in larger concentrations can be explained by the effects of secondary ice production (see section 5d). To conclude, although the $Z_H/Z_{DR}$ fall streak radar polarimetric boundary is not particularly clear, a rough threshold for distinguishing them could be $Z_H \sim 15\text{dBZ}$, $Z_{DR} \sim 1\text{dB}$.

**FIG. 8.** Probability distribution functions (y axis is percentage per bin) of different $Z_{DR}$ thresholds for in situ (a) ice number concentration (bin size 0.25 L$^{-1}$), (b) mean ice particle diameter (bin size 200μm), and (c) $Z_H$ (bin size 5dBZ). The data used for (a),(b) come from regions where the aircraft flew through $Z_{DR}/Z_H$ fall streaks. The data used for (c) come from the 3-GHz radar for ranges of 30-90 km and elevations >1 km (to avoid the ML and focus on fall streaks) between 1661:21 and 1928:28 UTC.
An example of the evolution of large-scale, moderately intense $Z_{DR}/Z_H$ fall streaks is illustrated in Fig. 9. The first fall streak started to form at ~1549 UTC (Figs. 9a,e), located at range = 100-110km and height = 4-6 km. In this region, a spatially more extensive but less intense, in terms of $Z_H$ ($Z_H < 17$dBZ), GC occurred (compared to the GC investigated in section 5b). At this stage, the $Z_{DR}$ fall streak was relatively modest, occasionally approaching 1-2.5 dB. The corresponding $Z_H$ structure was also relatively modest, exhibiting values of ~15dBZ, which was an enhancement of ~10dBZ over the $Z_{DR}$ fall streak. As the wind speed in the WCB appears to increase ~25 min later (1613 UTC; Figs. 9b,f), both $Z_{DR}$ and $Z_H$ fall streak signals become more obviously enhanced (up to 2.7dB and 24.5dBZ, respectively). The $Z_H$ fall streak coincided with a region of $Z_{DR} \sim 0$ dB and is first observed in a region of enhanced Doppler velocity ($> 20$ m $s^{-1}$). As this region was part of the WCB (which transported large amounts of liquid water), it is highly likely that ice crystals became intensely rimed and less oblate within it. Ten minutes later (~1623; Figs. 9c,g), the $Z_H$ fall streak was represented by $-0.6 < Z_{DR} < 1.5$dB and $Z_H > 15$dBZ, containing large (~984μm) aggregated and heavily rimed dendrites (Fig. 9i) in concentrations of ~1.4 L$^{-1}$ (Fig. 9m). The $Z_H$ fall streak was an almost-glaciated zone, typically exhibiting LWC $\sim0$ g m$^{-3}$ and IWC $>0.01$ g m$^{-3}$.

After a further 30 min (~1655), the fall streak was fully developed, demonstrating the largest length of all the observed fall streaks (~60 km) during the day and an expansion rate of ~17 m s$^{-1}$. Although the $Z_{DR}$ fall streak was characterized by $Z_H < 15$dBZ and $Z_{DR} > 1$ dB due to small concentrations of ice dendrites and plates (Figs. 9d,h), it was actually divided into two sub-fall streaks: one (Figs. 9d,h; range = 65-85 km, $Z_{DR} > 1.5$ dB, $T \sim -13^\circ$C) with small (~580μm) ice plates (Fig. 9j; frames 3-4) and another (Figs. 9d,h; range = 55-65 km, $Z_{DR} \sim 1.5$ dB, $T \sim -13.5^\circ$C) with large (400-1400μm) ice stellars/dendrites/dendrite aggregates (Fig. 9j; first two frames; Fig. 9n), both in small concentrations (<1 and < 5 L$^{-1}$, respectively). We speculate that the sub-fall streak, which contained ice stellars/dendrites, formed when small and light ice plates drifted away from the initial $Z_{DR}$ fall streak due to strong westerlies and remained within the WCB for a longer time. As a result, these plates grew into ice dendrites/stellars due to the high supersaturation regime at ~-
13°C (Fig. 9d) (Bailey and Hallett 2009). In the transitional zone between the \(Z_{DR}\) and \(Z_H\) fall streak, (Figs. 9d,h; range = 90 km, height = 3.5 km), small graupel particles mixed with slightly larger pristine plates were observed in small sizes (\(~460\mu m\)), but in larger concentrations (up to 8 L\(^{-1}\); Fig. 9j; last two frames). In total, although this enhanced \(Z_{DR}\) fall streak reached the ML retaining its characteristics, there was a decrease of \(Z_{DR}\) slightly above the ML, presumably due to aggregation.

An important conclusion that arises from both sections 5b and 5c(1) is that the slope of the fall streaks increased as the warm front was approaching. In particular, the slope of the \(Z_{DR}\) fall streak at 1927 UTC (Figs. 7a,e) was \(~5°\), compared to \(~3°\) and \(~7°\) at earlier (1656 UTC) and later (2007 UTC; Fig. 7d) times. It seems that the wind speed intensity of the WCB (discussed in sections 4 and 5a) and, thus, the wind shear intensity at its upper boundary, played an important role in the formation and shape of the fall streaks. Thus, stronger wind shear might cause stronger updrafts, forming larger and heavier hydrometeors in a shorter period of time, which fall faster toward the surface.
FIG. 9. RHIs of (a)-(d) $Z_H$ (dBZ) and (e)-(h) $Z_{DR}$ (dB). The grayscale line shows the aircraft track between 1621:13 and 1624:22 UTC in (c),(g) and 1652:39 and 1659:06 UTC in (d),(h) (white for the final and black for the starting position). The black isoline in (a)-(d) is for Doppler velocity equal to -20 m s$^{-1}$ and in (e)-(h) for $Z_H = 15$ dBZ. (i),(j) 2D-S images captured during the flight track at (c),(g) and (d),(h) at temperatures -7.8° and -12.9°C, respectively. (k),(m) IWC/LWC and ice number concentration/ice mean diameter, respectively, for the aircraft track in (c),(g). (l),(n) As in (k),(m), but for the aircraft track in (d),(h).
2) THE IMPACT OF THE Z_H FALL STREAK ON THE SURFACE PRECIPITATION

The identification of the Z_H fall streak is very important, as it is directly related to precipitation enhancement at the surface. In Fig. 10, we present the evolution of the Z_H/Z_{DR} fall streak dipole, which was described in section 5c(1), with the corresponding rain rate product from the Met Office operational radar network (NIMROD).

The Z_H fall streak remained elongated (length ~60 km) for ~30 min (1656-1717 UTC) before dissipating, exhibiting up to Z_H ~ 28dBZ. In the initial stages (1705 UTC; Figs. 10a,d), it appeared to be moving slantwise downward, being represented by Z_{H, 3GHz} > 20dBZ and Z_{DR} ~ 0 dB. As the Z_H fall streak evolved over the next hour, ice crystals moved slantwise downward, with aggregation and riming leading to greater and greater enhancements of the Z_H signal. At 1804 UTC, a region of significantly high Z_H (30-47dBZ) was observed at the ML (~ 0.7 km). Such enhanced Z_H values imply heavy aggregation due to higher temperatures (> 23.5°C; Fig. 2d) and, thus, “stickier” ice crystals. Comparing the surface rain rate graphs (Figs. 10i-l) with the RHIs, it seems that the surface precipitation is strongly affected by the Z_H fall streak. In particular, the rain rate increased (at range = 30 km; Figs. 10i-l) from < 1 to 5 mm h\(^{-1}\). This suggests that fall streaks that are initiated near cloud top and develop over time scales of approximately an hour have a significant impact on surface precipitation rates downstream.
FIG. 10. As in Fig. 9. (i)-(l) Surface rain rate (NIMROD) along the radar range of the RHIs in (a),(e); (b),(f); (c),(g); and (d),(h), respectively
Finally, although a copolar correlation coefficient $\rho_{HV}$ was not available in order to estimate the ML height (e.g., Giangrande et al. 2008; Boodoo et al. 2010), $Z_{DR}$ (2-4 dB) peaked around the ML at different heights in (range = 32 km) and out (range = 25 km) of the $Z_H$ fall streak (Fig. 11). In particular, at 1800 UTC, the ML was located at 0.8 km (Fig. 4a), which broadly agrees with the $Z_{DR}/Z_H$ peak (Brandes and Ikeda 2004; Houze 2014, p. 144) of the 25-km vertical profiles (Fig. 11, green lines). However, a slight lowering of the $Z_{DR}$ peak height was observed within the enhanced $Z_H$ region (by 0.2-0.4 km). The local depression of the melting layer was possibly caused by the melting of large aggregates (Stewart 1984; Stewart et al. 1984; Oraltay and Hallett 2005; Griffin et al. 2014). This is important, as ~0°C isothermal layers can be produced close to the surface, allowing (partially melted) ice to precipitate (Findeisen 1940; Szeto et al. 1988; Ryzhkov et al. 2011; Griffin et al. 2014).
d. Secondary ice in the \(Z_H\) fall streak

In this section, we try to analyze concurrent in situ and radar data in order to detect, determine, and investigate the characteristics of regions that appear to be greatly influenced by secondary ice processes.

Between 1912:54 and 1917:42 UTC, the aircraft passed through a region (range = 50-70 km) of low-medium \(Z_H\) (8-12 dB) and low \(Z_{DR}\) (0.5-1 dB), which was located at the entrance of the WCB. Although this region was located within a weak signal cloud top, we speculate that this region is a mature/dissipated \(Z_H\) fall streak. At the rear of this mature \(Z_H\) fall streak (range = 65-80 km), which was located within the WCB entrance (Doppler velocity > 20 m s\(^{-1}\)), the aircraft measured enhanced LWC and cloud droplet concentration (up to 0.17 g m\(^{-3}\) and 16.7 cm\(^{-3}\), respectively). In contrast, at the bottom of the WCB (range = 50-65 km), both cloud droplet concentration and LWC decreased by 10 cm\(^{-3}\) and 0.10 g m\(^{-3}\), respectively. This liquid water removal may imply intense riming, which is a mechanism assisting in ice multiplication (Mossop and Hallett 1974; Choularton et al. 1978, 1980). The occurrence of small (\(~300\mu m\) in average) ice columns (Fig. 12g; 2D-S frames 1-4) in high ice number concentrations (up to 37.8 L\(^{-1}\)) and the enhanced IWC (0.11 g m\(^{-3}\)) observed at temperatures ~ -4.8°C imply that Hallett-Mossop is the possible ice multiplication mechanism here. Some aggregated ice columns that were observed at range = 65-80km (Fig. 12g; CIP-100 frames 1-2) formed due to the occurrence of ice columns within a region of high LWC.

Farther to the east (Figs. 12c,f,h,i), a region of quasi-spherical rimed ice particles (Fig. 12g; 2DS frames 4-5, CIP-100 frame 3) and supercooled water drops mixed with some ice columns (range = 43-55km; T ~ -4°C; \(Z_H\) up to 28.6dB and \(Z_{DR}\) ~ 0.5-1dB) was observed, followed by another region of mostly ice columns (range = 31-43km; T ~ -5°C). As this region belonged to a short and newly formed \(Z_H\) fall streak originating in the core of a GC (range ~ 53km, altitude ~ 3km), processes such as riming and ice multiplication could be at early stages. This may explain the increased, but lower than the previous secondary ice region, ice concentrations (up to 14.1 L\(^{-1}\)). Decreased LWC (< 0.005 g m\(^{-3}\)) and cloud droplet concentration (< 0.4 cm\(^{-3}\)) could be measured due to riming process and the aircraft position (outside
the WCB). It should be noted that some pristine ice crystals from the Z_{DR} fall streak above (range = 40-50km) rimed (an example in Fig. 12; CIP-100 frame 4) and became heavier when they moved into the WCB. As a result, they might fall into the Z_{H} fall streak, which was finally characterized by small quasi-spherical (and possibly mixed phase) ice particles/supercooled water drops, but also of ice columns and ice lollies (Keppas et al. 2017).
FIG. 12. As in Fig. 9. (g) 2D-S and CIP-100 images. The time and range from the Chilbolton radar are indicated at the bottom of the figure in order to be comparable with the radar scans. (h) Green curve shows the cloud droplet concentration. The mean temperature along the aircraft track was $\sim -4.8^\circ C$. 
6. Conclusions

On 21 January 2009, multiple frontal zones affected the U.K. weather due to a maturing depression, which originally formed over the North Atlantic Ocean. In the work presented, the dynamics and microphysics of mixed-phase clouds, with embedded convective elements, associated with the warm front of this system are investigated, comparing high-resolution (0.28° beam) dual-polarization S-band radar with in situ aircraft data. The latter data collected by 2D-S probe (providing better resolution than 2D-C) were used together with CIP-100 probe data in order to provide a wider size range of measured particles. It should be noted that this is the first time that a warm front is comprehensively investigated with high-resolution data.

A few studies previously investigated warm fronts and GCs, comparing both in situ and radar datasets. Here, we discuss some of their results, which generally come to an agreement with the results of the present study. Herzegh and Hobbs (1980) found that large ice particles within GCs grew by riming, deposition, and aggregation as occurred in \( Z_H \) fall streaks. They also suggested that GCs work as feeders of ice to stratiform clouds below. According to Matejka et al. (1980), as GCs mature, they tend to be glaciated. They supported that surface precipitation is mainly associated with embedded convection in warm fronts. Hogan et al. (2002) demonstrated some evidence that embedded convection within a warm front is triggered by Kelvin-Helmholtz instability. However, they found that embedded convection was linked with narrow, vertical high-\( Z_{DR} \) zones of ice columns, which coincided with updrafts and was a feature that was not observed in the present study. Murakami et al. (1992) observed only a shallow layer of supercooled droplets above the warm frontal zone. Additionally, they found that GCs provided a favorable environment for ice crystals to grow rapidly, which agrees with Plummer et al. (2014). Finally, Plummer et al. (2015) noticed that ice mainly grew in regions below GCs, where enhanced moisture was observed.

As a conclusion, the general structure of the warm front is schematically summarized in Fig. 13, according to the airborne in situ and ground-based
polarimetric radar measurements. The main conclusions of the study are outlined as follows.

Regarding the macroscopic characteristics of the investigated warm front:

- The height of cloud tops gradually decreased from 6 to 4.5 km as the warm front was approaching. Cloud-top temperatures were, generally, > -25°C, which implies that the ice in the cloud system was formed via primary heterogeneous processes and potentially enhanced by secondary multiplication processes.
- The height of the 0°C level, or the melting layer (ML), was first located at 0.7 km, rising by almost 1 km after the passage of the warm front.
- The horizontal dimension of the cloud mass was larger than 120 km, the maximum range of the RHIs presented in this work.

The cloud mass consisted of distinctive features that demonstrated definite radar polarimetric and/or microphysical characteristics. These features are:

- The warm conveyor belt (WCB), initially depicted by a group of multiple zones of enhanced Doppler velocity. These multiple zones merged into a single one, which decreased in altitude as the warm front moved farther inland, away from the coast. High radar Doppler velocities (20-30 m s\(^{-1}\)) and radiosonde wind speeds (up to 26 m s\(^{-1}\)) were recorded within the WCB, presenting, essentially, a fair agreement for wind directions along the radar azimuth. The WCB was vitally important for transporting large amounts of liquid water (up to 0.37 g m\(^{-3}\)) into the system, offering at the same time an ideal regime for secondary ice production at temperatures ~ -6°C.
- The generating cells (GCs), formed at the unstable (based on potential temperature profiles) layer above the WCB, where vertical wind shear triggered Kelvin-Helmholtz instability. The GCs were represented by a core of high \(Z_H\) (10-33 dBZ)/~0-dB \(Z_{DR}\) and a shell of low \(Z_H\) (<10 dBZ)/high \(Z_{DR}\) (up to 4 dB).
- The ice fall streaks, formed as GCs were moving into a sheared flow caused by the WCB. The strong westerlies “converted” the GCs’ high-\(Z_{DR}\) shells and high-\(Z_H\) cores into an ice fall streak, which consisted of two individual fall streaks with different polarimetric characteristics: a \(Z_{DR}\) and a \(Z_H\) fall streak, respectively. Ice fall streaks exhibited lengths of up to 60 km and slope of 3°.
which increased with time (up to 7°) due to the intensification of the GCs. The $Z_{DR}$ fall streaks were divided into two low-$Z_H$ (<15 dBZ) subzones: (i) a zone of ice plates with $Z_{DR} > 1.5$ dB and (ii) a zone of stellar/dendrites with $Z_{DR} \sim 1.5$ dB, which might form from small ice plates that were moving from GC shells to the high supersaturated regions of the WCB. As icing can be harmful for aviation in dendritic ice regions, this can help to improve existing icing detection algorithms (Serke et al. 2008; Pitertsev and Yanovsky 2011; Ellis et al. 2012). Beneath the $Z_{DR}$ fall streaks, $Z_H$ fall streaks were represented by high $Z_H$ (>15dBZ) and $Z_{DR} \sim 0$ dB, consisting mostly of rimed and/or aggregated ice crystals in variable sizes (from < 500 to 3600μm) and concentrations (from < 4 to > 10 L$^{-1}$). The $Z_H$ fall streaks that form higher up in the clouds can affect the surface precipitation. The time and spatial evolution depends on the profile of wind direction and speed in the clouds. As an illustration, a case where a $Z_H$ fall streak enhanced the surface precipitation by 4 mm h$^{-1}$ an hour after its genesis was presented.

- The secondary ice regions, which presented large ice concentrations (up to 37.8 L$^{-1}$) and mostly consisted of ice columns. Such regions were located within the intersection region of the $Z_H$ fall streaks with the WCB, where riming was intense at temperatures ~ -5°C, and liquid water was dramatically depleted, potentially affecting the cloud lifetime. In the present paper, both heterogeneous ice nucleation and secondary ice formation were observed in mixed-phase clouds, which are highly uncertain processes. Heterogeneous ice nucleation plays an important role in the formation of GCs, as new ice forms from freezing liquid water transported aloft along the rear updraft region of GCs. As in situ measurements within GCs and ice fall streaks are limited, more observations should be performed aimed at investigation of the cloud processes (aggregation, riming, and ice multiplication) in order to obtain a deeper understanding of cloud microphysics. Except for observing real GCs, they could also be created and observed in laboratories. Comprehensive datasets, which could describe the structure and evolution of such clouds, could be used for the validation of complex microphysics models (Stoelinga et al. 2003; Keeler et al. 2016a).
FIG. 13. A schematic representation summarizing the investigated processes and polarimetric signatures of the frontal clouds observed.

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CHAPTER FIVE

ICE LOLLIES: AN ICE PARTICLE GENERATED IN SUPERCOOLED CONVEYOR BELTS

S. CH. Keppas, J. Crosier, T.W. Choularton and K.N. Bower

The following chapter has been published in the American Geophysical Union Geophysical Research Letters. Although this paper was published prior to the one in chapter four, we feel that it should be placed here, because chapter four is a comprehensive analysis of the warm front on 21 January 2009, while the paper in chapter five focuses on the formation of a specific ice particle within the same warm front. S. CH. Keppas carried out the analysis and wrote the paper. J. Crosier and T.W. Choularton assisted with the analysis phase and supervised the research. K. N. Bower collected airborne measurements.
Ice-Lollies: An ice particle generated in supercooled conveyor belts

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Abstract

On 21st January 2009, a maturing low pressure weather system approached the UK along with several associated frontal systems. As a part of the APPRAISE-Clouds project, an observational research flight took place in southern England, sampling the leading warm front of this system. During the flight, a distinctive hydrometeor type was repeatedly observed which has not been widely reported in previous studies. We refer to the hydrometeors as “drizzle-rimed columnar ice”, or “ice-lollies” for short due to their characteristic shape. We discuss the processes that led to their formation using in-situ and remote sensing data.

1 Introduction

Remote sensing and in-situ observations have made a significant contribution to the study of cloud microphysical processes. Locatelli and Hobbs [1987] provided a detailed analysis of the structure of a warm front using radar and radiosonde data, identifying the typical patterns in temperature, wind and precipitation, including the existence of non-uniform rainbands. Browning and Roberts [1994] further developed the understanding of frontal cyclones using model output and satellite images, and discussed the dynamical factors leading to the formation of various cloud features.

Dearden et al [2016] used the WRF numerical model to simulate two summer cyclones to gain an understanding of sensitivities to ice microphysics, including the role of primary vs secondary ice formation. Primary ice formation describes the nucleation of new ice particles via homogenous freezing of liquid droplets at cold
temperatures (<-37°C, Koop et al., 2000), or via heterogeneous processes involving Ice Nuclei (IN, Vali et al., 2015). Secondary ice formation processes, often referred to as “ice multiplication processes”, increase the number of ice particles in clouds by fragmenting existing ice particles in various ways, such as freezing and shattering of water drops [Johnson and Hallett, 1968; Pruppacher and Schlamp, 1975; Lawson et al., 2015], rime splintering [Mossop and Hallett, 1974] fragmentation following ice crystal collisions [Hobbs and Farber, 1972; Vardiman, 1978] and fragmentation of ice due to evaporation and melting [Knight, 1979]. Dearden et al. [2016] found that in deep frontal systems, precipitation was dominated by primary ice production mechanisms, and that secondary processes had no significant impact.

In contrast to this modelling study, observations presented in Crosier et al [2011 and 2014] and Lloyd et al [2014] showed that secondary ice processes can dominate ice crystal formation in a wide variety of frontal clouds. These studies implicate the Hallett-Mossop (H-M) process [Mossop and Hallett, 1974], which is thought to be active in the temperature region -3°C and -8°C, as being the source of the secondary ice particles - manifesting as abnormally large numbers of pristine columnar ice crystals at relatively warm temperatures. The leading theory describing the H-M process is presented by Choularton [1980], which suggests that supercooled droplets with diameter >24μm explosively freeze when they come into contact with pre-rimed ice particles. The explosive freezing is thought to occur as a result of the formation and subsequent fracturing of an ice crust which forms on the surface of the particle - the subsurface water freezes after the crust forms which imparts stress on the crust, resulting in fracturing. The fracturing of the surface crust is thought to be responsible for the generation of small splinters of ice which typically grow into observable columnar ice particles due to the temperature at which the whole process operates. As the H-M processes requires large liquid water droplets, the process is most active in convective regions where large droplets can be generated. Predicting the formation of convective features on appropriate scales may impose a limit on the ability of numerical models to assess the importance of the H-M process, in addition to fundamental uncertainties associated with the H-M process itself.
In this paper, we present observations of a warm front which is sampled extensively. We present particle imagery showing a distinctive type of ice particle which was repeatedly observed in the warm frontal cloud. This ice particle has never been observed in significant concentrations before. We discuss the microphysical and dynamical processes which lead to the formation of these ice particles through the use of in-situ aircraft observations and radar data.

2 Instrumentation and data

Measurements of cloud microphysical properties were collected on 21st January 2009 between 15:00 UTC and 20:30 UTC as part of the APPRAISE-Clouds (Aerosol Properties, PRocesses And InfluenceS on the Earth’s climate) project. On this day a low pressure system approached the British Isles from the west. The system consisted of multiple frontal boundaries, with multiple rainbands affecting the country, especially in the SW regions of England and Wales. The rainbands consisted of mixed-phase clouds with some embedded convective elements. Observations were collected during the passage of the leading warm front.

In situ measurements were collected on board the UK Facility for Airborne Atmospheric Measurement (FAAM) BAe-146 aircraft, while remote-sensing measurements were obtained from the Chilbolton Facility for Atmospheric and Radio Research (CFARR, located in Southern England; 51.15°N, 1.44°W). The BAe-146 aircraft obtained in-situ observations along an azimuth of approximately 255° relative to the CFARR radar site, at horizontal distances ranging from 0 km (overpassing CFARR) to 100 km. The Chilbolton Advanced Meteorological Radar (CAMRa) system continuously scanned along the same 255° azimuth, leading to co-located in-situ and radar observations of the frontal cloud. The CAMRa is a dual-polarisation Doppler radar operating at 3 GHz, with no significant two-way attenuation (< 1 dB km⁻¹) even in cases of heavy precipitation (e.g. 16 mm h⁻¹) [Curry, 2012]. It has a large 25-m antenna providing high-resolution data (0.28° beam width, 0.3 km in range) [Goddard et al., 1994]. The CAMRa provides information on the microphysical properties of clouds: reflectivity factor (Z_H) is proportional to the square of the total hydrometeor mass [Doviak et al., 1979; Brown
and Wood, 2007]; differential reflectivity ($Z_{\text{DR}}$) provides information on the shape of hydrometeors with larger values for oblate and horizontally orientated 2D-planar particles, and low values for quasi-spherical particles [Straka et al., 2000; Wolde and Vali, 2001; Kennedy and Rutledge, 2011; Andric et al., 2013; Bechini et al., 2013; Kumjian, 2013; Schuur et al., 2014; Moisseev et al., 2015; Schrom et al., 2015; Schrom and Kumjian, 2016]; doppler velocity provides dynamical information by characterizing the velocity of the hydrometeors along the axis of the radar beam [Brown and Wood, 2007].

The BAe-146 aircraft was equipped with a suite of microphysical probes to measure a wide range of hydrometeors (see Crosier et al. [2014] and references therein). To summarize the key instrumentation: a CDP (Cloud Droplet Probe) Mie scattering probe was used to measure the number and size of cloud droplets [Lance et al., 2010]; 2D-S (Two-Dimensional Stereo), CIP-15 and CIP-100 (Cloud Imaging Probe) Optical Array Probes (OAP) were used to provide shadow imagery of particles from 10-1280, 15-960 and 100-6200 µm respectively [Knollenberg et al., 1970; Lawson et al., 2006]; ambient temperature was measured on a Rosemount inlet.

Liquid Water Content (LWC) was estimated from the CDP cloud droplet number size distributions. Number size distributions for larger particles were calculated from the OAP data according to Crosier et al. [2011, 2014]. OAP images were filtered using an inter-arrival time threshold to remove artefacts associated with shattering of large particles on the OAP hardware which are measured as abnormally closely spaced clusters [Field et al., 2006]. Particle size was estimated from images as the diagonal of the bounding box. Ice Water Content (IWC) was estimated from the OAP number size distributions (2DS and CIP-100) using estimates of particle mass following Heymsfield et al. [2007]. There are significant uncertainties of the order 50% which arises from the diversity in OAP image processing options. In this study we do not focus on quantitative number or mass concentrations. Our analysis focuses on particle imagery, and our use of derived quantities is purely qualitative.
Figure 1. (a) Schematic representation of the formation of an ice-lolly and (b) of ice-lollies with spikes on edge (see section 3.3 for the description). (c) Ice-Lollies observed by the 2DS as a function of temperature (left axis in °C) and altitude (right axis in km).

3 Ice-Lolly formation

During the collection of in-situ data on 21st January 2009, a significant number of the hydrometeors observed exhibited a characteristic shape not widely seen in any previous studies with the exception of one isolated example reported by Korolev et al. [2004]. The hydrometeors in question appear to be a combination of single columnar pristine ice crystals (needle or column) with a single drizzle sized water droplet (typically 300µm diameter). We call these hydrometeors “drizzle-rimed columnar ice”, or “ice lollies” due to their similarities in shape. Example images of ice-lollies from the 2DS probe are shown in figure 1c. Labelling and understanding different hydrometeor types gives insight into processes active in the cloud system. In our specific case the ice lollies demonstrate interactions between warm rain, primary ice and secondary ice processes. In sections 3.1, 3.2 and 3.3, we discuss the source of the liquid water droplets, the source of the pristine columnar ice, and the formation of ice lollies respectively.
3.1 Liquid water in the Warm Conveyor Belt

On 21 January 2009, a Warm Conveyor Belt (WCB, *Browning and Roberts* [1994]) passed over the southwest regions of the UK, lifting warm moist air from the Atlantic Ocean which led to widespread cloud formation. The WCB is well depicted by the radar Doppler velocity parameter in figure 2b as a slanted region of high velocities (~ -20 m s⁻¹). Near the radar, the WCB is located at higher altitudes (~3km altitude), whereas further away to the southwest the WCB is located close to the surface (~1.5km altitude). The aircraft flew through the WCB (figure 2b, from left to the right) and surrounding regions, and data from one pass through the WCB at T=-5°C is shown in figure 2a, with the associated aircraft flight track highlighted in figure 2b. In the core of the WCB, which the aircraft passed through between 65-80 km from the radar, significant mass concentrations of liquid water were observed (>0.15g m⁻³), whilst relatively little ice mass was observed (generally <0.025g m⁻³). As the aircraft exited the core of the WCB and sampled below the overlying elements of the WCB (30-65km from the radar), the mass concentration of liquid water reduced greatly (<0.01g m⁻³) and the mass concentration of ice increased, occasionally exceeding 0.05g m⁻³). Hence, it seems that the WCB was feeding the cloud system with water vapor, and directly led to the formation and growth of the water droplets. During the same measurement period, water droplets with diameters <12μm and >24μm were observed with average number concentrations of 0.13cm⁻³ and 1.99cm⁻³ respectively (with peak number concentrations of 0.27 and 4.94cm⁻³ respectively). According to current understanding, the existence of those specific sizes of water drops is a key ingredient for the ice column formation via the H-M process, and so for the ice-lolly formation as will be discussed in section 3.3.
Figure 2. (a) LWC, IWC and ice number concentration, (b) Doppler velocity (dashed line for -20 m s⁻¹), (c) reflectivity factor (Z_H), (d) differential reflectivity (Z_{DR}). Figures (b) to (d) refer to the period between 19:16:12 and 19:17:42 UTC at an azimuth of 255°, pointing roughly WSW. In figure 2b negative Doppler velocity values indicate flow towards the radar. The grayscale line shows the aircraft track between 19:11:26 and 19:17:42 (black for the starting position and white for the final position). Figure 2a corresponds to the aircraft path of figure 2b, which is from right to left.
3.2 The origin of the pristine columnar ice

Particle imagery shows that pristine planar ice crystals were present near cloud top (~4 km, ~ -12.5°C) (Figure 3a). Average ice particle number concentration and size in this region are ~2 L\(^{-1}\) and ~670μm respectively. This observed crystal habit is consistent with the expected habit based on the ambient temperature [Kobayashi, 1961]. These crystals are formed via heterogeneous processes due to the relatively modest cloud top temperatures [Rogers and Yau, 1989]. The radar can identify regions of cloud with significant concentrations of planar ice crystals, as these types of particle have strong differential reflectivity signals (Z\(_{DR}\) >2dB) due to their relatively 2-dimensional structure [Seliga and Bringi, 1978; Ryzhkov and Zrnic, 1998; Kennedy and Rutledge, 2011; Andric et al., 2013; Moiseev et al., 2015]. Z\(_{DR}\) features were frequently observed near cloud top which is consistent with ice being formed in convective generating cells near cloud echo top (Kumjian et al., 2014). The association of planar ice crystals with high values of Z\(_{DR}\) was found to be valid near cloud top based on aircraft observations. An example of a similar Z\(_{DR}\) feature at cloud top can be seen in Figure 2d. Below the high Z\(_{DR}\) cloud tops, slanted zero Z\(_{DR}\) (typically between -0.5 and 0.5dBZ) zones were observed. Within these zones more aged/processed ice particles, such as rimed/aggregated ice crystals, were observed in variable concentrations between 1-10 L\(^{-1}\) and variable average sizes between 300μm (for rimed) and 800μm (for aggregates). Some pristine ice crystals high Z\(_{DR}\) regions near cloud top appear to be transported to these zones and are observed in a more processed state (rimed and/or aggregated) as shown in some of the imagery in Figure 3b, c.

Observations at lower levels (T ~ -6.2°C, altitude ~2.3km), which represent more aged/processed ice particles relative to the young/pristine particle observed near cloud top, were observed in similar number concentrations (~9 L\(^{-1}\)) and size (337 μm). These observations were collected in a region of low Z\(_{DR}\), with particle imagery (Figure 3d) supporting the concept of increased processing via riming and aggregation particles precipitate down towards the surface.

At the same altitude and temperature (approx. -6°C), large numbers of pristine columnar ice particles were observed in mixed phase regions, where large amounts
of liquid water were observed. Imagery of these pristine columnar ice particles and the individual water droplets (shown as single pixels due to the coarse resolution of the probes) are shown in figure 3e. Further east (figure 2a) at the same altitude, the average ice crystal number concentration is significantly enhanced relative to regions above the WCB with no liquid water, with a mean value of 20.7L^{-1} (peak values up to 37.8L^{-1}) at 50-70km and 8.9L^{-1} (peak values up to 14.1L^{-1}) at 31-43km range. In these regions mixed ice columns, ice-lollies and rimed crystals were observed in the imagery. The increased number concentrations and relatively small and pristine nature of the observed columnar ice suggests a multiplication mechanism has been active. As mentioned in section 3.1, the drop size distribution is suitable for the H-M rime splintering mechanism to operate, as both <12\mu m and >24\mu m droplets (required for the splinter production) were present at within the relatively narrow temperature regime where the H-M process is active \[Mossop\ and\ Hallett,\ 1974;\ Mossop,\ 1976;\ Choularton\ 1978;\ Choularton,\ 1980\]. This interaction would result in the large number of pristine columnar ice crystals observed.

To summarize, it appears that pristine columnar ice particles are formed as a result of the H-M process. The H-M process is facilitated by the presence of ice particles which form in generating cells near cloud tops, and by liquid water which is formed in the core of the WCB.

3.3 The formation of Ice-Lollies and their polarimetric radar properties

As mentioned in sections 3.1 and 3.2, columnar ice crystals were produced by ice multiplication processes and they were detected within a mixed-phase region, which consisted of ice columns and water droplets. Here, we try to demonstrate that the collision between these water drops and ice columns, and thus the formation of ice-lollies (figure 1a), was possible. In general, the columnar part of ice-lollies was >250\mu m (up to 1400\mu m), while the attached supercooled droplet diameter was >100\mu m (up to 700\mu m). According to Bohm [1992; 1994], such particles would collide with efficiencies >70\%. In addition, the ratio between supercooled droplet and ice column terminal velocities \[Mitchell,\ 1996\] for the
observed range of sizes were mostly >1, which also supports particle collisions. It should be highlighted that ice columns (for Reynolds number ≤50, thus laminar flow) fall with their long (c-) axis perpendicular to their fall direction, which explains why the spherical part was in the middle of some ice-lollies (figure 1c) [Jayaweera and Mason, 1965].

An alternative mechanism, as suggested by Korolev et al. [2004], would be for the long columnar component of an ice-lolly being as a result of depositional growth on a frozen water droplet. However, depositional growth on a frozen droplet at these temperatures would hypothetically result in something resembling a bullet rosette and not a frozen droplet with a single protrusion which is what we observe. Also, evidence for the “drizzle riming” mechanism which we propose is strongly supported by the occurrence of large numbers of pristine columnar ice particles in the vicinity of ice-lollies.

The 2DS probe imaged many ice-lollies (figure 3f) during the scanning period of the radar images (figures 2b, c, d), only a small number of which are shown. In this paragraph we present some radar polarimetric characteristics of the ice-lollies. When the aircraft flew through a high $Z_H$ region ($Z_H>15$dBZ) (figure 2c), low ice-lolly concentrations were measured (~0.4L$^{-1}$). However, as the aircraft was flying towards a region of moderate $Z_H$ (7-15 dBZ), larger concentrations were observed (~3.8L$^{-1}$). Typically, ice-lollies were found in regions with $Z_{DR}$ values between 0.3 and 1.6 dB (figure 2d). The elongated shape of the ice-lollies can justify such values, as shown later in the text. However, regions of 0.3 dB are present due to the existence of less oblate particles (such as large aggregates or quasi-spherical heavily rimed crystals/graupels), which were probably dominating both $Z_H$ and $Z_{DR}$ signals. Using the Discrete Dipole Approximation (public-domain code DDSCAT) [Draine and Flatau, 1994] for particles consisted of a column and a sphere (thus ice-lollies) in various observed size combinations, we calculated the theoretical $Z_{DR}$ of ice-lolly particles. Ice-lollies could lead to $Z_{DR}$ ranging between 0.2 and 3.2dB depending on their specific geometry. In particular, ice-lolly $Z_{DR}$ seems to be dependent primarily on the ratio between the column width and the drop diameter and, secondarily, on the ratio of column length to column width. In our observations the ratio of column width to drop diameter (CWDD) ranged from 0.2
to 0.75, with a typical value of ~0.4. The ratio of column length to width ranged between 2.5 and 10, with a typical value of ~5. For ice-lollies with relatively large droplets frozen onto relatively small columns, with CWDD = 0.2, $Z_{DR}$ values range from 0.2dB to 0.5dB, with longer columns generating the larger $Z_{DR}$ values. For the case of relatively small droplets frozen onto relatively large columns, with CWDD = 0.75, $Z_{DR}$ values are significantly higher, and range from 2dB to 3.2dB, again with the longest columns generating the largest values. Theoretical $Z_{DR}$ values for the most typically observed shape of horizontally oriented ice-lolly are between 1.5-2.2dB. Jayaweera and Mason [1963] showed that columns (diameter=160μm) capped by hexagonal plates (e.g. diameter<=300μm) fall horizontally aligned if the column length is >1000μm. Otherwise, these particle types fall vertically. Capped columns have the most similar geometry to an ice-lolly, and so are likely to behave in a similar manner. However, there is uncertainty in fall orientation due to lack of data.

In general, ice-lollies were found above the melting layer and in the vicinity of the WCB (in or below it), where temperatures were between 0 and -6°C. Nevertheless, some deformed and melted ice-lollies were observed between 0 and 2°C (figure 1c). It is worth noting that both columnar and spherical parts of ice-lollies became thicker as the temperature increased, probably due to growth from vapour diffusion.
Figure 3. (a) Pristine crystals and snowflakes from high $Z_{DR}$ regions near cloud top ($T=-12.5^\circ C$, Altitude=\sim 4km), (b) rimed and aggregated ice crystals from low $Z_{DR}$ regions below cloud tops ($T=-7.5^\circ C$, Altitude=\sim 2.5-3km), (c) rimed crystals from low $Z_{DR}$ regions below cloud tops ($T=-7^\circ C$, Altitude=\sim 2.6km), (d) aged rimed ice crystals from low $Z_{DR}$ regions close to ice-lolly regions ($T=-6.2^\circ C$, Altitude=\sim 2.3m), (e) ice columns and liquid droplets ($T=-5^\circ C$) and (f) ice-lollies observed at mid-levels ($T=-5^\circ C$, 2km altitude). The red square includes some examples of the ice-lollies with side planes (see explanation in the text) imaged by the 2DS. The horizontal dimension of each strip of images is equal to 1.28 mm. In (b), (c) and (d) CIP-100 strips dimension is equal to 6.4mm.
The growth of the columnar part of the ice-lolly appears to be more complicated in some cases. In general, needles and columns can be produced at the same temperatures (~ -5°C), but in different super-saturation conditions [Kobayashi, 1961; Magono and Lee, 1966; Bailey and Hallett, 2009]. However, some ice-lollies were observed with some side planes (growing from the spherical part of some ice-lollies) facing away from the attached column (examples are depicted in figure 3c - in red squares) at temperatures between -4 and -6°C. We speculate that as the frozen droplet breaks due to H-M process or shattering due to freezing, the surface at the breakage point (where the side planes were developed on the frozen drop) becomes roughened. Then, these protrusions grow into needles/spikes due to diffusion (figure 1b). Broadly, ice multiplication can potentially occur during the formation of an ice-lolly as explained above. However, these particular ice-lollies could be a visual evidence of ice multiplication occurrence.

4 Conclusion

On 21 January 2009, a maturing low pressure system over North Atlantic moved towards the British Isles. A warm front associated with the depression passed over the British Isles and was sampled intensively and consisted of large regions of mixed-phase clouds. In these clouds, we observed a hydrometeor type, ice-lollies, for the first time in significant concentrations, which form due to a variety of microphysical processes. Ice-lollies were also observed in significant concentration during BAE146 flight on 23 September 2016 within a warm conveyor belt system. However, the data were collected over the North east Atlantic Ocean, with no supporting radar coverage, which complicates in-situ data interpretation. Aside from these two cases, ice-lollies have not been seen on other research flights. Thus, future missions in warm conveyor belts spanning the H-M zone would provide insight into the importance of ice-lolly formation. The microphysical processes that led to the ice-lollies formation are described and summarized, as in figure 4, as follows:

- Widespread supercooled cloud was formed by a large scale WCB circulation.
• Planar ice crystals and snowflakes were formed via heterogeneous processes near cloud top, and fell through the core of the WCB with contain significant amounts of supercooled liquid water.

• Riming occurring in the WCB led to large numbers of columnar ice particles in the HM secondary ice production zone.

• Ice-lollies were formed due to the collision of water drops with columnar crystals within the H-M region and were observed at temperatures between -6 and 0°C.

• Some ice-lollies appear to exhibit some side planes facing away from the attached column. This could be a visual evidence of ice multiplication occurrence.

• The freezing of supercooled water is involved in the formation of ice-lollies. Also ice-lollies can act as rimers enhancing ice multiplication mechanisms, removing further liquid water from the clouds. These processes can impact on cloud lifetime and precipitation formation.

• Ice-lollies in isolation demonstrate high $Z_{DR}$ due to their elongated shape. Ice-lollies were observed in significant concentrations within regions with $Z_{DR}$ ranging between 0.3-1.6 dB, albeit not in complete isolation from other hydrometeor types. Such values could also be representative of ice-lollies and were confirmed by calculations using the Discrete Dipole Approximation.

• Some ice-lollies were also still found in areas where quasi-spherical graupels/heavily rimed crystals or water drops were the dominant hydrometeor type. In these areas, $Z_{DR}$ tends to be much lower (~0 dB).
Figure 4. A schematic representation of the processes that occurred for the ice-lollies formation. For simplicity and clarity reasons aggregates and planar crystals are not repeated in lower levels.

Acknowledgements and data

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CHAPTER SIX
EVALUATING A HYDROMETEOR CLASSIFICATION ALGORITHM USING IN-SITU DATA. INTRODUCING AN ALTERNATIVE APPROACH.

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We aim to submit this final chapter to the Atmospheric Measurement Techniques (AMT) journal. S. CH. Keppas carried out the analysis and wrote the paper. J. Crosier and T.W. Choularton assisted with the analysis phase and supervised the research. K. N. Bower collected airborne measurements. R. R. Neely processed the radar data.
Evaluating a Hydrometeor Classification Algorithm using In-Situ data.
Introducing an alternative approach.

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Abstract

In this study, the NCAR Hydrometeor Classification Algorithm (HCA) for S-band radars is evaluated using in-situ data collected during the passage of a warm front across the southern UK on 21 January 2009. A detailed evaluation and analysis of the HCA results is conducted using in-situ airborne observations of microphysical properties alongside particle imagery from 2DS and CIP-100 Optical Array Probes (OAPs). In order to evaluate the HCA objectively, we firstly suggest the use of three in-situ parameters calculated from OAP imagery, (weighted shape factor, weighted area ratio and weighted transparency factor) and we conduct a statistical analysis of these parameters. Then, we evaluate the ability of the HCA to distinguish between four different ice hydrometeor classes: a. dry snow, b. wet snow, c. irregular crystals and d. ice crystals, making use of the in-situ parameters,
but also of the probe imagery. It seems that weighted shape factor may be used for distinguishing regions of dry snow (with values between 2 and 7) from regions of pristine ice crystals (~3.5 or larger depending on their shape), while riming zones are characterised by certain thresholds of weighted area ratio (~0.6) and transparency depending on the riming intensity (up to 0.5 in cases of thin crystals). Wet snow observations exhibit a wide range of values due to the presence of spherical liquid droplets and the shape complexity of semi-melted ice crystals and aggregates. Although merged (2DS+CIP-100) parameters seem to produce satisfactory results, parameters calculated from 2DS can provide higher detail in cases that particles are small (<1000μm; e.g. pristine ice crystal regions). Whilst the HCA seems to identify correctly most of the content of the frontal clouds, the accuracy of the results around the melting layer are unclear.

1. Introduction

Microphysical parametrisations, which are used by climate and numerical weather prediction models to simulate the cloud microphysical processes, can affect the predictive model accuracies regarding the intensity or the form of precipitation and the frontal structures [e.g. Forbes and Clark, 2003; Dierer et al., 2009; Efstatiiou et al., 2013; Milbrandt and Morrison, 2013; Gettelman and Morrison, 2015]. Simulating clouds is of high importance, as they make a significant contribution to many earth-atmosphere system processes, such as radiative transfer and precipitation formation [e.g. Schiffer and Roscow, 1983; Stowe et al., 1995; Del Genio et al., 2005; Patra et al., 2005; Slingo, 1990; Stephens, 2005]. As microphysical parametrisations become more and more sophisticated, they need to be validated with representative and accurate observations (e.g. cloud optical depth, cloud water content, melting layer height etc.) [Bauer et al., 2011]. The accuracy of the observations, which are used as input for the numerical models, is also valuable for the minimisation of the model output biases [Lorenz, 1963]. Weather radars can provide information that describes cloud properties and rain rate [e.g. Seliga and Bringi, 1978; Jameson, 1985; Straka et al., 2000; Bringi and Chandrasekar, 2001; Wolde and Vali, 2001; Kennedy and Rutledge, 2011; Andrić et al., 2013; Kumjian, 2013; Moisseev et al., 2015; Schrom and Kumjian, 2016] in
almost real time. Dual-polarisation capabilities have made significant contributions to the improvement of radar data quality [e.g. Ryzhkov and Zrnić, 1995; Brandes et al., 2002; Seo et al., 2015]. The data mentioned above can be used for the development of new microphysical schemes, the validation of numerical models [e.g. Hogan et al., 2006; Salonen et al., 2008; Jankov et al., 2009; Shi et al., 2010; Chu et al., 2014; Pytharoulis et al., 2016; Putnam et al., 2017] or data assimilation [Xiao and Sun, 2007; Li and Mecikalski, 2010; Wang et al., 2013; Carlin et al., 2016].

Dual-polarisation radar data can be used to gain insight into the shape, phase and orientation of hydrometeors [e.g. Hall et al., 1984; Illingworth et al., 1987; Hubbert et al., 1998; Smith et al., 1999; Ryzhkov et al., 2005b; Kumjian and Ryzhkov, 2008; Kumjian, 2013]. This has led to the development of Hydrometeor Classification Algorithms (HCA), which process polarimetric radar data in order to classify the various hydrometeor targets observed [e.g. Vivekanandan et al., 1999; Liu and Chandrasekar, 2000; Straka et al., 2000; Zrnić et al., 2001; Park et al., 2009; Al-Sakka et al., 2013; Thompson et al., 2014; Bechini et al., 2015; Grazioli et al., 2015]. HCAs can provide an effective way of illustrating and interpreting radar data, which assist the development of insight into cloud microphysical processes.

Multiple benefits can be delivered of using a HCA that successfully identifies regions of different hydrometeor types and boundaries between them. According to Giangrande and Ryzhkov [2008], a successful hydrometeor type classification can be the key for choosing the appropriate semi-empirical relations to retrieve the precipitation rates [e.g. Vivekanandan et al., 1994; Mitchell, 1996; Ryzhkov et al., 2005a; Wolfe and Snider, 2012]. For instance, enhanced aggregation processes in the clouds could be associated with higher precipitation rates at the surface [Kennedy and Rutledge, 2011; Bechini et al., 2013; Keppas et al., 2018]. In addition, the discrimination between different ice crystal habits can provide information about the cloud microphysics (e.g. saturation regime in the clouds [Kobayashi, 1961; Magono and Lee, 1966; Bailey and Hallett, 2009] or the presence of secondary ice production [Mossop and Hallett, 1974; Mossop, 1976; Mossop, 1978; Choularton 1978; Choularton, 1980]). Finally, as HCAs can be used in an operational basis by national meteorological services [Chandrasekar et
al., 2013], they can be a valuable tool for providing nowcast information. This information is usually associated with visibility in the troposphere [Rasmussen et al., 1999], short-term forecasts [Lin et al., 2005] and aircraft icing conditions [Vivekanandan et al., 2001], which are also associated with civil protection systems and aviation safety.

Several studies have validated HCAs using airborne in-situ particle imagery, typically using an Optical Array Probe (OAP). This typically involves inspection of particle imagery for a quick visual assessment of HCA performance [e.g. Liu and Chandrasekar, 1999; Lim et al., 2005; Ribaud et al., 2015]. Other studies have used surface rain gauges/disdrometers and hail reports for the same purpose [e.g. Steinert et al., 2014; Kouketsu et al., 2015; Cunha et al., 2015; Ortega et al., 2016]. Although algorithms which perform habit classification of particle imagery and which may aid HCA evaluation are available [Lawson et al., 2006b; Um and McFarquhar, 2009; Lindqvist et al., 2012], they are in need of validation and typically require high resolution grayscale images which are in focus, like those recorded by the Cloud Particle Imager – CPI. The HCA used in this study has been validated by Barthazy et al. [2001], who used a ground-based optical spectrometer and formvar slides to observe hydrometeors, and Martini et al. [2015], who compared OAP imagery of cloud particles with the HCA results in two different winter cases of stratiform clouds and stratiform with embedded convection. Both studies report that the HCA was able to efficiently distinguish particles, which were consistently recorded by in-situ probes. Barthazy et al. [2001] noticed some misclassification between oblate irregular ice crystals and ice needles due to similar fuzzy thresholds. In addition, Martini et al. [2015] highlight the difficulty associated with identifying partially melted, refrozen, heavily rimed and aggregated particles in OAP images, which makes the HCA validation process difficult.

The objectives of the present study are to: a. present the result of applying NCAR’s HCA (based on Vivekanandan et al. [1999]) to data collected on 21 January 2009 from a high resolution S-band radar, b. evaluate the HCA results using OAP imagery data in a purely statistical and objective manner, and c. propose an HCA evaluating scheme based on various in-situ observations such as shape factor, area
ratio, transparency of hydrometeors, number concentration of small and large particles and LWC, which are all calculated from data collected in clouds.

2 Instrumentation and methodology

The present study uses data obtained during the NERC-funded Aerosol Properties, PRocesses And InfluenceS on the Earth’s climate (APPRAISE)-Clouds project. This programme took place in the UK during 2007-2010. The objective of this project was the comprehensive investigation of the role played by aerosols and various microphysical processes in the mixed-phase clouds of numerous North Atlantic Ocean weather systems [Crosier et al., 2011; Crawford et al., 2012].

a. Instrumentation
i. In-situ data

The in-situ measurements of cloud properties used in this study were collected by the UK- Facility for Airborne Atmospheric Measurements (FAAM) BAe-146 aircraft on 21 January 2009, in the framework of the APPRAISE-Clouds project. During the B424 flight, the aircraft sampled mixed-phase clouds that developed within a warm front, which was passing over southern England. The aircraft performed a series of profiles, saw-tooth profiles and level runs for around 5 hours (1500-2000UTC). The aircraft was constantly sampling along an azimuth of 255° within ~100km from the Chilbolton Facility for Atmospheric and Radio Research (CFARR; 51.15°N, 1.44°W; near Chilbolton, Hampshire, UK) to obtain near co-located radar and in-situ measurements. Between 0700-1800UTC hourly radiosondes were launched from the CFARR ground site providing the temperature profile of the atmosphere above.

The key instrumentation attached to the FAAM aircraft and used in this work was described by Crosier et al. [2014], and includes a 2D-S (Two-Dimensional Stereo) Optical Array Probe (OAP), a CIP-100 (Cloud Imaging Probe-100) OAP and a Cloud Droplet Probe (CDP). The latter uses Mie scattering theory to determine the drop size distribution (over the size range 2-50µm), from which a cloud liquid
water content (LWC) is calculated [Lance et al., 2010]. The 2D-S and CIP-100 were used to provide shadow imagery of the observed particles with sizes in the range 10-1280μm and 100-6400μm respectively [Knollenberg et al., 1970; Lawson et al., 2006a]. For OAPs, particle number concentration and diameter (calculated from the diagonal of the bounding box around the recorded particle image) was estimated according to Crosier et al. [2011, 2014]. Artefacts associated with large ice particle shattering on the probes were reduced by applying an inter-arrival time threshold according to Field et al. [2006].

Regarding the ability of the probes to classify different types of hydrometeors, it should be noted that not all of them can be distinguished efficiently. The 2D-S probe can clearly illustrate large droplets and pristine unrimed (or slightly rimed) crystals. However, large particles may be missed due to the small size range of the probe. In contrast, the CIP-100 can be used to identify large aggregates or large dendrites, but it cannot provide clear images of smaller particles such as pristine ice due to the coarse resolution. Finally, neither 2D-S nor CIP-100 can provide efficient images for distinguishing between wet snow (aggregated crystals with liquid edges), dry snow (aggregated crystals) and heavily rimed crystals, as these three types may present similar shapes and sizes.

ii. Polarimetric radar data

The Chilbolton Advanced Meteorological Radar (CAMRa) performed Range Height Indicator (RHI) scans every ~90 seconds along the 255° azimuth to allow comparison with in-situ data collected by the FAAM aircraft flying along the same radial. CAMRa is an S-band dual-polarisation Doppler radar operating at 3GHz (λ=10cm). CAMRa is equipped with a 25m antenna, which enables high resolution data (range resolution=0.3km, beam angle=0.28°) with insignificant two-way attenuation issues (<1dB km⁻¹) even in cases of high rain rates (e.g. 16mm h⁻¹) [Goddard et al, 1994; Curry, 2012]. The particularly small beam angle results to a small beam width. As an illustration, in figure 5 the beam width is ~147m at 30km range widening to ~537km at 110km range.
The polarimetric variables provided by CAMRa and used by the HCA in this study are: a. equivalent reflectivity factor ($Z_{H}$) [Doviak et al., 1979], b. Doppler radial velocity ($V_{RAD}$) [Browning and Wexler, 1968], c. differential reflectivity ($Z_{DR}$) [Seliga and Bringi, 1976], d.) linear depolarization ratio ($L_{DR}$) [Fukao, 2014] and e.) differential phase shift ($\Phi_{DP}$) [Sachidananda and Zrnić, 1986]. Specific differential phase shift ($K_{DP}$) was then derived from the observed $\Phi_{DP}$ [Bringi and Chandrasekar, 2001]. In particular, after unfolding $\Phi_{DP}$, a finite impulse response filter is first applied to $\Phi_{DP}$ to smooth it. The effect of the filter is a combination of the filter length and the number of iterations it is applied. Based on the suggested NCAR setting, a filter that is 10 gates long is applied two times to the data. Further validation of the data includes calculation of standard deviation in the circle (to take account if folding is present) and jitter of $\Phi_{DP}$ (mean change in angle between successive $\Phi_{DP}$ measurements in range). Then, if standard deviation and jitter of $\Phi_{DP}$ are less than 20 and 25, the data is valid. Finally, $K_{DP} < -0.1$ is set to 0. Note that the correlation coefficient $\rho_{HV}$ was not available from CAMRa observations during the 2009 observing period.

b. HCA Validation methodology

In the present work, we introduce three parameters retrieved from the in-situ particle imagery, which are used alongside microphysical measurements of cloud properties to quantitatively evaluate an HCA. The parameters derived from the particle imagery are the weighted shape factor (SF), the weighted area ratio (AR) and the weighted transparency (TR), as described below. In addition, we use mean particle diameter, LWC (calculated from the CDP), and total number concentrations of particles with size $<1000\mu m$ ($N_{<1000}$) and $>1000\mu m$ ($N_{>1000}$). We compare the in-situ parameters with the HCA results extracted along the aircraft flight track using the RHI which was obtained during the same period, resulting in maximum time offset of approximately 90 seconds.
i. **Shape factor (SF)**

The first parameter that we use to quantify the shape of cloud particles is the shape factor [Crosier et al. 2011]. To calculate this parameter, we first calculate the shape factor for each particle recorded by an OAP using equation 1:

\[
SF = \frac{P^2}{4 \pi A} \tag{1}
\]

where \( P \) is the perimeter and \( A \) the projected area of the image. As we are examining the external boundary of the image, \( P \) and \( A \) are calculated after any internal void pixels have been filled. A perfectly imaged circle will have a perimeter \( P = 2 \pi r \), and area \( A = \pi r^2 \) (\( r \) is the radius of the circle), resulting in a SF of 1. Particles with more complex structure exhibit SF values greater than 1. The general behavior of SF for different ice particle types is shown in Fig 1a. Particles which had a SF < 1 due to pixilation effects were rounded up to have the SF =1. It should be highlighted that for the calculation of the perimeter and area, the algorithm searches for the first voxel that belongs to a particle growing, then, the particle from that seed.

ii. **Area ratio (AR)**

The area ratio, as introduced by Heymsfield and Miloshevich [2003], is an alternative parameter to help distinguish between circular and irregular particles (Fig 1b). AR is the ratio between the projected area of the image and the area of the smallest enclosing circle which fully encapsulates the image. In this study, we use the maximum dimension to approximate the diameter of the smallest enclosing circle. Area ratio can be calculated making use of an equation of the form:

\[
AR = \frac{4 A}{\pi D_{\text{max}}^2} \tag{2}
\]

where \( D_{\text{max}} \) is the maximum distance between perimeter pixels, and \( A \) is again the measured particle area after filling internal voids.
Area ratio varies between 0 and 1, with values close to unity referring to near circular images, which suggest pseudo-spherical particles are present (e.g. heavily rimed crystals, ice plates). Images with higher irregularity (e.g. ice columns, stellars) exhibit lower area ratio. **Fig 1b** shows indicative area ratio values of a hexagonal plate and a stellar. In the present work, we aim to evaluate only the performance of the HCA in recognizing ice particles. As small liquid droplets exhibit spherical shapes [Wakimoto and Bringi, 1988], we used an area ratio threshold (<0.85) in order to exclude small spherical liquid droplets. An 80μm size threshold was also applied, as a large portion of liquid droplets present small sizes in mixed-phase clouds [e.g. Marshall and Palmer, 1948; Lloyd et al., 2015].

iii. Transparency (TR)

We finally used the transparency factor, which is calculated as the ratio of the number of void (or “off”) pixels inside the image boundary, to the total number of “on” and “off” pixels within the image boundary (**Fig 1c**). An image which is characterised by transparency equal to 0, is assumed as a dense particle (e.g. heavily rimed crystal), while an image that demonstrates larger values may be indicative of more transparent pristine ice (e.g. thin hexagonal ice plates). It should be highlighted that a. even thin ice particles may be considered as thick depending on their angle relatively to the laser beam of the probe and b. out of focus liquid droplets can appear as doughnut shaped [Korolev, 2007], thus as transparent particles. We try to eliminate the latter issue by making use of the area ratio threshold mentioned previously, as these out-of focus particle also tend to generate more circular images.

iv. Weighting and Merging Method

In order to enhance the ability of the parameters described previously to distinguish between particles not only of different shapes, but also of different sizes, we used a particle diameter-concentration weighting method. In addition, we unified the parameters coming from the two probes. Both the weighting methodology and the
merging methodologies are described in this section. A weighted parameter (WP) is calculated by equation 3:

$$WP_a = \frac{\sum_{i=1}^{n}(PAR_i \ D_i^a \ C_i)}{\sum_{i=1}^{n}(D_i^a C_i)} \quad (3)$$

where n is the number of particles within the sample volume, PAR is the parameter of interest (e.g. AR, SF or TR), D the diameter and C the equivalent concentration for each particle (which is 1/SV, where SV is the sample volume for each particle [Knollenberg, 1970]), with the subscript i representing as summation across the sampled particle population, and α representing the magnitude of the diameter weighting. After analyzing results for α=0-4, it seems that there are no significant differences. Thus, we use α=4 as the study focuses on ice particles [Hogan et al., 2006]. We apply this weighting method to all three parameters of shape factor, area ratio and transparency and we refer to them throughout the text as SF₃, AR₃, TR₃ respectively. The probe from which the data were used is identified by the subscripts “CIP” and “2DS”, while “MRG” refers to the merged product from the two datasets.

In order to obtain a total SF₃ and AR₃ product from both 2DS and CIP-100 probes, we merged the two datasets. For the merging method, a threshold of 1000μm was set. Then, SF₃, AR₃ and TR₃ were calculated for every single particle using 2DS data if the particle size was <1000μm and CIP-100 in case of particle size ≥1000μm. Finally, data were averaged into 1 second time bins, which should be taken into account when comparing with vertical radar scans that are completed within 90 seconds (considering that the aircraft speed was 100m s⁻¹ and the horizontal resolution of the radar data is 300m, each radar voxel corresponds to 3 1-second probe samples).

v. The hydrometeor classification algorithm (HCA)

Hydrometeor classifications were performed using NCAR’s HCA for S-band radar (also referred to as a particle identification or ‘PID’ algorithm) [Vivekanandan et al. 1999]. HCAs using multiple mathematical methods and diverse degrees of
complexity have been developed for radars of various wavelengths since the 1990s (including Straka and Zrnić, 1993; Vivekanandan et al., 1999; Straka et al., 2000; Liu and Chandrasekar, 2000; Zrnić et al., 2001; Lim et al., 2005; Marzano et al., 2007; Park et al., 2009; Snyder et al., 2010; Bechini and Chandrasekar, 2015).

Fuzzy logic has been the basis for a majority of HCAs [Vivekanandan et al., 1999; Liu and Chandrasekar, 2000, Zrnić et al., 2001; Lim et al., 2005; Marzano et al., 2007; Park et al., 2009; Snyder et al., 2010; Bechini and Chandrasekar, 2015]. Instead of simply applying thresholds, this method allows for smooth and overlapping transitions to be established in the relationships of the observed boundaries of polarimetric variables among hydrometeor types.

The NCAR PID uses a fuzzy logic approach with two-dimensional piecewise membership functions to assign the most dominant contributor to the radar signal in a given voxel to one of 17 classes (14 hydrometeor classes with 3 additional classes to identify regions of flying insects, ground clutter and second trip echoes) [Vivekanandan et al., 1999]. The HCA in the present study has been modified to discriminate between the following 15 hydrometeor classes: 1. Ground Clutter, 2. Cloud, 3. Drizzle, 4. Light Rain, 5. Moderate Rain, 6. Heavy Rain, 7. Supercooled Droplets, 8. Graupel and Rain mixture, 9. Rain and Hail mixture, 10. Graupel/Small Hail, 11. Hail, 12. Wet Snow (melting aggregates), 13. Dry Snow (aggregated/heavily rimed crystals), 14. Ice Crystals (pristine ice crystals), 15. Irregular Ice crystals (rimed crystals).
Fig 1. Schematic representation and explanation of a. the shape factor (SF), b. the area ratio and c. the transparency factor. The ice particle images in Fig 1c were captured by the 2DS probe.

The algorithm is a standard part of NCAR’s LIDAR RADAR Open Software Environment (LROSE). The code and definitions of standard membership functions are available at https://github.com/NCAR/lrose-core [Heistermann et al.,]
2015]. The original S-band HCA is described by Vivekanandan et al. [1999] and has been extensively utilised for studying cloud microphysics with S-band radars [e.g. Barthazy et al., 2001; Richard et al., 2003; Evans et al., 2005; Richard and House, 2007; Deierling et al., 2008; Plummer et al., 2010; Pujol et al., 2011; Houze, 2012; Powell and Houze, 2013; You et al., 2014; Martini et al., 2015; Kalina et al., 2016; Schultz et al., 2017].

The standard input to the HCA includes horizontal reflectivity ($Z_H$), differential reflectivity ($Z_{DR}$), correlation coefficient ($\rho_{HV}$), specific differential phase ($K_{DP}$), linear depolarisation ratio ($L_{DR}$), temperature, the standard deviation of the differential propagation phase shift ($\sigma(\Phi_{DP})$) and the standard deviation of the differential reflectivity ($\sigma(Z_{DR})$) (for a review of these variables, see Kumjian, 2013). Standard deviation were calculated along an individual ray of the radar observations and calculation considered 9 gates centered on the current gate of interest. As $\rho_{HV}$ was not available from CAMRa observations during the 2009 observation period, it was omitted from the classification process. All other membership functions and weightings remain the same from the NCAR PID algorithm for the standard results shown. Temperature profiles were provided by radiosonde launches coincident with CAMRa and within one hour of the cases shown here.

vi. The validation method

In order to validate the ability of the HCA to classify the hydrometeors within the clouds, we calculated the 5th and 95th percentiles of the merged SF$_4$, AR$_4$ and TR$_4$ for each of the four different hydrometeor classes identified. Then, these values were used as bottom and upper thresholds for each hydrometeor class (table 1). A simplistic Probe Classification Algorithm (PCA) uses these thresholds to classify the 1-second sample into one of the four classes. However, as thresholds of different classes overlap each other, it is possible that a sample can fall into more than one classes. Finally, a “hit” is counted if HCA matches any of the PCA probe categories, and a “miss” if it does not.
Dry Snow | Wet Snow | Pristine Crystals | Irregular Crystals  
---|---|---|---
**SF₄** | 1.8 - 7 | 1.3 - 5.5 | 1.6 - 4.7 | 1.7 - 7.1  
**AR₄** | 0.29 – 0.58 | 0.21 – 0.76 | 0.22 – 0.59 | 0.27 – 0.56  
**TR₄** | 0.002 – 0.21 | 0.001 – 0.08 | 0.003 – 0.13 | 0 – 0.19  

Table 1. The 5th and 95th percentiles of SF₄, AR₄ and TR₄ used as thresholds for classifying observations to one or more of the hydrometeor classes.

3 Results

i. The synoptic condition

On 21 January 2009, the FAAM Bae146 aircraft flew within mixed phase-clouds associated with the leading edge of a warm front, which was associated with a deep extratropical low-pressure system, passing over the UK [Fig 2; Keppas et al. 2017; Keppas et al., 2018]. According to the HCA, the dominant hydrometeor classes most frequently observed above the melting layer, in the approximate region of the aircraft track, were wet snow, dry snow, pristine and irregular ice crystals (e.g. Fig 6a). It should be noted that the drizzle and light rain observations were few in number as the aircraft mostly flew above or around the melting layer. Therefore, we do not assess the performance of the HCA with respect to drizzle and rain.
Fig 2. Temperature (°C; colour shading and dashed contours) at the 850mb pressure level and mean sea level pressure (mb; white contours) from the ECMWF Re-Analysis (ERA Interim) (0.125° resolution) on 21/01/2009 at 18UTC. The warm front boundary is depicted by the red line, while the aircraft flight region is presented by the pink line.

ii. Linking in-situ measurements with the hydrometeor classes

At this point, we analyse the in-situ parameters (SF₄, AR₄ and TR₄, N<1000, N≥1000, LWC and mean diameter), which will be used for the evaluation of the HCA later. In this section, we discuss all the parameters produced from both CIP-100 and 2DS probes in order to investigate their changes through the different hydrometeor classes. However, for the purposes of this analysis, we use the merged product of SF and AR, as the particle size distributions for 2DS and CIP-100 overlap sufficiently (Fig 3h). It should be noted that SF₄, AR₄ and mean diameter (averaged diameter of particles observed within 1 sec) were retrieved from 2DS and CIP-100 separately, and also after merging the two datasets. TR₄ was only retrieved from the 2DS dataset, N>1000 from CIP-100, and N<1000 from 2DS data.

Before proceeding to the analysis of such data, the differences between the 2DS and CIP-100 probes should be discussed in order to understand the drawbacks but also the benefits gained from using both these probes. The CIP-100 has lower resolution, which means that small semi-irregular crystals (e.g. stellars) may appear as circular particles (meaning that SF₄ and AR₄ will be higher and lower respectively than they would be expected). However, this probe can capture large particles with sizes up to 6400μm, such as large aggregates, due to its large sample area. In contrast, the 2DS probe is qualified with high resolution. Thus, details on the shapes of ice crystals (e.g. stellars or rimed crystals) can be visible affecting SF₄ and AR₄. This is the reason that TR₄ is calculated from 2DS data. However, the sample area of the probe does not allow particles >1280μm to be captured.

Following this discussion and taking into account the merging method which is described in section 2.iv., SF₄, AR₄ and TR₄ may present intense peaks and troughs along a flight track (see section 3.iii.). As these parameters are calculated as a weighted average of the particles observed in 1 second, it is possible that there are particles within this 1 second sample that may significantly differ from the average.
As an illustration, large particles make a greater contribution to the calculation of SF<sub>4</sub> comparing to smaller ones.

Probability distribution functions of 1 Hz means of these parameters for the entire flight are shown in Fig 3, and are separated into the four HCA classes (wet snow, dry snow, irregular crystals, ice crystals). In general, SF<sub>4,CIP</sub> may be used for distinguishing dry snow from the other three classes (Fig 3b). In particular, one of the most interesting features of Fig 3b is the wide range of SF<sub>4,CIP</sub> that the dry snow class exhibits. Approximately 54% of this class has a SF<sub>4,CIP</sub> ≤ 3, with the remaining 46% having SF<sub>4,CIP</sub> > 3. This occurs due to the wide range of shapes and sizes that dry snow aggregates can exhibit. It should be noted that dry snow aggregates can present larger SF<sub>4</sub> compared to wet snow aggregates, because the crystals that constitute dry aggregates can partially retain their shapes, while liquid water on melting/rimed aggregates tend to make crystals appear as quasi circular. Irregular crystal measurements also exhibit low SF<sub>4,CIP</sub> (≤ 3 by 90%), possibly due to the quasi-circular shape of the heavily rimed or aggregated crystals. In addition, the relatively coarse pixel size of the CIP-100 probe (100μm) may artificially reduce the measured SF<sub>4</sub>. Although the ice crystal class is also dominated by small SF<sub>4</sub> values, a significant fraction of measurements (~29%) exhibit larger values with SF<sub>4,CIP</sub> > 3. This suggests that particles with more irregular shapes (e.g. ice dendrites, stellar, irregular, elongated or larger aggregates) may be present.

Regarding SF<sub>4,2DS</sub> (Fig 3a), all of the hydrometeor classes exhibit similar SF<sub>4,2DS</sub> ranges (1 ≤ SF<sub>4,2DS</sub> ≤ 4), which comes in contrast with the SF<sub>3,CIP</sub> measurements (e.g. a significant amount of dry snow exhibits SF<sub>4,CIP</sub> >3). This can be explained by the fact that the 2DS probe size range is lower and the resolution finer than the CIP-100. This means that the shape of smaller particles can be determined in greater detail (this is the reason that irregular crystals can also exhibit SF<sub>4,2DS</sub> >2.5). However, larger particles may be excluded, or their shape may be incorrectly estimated (e.g. 75% of dry snow measurements refer to SF<sub>4,2DS</sub> ≤ 3), which has a significant impact on SF according to equation 2. Finally, 78% of the wet snow measurements exhibit values between 1 to 3 for similar to those for dry snow, but also due to the presence of small circular droplets.
When merging the two datasets (CIP-100 and 2DS), the signatures provided by both the probes can be retained (Fig 3c). Irregular crystals and wet snow mostly (by 79% and 72% respectively) demonstrate $1 \leq \text{SF}_{4,\text{MRG}} \leq 3$. However, ice crystals and dry snow present $\text{SF}_{4,\text{MRG}} > 3$ by $\sim32\%$ and $\sim49\%$ respectively, highlighting the presence of highly irregular particles.

In Fig 3d-f, the AR distributions for the CIP-100 and 2DS probes are shown. Wet snow class seems to present significantly different characteristics for the two different probe datasets. Particularly, 70% of the CIP-100 measurements are linked with small $\text{AR}_{4,\text{CIP}} (\leq 0.35)$, but 54% of the 2DS observations are linked with larger $\text{AR}_{4,\text{2DS}} (>0.45)$ (Fig 3e). This happens because the CIP-100 is able to capture entire wet aggregates, which can take both irregular and non-circular shapes (e.g. ice column aggregates or irregular melting aggregates). Pristine ice crystal class also presents somewhat different traits in the two datasets. The majority ($\sim86\%$) of the measurements of this class exhibit $\text{AR}_{4,\text{CIP}}$ between 0.3 and 0.5 due to the coarse resolution of the probe and the usually small sizes of pristine ice crystals (e.g. Fig 7c, e). Instead, the 2DS probe is able to capture such particles in higher detail and, thus, $\text{AR}_{4,\text{2DS}}$ distribution presents a wider range of values for this class (62% of the measurements are characterized by $0.35 \leq \text{AR}_{4,\text{2DS}} \leq 0.6$). Finally, dry snow and irregular crystals show similar characteristics in both probes mainly exhibiting area ratio $\sim0.45$ and $\sim0.5$ respectively (Fig 3d-e). It seems that irregular crystals may be distinguished from dry snow and pristine ice crystals when merging the datasets from the two probes. More than half ($\sim55\%$) of the measurements for irregular crystals demonstrate $0.45 \leq \text{AR}_{4,\text{MRG}} \leq 0.65$, while dry snow and ice crystal observations mostly (84% and 78% respectively) exhibit lower $\text{AR}_{4,\text{MRG}} (\leq 0.45)$ (Fig 3f).

Finally, TR is calculated only using the high resolution 2DS images, as internal “voids” (un-triggered pixels) were not observed in the CIP-100 images. About 37% of the measurements in HCA ice crystal regions exhibit $\text{TR}_4 \geq 0.04$ mainly due to the presence of thin planar crystals (Fig 3g). A similar proportion (40%) of dry snow measurements refer to values $\geq 0.04$, resulting from aggregates consisting of thin and un-rimed pristine ice crystals which can contain gaps between the aggregates constituent parts.
Fig 3. Probability Distribution Functions of four different hydrometeor classes for (a-c) the weighted shape factor (SF), (d-f) weighted area ratio (AR) and (g) weighted transparency (TR). Graphs (a), (d), (g) refer to the 2DS dataset, graphs (b) and (e) refer to the CIP-100 dataset and graphs (c) and (f) refer to the average between the 2DS and CIP-100 data. For these graphs, particles with AR > 0.85, SF < 1.2 and diameter <80 μm were removed in order to exclude spherical liquid droplets. Figure h shows the particle size distribution for CIP-100 and 2DS.

In addition to parameters related to particle image characteristics, as mentioned above, we also consider the characteristics of other microphysical observations, such as, N<1000, N>1000, the LWC and the mean particle diameter. In order to understand the trends of the first three parameters we used logarithmic scale in x axis for the presentation. However, due to logarithmic scale, size bin widths are wider for larger logarithmic cycles. In order to normalise the data we converted probability into probability density by dividing the probability for each parameter with the logarithm of the beam width. Here, we briefly describe some general trends of these parameters. Around 0.38 of wet snow regions were found consist of relatively larger number of smaller spherical droplets (N<1000 ≥ 6 L⁻¹). All the other classes are mostly (> 0.7) characterised by lower concentrations of small particles (N<1000 < 6 L⁻¹). It should be highlighted that there is a possibility density of ~0.03 of rimed crystal measurements to demonstrate large concentrations (N<1000 ≥ 40L⁻¹)
possibly because ice multiplication may occur within such regions (Fig 4a). In addition, ~44% of wet snow measurements are linked with regions of high LWC ($\geq 0.1 \text{ g m}^{-3}$) due to the presence of liquid water in the melting layer. Moreover, 90% of wet snow measurements were found to have mean diameter $<700 \mu\text{m}$ being separated from all the other classes, which are characterised by a wider range of particle diameters (80-1500$\mu\text{m}$). It should be noted that 24% of dry snow observations coincide with particle diameters $>1300\mu\text{m}$. This can be linked with the fact that 46% of dry snow measurements also appear to have large $N_{>1000}$ $(\geq 0.6 \text{ L}^{-1})$, while irregular, pristine ice crystals and wet snow are mostly associated with $N_{>1000} \leq 1 \text{ L}^{-1}$ (81%, 73% and 95% respectively).

![Fig 4](image)

**Fig 4.** Probability density (probability divided by the logarithm of the size bin) of total number concentration of particle sizes a. <1000$\mu\text{m}$, b. >1000$\mu\text{m}$, c. LWC. In d. the probability distribution function of the mean diameter of particles is shown.

iii. Validating the HCA

In this section, we compare radar and in-situ data, which were collected along three different level runs, in order to test the performance of both the HCA and the evaluation technique of the parameters retrieved by the OAPs.

A comprehensive analysis and interpretation of the in-situ and radar data collected within a warm front, which passed through the UK on 21 January 2009, has been conducted by Keppas et al. [2018]. Here, we briefly summarise the most important and relevant (to this work) features (Fig 5), which are the $Z_{\text{DR}}$ fall streak, the $Z_{\text{H}}$
fall streak and the Generating Cells. A Z$_{\text{DR}}$ fall streak is a slant zone depicted by high Z$_{\text{DR}}$ values (>1.5dB) mainly consisting of pristine ice crystals (dendrites, stellars and plates), while a Z$_{\text{H}}$ fall streak, which is located below the Z$_{\text{DR}}$ fall streak, is a zone of high Z$_{\text{H}}$ (>15dBZ) consisting of heavily rimed and aggregated ice crystals. Both fall streaks originate in a Generating Cell, which is a convective cell embedded in a mass of stratiform clouds. The formation of Generating Cells is associated with the warm conveyor belt, which can transport humid air aloft and trigger Kelvin-Helmholtz instability due to vertical wind shear. On 21 January 2009 fall-streaks developed at an angle from west to east due to wind shear.

**Fig 5.** RHI depicting a. the differential reflectivity (Z$_{\text{DR}}$) factor and b. the reflectivity factor (Z$_{\text{H}}$) (shaded) for the period 17:00:54-17:02:24. In a. the black contour corresponds to Z$_{\text{H}}$=15dBZ and the blue contour (doppler velocity=-20m s$^{-1}$) is used to highlight the warm conveyor belt position.

In **Fig 6a**, we present the HCA result in RHI mode together with the track of the aircraft collecting data along a level run. A zone of dry snow is observed (1km<height<3.5km, 68km<range<92km, temperature ~ -7.6°C) alongside a
region of irregular crystals (1km<height<5km, range < 70km). The dry snow region coincides with a Z\text{H} fall-streak like the one depicted in Fig 5 (range 50-110km, altitude 2-4km). Both 2DS and CIP-100 probes recorded aggregated dry snow along this dry snow zone (Fig 6(g-h); frames 8-10). A transitional region from dry snow to irregular ice crystals is identified by the HCA at range = 68km and altitude ~2.5km. This coincides with a decrease in mean diameter of the particles (from >1200 to 600-900\mu m; Fig 6c) and the presence of quasi-spherical rimed crystals (Fig 6(g-h); frame 7) instead of highly irregular aggregated crystals (Fig 6(g-h); frame 8). The merged mean diameter exhibits a similar decrease in this region. For the remaining section of the aircraft track within irregular ice (ranges <70km), OAPs recorded rimed crystals (Fig 6(g-h); frames 1-7) agreeing with the HCA output.

The dry snow region is characterised by SF\text{4} values between 1.9-4.5 and AR\text{4,CIP} ~0.4-0.5 (Fig 6b). TR\text{4} demonstrates a wide range of values (0-0.4), because snow aggregates may appear as solid and thick crystals (Fig 6g; frame 9) or as crystals with thinner edges and/or gaps between the aggregated crystals (Fig 6g; frame 10). Moreover, low N\text{<1000} (<5L\text{^{-1}}) and high N\text{>1000} (up to 3L\text{^{-1}}) are observed. Interestingly, N\text{>1000} increases from ~1L\text{^{-1}} to ~3L\text{^{-1}} as the aircraft moves to lower altitudes within the dry snow (Fig 6e) implying some intensification of aggregation/riming processes.

The irregular crystal region (39km<range<68km) is characterised by SF\text{4} ~1.9-4, AR\text{4} ~0.27-0.53. Slightly decreased SF\text{3} values comparing to the dry snow region (Fig 6d) are produced due to smaller sizes and quasi-circular rimed ice crystals. AR\text{4} does present a slightly increased median by 0.01 (Fig 6e)and 75\text{th} percentile by 0.015. In contrast, in case of irregular crystals, TR\text{4} does not exceed 0.1 (with median ~0.015) (Fig 6b, f). The thickness and circularity of the heavily rimed ice crystals can explain the slightly larger AR\text{4} and lower TR\text{4} comparing to the dry snow region. In this region, N\text{<1000\mu m} presents values between 1.3-10.8L\text{^{-1}}, significantly larger than in the adjacent dry snow region. Concentrations of larger particles decrease to less than 1L\text{^{-1}}, which are significantly lower than was found in the dry snow region. Particle imagery suggests aggregation and riming processes have not been rapidly occurring in this region (Fig 6c).
In this level run, the HCA matches one of the classes ascribed by the PCA by 93% (this is shown qualitatively in the blue-red shaded bar in Fig 6b). Disagreement only appears in the dry snow region due to a few TR4 observations that exceed the threshold of 0.21 (Table 1). In Fig 6g (frame 10), a rimed crystal (bottom) and an aggregated larger crystal (top) appear to include large void regions which may produce larger TR4 values.

Between 16:49:21-16:59:06UTC the aircraft carried out another level-run at a height of 2-3km observing a long (~50-60km) Z_{DR}/Z_{H} fall streak at temperatures ~ -13.2°C, depicted by the HCA as separate layers of dry snow and pristine ice crystals. In addition, at ranges 30-50km, the HCA identifies a region of irregular ice crystals, which, generally, agrees with the in-situ measurements (Figs 7(g-h); frames 1-3). However, the increased particle diameters (up to 1500μm; Fig 7c) may imply the presence of some aggregates. The ice crystal region at ranges between 64-88km is characterised by gradually smaller (250-1000μm) pristine ice crystals (both dendrites/stellars and plates; Figs 7(g-h); frames 4-11). Finally, when the aircraft moves into the zone of dry snow (dark green; 88km<range<98km), it measures large (up to 1350μm) aggregated and heavily rimed ice crystals (Figs 7g-h; frames 12-13). The boundary between pristine ice crystal and the dry snow zone is characterised by the presence of some slightly rimed pristine dendrites/stellars (Fig 7g; Frame 11) and some aggregated crystals appear together with some smaller ice particles (Fig 7h; Frame 11). An increase in particle diameter from ~900μm to ~1100μm also appears when moving into the irregular crystal region (Fig 7c).
Fig 6. A. Radar RHI scan showing the HCA result at 16:19:34-16:21:04 UTC. The horizontal grayscale line shows the aircraft track between 16:16:16-16:24:31 (white for the final, and black for the starting position). The dashed black line indicates the 0°C level. b. The weighted shape factor (SF₄, red contour), area ratio (AR₄; cyan contour) and transparency (TR₄; purple contour) c. the total number concentration of particles with mean size <1000μm (orange contour) and >1000μm (red contour) and the mean particle diameter (yellow contour) along the aircraft track presented in a. The vertical blue lines are used to locate the position that frames in g. and h. were captured along the aircraft track. The upper shaded horizontal line at the top of the graphs (b-e) indicates the HCA result along the aircraft track, while the lower bar in b. shows whether PCA classification agrees (blue) with HCA results or not (red). d-f. Whisker graphs presenting the distribution of SF₄, AR₄ and TR₄ values (5, 25, 75, 95 percentiles and median with red line) within the different hydrometeor classes as identified by the HCA along the aircraft track. Cloud particle images obtained from f. 2DS and g. CIP-100. The positions where the images were captured are indicated by image strip numbers on Fig 6b. The RHI scan a. does not refer to the entire period of the aircraft track but is presented in order to illustrate the most representative HCA result. The mean temperature along the aircraft track was ~ -7.6°C.
The irregular ice crystal region (35-55km) is characterised by small SF$_4$ (~2) with high peaks of SF$_4$ (up to 4.4) (Fig 7b), AR$_4$ around 0.53 (up to 0.6) (Fig 7c, e) and high TR$_4$ (up to 0.54 with 95$^{th}$ percentile of 0.29) (Fig 7c, f). In this region, ice particles appear with more circular shapes (Fig 7g-h; frames 1-3) explaining the aforementioned increased comparing to the other two regions AR$_4$ values (Fig 7e). TR$_4$ demonstrates increased values due to the presence of thin planar crystals (Fig 7g-h; frames 2-3), which are mixed with rimed crystals, with thin edges. Essentially, this level run is closer to the cloud tops, where pristine ice crystals can be more easily found in their initial form and shape.

One of the traits of the Z$_{DR}$ fall-streak (which generally coincides with the pristine ice crystal region of the HCA) is that it is divided into an upper zone of ice dendrites/stellars and a subjacent zone of hexagonal ice plates [Keppas et al., 2018]. The HCA scheme used in this study cannot distinguish between these two habits of ice. However, the in-situ data can assist in the analysis of such levels of detail. Passing to the ice crystal region (57-88km), generally, this region is characterised by slightly higher SF$_4$ (~2.5), lower AR$_4$ ~0.47 and narrower ranged TR$_4$, compared to the irregular ice region (Fig 7b, d, e, f). However, this region is divided into one with stellars/dendrites (within ranges 57-64km) and another with ice hexagonal plates (within ranges 64-88km). In order to further analyse this region, we use SF$_4$ and AR$_4$ produced from the 2DS probe data. In figure 7i, the observations from the planar region, which exhibit large concentrations (5-19 L$^{-1}$) of particles <1000μm, tend to present narrow ranged SF$_4,2DS$ (~3-4) and high AR$_4,2DS$ (~0.4-0.6). In contrast, the region of stellars/dendrites, which is characterised in this case by much lower concentrations of particles <1000μm (~2 L$^{-1}$), appears with a wider range of AR$_4,2DS$ and SF$_4,2DS$. The wider range indicates a variety of shapes due to the orientation or the degree of riming of the observed particles (e.g. Fig 7g-h; frame 6). In addition, TR$_4$ tends to present a broader range of values with many lows and highs (up to 0.27) within the stellar region due to the transparency of the crystals (Fig 7f; frame 5), while in the hexagonal plate region TR$_4$ obtains values ~0.03 because of slightly rimed hexagonal plates (Fig 7f; frames 6-10). Finally, the dry snow region (88-98km) represents, in contrast to the previous level run (Fig 6), increased SF$_4$ (3.6-7.9) with a median ~5.2 (Fig 7b, d).
Within this region, AR₄ presents similar or significantly decreased values (0.27-0.4 with a median ~0.36). This occurs due to the larger size and degree of irregularity of the snow aggregates. TR₄ exhibits again some peaks between 0.1-0.26 due to the gaps between the aggregated crystals and the presence of thin dendrites (Fig 7f; frames 12-13). Similarly to the previous level run (Fig 6), the dry snow region demonstrates high Nₜ>1000 (up to 6.2 L⁻¹) due to large aggregates. It should be highlighted that the transition between the pristine ice crystal and the dry snow region is fairly smooth (SF₄, mean size and TR₄ gradually increase). However, these two zones demonstrate totally different characteristics and, thus, a dramatic change would be expected in these parameters, as the only source of larger snow aggregates is the dry snow zone, which is located under the ice crystal zone. Consequently, this may reveal that entrainment occurs between the two zones/fallstreaks.

In this level run, HCA classification agrees with the PCA results by 77%. Most of the disagreement is detected in the irregular crystal region due to high TR₄ that some unrimed transparent crystals present (Fig 7g; frame 3) and in the zone of stellars/dendrites, where particles as previously mentioned exhibit larger TR₄ than the rest of the pristine crystal region.

The third level run that we present in this work was carried out close to the melting layer at a temperature ~0°C (height ~1km). Regions between 38-55km, 57-65km and 10-20km (Fig 8a) according to the HCA fall into the wet snow class. Along these ranges, CIP-100 recorded snow aggregates (8(f-g); frames 1-2, 4-6, 8-9) of ~1000μm in diameter (peaking at ~1750μm) within a region of significant LWC (0.2-1g m⁻³) (Fig 8d) comparing to other upper regions of clouds. However, a region at range = 55-57km is classified by the HCA as dry snow, while it is dominated by a mixture of snow aggregates/liquid droplets (Fig 8f-g; frames 2-3) and liquid water content ~0.5g m⁻³. In addition, focusing on figures 8e-g, the observations of this region seem to present similar traits. There are several reasons for this misidentification: a. the region is dominated by spherical liquid water droplets, which produce low ZDR affecting the HCA result, b. there are no ρHV data available, which might assist in a more effective representation of the cloud region around the melting layer and c. the dry snow region is represented by just a few
pixels within a larger region of wet snow and, as previously stated, the aircraft and radar beam may not be exactly collocated due to the difference of volumes that aircraft and radar sample at the same time.
Fig 7. Similar to Fig 6. a. but here the RHI scan refers to 16:51:00-16:52:31UTC and aircraft track refers to 16:49:21-16:59:06UTC. The mean temperature along the aircraft track was ~ -13.2°C. Graphs i. and j. show shape factor versus area ratio calculated only from 2DS probe data, while shaded scale is for concentration of particles <1000μm.
At ranges 20-35km, the aircraft flew along the boundary between two zones, within a wet snow region, but in close proximity to an overlying region of irregular ice crystals. The 2DS recorded ice columns, ice-lollies [Keppas et al., 2017] and supercooled droplets and the CIP-100 observed also some ice column aggregates (Fig 8(h-i), frame 7). As ice crystals dominate scattering in this region and LWC retrieved by the CDP is ~0g m\(^{-3}\) (Fig 8d), the region can be assumed to be a dry ice or irregular ice crystal (due to the presence of ice-lollies) region. However, it is hard to judge if these ice crystals are rimed or not. The irregular crystal region (20-35km) presents higher SF\(_4\) (~3.5) than in previous two runs because of highly irregular ice columns and ice-lollies (Fig 8(h-i); frames 6-8). For the same reason AR\(_4\) is low (0.14-0.3).

Within the wet snow region, SF\(_4\) exhibits values between 1.5 and >5 obtaining the highest values for large complex melted aggregates (Fig 8h; frame 5) and the lowest for melted ice columns mixed with spherical droplets/melted graupel (Fig 8h-i; frame 9). AR\(_4\) exhibits a wide range of values (~0.21-0.56) due to the presence of irregular or quasi-circular aggregated snow (Fig 8b, f). TR\(_4\) demonstrates low values (<0.1), because melted and rimed crystals appear in 2DS images as optically thick and non-transparent particles. The small spots of pristine ice crystals along the wet snow region may be ignored as they are limited in space.

In this third run, it seems that SF\(_4\) significantly changes when moving from rimed ice columns/ice-lollies to large ice column aggregates (from ~3.5 to ~4.5 with peaks; Fig 8b, h; frames 5, 7). However, SF\(_4\) can also be similar (~3) for rimed and melted ice mixed with melted graupel (Fig 8h; frames 8-9). AR\(_4\) can show increased values (0.4-0.55) for graupel mixed with small droplets but decreased (0.2-0.4) when graupel is mixed with irregular ice columns. It should be noted that along this level run, high N\(_{<1000}\) (1-28 L\(^{-1}\)) is observed due to the presence of liquid droplets around the melting layer, while N\(_{>1000}\) stayed around 0.5 L\(^{-1}\) (Fig 8c). Within the irregular ice crystal region, N\(_{<1000}\) attained values ~5-10 L\(^{-1}\) as in previous level runs. The very large N\(_{<1000}\) in the region between 10-21km may due to the presence of rimed ice particles within a mature Z\(_H\) fall streak.
For this level run, 77% of the observations exhibit the same HCA and PCA result. Most of the discrepancy appears in the dry snow and irregular crystal region. In the dry snow, the main reason is the limited range of AR₄ comparing to this of the wet snow (table 1). AR₄ may be fairly low for very irregular melting aggregates (Fig 8h; frame 5) or much larger for almost melted particles and liquid droplets (Fig 8h; frame 3). The irregular crystal region presents significantly different characteristics examining in contrast to the other two level runs. SF₄ is high (up to 7.7) and AR₄ is low (median is ~ 0.27; Fig 8f), while TR₄ is ~0 due to the presence of thick and long ice columns and ice-lollies. For this reason, in this investigated case, such particles are assumed as outliers for the irregular crystal class.
Fig 8. Similar to Fig 6. a. RHI scan refers to the period 19:50:07-19:51:38UTC and the aircraft track refers to the period 19:48:00-19:55:24UTC. In d. the red curve indicates liquid water content along the aircraft track. The mean temperature along the aircraft track was $\sim 0^\circ$C.
4 Conclusions

In this study we use parameters calculated from airborne in-situ optical array probes (OAPs) and a Mie scattering cloud droplet probe (CDP), in order to evaluate the ability of the NCAR’s Hydrometeor Classification Algorithm (HCA) to distinguish between the following ice hydrometeor classes: a. dry snow, b. wet snow, c. irregular crystals and d. ice crystals. The data were collected during the passage of a warm front across the UK on 21 January 2009. We examined three level runs in detail, two of which crossed fall streaks, while the other was carried out slightly above the melting layer. As a result of the analysis, the HCA results seem to be sensible with respect to the shapes and the location of the various classes. However, small discrete regions close to the melting layer, which were expected to fall into the wet snow category, were classified as dry snow by the HCA. Possible reasons for this are a potential mismatch in the aircraft and the radar sample volumes, the lack of $\rho_{HV}$ data and the presence of small spherical droplets.

The comparison of the in-situ parameters and the HCA which was conducted a. collectively for the entire flight and b. along three separate level runs by examining time series leads to the following conclusions:

$Z_H$ fall streaks were classified as zones of dry snow by the HCA. These zones were characterised by $1.9 < SF_4 < 7.9$, $0.27 < AR_4 < 0.5$ and $TR_3$ up to 0.35. The parameter that mainly characterised this class was $N > 1000$. Only in dry snow regions did $N > 1000$ reach 2-5 L$^{-1}$.

$Z_{DR}$ fall streaks were classified by the HCA as regions of ice crystals. $SF_4$ identified the boundary between the pristine ice and dry snow regions (from $SF_4 \sim 2.5$ in the ice crystal region to $\sim 5$ in the dry snow region). $SF_4$ changed due to the presence of large particles, which were gradually dominated the sample volume of the probes in the dry snow region. Additionally, the two sub-zones (classified as containing planar and dendritic/stellar according to OAPs) of the $Z_{DR}$ fall streak were better described by the 2DS measurements and retrievals, as the mean size of the crystals (500-1000μm) was lower than the upper limit of the size that the probe can measure. In particular, the planar region was characterised by $SF_{4,2DS} \sim 3-4$ and $AR_{4,2DS} \sim 0.4-0.6$, while in the dendritic/stellar region $SF_{4,2DS}$ and $AR_{4,2DS}$ presented wider range of values (1.5-11.7 and 0.2-0.6 respectively). $TR_4$ seems to attain larger values (up to 0.27) within the dendritic/stellar region. $N < 1000$ reached up to 18.6 L$^{-1}$ in the hexagonal plate region possibly due
to the reduced presence of supercooled droplets mixed with new pristine ice, remaining much lower in the dendritic/stellar region.

Riming regions were determined by $SF_4 \sim 1.9-4.4$, and slightly increased $AR_4$ (up to $\sim 0.6$) in cases of heavily rimed crystals due to the more spherical shapes of the crystals. When rimed ice columns were present (such as around the melting layer), $SF_4$ exhibited larger values ($\sim 5$ with peaks and lows) and $AR_4$ fairly low (0.14-034). $TR_4$ may reach large values (up to 0.54) due to the presence of transparent ice crystals with little or no rime within regions of light riming. Such regions were mainly consisted of cloud particles <$1000\mu m$ in size.

Finally, the wet snow region was better described by either large $SF_4$ (with peaks $>5$) and/or wide ranged $AR_4$ ($-0.21$-$0.56$). It should be highlighted that melting ice can have different shapes depending on the different original ice habits and, thus, all of these parameters can significantly vary. $TR_4$ remained low ($\sim 0.05$) due to opaque crystals. The presence of high LWC (up to 1g m$^{-3}$) and large $N_{<1000}$ (up to 28 L$^{-1}$) due to the presence of liquid droplets may be used to characterise wet snow regions.

Running a simple algorithm (PCA), which uses thresholds obtained from 5$^{th}$ and 95$^{th}$ percentiles of $SF_4$, $AR_4$ and $TR_4$ (from the entire dataset), to classify the 1-second sample into one or more of the four hydrometeor classes, it seems that PCA and HCA agreed by 77-93%. Problematic regions were spotted within riming regions when ice columns and ice-lollies (thus lower than expected $AR_4$) or unrimed crystals were present and within dry snow in cases of highly irregular aggregates, which included more void pixels (and thus higher $TR_4$ than expected).

In general, $SF_4$, $AR_4$ and $TR_4$ exhibit significant changes for different regions of ice cloud, but we cannot determine thresholds for a specific hydrometeor type, as the range of the values that a parameter demonstrates depends on a variety of factors. As an illustration, irregular (thus rimed and/or aggregated) crystals can demonstrate a wide range of $AR_4$ depending on the initial shape of the ice crystals, which then rimed. However, grouping total number concentrations into small ($<1000\mu m$) and large ($>1000\mu m$) particles may provide a clearer signature, especially, for the identification of dry snow and regions that contain supercooled droplets. Future work may include a statistical/fuzzy analysis of these and other in-situ parameters for other cases to generate an objective and robust tool for distinguishing of hydrometeor types and evaluating HCAs.
5 Abbreviations

A  projected area of a particle
AR  weighted area ratio
D  particle diameter
C  particle concentration
FAAM  Facility for Airborne Atmospheric Measurements
HCA  Hydrometeor Classification Algorithm
K_{DP}  specific differential phase
LROSE  LIDAR RADAR Open Software Environment
N_{<1000}  Number concentration of particles with size <1000μm
N_{>1000}  Number concentration of particles with size >1000μm
PCA  Probe Classification Algorithm
PID  Particle Identification
P  particle perimeter
PAR  in-situ parameter referring to one of SF, AR and TR
SF  weighted shape factor
TR  weighted transparency
Z_{DR}  radar differential reflectivity
Z_{H}  radar reflectivity factor
\rho_{HV}  correlation coefficient between horizontally and vertically polarized radar signal
\sigma(Z_{DR})  standard deviation of radar differential reflectivity
\sigma(\Phi_{DP})  standard deviation of differential phase shift
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References


CHAPTER SEVEN

CONCLUSIONS AND FURTHER WORK

7.1 Overview and Conclusions

The present thesis focuses on the comprehensive and detailed analysis of cloud microphysics of the warm front, which affected the UK weather on 21 January 2009. The entire work, consisting of three papers, is based on both airborne in-situ and remote sensing measurements, which were collected during the APPRAISE-Clouds project. The main focus was to investigate the microphysical processes within the warm front, in the high detail that the airborne probes and the S-band Chilbolton radar allow, and strengthen the link between remote sensing and in-situ data of cloud properties. The first two papers make use of the measurements in a number of ways in order to expand our knowledge on the frontal cloud processes. Although the third part of the present work refers to the same warm front case, it focuses mainly on the evaluation of an algorithm (HCA) that classified the different regions of the clouds, and the methodologies that could be used for this reason. The analysis in the latter work was benefited by the fact that significant knowledge of the structure of the frontal clouds had already been obtained during the first two papers. Briefly, the present document, through its individual parts, aims to report and compare observations of a warm front in order to investigate the cloud structures and microphysical processes and provide further insight in their impacts on the precipitation.

• In the next lines, we summarise the most important conclusions and results from all the three studies that compose this thesis. The first part of the conclusions refer to the structures and the characteristics of the warm front: The cloud mass was characterised by cloud tops at ~6km and temperature > -25°C allowing primary heterogeneous and secondary multiplication processes. The height of the cloud tops decreased by 1.5 km, while the height of the melting layer increased by 1km due to the passage of the frontal boundary.

• The main features that were recognised in the warm front cloud mass structure are the warm conveyor belt (WCB), the generating cells (GCs) and two different ice fall streaks with different microphysical and polarimetric characteristics.
• The **WCB** was the main mechanism that caused the formation of massive mixed-phase clouds transporting large amounts of liquid water aloft (up to 0.37g m\(^{-3}\)). The WCB was represented by high Doppler velocities (20-30m s\(^{-1}\)) and actual wind speeds (up to 26m s\(^{-1}\)) and temperatures \(\sim -6^\circ\text{C}\) offering an ideal environment for secondary ice production. The zone of high Doppler velocities that characterised the WCB decreased in altitude, as the warm front moved further inland away from the coast.

• The **Generating Cells** (GCs) demonstrated a core of high-\(Z_H\) (10-33dBZ)/\(\sim 0\text{dB} \ Z_{DR}\) and a shell of low-\(Z_H\) (<10dBZ)/high-\(Z_{DR}\) (up to 4dB). These convective turrets formed at the unstable, based on potential temperature profiles (**chapter 4, section 4**), layer above the WCB, where vertical wind shear triggered Kelvin-Helmholtz instability.

• As GCs were entering the sheared flow caused by the WCB, two different ice fall streaks formed demonstrating length of up to \(\sim 60\text{km}\) and slope of \(\sim 3^\circ\). The length and slope, as depicted along the RHIs, tended to decrease and increase respectively as the warm front was approaching the UK inland, due to the intensification of the processes within the GCs. However, ice fall streaks may present a different orientation in the space, not captured by the RHIs) as the wind was veering with altitude and time. The low \(Z_H\) (<15dBz) \(Z_{DR}\) fall streaks originated in GC shells being consisted of a zone of hexagonal ice plates (\(Z_{DR}>1.5\text{dB}\)) and another zone of stellars/dendrites (\(Z_{DR} \sim 1.5\text{dB}\)), which might form from small ice-plates that were moving from GCs boundaries to the high supersaturated regions of the WCB. As GC cores were moving into the WCB, \(Z_H\) fall streaks formed being represented by \(Z_{DR}\sim 0\), \(Z_H\sim 15\text{dB}\). Within such fall streaks, various sized (from <500\(\mu\)m to 3600\(\mu\)m) rimed/aggregated ice crystals in various concentrations (from <4L\(^{-1}\) to >10L\(^{-1}\)) were measured. The \(Z_H\) fall streaks were associated with enhanced surface precipitation. A particular \(Z_H\) fall streak, which was analysed, enhanced the surface precipitation by \(\sim 4\text{mm h}^{-1}\) an hour after its genesis. This fact highlights the way that precipitation can be affected by the profile of wind direction and speed in the clouds.

The WCB played an important and unique role in the ice multiplication process that occurred within the clouds. As previously stated, the WCB transported large amounts of liquid water aloft. As the average temperature at the level that the WCB presented the maximum wind speeds was \(\sim -6^\circ\text{C}\), it seems that this humid stream created an ideal environment for ice multiplication via the Hallett-Mossop process. This can be assumed by the presence of large ice concentrations (up to 37.8L\(^{-1}\)) of mostly ice columns within the intersection region of the \(Z_H\) fall-streaks with the WCB.
The WCB played also an important role via the Hallett-Mossop process in the formation of ice-lollies, which is presenting in chapter 5 [Keppas et al., 2017]. Planar ice crystals and snowflakes falling from the cloud tops through the WCB core, which contained significant amounts of supercooled water, rimed. Because of the temperature, the Hallet-Mossop multiplication was activated producing high concentrations of ice splinters, which grew into ice columns. The shape of the ice-splinters was determined by the temperature (~ -5°C; Kobayashi, 1961; Magono and Lee, 1966; Bailey and Hallett, 2009). Then, the newly formed ice columns collided with supercooled droplets forming ice-lollies. The presence of ice-lollies can be an indication of ice multiplication occurrence. Ice-lollies can also act as rimers enhancing ice multiplication mechanisms, removing further liquid water from the clouds. These processes can impact on cloud lifetime and precipitation formation. The involvement of ice-lollies in further ice multiplication may be proved by the fact that some of these particles appear to exhibit some side planes on the frozen droplet part facing away from the attached column. Ice-lollies in isolation can demonstrate high $Z_{DR}$ values due to their elongated shape confirmed by calculations using the Discrete Dipole Approximation. In clouds, they were observed within regions of $Z_{DR} \sim 0.3-1.6$dB not in complete isolation from other hydrometeor types, but also in regions of mostly quasi-spherical graupels/heavily rimed crystals or water drops with $Z_{DR} \sim 0$dB.

Finally, we used the NCAR’s Hydrometeor Classification Algorithm (HCA) to distinguish between the following ice hydrometeor classes: a. dry snow, b. wet snow, c. irregular crystals and d. ice crystals within $Z_H/Z_{DR}$ fall streaks and around melting layer that were observed during the warm front on 21 January 2009. The HCA was evaluated using all the available in-situ data and derivative parameters (shape factor (SF), area ratio (AR), transparency (TR), mean diameter, LWC, concentration of groups of particles ($N_{>1000}, N_{<1000}$)), which can provide details about the shape, the size, the transparency and the concentration of cloud particles. The comparison between the radar and the in-situ data was conducted collectively for the entire flight and along three separate level runs. Although we examine the performance of the HCA, we mainly focus on the ability of the in-situ parameters to evaluate the HCA performance. Here we summarise the measurements within the different cloud features:

- $Z_H$ fall streaks were classified as dry snow zones by the HCA. These zones were characterised by ice particles of medium-high shape irregularity and transparency
(1.9<SF4<7.9, 0.27<AR4<0.5 and TR4 up to 0.35). It should be noted that only in dry snow regions did N_{<1000} reach 2-5 L^{-1}.

- **Z_{DR}** fall streaks were classified by the HCA as regions of pristine ice crystals. SF4 identified a boundary between the pristine ice and dry snow regions (from SF4 ∼2.5 in the ice crystal region to ∼5 in the dry snow region). SF4 changed due to the presence of large particles, which were gradually dominated the sample volume of the probes in the dry snow region. Additionally, the two sub-zones (classified as containing planar and dendritic/stellar according to OAPs) of the Z_{DR} fall streak were better described by the 2DS measurements and retrievals, as the mean size of the crystals (500-1000μm) was lower than the upper limit of the size range of the probe. In particular, the planar region was characterised by SF4_{2DS} ∼3-4 and AR4_{2DS} ∼0.4-0.6, while the dendritic/stellar region exhibited a wider range of SF4_{2DS} (1.5-11.7) and AR4_{2DS} (∼0.2-0.6). TR4 attained larger values (up to 0.27) within the dendritic/stellar rather than in planar ice region. The presence of supercooled droplets mixed with new pristine ice caused increased N_{<1000} (up to 18.6 L^{-1}) within the planar ice region.

- **Riming regions** consisting of heavily rimed crystals were determined by small SF4 ∼1.9-4.4, and slightly increased AR4 (up to ∼0.6) due to their quasi-circular shapes. In cases of rimed ice columns (e.g. in regions around the melting layer), SF4 exhibited larger values (∼5 with peaks and lows). In cases that ice particles with no or light riming were present, TR4 reached large values (up to 0.54). In such regions we mainly observed cloud particles <1000μm in size.

- Finally, the wet snow region was better described by by either wide ranged AR4 (∼0.21-0.56) and/or large SF4 (with peaks ∼5). It should be highlighted that SF4 and AR4 can significantly vary within wet snow regions, because melting ice can appear in different shapes depending on the different original ice habits. TR4 remained low (∼0.05) due to the opacity of the wet snow. In general, wet snow seems to be characterised by high LWC (up to 1 g m^{-3}) and N_{<1000} (up to 28 L^{-1}) due to the presence of liquid droplets.

A simple algorithm (PCA) using thresholds obtained from 5th and 95th percentiles of SF4, AR4 and TR4 (from the entire dataset), to classify the 1-second sample into one or more of the four hydrometeor classes was run in order to compare the HCA results with the airborne observations. The agreement between HCA and PCA reached 77-93%. Problematic regions were spotted within riming regions when ice columns and ice-lollies (thus lower than expected AR4) or unrimed crystals were present and within
dry snow in cases of highly irregular aggregates, which included more void pixels (and thus higher TR4 than expected).

Although SF4, AR4 and TR4 present significant changes for regions of clouds with different ice particles, thresholds cannot be specified for a specific hydrometeor type, as the range of the values of a parameter depends on a variety of factors, such as the initial shape of a crystal, the degree of riming/aggregation, the presence of liquid droplets etc. Clearer signatures may be provided by the parameters N<1000μm and large N>1000μm for the dry snow class or when supercooled water droplets are present.

7.2 Discussion and Future Work

In the present work, a comprehensive analysis of a warm front is presented. Both heterogeneous ice nucleation and secondary ice formation were observed in mixed-phase clouds, which are highly uncertain processes. Heterogeneous ice nucleation played an important role in the formation of cloud tops of the investigated warm front, including the GCs. One of the most important features that was investigated is the WCB. Liquid water was transported aloft, along WCB and rear GC updraft regions, freezing and forming new ice. The WCB played a crucial role not only in the formation of massive stratiform clouds, but also in the development of GCs and the formation of ice-lollies through ice multiplication.

As in-situ measurements within GCs, ice-fall-streaks are limited, more observations should be performed aimed at investigation of the cloud processes (aggregation, riming, ice multiplication) in order to obtain a deeper understanding of cloud microphysics. Except for observing real GCs, they could also be created and observed in labs. Comprehensive datasets, which could describe the structure and evolution of such clouds and microphysical processes could be used for the validation of complex microphysics models [Stoelinga et al., 2003; Keeler et al., 2016a]. Similarly, as ice-lollies have been observed only twice, during BAE424 flight on 21 January 2009 and BAE146 flight on 23 September 2016. However, only in the first flight, ice-lollies were observed in significant concentrations and there was supporting radar coverage. Future missions in warm conveyor belts spanning the HM zone would provide insight into the importance of ice-lolly formation.
7.3 Publications

In the following list, all the publications associated with the work presented in this thesis are mentioned.


References


Morrison, Hugh, et al. "Possible roles of ice nucleation mode and ice nuclei depletion in the extended lifetime of Arctic mixedphase clouds." Geophysical research letters 32.18 (2005).


