Observations of the Origin and Distribution of Ice in Cold, Warm, and Occluded Frontal Systems during the DIAMET Campaign

G. Lloyd, C. Dearden, T. W. Choularton, J. Crosier, and K. N. Bower
Centre for Atmospheric Science, University of Manchester, Manchester, United Kingdom

(Manuscript received 18 December 2013, in final form 8 May 2014)

ABSTRACT

Three case studies in frontal clouds from the Diabatic Influences on Mesoscale Structures in Extratropical Storms (DIAMET) project are described to understand the microphysical development of the mixed phase regions of these clouds. The cases are a kata-type cold front, a wintertime warm front, and a summertime occluded frontal system. The clouds were observed by radar, satellite, and in situ microphysics measurements from the U.K. Facility for Airborne Atmospheric Measurements (FAAM) research aircraft. The kata cold front cloud was shallow with a cloud-top temperature of approximately $213^\circ$C. Cloud-top heterogeneous ice nucleation was found to be consistent with predictions by a primary ice nucleation scheme. The other case studies had high cloud tops ($>240^\circ$C) and despite no direct cloud-top measurements in these regions, homogeneous ice nucleation would be expected. The maximum ice crystal concentrations and ice water contents in all clouds were observed at temperatures around $-5^\circ$C. Graupel was not observed, hence, secondary ice was produced by riming on snow falling through regions of supercooled liquid water. Within these regions substantial concentrations ($10$–$150 \text{ L}^{-1}$) of supercooled drizzle were observed. The freezing of these drops increases the riming rate due to the increase in rimer surface area. Increasing rime accretion has been shown to lead to higher ice splinter production rates. Despite differences in the cloud structure, the maximum ice crystal number concentration in all three clouds was $100 \text{ L}^{-1}$. Ice water contents were similar in the warm and occluded frontal cases, where median values in both cases reached $0.2$–$0.3 \text{ g m}^{-3}$, but lower in the cold front case.

1. Introduction

In the midlatitudes extratropical cyclones can dominate the synoptic situation and, therefore, the conditions experienced at the surface. Occasionally these systems bring conditions that cause substantial damage to property and significant loss of life, primarily through strong winds and excessive rainfall (Leckebusch et al. 2007; Ulbrich et al. 2001, 2003a,b). Observations of the large-scale characteristics of these storms have led to a number of models describing their formation and progression, most notably the Norwegian (Bjerknes 1919; Bjerknes and Solberg 1922) and Shapiro–Keyser models (Shapiro and Keyser 1990).

The dominant features of these midlatitude storms are the contrasting air masses forming the frontal boundaries that can bring abrupt changes in weather conditions.

The Norwegian model fails to represent the many variations within a cyclones structure on a smaller scale (Browning 1990). Matejka et al. (1980) reported numerous bands of precipitation associated with the frontal systems of midlatitude cyclones as part of the Cyclonic Extratropical Storms (CYCLES) project. A distinctive example of this variability between frontal systems is the existence of narrow cold frontal rainbands (NCFR) and wide cold frontal rainbands (WCFR; Matejka et al. 1980;
Houze and Hobbs 1982; Crosier et al. 2014). Often the terms “ana” and “kata” (Bergeron 1937) are used to describe different types of cold frontal systems. The an-type cold front has a rearward component to the warm conveyor belt (WCB) flow relative to the surface front. This produces rearward-sloping ascent and the generation of a broad area of precipitation known as the WCFR. However, ascent of warm air forced by the surface cold front can lead to more vigorous convection over a narrow area up to an altitude of around 3 km (Browning and Reynolds 1994). Kata cold fronts are the result of dry air intrusions, characterized by a region of air descending from the tropopause region that can overrun the surface front (Browning 1997). This prevents rearward ascent of warm air behind the cold front as the dry air intrusion causes a descending motion in all but the lowest layers of the frontal system (Sansom 1951). It also leads to forward motion of the air above the warm conveyor belt relative to the movement of the cold front. This produces forward-sloping ascent and causes the main band of precipitation to become situated ahead of the surface cold front (Browning 1986).

Matejka et al. (1980) investigated the microphysics of different frontal bands and, in some cases, highlighted the importance of microphysical and mesoscale convective processes that had a direct influence on the formation and spatial distribution of precipitation. For example, below the narrow convective updraft of an NCFR, precipitation is suppressed through suspension or upward movement of hydrometeors. Crosier et al. (2014) found that production of small secondary ice crystals through rime splintering in an NCFR accounted for a significant fraction of the ice water content (IWC) and suggested this is important in altering the precipitation budget and the distribution of latent heat release. The reader is referred to Dearden et al. (2014) for a consideration of diabatic heating and cooling rates derived from some of the microphysics data that will be presented here. In warm frontal systems Hobbs and Locatelli (1978) examined precipitation cores within the frontal rainbands and found they were linked to ice crystals produced at higher altitudes in “generating cells” that seeded the stratiform cloud below.

The Diabatic Influences on Mesoscale Structures in Extratropical Storms (DIAMET) project (Vaughan et al. 2014), part of the Storm Risk Mitigation (SRM) aims to advance our understanding of the mesoscale features within midlatitude storms in order to improve accuracy in predicting them. The overarching theme of the DIAMET campaign is to consider the role of latent heating in the generation of mesoscale potential vorticity (PV) and moisture anomalies in cyclonic storms and the impact these may have on the weather. The aims of the campaign most relevant to the work presented here are the measurement of microphysical properties and variability in mesoscale structures. Here we present data from three intensive observation periods (IOPs) of a cold front, warm front, and occluded frontal system during the DIAMET campaign. We discuss the large-scale features associated with these systems and analyze the dominant microphysical processes that were observed in each case.

2. Methodology

During autumn–winter 2011 and spring–summer 2012 a total of 14 IOPs involving ground-based and in situ measurements were carried out as part of the DIAMET project. In all cases in situ measurements were provided by the U.K. Facility for Airborne Atmospheric Measurements (FAAM) BAe-146 aircraft. The strategy used to explore the microphysics of each frontal system involved flying a combination of constant altitude runs with profiled ascents and descents that aimed to capture the vertical structure of each system. Particular attention was paid to the temperature zone between −3° and −8°C where secondary ice production (SIP) takes place (Hallett and Mospop 1974; Crosier et al. 2011; Crawford et al. 2011).

Instrumentation used to measure cloud microphysics on board the BAe-146 included the Cloud Imaging Probe models 15 and 100 (CIP-15 and CIP-100, Droplet Measurement Technologies, Boulder, Colorado; Baumgardner et al. 2001), the Cloud Droplet Probe (CDP-100 Version 2, also DMT; Lance et al. 2010), and the Two-Dimensional-Stereoscopic (2D-S, Stratton Park Engineering Company, Inc., Boulder, Colorado) Probe (Lawson et al. 2006). The CIP-15 and CIP-100 optical array probes consist of 64 element photodiode arrays with element resolutions of 15 and 100 μm, respectively. We analyzed data from measurements of particle size over the range 15–930 μm (CIP-15) and 100–6200 μm (CIP-100). The 2D-S also uses an optical array but has 128 elements each of 10–μm resolution, capable of imaging particles between 10–1280 μm. Particle size histograms were provided at 1 Hz for analysis.

It has been recognized for some time that ice breakup on the inlets of probes creates errors in size distribution and number concentration data that need to be corrected (Field et al. 2006). The magnitude of particle enhancement attributed to shattering is dependent on the probe type and the microphysical properties of the cloud being measured. Jensen et al. (2009) found 2D-S concentrations were enhanced by about an order of magnitude when natural concentrations of small cloud ice were low, but negligible when these concentrations were high. During this campaign the CIP-15 and CIP-100 optical array probes were fitted with Korolev-designed tips (Korolev et al. 2011) to minimize shattering of ice particles on the probe housing, which can potentially cause artifacts in the measured size.
distributions. In addition, software processing removed particle shattering events through the use of interarrival time (IAT) filtering (Crosier et al. 2014; Field et al. 2006). It should be borne in mind that not all shattering artefacts may be accounted for with IAT filtering.

Size distribution measurements from the CIP-15 and CIP-100 were merged, providing continuous size distribution data sets covering the size range \(15–6200\ \mu\text{m}\). Where there was significant disagreement between the two probes in their respective size overlap regions, preference was given to the CIP-15 dataset, due to its greater resolution (15 \(\mu\text{m}\)) compared to the CIP-100 (100 \(\mu\text{m}\)). The CIP-15 and CIP-100 datasets were filtered using an IAT threshold of \(1 \times 10^{-4}\) s. Tests to examine the sensitivity of ice particle concentrations to different IAT thresholds showed some evidence of shattering artifacts remaining in the size distributions when the IAT threshold was relaxed and lowered to \(1 \times 10^{-5}\) s or IAT

---

**Fig. 1.** Met Office surface pressure charts for the (a),(b) cold front case; (c),(d) the warm front case; and (e),(f) the occluded frontal case: (left) 1200 and (right) 1800 UTC. (Source: Met Office.)
filtering was turned off completely. As a result a threshold value of \(1 \times 10^{-4}\) s was maintained and used in this work.

The 2D-S probe has a significantly higher resolution and frame rate than the CIP-15 and CIP-100 probes enabling it to provide more detailed images of hydrometeors. The higher resolution (at 10\(\mu\)m) and faster response (>10 times faster) makes the 2D-S generally more suitable for use in the classification of ice particle habits and in discriminating these from water drops in each of the cases described here, particularly as a function of altitude and for rapidly changing regions of glaciation. Identification of liquid droplets and ice crystals is based on analysis of particles shape and is described in Crosier et al. (2011). The ability to discriminate between particle phases allowed for calculation of the ice water content using the Brown and Francis (1995) mass–dimensional relationship. Secondary ice production through the Hallett–Mossop (H–M) process is a known powerful mechanism of glaciation that occurs when cloud temperatures are between –3°C and –9°C, liquid water is present, and existing ice particles come into contact with the liquid droplets to produce riming. In these conditions we frequently observe large increases in ice crystal number concentrations. To accurately capture the high concentrations of small ice crystals naturally present in the (H–M) temperature zone, an IAT threshold of \(1 \times 10^{-5}\) s (as suggested by Crosier et al. 2014) was used. Lowering the threshold to this value for all 2D-S data, increases the possibility that shattered particles could have contaminated the dataset and artificially increased number concentrations, particularly in regions where larger more delicate ice crystals were seen to have developed. This possibility was investigated by examining IAT frequency plots and testing for changes in ice number concentrations when varying the IAT threshold to \(1 \times 10^{-4}\) s and also by turning the filtering technique off, allowing all particles to be accepted. IAT frequency plots suggested a limited contribution from a shattering mode and that a threshold of \(1 \times 10^{-5}\) s was appropriate.

**FIG. 2.** Radar-derived precipitation rate for the cold front case. (a)–(c) Precipitation over the southern United Kingdom; (d)–(f) zooms in on the region west of Chilbolton with the track (red trace) of the BAe-146 aircraft overlaid: (left) 1400, (middle) 1500, and (right) 1600 UTC. (Source: Met Office.)
FIG. 3. MSG cloud-top temperature products for the (a),(b) cold front; (c),(d) warm front; and (e),(f) occluded front. Outline of the United Kingdom in white. (Source: EUMETSAT, second generation.)
Cloud particles in this study were selected for analysis using the “center in” method (Heymsfield and Parrish 1978), in which the probe’s sample volume does not decrease with increasing particle size as is the case with the “entire in” acceptance technique. This allows greater numbers of larger particles to be measured within a given sample, improving counting statistics. The size of the particles (\(D_p\)) was calculated by taking the average of the maximum size in the along (\(D_y\)) and across (\(D_x\)) directions. The sizing method is less sensitive to elongated particles than using just \(D_x\) and allows more accurate representation of particle size distributions in regions where elongated particles such as columns are the dominant growth habit (Crosier et al. 2014). In addition to the cloud microphysics probes used, the aerosol size distribution was measured using a Passive Cavity Aerosol Spectrometer Probe 100-X (PCASP; Rosenberg et al. 2012) that measures particles over the range 0.1–0.3-\(\mu\)m diameter. A limited amount of information from this instrument was used to challenge a primary ice nuclei (IN) parameterization (DeMott et al. 2010) for the cold front case.

Ground-based remote sensing measurements for the cold front IOP, were provided by the Chilbolton Advanced Meteorological Radar (CAMRa; Hogan et al. 2003) at the Chilbolton Facility for Atmospheric and Radio Research (CFARR) in Hampshire, southern England (51.14\(^\circ\)N, 1.44\(^\circ\)W). This is a fully steerable 3-GHz dual-polarization Doppler radar. During this flight the BAe-146 aircraft flew along a radial of 240\(^\circ\) while CAMRa performed range–height indicator (RHI) scans along the same radial. Other data collected at CFARR to monitor the passage of the cold front included atmospheric pressure, temperature, wind speed, wind direction, and rainfall rate as measured by a rapid response drop counting rain gauge (Norbury and White 1971).

3. Frontal cases

The microphysics and larger-scale characteristics of the three cases will now be described in detail. These were chosen to represent three distinct frontal types observed during DIAMET. The cold front case provided an excellent opportunity to use in situ and ground-based remote sensing techniques to observe a frontal system in a state of transition. Warm frontal systems are characterized by areas of broad, predominantly light precipitation. The one presented here provided a case in which detailed microphysics measurements of the vertical structure of the system were made. The selection of a summertime occluded front, which had a much higher freezing level, allowed us to compare with the microphysics of the other two wintertime cases.

a. Cold frontal case

On 29 November 2011, Met Office surface pressure charts showed a large depression northeast of Iceland with a central pressure of 961 mb (Fig. 1). A secondary low pressure system developed, pushing a cold front from west to east across the United Kingdom, which eventually reached mainland Europe by 0000 UTC 30 November. This cold front was particularly active over northern England with a squall line and damage caused by a tornado reported in the greater Manchester area.

The aircraft departed Exeter Airport (50.44\(^\circ\)N, 3.24\(^\circ\)W) at ~1400 UTC and arrived overhead Chilbolton, England, at ~1430 UTC. A number of constant altitude and stepped profiled runs were repeatedly performed on a radial of ~240\(^\circ\) from the CFARR location. The aircraft made measurements for ~90 min before landing back at Exeter Airport around 1600 UTC. Examination of the Met Office rainfall radar composite (Fig. 2) shows that the frontal system exhibited both
a WCFR and NCFR earlier in the day, which was eventually dominated by the NCFR with no broad area of precipitation. The NCFR exhibited precipitation rates up to 32 mm h\(^{-1}\). Analysis of the in situ microphysics measurements, meteorological data, remote sensing, and satellite products all support the assumption of a transitioning frontal system. Meteosat Second Generation (MSG) cloud-top height and temperature products showed that at 1000 UTC cloud-top temperatures were close to \(\sim 30^\circ C\) over a wide area (Fig. 3a). As the front moved over land, cloud-top temperatures increased significantly to around \(-15^\circ C\) (Fig. 3b). The increase in cloud-top temperatures was associated with the lowering of cloud-top heights through suppression of convection by the descending dry air overrunning the cold front. The 1200 UTC radiosonde from Larkhill, England, just ahead of the frontal rainband, revealed this dry layer between \(\sim 850\) and 600 mb (Fig. 4c). In addition to analysis of radiosonde data, the aircraft deployed 13 dropsonsdes off the coastline of southwest England. Dropsonde data from \(\sim 0900\) UTC revealed a saturated profile to 600 mb before the dewpoint fell to \(-60^\circ C\) at 550 mb (Fig. 4a). As the front approached and moved overland dropsonde data from \(\sim 1240\) UTC revealed the appearance of a dry layer between 800 and 650 mb with dewpoints in this region falling to \(-30^\circ C\) (Fig. 4b).

The passage of the front over CFARR was well represented in the meteorological data recorded at the site. Surface pressure fell to 994 mb at \(\sim 1500\) UTC before recovering following the passage of the cold front while winds veered westerly and the temperature fell \(-5^\circ C\) by 1800 UTC as the cold air mass advanced eastward. The precipitation rate at CFARR, measured by the droplet counting rain gauge, reported values typically below 5 mm h\(^{-1}\). And in this case the precipitation arrived before the surface front, which is consistent with the characteristics of a kata-type cold front. The front appeared to be in the later stages of transition from ana to kata type as there was some evidence of slantwise ascent behind the front from the radar data shown in Fig. 5.

The RHI scans from CAMRa showed a broad convective region (Fig. 5a) between \(\sim 60\) and 80 km advancing toward Chilbolton that eventually split into two separate convective elements (Fig. 5d) with the second of the two regions, around 50 km from CFARR, becoming the most active. This development is supported by the rainfall radar sequence that showed a broader region of precipitation (Figs. 2a,d) that split into two bands (Figs. 2b,e) as the front approached CFARR. The dominant convective feature is represented in the CAMRa data by a narrow band of high radar reflectivity up to 50 dBZ (Fig. 5d), \(\sim 50\) km from CFARR. This feature coincided with the passage of the surface cold front and had an associated signature in the unfolded radial velocity field (Fig. 5f) that represents scattering hydrometeors being transported vertically by the convection driven by the advancing cold air mass. Above the frontal uplift, a region of high differential reflectivity, \(\sim 4\) dB, developed between \(-60\) and 90 km from Chilbolton (Fig. 5b) and grew as the front advanced eastward (Fig. 5e).

The aircraft made several penetrations through this frontal system, as summarized in Table 1. During
segment A1 the aircraft made two profiled descents and one straight and level run. The altitude of the aircraft varied from ~5.2 to 3.7 km and temperatures increased from −16° to −9°C. During this period the aircraft made some brief penetrations of cloud tops in the frontal system. The flight path during these cloud-top penetrations intersected regions where CAMRa detected high differential reflectivity. The in situ microphysics measurements revealed mixed phase, glaciated, and supercooled conditions with swift transitions between them (Fig. 6). The 2D-S reported ice crystal number concentrations of up to ~1 L⁻¹ likely achieved through a combination of heterogeneous ice nucleation in the relatively high cloud-top temperatures of around −13°C and the effects of H–M SIP occurring below. This cloud-top temperature is only ~5°C lower than that in the H–M temperature zone that lies between −3° and −8°C (Hallett and Mossop 1974).

Radar imagery from CAMRa revealed regions of high reflectivity representing convective elements that provided a lifting mechanism that had been generated by the advancing cold front (Fig. 5a). The 2D-S particle imagery showed that at cloud top, columnar ice crystals that had presumably been lifted from an H–M-generating region below, where high concentrations were being produced through SIP. Interestingly, the images also revealed the presence of heterogeneously nucleated ice crystals, in the form of pristine plates (Fig. 7a). The existence of columnar ice crystals within cloud tops might be expected to lead to an enhancement in the ice crystal number concentration there above that expected from primary heterogeneous ice nucleation alone. This is because first ice within the cloud tops is initiated through primary ice nucleation but subsequently these concentrations may be enhanced as ice crystals produced by SIP are lifted from an H–M zone below into the cloud tops. However in this case data measured by the PCASP was used to calculate the expected IN concentration according to the parameterization of DeMott et al. (2010) and found the predicted ice crystal concentrations to be about a factor of 2 lower than mean concentrations measured in a cloud-top region during a ~15-s period by the 2D-S (~0.32 L⁻¹). The PCASP iced up during the flight, limiting the data we could use. The mean aerosol concentration >0.5 μm was calculated from a 15-s time period below cloud base (~9.3 cm⁻³).

CDP-derived liquid water contents (LWCs) of 0.1–0.15 g m⁻³ were present within liquid regions at cloud top. During flight segment A2 the aircraft performed a profiled descent from ~3.6 to 1.8 km where temperatures increased from −7° to 1°C, intersecting the H–M temperature zone where the imaging probes observed large needle and columnar ice crystals in much higher concentrations than those observed at cloud top (~80 L⁻¹; Fig. 7b). During a straight and level run, within flight segment A3, at 1.8 km, outbound from Chilbolton, temperatures fell from ~+1° to −2°C and the aircraft passed through the dominant convective feature, where LWCs reached 0.8 g m⁻³, the highest recorded value at any time during the mission. Imagery from the 2D-S revealed small liquid droplets in this region (Fig. 7d) and the CDP reported a mode droplet diameter of ~25 μm in concentrations ~100 cm⁻³. (Information about droplet concentration and size as a function of altitude can be found in Fig. 22.) In this case concentrations were generally <100 cm⁻³ with median diameters between ~15 and 30 μm. Analysis
of the LWC profile as a function of altitude (Fig. 8a) showed that significant ($>0.01 \text{ g m}^{-3}$) LWCs reached right up to cloud top at $\sim 4$ km. It may be significant that close to the region where the highest concentration of columns were observed, larger supercooled drops (greater than 80-µm diameter) were present in concentrations of up to $\sim 150 \text{ L}^{-1}$, the concentrations of drizzle fell to $\sim 1$–$30 \text{ L}^{-1}$ in regions where columnar crystals were in concentrations of $\sim 80 \text{ L}^{-1}$ (Fig. 9a). Crawford et al. (2011) showed the importance of warm rain processes leading to the formation of droplets of diameter $>80 \text{ µm}$ in generating high concentrations of ice crystals. Figure 9a has been modified to exclude ice number concentrations in the time period between $\sim 1500$ and $\sim 1511$ UTC. Examination of 2D-S probe imagery revealed misclassification of liquid droplets as ice particles during this period. After this point, ice crystals were observed in the probe imagery and these data have been included. It is possible that the freezing of the droplets in this case played an important role in the generation of the high ice concentrations in this region.

Matejka et al. (1980) described the microphysics of convective updraft regions within an NCFR as consisting of small growing cloud droplets that were initiated through vertical ascent forced by the surface front. Significant liquid water ($>0.01 \text{ g m}^{-3}$) was associated with this region along with low ice particle number concentrations when compared to areas immediately adjacent to the convective element. The cold front in this study displayed similar characteristics, with a high LWC ($\sim 0.8 \text{ g m}^{-3}$) and, for the most part, low ice number concentrations. In this particular instance the highest concentrations of ice crystals were observed ahead of the convective feature (Fig. 10) and were likely to have been produced through rime splintering (Hallett and Mossop 1974). The convective element created by the movement of the cold front from west to east generated supercooled liquid water up to the cloud tops of this frontal system. In these regions small, pristine ice crystals increased in size through riming and vapor growth, which then began to descend and aggregate. The in situ image data clearly showed ice particles in this region were aggregating and transitioning to snow particles while CAMRa recorded high differential reflectivity values during RHI scans in this same region. As the crystals continue to fall, the ZDR signal decreases due to the particles acquiring more complex shapes through changes in habit and the process of aggregation (Fig. 5e). Ice number concentration as a function of altitude from the 2D-S shows a steady order of magnitude increase in median ice crystal number concentrations from $\sim 1$ to $10 \text{ L}^{-1}$ as the aircraft descended from cloud top into the H–M temperature zone (Fig. 11a). Here, the snow particles came into contact with supercooled liquid water droplets, leading to SIP through rime splintering. Size distributions from the merged CIP-15 and CIP-100 dataset (Fig. 12a) reveal the high concentrations of small ice crystals $\sim 100 \text{ µm}$ present during segment A2 while flying through an active area of SIP. This contrasts with lower concentrations of small ice crystals $\sim 100 \text{ µm}$ present during segment A1. Note that the ice particle size distributions include all particles classified as ice even in the presence of supercooled drops.

Because of the relatively warm cloud tops, this system does not display the typical reduction in ice number concentrations with decreasing altitude, associated with
FIG. 7. The 2D-S imagery from runs (a) A1; (b), (c) A2; and (d) A3 (see Table 1). The red line in the bottom left corner of (a) indicates probe array width (see Table 1).
Fig. 8. (left) Liquid water content from the CDP and (right) ice mass as a function of altitude for the (a),(b) cold front; (c),(d) warm front; and (e),(f) occluded front showing 25th, 50th, 75th percentiles, whiskers to 10% and 90%, and outliers between 95% and 100%.
aggregation, which might be expected. The cloud-top temperature of $\sim -13^\circ$C in this case is close to the main region of secondary ice production between $-3^\circ$ and $-9^\circ$C. The ice particles generated in this temperature zone dominate the ice crystal number concentration throughout the depth of the cloud. Percentile plots of ice water content as a function of altitude (Fig. 8b) show that the highest contribution to the ice water content occurred at $\sim 3$ km, where higher concentrations of ice crystals were observed than in any other region of the cloud system.

Crosier et al. (2014) observed a lack of graupel when investigating an NCFR and suggested that snow particles were the likely sites that produced the riming required for secondary ice production. In this case graupel was also not observed in regions where ice enhancement through secondary ice production was taking place, or in the areas surrounding them. The presence of a component of forward motion relative to the kata cold front in this case is likely to have directly influenced the spatial distribution of precipitation. Growing ice crystals at
cloud top were pushed ahead of the convective feature before descending (Fig. 5e), leading to glaciation of this region through depletion of liquid water droplets via the Wegner–Bergeron–Findeisen process and riming. 2D-S probe imagery reveals snow particles in close proximity to liquid droplets (Fig. 7c). The riming that took place on these snow particles, in the temperature range between \(-3^\circ\) and \(-9^\circ\)C, is the likely mechanism by which ice crystal number concentrations were enhanced ahead of the surface cold front. Further details of the cloud ice spectra and their parameterization in models together with detailed modeling of the diabatic processes in this case may be found in the companion paper Dearden et al. (2014).

b. Warm frontal case

At 0000 UTC 12 December 2011, Met Office surface pressure analysis charts showed a very unsettled North Atlantic with several low pressure centers (Fig. 1c). By 0000 UTC 13 December an intense midlatitude cyclone had developed with a central pressure of 952 mb lying to the west of Scotland (Fig. 1d). Satellite imagery (Figs. 13c, d) revealed a characteristic comma pattern to the cloud structure associated with this midlatitude cyclone. Eventually the warm front passed across southern England and the system matured, with an occlusion forming over northern parts of the United Kingdom. The focus of this study was the warm front, and to investigate this, the BAe-146 operated off the coasts of southwest England and south Wales as the front approached land.

Radar images showed a broad area of rainfall consisting of predominantly light precipitation with intensities typically around 5 mm h\(^{-1}\) (Fig. 14). MSG products showed cloud-top heights over an extensive area to be about 8.5 km with temperatures at this altitude around \(-60^\circ\)C (Figs. 3c,d). To study the evolution within these features, a series of profiled descents and runs at constant altitude were performed from approximately 8.5-km altitude, which are summarized in Table 1. Based on MSG cloud-top height products, 8.5 km was \(\sim 2\) km below cloud top. Some of the dropsonde data from the aircraft behind the frontal rainband showed an inversion around 700 hPa, likely representing the warm sector air that had been lifted over the colder air ahead of the warm front (Fig. 15b).

During the first profiled descent (B1) the temperature rose from \(-43^\circ\)C at 8.5 km, to \(-13^\circ\)C at 4.5 km. The 2D-S measured ice crystal number concentrations at the beginning of the profile between 25 and 30 L\(^{-1}\), with small irregular ice crystals with mode diameters between 200 and 300 \(\mu m\) (Fig. 16a) that very likely formed through homogeneous ice nucleation at temperatures below \(\sim 37^\circ\)C. Size distributions for the ice particles only (Fig. 12b) from the merged CIP dataset showed the absence of any larger ice crystals at this altitude. The aircraft then performed a SLR (B2) at 4.7 km where temperatures were between \(-10^\circ\) and \(-13^\circ\)C. No liquid water was detected at this altitude where ice crystal number concentrations were between 5 and 10 L\(^{-1}\). There was evidence consistent with heterogeneous ice nucleation at this altitude, in the form of pristine platelike ice crystals (Fig. 16b). The cloud ice microphysical structure, however, was dominated by aggregates of ice crystals falling from higher in the cloud. The increase in size, in the absence of liquid water, was mostly due to aggregation as the ice crystals fell, leading to a reduction in number concentration but a general increase in size to a mode diameter of \(\sim 650 \mu m\), as illustrated in Fig. 12b. Reductions in ice crystal number concentrations due to aggregation have been described and quantified in previous studies (Cardwell et al. 2003;
FIG. 11. Percentile plots as a function of altitude for (left) ice number concentration (L⁻¹) and (right) mean diameter (μm) of ice particles. (a),(b) cold front case; (c),(d) warm front case; and (e),(f) occluded frontal case showing 25th, 50th, 75th percentiles, whiskers to 10% and 90%, and outliers between 95% and 100%.
Crosier et al. 2014) and this is very likely to be the mechanism observed here. The ice particle size distributions recorded during run B2 (Fig. 12b) showed a decrease in smaller ice crystal concentrations compared to the high-altitude run (B1) and the presence of higher concentrations of larger aggregated ice crystals ~650 μm as well as the appearance of ice particles >1000 μm in size.

Percentile plots of ice crystal number concentration (Fig. 11c), mean diameter (Fig. 11d), and ice water content (Fig. 8d) as a function of altitude show the change in the microphysics between runs B1 and B2. During the highest altitude run (B1) at a temperature of −24.38°C ice crystal number concentrations ~25 L−1 were observed. With decreasing altitude the process of aggregation gradually increased the mean size while some vapor growth of the ice crystals present increased the IWCs. Around 4 km (−15°C) a more significant increase occurred as the median ice crystal diameter increased to ~650 μm and the IWC rose to ~0.2 g m−3, probably due to the presence of liquid water at this altitude. Rimming also contributes to the increase in size, as the existing growing particles sweep out the supercooled liquid water droplets. Measurements of the droplet concentrations and sizes made by the CDP (Fig. 22) showed low concentrations of droplets in this case, with maximum median values ~10 cm−3. Median droplet diameters were <30 μm.

At 2.8 km, the aircraft penetrated the H–M temperature zone (B3), with temperatures ranging from −4°C to −7°C. The microphysical structure in this region was much more variable than experienced at higher altitudes (Figs. 11c,d). Changing conditions were observed, including the presence of aggregated ice crystals that had descended from above and the presence of liquid water as detected by the CDP. During this run, high ice crystal number concentrations of a columnar nature up to about 80 L−1 were detected (Fig. 16c). As in the cold front case above, significant concentrations of drops in excess of 80-μm diameter were observed by the 2D-S in concentrations of around 15–30 L−1 in association with the columnar crystals (Fig. 9b). The enhancement in columnar crystal concentration is likely due to secondary ice production taking place through the H–M process, as ice crystals falling from above sweep out the supercooled liquid water droplets and grow through riming. As in the cold front case, a possible role for the larger drops needs to be considered. Analysis of the LWC profile measured from the CDP (Fig. 8c) showed that the less intense convection associated with these types of fronts generally prevents significant liquid water penetrating to above the ~−5°C level. However in this case significant LWC was still seen to exist up to around 3 km, which was within the H–M temperature zone. The percentile plots representing ice number concentration (Fig. 11c), mean ice particle diameter (Fig. 11d), and ice water content (Fig. 8d) as a function of altitude reveal that at around 3 km, where temperatures increase to ~−5°C, higher concentrations of ice particles (~80 L−1) and higher ice water content values (~1.2 g m−3) were observed than anywhere else within the warm frontal system.

c. Occluded frontal case

On 15 August 2012 the aircraft performed a number of in-cloud runs to investigate the microphysical structure of an occluding frontal system over and off the coast of Northern Ireland. The synoptic situation over the United Kingdom was notably unsettled for the time of year with a deep low pressure system positioned just south of Ireland at 1200 UTC (Fig. 1e). Warm and cold frontal systems pushed north across England and Wales with the occluding sector crossing Northern Ireland (Fig. 1f).
FIG. 13. MSG visible satellite imagery for the (a),(b) cold front; (c),(d) warm front; and (e),(f) occluded front.
(Source: EUMETSAT, second generation.)
Analysis of visible satellite imagery showed a mature cyclone in the process of occluding (Figs. 13e,f). The occluded front was eventually positioned through Northern Ireland, Wales, and England while farther to the south and west the satellite imagery showed a well-defined trailing cold front.

Dropsonde data from the aircraft (Fig. 17c) show that as the aircraft traveled farthest south, behind the surface front (Fig. 17c), a marked dry sector was observed around 600 hPa and water vapor imagery showed this to be dry air wrapping around the cyclone center behind the cold front, corresponding to the dark band in Fig. 18 lying across Ireland and wrapping around the center of the low pressure system. Radar precipitation rates initially showed a broad frontal system with significant variation in the structure and intensity of precipitation across Northern Ireland (Fig. 19). By 1500 UTC an intense squall line with precipitation rates reaching 100 mm h\(^{-1}\) stretched across Northern Ireland, through the Irish Sea and over England. Cloud tops over Northern Ireland were as cold as \(-52^\circ C\) (Figs. 3e,f). The FAAM BAe-146 flew microphysics legs at varying altitudes over Northern Ireland and the sea to the north. During one SLR run at \(3\) km (C4) the aircraft passed directly through a narrow region of intense precipitation where nearby radar reflectivity indicated rainfall rates of \(\sim 50\) mm h\(^{-1}\). However, the higher altitude of the aircraft in relation to the height at which the radar reflectivity actually represented meant that no direct association between the microphysical properties and regions of high precipitation rates could be established.

During profiled climb C1 the aircraft reached an altitude of 7.8 km, the highest altitude at which measurements were made during the flight. Temperatures in this region were around \(-25^\circ C\), and so a significant distance beneath cloud top. This is confirmed by MSG cloud-top height products that suggest cloud tops of around 9 km.
Here small irregular ice crystals were observed (Fig. 20a) in concentrations between 5 and 10 L\(^{-1}\) and with a mode diameter between 200 and 300 \(\mu\)m. There was also evidence of liquid water as the CDP detected a liquid water mass of \(\sim 0.1 \text{ g m}^{-3}\). This was supported by 2D-S imagery that revealed small spherical particles together with ice crystals. In this environment, where ice crystals are observed together with liquid droplets, growth of the ice phase will take place through vapor diffusion. The ice crystals present are also likely to aggregate, which will further increase particle size. The average size distribution from the merged CIP datasets from flight segment C1 (Fig. 12c) shows a noticeable absence of any larger particles over 1000 \(\mu\)m at this altitude. During profiled descent (C2) the aircraft descended from 6 to 3 km and temperatures rose from \(-15^\circ\) to 0°C. Initially large irregular aggregates of ice crystals were observed together with smaller pristine plates and sectored plates that were likely produced through heterogeneous ice nucleation (Fig. 20b).

Percentile plots of ice concentration (Fig. 11e), mean diameter (Fig. 11f), and ice water content (Fig. 11f) as a function of altitude show that between \(-8\) and 6.25 km the ice microphysical structure changed little. However at \(-6\) km there was a significant increase in median IWC to \(\sim 0.2 \text{ g m}^{-3}\), and a decrease in median ice number concentrations from about 5 to 2 L\(^{-1}\). These measurements highlight a transition from smaller ice crystals, closer to cloud top during run C1, to an increasing number of much larger ice crystals observed during C2. The merged CIP size distribution during run C2 also shows the reduction in concentrations of smaller ice crystals with a corresponding increase in the number of larger particles \(>1000 \mu\)m (Fig. 12c). The number concentrations of the crystals during run C2 vary from just a few per liter to, on occasion, 15 L\(^{-1}\). A number of processes probably contributed to the mix of crystal habits and total ice number concentrations observed at each level during C2 including: (i) aggregates of ice crystals formed at higher altitudes, (ii) heterogeneous ice nucleation at this altitude, and (iii) columns produced through secondary ice production in the H–M temperature zone below. During the flight, crystals were observed that had changed their growth regimes during their lifetime. Ice crystals were occasionally observed that had initially followed a growth regime in temperatures around \(-5^\circ\), developing into columns, which were then captured by convective elements within the frontal system, and lifted to an environment where temperatures were significantly lower. At this point the growth regime changed, and the columns developed platelike structures on the ends of the original columnar ice crystals, forming capped columns (as illustrated in Fig. 21).

As the temperature increased during profile C2 the aircraft encountered pristine columnar ice crystals created through SIP in enhanced concentrations of around 40 L\(^{-1}\). This was repeated in a subsequent profiled ascent where ice number concentrations again rose to around 40 L\(^{-1}\) in the H–M temperature zone. These ice crystals were seen together with larger aggregates in the presence of LWCs close to 1 g m\(^{-3}\). Analysis of the LWC profile as a function of altitude for the entire science flight (Fig. 8e) showed significant amounts of liquid water penetrating to around 5 km where mean temperatures were \(\sim -10^\circ\). Analysis of CDP data as a function of altitude (Fig. 22) showed median droplet concentrations up to \(\sim 100 \text{ cm}^{-3}\) and periods in which high concentrations \(>500 \text{ cm}^{-3}\) were observed. The droplet diameter was generally small with median values \(<\sim 15 \mu\)m. During a subsequent run, (C3), the aircraft maintained a constant altitude within the H–M
temperature zone and ice particle number concentrations as high as 100 L$^{-1}$ were recorded (Fig. 20c). Crystal habits included larger aggregates in the presence of liquid and droplets $>80 \mu$m in diameter (Fig. 9c) in concentrations between $\sim 30$ L$^{-1}$, with liquid water contents as high as 1 g m$^{-3}$. Figure 11 shows percentile plots of ice crystal number concentrations, ice crystal mean size, and ice water content from this case. It was observed that during descent into the H–M temperature zone, at $\sim 4$ km, mean ice crystal number concentrations increased to around 10 L$^{-1}$, while median ice crystal diameter fell to $\sim 300 \mu$m. Periods of secondary ice production increased maximum concentrations in this region to $\sim 100$ L$^{-1}$. Percentile plots of the ice water content as a function of altitude (Fig. 8f) show that the highest median ice water contents (0.25 g m$^{-3}$) occurred at $\sim 5.5$ km, where temperatures were $\sim -10^\circ$C. The ice particle size distribution from C3 (Fig. 12c) again showed the highest concentrations of small ice crystals in this region as well as the larger aggregates that were present. All of these observations are consistent with an active H–M zone, enhancing the ice number concentration significantly to above that observed in other regions of this occluded frontal system. Figure 23 shows an example of ice particle size distributions from the 2D-S probe for runs C1 to C3 and
confirms the change in the diameter of ice crystals with a reduction in altitude.

4. Discussion

In situ microphysics measurements showed that in all the cases investigated the production of secondary ice, likely to have been initiated by the H–M process, was evident around −5°C. In these regions the concentration of ice crystals as observed by the 2D-S reached 140 L⁻¹. Examination of the LWC profile with altitude reveals that in the cold frontal and occluded frontal cases liquid water was able to penetrate to significantly greater altitudes above the freezing level when compared to the
warm front case, with the highest altitude and mass concentrations of liquid water observed in the occluded frontal system. When comparing the summer occluded frontal case to both the winter warm front and the winter cold front cases, it was found that liquid water mass was significantly higher in the summer case (Fig. 8e). Examination of the CDP droplet size spectra showed that this increase over the other two cases could be attributed to an increase in the number concentration of small liquid droplets of mode diameter \( \approx 10 \, \mu m \). In the winter cases droplet concentrations were lower (generally \(<100 \, cm^{-3}\)) than in the summer case where concentrations reached \(>500 \, cm^{-3}\). Median droplet diameter in the winter cases reached \(\sim 25 \, \mu m\), which was greater than in the summer cases where values were generally \(<15 \, \mu m\) (Fig. 22). Despite differences in the distribution of liquid water in each case, significant liquid water was always observed in the H–M temperature zone, where its presence is critical to the powerful secondary glaciation process that takes place through the accretion of supercooled liquid droplets on riming particles. Although the freezing level of the occluded front was \(\sim 3.5 \, km\) (1.5 km higher than the other cases) the convection generated by this system was sufficient to produce significant liquid water to an altitude of \(\sim 5 \, km\) where the temperature of the environment was still within the H–M zone. No evidence for graupel was observed in the H–M temperature zone or adjacent regions on any of the flights. However, there were frequent encounters with rimed aggregates of various ice crystals together with supercooled liquid water in the temperature range \(-3^\circ\) to \(-8^\circ\)C.

It may be significant that larger supercooled drops (in excess of 80-\(\mu m\) diameter) were found in high concentrations in association with the high concentrations of columnar ice crystals and in the cold front case up to 150 L\(^{-1}\) in a region of supercooled liquid water close to the region of high ice crystal concentrations. Such concentrations of supercooled drops have been reported by Crawford et al. (2011) in a rapidly developing convective
The application of a detailed cloud microphysical model in this paper showed that the supercooled drops played a key role in the rapid glaciation of the cloud. The mechanism was that ice particles falling into the Hallett–Mossop zone produced ice splinters that were then captured by the supercooled drops causing them to freeze and become rimers themselves producing further splinters. It was found that this mechanism was able to explain the rapid glaciation of this cloud in the observed time scale.

In the cold front case, relatively warm cloud-top temperatures led to initiation of the ice phase through heterogeneous nucleation. In the remaining cases cloud-top temperatures were $<\sim 37^\circ$C. In this environment it is likely homogeneous ice nucleation contributed to the glaciation of the cloud, although heterogeneous ice nucleation via deposition cannot be ruled out. In the cold front case, ice number concentrations of typically up to $\sim 1 \text{ L}^{-1}$ within cloud tops were observed, which is in reasonable agreement with the predicted primary IN using the Cooper scheme, as reported in DeMott et al. (2010). However, the observed concentrations are two orders of magnitude higher and an order of magnitude lower than predicted by the Fletcher and Meyers schemes, respectively (DeMott et al. 2010). Concentrations elsewhere in the cloud reached $\sim 80 \text{ L}^{-1}$, which is one to two orders of magnitude greater than the predicted primary IN concentration.
In the other two cases, cloud-top microphysical regions were well above the H–M temperature zone. There was strong evidence of aggregation processes being important, particularly in the warm front case, resulting in a steady decrease in ice crystal number concentration above the altitude of the −5°C level. For example, observations of ice crystal number concentration (Fig. 11c) and mean size (Fig. 11d) as a function of altitude from the warm front case show a well stratified microphysical system with a continuous decrease in ice crystal number concentration and a corresponding increase in mean particle size (with decreasing altitude) above the H–M temperature zone. Although the summertime occlusion displayed a similar trend there is much greater variability in these microphysical parameters and this is likely due to strong convection causing intermittent lifting of high concentrations of these small ice crystals to greater altitudes. In the summer case there was also a substantial depth of cloud below the freezing level and, hence, warm rain processes will have contributed to the precipitation observed at the surface although no observations were made to quantify this.

In each case, ice water content as a function of altitude (Fig. 8) was calculated using the merged dataset from the CIP-15 and CIP-100. The warm front case initially exhibited little change in ice water content with decreasing altitude, from ~8 to ~5 km. Below this level a significant increase in ice water content was observed around 4 km. Percentile plots of liquid water content as a function of altitude show significant liquid water to below 3 km. Ice crystal mass was increasing through the riming process and vapor growth with some of this increased mass lifted to higher altitudes where a notable increase in ice water content was observed just above 4 km. In the summer occlusion the significant increase in ice water content occurred much more quickly, taking place at around 6 km. Again this was ~1 km above the level where significant liquid water content was observed, suggesting that greater ice water content may have been lifted from regions below.

The depth of cloud in the cold frontal case appears to have significantly affected the ice water content, with maximum median values of ~0.025 g m⁻³ compared to values of 0.25 and 0.3 g m⁻³ in the occluded front and warm front, respectively. This may be because ice particles have relatively little time in an environment supersaturated with respect to ice, leading to less growth potential through the Bergeron–Findeisen process. In all three cases the highest ice water contents were observed where environmental temperatures were ~−5°C, where secondary H–M ice production through rime splintering is most efficient.

Radar data from CAMRa (Fig. 5) allowed us to observe the evolution of the cold front as it made progress eastward. The surface cold front produced a dominant convective feature (~50 dBZ) by ~1500 UTC that delivered significant liquid water to cloud top where the initiation of the ice phase in regions above this feature took place through primary ice nucleation. This was well represented in the ZDR field with high values being returned by pristine platelike crystals. This region of glaciation expanded over time and pushed ahead of the convective feature before the growth of the hydrometeors through riming, aggregation and vapor diffusion took place as they descended into the H–M zone below. By this time the characteristics of the front strongly suggest a kata-type system and the unfolded radial velocity field from CAMRa suggested an element of forward motion in the cloud tops relative to the lowest layers of the front. This is likely to have pushed the ice crystals initiated here ahead of the surface cold front.

5. Conclusions

With the use of a suite of cloud microphysics instruments, remote sensing measurements, and meteorological data we...
have described and compared the microphysical structures related to the mesoscale characteristics of three contrasting frontal systems.

- Ice formation processes in each case were different. The warm and occluded front presented cloud-top temperatures cold enough (\(-37^\circ C\)) for homogeneous and heterogeneous ice nucleation to occur, while the ice phase in the low altitude, relatively warm cloud tops (\(-13^\circ C\)) of the kata-type cold front, was initiated through heterogeneous ice nucleation only.
- The relatively warm cloud-top temperatures of the cold front meant that cloud top was in close proximity to the H–M temperature zone below, leading these two areas to become coupled with the ability to directly influence each other. This is highlighted by the presence of columnar ice crystals produced through rime splintering being present at cloud top, as a result of being lifted up from below. However, ice number concentrations at cloud top, measured by the 2D-S, were in reasonable agreement (within a factor of 2) with the predicted primary IN using the DeMott et al. (2010) parameterization. Concentrations of ice particles elsewhere in the cold front reached \(80 L^{-1}\), one to two orders of magnitude greater than the predicted primary IN concentration at cloud top. This enhancement was due to secondary ice production.
- In all three cases the 2D-S shadow imaging probe measured the greatest ice number concentrations and ice water contents in the H–M temperature zone \(-5^\circ C\). Ice water contents were calculated using the Brown and Francis (1995) mass–dimensional relationship. The integrated size distribution from the CDP was used to calculate liquid water contents and concentrations of larger droplets \(80 \mu m\) were determined by the 2D-S shadow imaging probe. In all cases significant liquid water contents (>0.01 g m\(^{-3}\)) and concentrations (\(10–150 L^{-1}\)) of larger supercooled drops \(80-\mu m\) diameter were observed and this is likely to have facilitated the rime-splintering mechanism and contributed directly to generating these high mass and number concentrations.
- The dynamics of the kata cold front had a direct influence on the distribution of precipitation and location of the most significant area of SIP. This may have occurred as the component of forward motion in these types of fronts carried ice crystals at cloud top ahead of the surface cold front generating the observed patterns. Vertical motions were inferred for the cold front using data from the 3-GHz

FIG. 22. (left side of panels) Droplet concentration and (right side) diameter as a function of altitude from the CDP for the (a) cold, (b) warm, and (c) occluded front cases. Droplet concentrations include 25th, 50th, and 75th percentiles, whiskers to 10% and 90%, and outliers (green circles) between 95% and 100%. Droplet diameter represented by median values as a function of altitude.
advanced meteorological radar at CFARR. Velocity data from the aircraft were unavailable due to the probe icing up.

- Heavily rimed snow particles were frequently observed and are likely to have provided the sites on which rime splintering took place.

- The most extensive study to date into splinter production as a function of rimer velocity covered the range 1.5–12 m s$^{-1}$ (Saunders and Hosseini 2001). However, the terminal velocity of snowflakes composed of ice crystals exhibiting various growth habits has been shown to be outside of this range (<1.5 m s$^{-1}$) (Heymsfield 1972; Langleben 1954). In all the cases presented here it is probable that accretion of supercooled water on snow particles traveling at low velocities is the mechanism by which ice crystal number concentrations are enhanced above levels observed anywhere else in each system. Therefore, experimental studies of SIP at low riming velocities are needed to quantify this process. The role of supercooled larger drops in the H–M temperature zone also needs further investigation.

**Acknowledgments.** This work was supported by the Natural Environment Research Council and the Met Office. The core aircraft data were contributed by the Facility for Airborne Atmospheric Measurement.

**REFERENCES**


