Chapter 4

Ice Fog: The Current State of Knowledge and Future Challenges

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ABSTRACT

Ice fog is a natural, outdoor cloud laboratory that provides an excellent opportunity to study ice microphysical processes. Ice crystals in fog are formed through similar pathways as those in elevated clouds; that is, cloud condensation or ice nuclei are activated in an atmosphere supersaturated with respect to liquid water or ice. The primary differences between surface and elevated ice clouds are related to the sources of water vapor, the cooling mechanisms and dynamical processes leading to supersaturation, and the microphysical characteristics of the nuclei that affect ice fog crystal physical properties. As with any fog, its presence can be a hazard for ground or airborne traffic because of poor visibility and icing. In addition, ice fog plays a role in climate change by modulating the heat and moisture budgets. Ice fog wintertime occurrence in many parts of the world can have a significant impact on the environment. Global climate models need to accurately account for the temporal and spatial microphysical and optical properties of ice fog, as do weather forecast models. The primary handicap is the lack of adequate information on nucleation processes and microphysical algorithms that accurately represent glaciation of supercooled water fog. This chapter summarizes the current understanding of ice fog formation and evolution; discusses operating principles, limitations, and uncertainties associated with the instruments used to measure ice fog microphysical properties; describes the prediction of ice fog by the numerical forecast models and physical parameterizations used in climate models; identifies the outstanding questions to be resolved; and lists recommended actions to address and solve these questions.

1. Chapter overview

Ice fog has taken on a number of names since early times. According to the AMS Glossary of Meteorology ([American Meteorological Society 2017](#)), ice fog is also known as ice crystal fog, frozen fog, ice crystal haze,

Arctic mist, frost fog, frost flakes, air hoar, rime fog, or pogonip. Likewise, there are dozens of other names, in other languages, for this phenomenon. Strictly speaking, however, since we will be discussing fog that consists only of suspended ice crystals, for simplicity sake, the term *ice fog* will be the term used from here forward.

On a local basis, in populated regions, diminished visibility caused by ice fog presents a potential hazard on highways, in marine environments, and at airports.

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Local climate can also be impacted by ice fog because it alters the surface albedo. Likewise, on a global scale, there are large uninhabited areas, particularly in high latitudes and polar regions, where ice fog plays an important role in altering the climate by modulating the heat and moisture fluxes in the surface layer and lower troposphere (Curry et al. 1996; Beesley and Moritz 1999). During Arctic winters when temperatures fall well below -30°C and relative humidity with respect to liquid water (RH_w) exceeds 80%, even a shallow layer of ice fog will significantly affect the surface energy budget (Blanchet and Girard 1995; Curry et al. 1990, 1996). Sea ice thickness and snow cover also are impacted because of ice fog's interaction with radiation (Curry and Ebert 1990), and the formation of ice crystals serves to remove water vapor from the lower troposphere (Curry 1983; Blanchet and Girard 1994, 1995). Girard and Blanchet (2001a,b) estimate that diamond dust (freely falling larger ice crystals) may increase the downward flux of infrared radiation at the surface by as much as 60 W m^{-2} during the wintertime. The actual magnitude of ice fog's climate effect remains largely unknown because current climate models still do not accurately represent the temporal or spatial formation of fog or its microphysical and optical properties.

Our understanding of how ice fog forms, evolves, dissipates, and impacts our environment is complicated by the same factors that have challenged our general understanding of all clouds with ice, that is, insufficient information on 1) sources and properties of ice nuclei (IN); 2) atmospheric cooling rates and the dynamics that drive them; 3) morphology and terminal velocities of ice crystals, particularly at sizes less than $50\text{ }\mu\text{m}$; 4) growth by diffusion and aggregation of ice; 5) ice crystal optical properties; and 6) Earth's surface characteristics, such as moisture and albedo. Much of the missing information is not only due to the paucity of observations that have been made in ice fog but also because of the limitations and uncertainties of the sensors that gather the data.

In the remainder of this chapter the reader will be introduced to how ice fog forms and evolves, accompanied by what we currently know about its microphysical properties, based upon those measurements that exist (section 4b). Then there is a description of the instruments that are used for making in situ and remote sensing measurements (section 4c). The impact on the environment is explored with process models along with regional and global climate simulations. Some of these models are described in section 4d. The final section highlights the obstacles that continue to hinder our understanding of ice fog and the future studies that will be needed to overcome them.

2. Ice fog formation, evolution, and microphysical properties

Ice fog consists of a suspension of ice crystals at temperatures generally lower than -20°C . Its classification as a fog is somewhat subjective, but operationally, that is, when used to forecast driving or flying conditions, it is defined by a visibility less than 1 km. Visibility is one property that distinguishes ice fog from another surface cloud type called "diamond dust" that forms under similar temperature and humidity regimes, but whose visibility is much higher than ice fog. Figure 4-1a shows ice fog in Fairbanks, Alaska (Schmitt et al. 2013), and Fig. 4-1b, in Yellowknife, Northwest Territories, Canada (Gultepe et al. 2014).

The term "diamond dust" is sometimes lumped together with ice fog; however, diamond dust and ice fog are differentiated mostly by the size and concentration of the ice crystals. Girard and Blanchet (2001a) distinguish diamond dust from ice fog according to the upper size threshold d and the number concentration N_i . For diamond dust, they used $d > 30\text{ }\mu\text{m}$ and $N_i < 4000\text{ L}^{-1}$, and for ice fog, $d < 30\text{ }\mu\text{m}$ and $N_i > 1000\text{ L}^{-1}$. Nevertheless, in most of the studies that have been reported [section 4b(2)], the sizes and concentrations of ice fog crystals frequently fall outside these ranges, so that the separation between diamond dust and ice fogs remains somewhat blurred.

The formation of ice fog, its growth, and its eventual dissipation are linked to complex interactions between atmospheric motions (horizontal and vertical) and thermodynamics (temperature and water vapor) coupled with cloud condensation nuclei (CCN) and IN. These processes are described in greater detail in the following subsections.

a. Meteorological conditions

Ice fog most frequently forms at temperatures $T < -20^{\circ}\text{C}$. In the Arctic, for example, the main mechanism for ice fog formation is the advection of warm, moist air from midlatitudes followed by radiative cooling at constant pressure (Curry et al. 1990). Similarly, ice fog forms over the Antarctic by the same mechanism (Bromwich 1988). There are, however, other atmospheric drivers that lead to the rapid cooling and increase in relative humidity that are necessary precursors to ice formation: (i) orographic lifting of humid air (Lax and Schwerdtfeger 1976), (ii) condensation over open leads in the sea ice (Ohtake and Holmgren 1974; Schnell et al. 1989; Gultepe et al. 2003), and (iii) humidity excess due to water vapor released from cities (Benson 1970).

The other important meteorological feature that usually precedes and then accompanies evolution of ice

FIG. 4-1. Images of ice fog in (a) Fairbanks and (b) Yellowknife (used by permission of AMS).

fog is the formation of a temperature inversion and deep stable layer above the surface caused by the surface radiative cooling (Wexler 1936, 1949; Oliver and Oliver 1949; Robinson and Bell 1956; Ohtake 1967; Bowling et al. 1968; Wendler 1969; Gultepe et al. 2007). This layer is typically formed under conditions of low winds and cloud-free skies. The depth of the inversion can vary from less than 100 m to more than 1 km. Schmitt et al. (2013) showed the top of the fog layer (Fig. 4-1a) defines the top of the inversion that is less than 100 m in depth. The ensemble of T and dewpoint (T_d) vertical profiles shown in Fig. 4-2 illustrates the range of inversion depths compiled during the 2011 Fog Remote Sensing and Modeling–Ice Fog (FRAM-IF) project (Gultepe et al. 2015). The inversion will eventually erode because of turbulent mixing, heating of the surface by solar radiation, or advection bringing warmer air masses.

Not only are low temperatures needed to initiate the formation of ice fog, but the rate at which the air cools is also a critical factor. The aerosol particles that activate as CCN or IN will do so only when the atmosphere has a relative humidity with respect to ice (RH_i) that is greater than 100%, that is, supersaturated. The rate at which the air cools determines the initial supersaturation (SS) prior to the formation of supercooled water droplets or ice crystals. The magnitude of SS largely

determines the number of nuclei that are activated along with their subsequent growth by vapor deposition. Accurately forecasting the cooling rate and RH_i is a major challenge. This is due to the multiple factors that contribute to how rapidly the temperature decreases (e.g., surface characteristics that impact the terrestrial infrared radiation and radiative cooling, the depth, T , and RH of advected air masses) and the chemical and physical properties of potential cloud particle nuclei, to name just a few.

b. Ice crystal formation, evolution, and microphysical properties

Figure 4-3 provides a conceptual framework for understanding the pathways through which the crystals in ice fog form. In simplest of terms, the ice crystals that form in fog are generated through either homogeneous or heterogeneous nucleation. In the former case, water droplets are formed from activated CCN and then freeze homogeneously when the temperature decreases below -38°C . In the second case, ice crystals can form following a number of pathways that are less understood than that of homogeneous nucleation, that is, immersion freezing, contact freezing, or deposition ice nucleation. Immersion and contact freezing occurs when an activated water droplet freezes at temperatures warmer

FIG. 4-2. An ensemble of temperature (red) and dewpoint (green) profiles from radiosonde measurements during January 2011 in Yellowknife. The thick lines are for 0000 UTC 13 Jan 2011.

than -38°C (Gultepe 2015). In the former case, the droplet freezes because of the composition of the nuclei on which it condensed, whereas in the second case, freezing is caused by the supercooled droplet coming into contact with an IN particle. Deposition nucleation is the activation of an ice crystal by the direct transfer of water vapor to the surface of an IN particle in the absence of liquid water. Chapters 1 (Kanji et al. 2017) and 8 (Cziczo et al. 2017) provide detailed information on IN properties and techniques for measuring them. The interested reader is encouraged to access these chapters for a review of the current state of knowledge on IN.

An open question at this time concerns the process that is the most common for ice fog formation: the formation of a fog of supercooled droplets followed by a transition to ice crystals or the formation of ice crystals directly by heterogeneous ice nucleation. The frequency

of ice fog that forms at T higher than -30°C suggests that ice crystals may be forming on IN that are active at these temperatures (Gultepe et al. 2007). Early observations, however, provide evidence that super cooled liquid clouds often precede the ice fog formation, as is discussed in the following section.

1) ICE CRYSTAL FORMATION AND EVOLUTION

In a review of Arctic aerosols, Garrett and Verzella (2008) summarize the history of observations that have been made of the various types of particles that are found in this region. Much of the early research was propelled by the need to understand the nature of Arctic haze that caught the attention of researchers in the late 1940s. Mitchell (1957) observed this haze from aircraft and deduced from its color that the particles could not be much larger than $2\text{ }\mu\text{m}$ in size. Subsequent

FIG. 4-3. A conceptual drawing of the pathway by which crystals form during the evolution of ice fog. A warm, moist air mass arrives over a region of surface that then is radiatively cooled. Some ambient particles like sulfates (yellow), haze droplets (blue), bioaerosols (green), dust (red), or other aerosol types can serve as CCN or IN. As the air cools, some of the CCN activate as water droplets. Further cooling leads to ice crystal formation through immersion or contact freezing. Any remaining supercooled droplets freeze homogeneously when the temperature decreases below -38°C . In the absence of liquid water, some types of IN will activate as ice crystal by deposition nucleation.

studies have now shown that there are myriad types of aerosol particles in Arctic haze and also deposited on the surface that cannot be attributed to local sources, for example, dust with varying compositions of metals that can only come from combustion, soot made up of varying fractions of organic and elemental carbon, and bioaerosols that have been traced to local blooms of algae. Other bioaerosols are likely transported from more distant sources, along with the pollutants that are associated with industrial processes.

As discussed by Kanji et al. (2017, chapter 1, and references therein), there has been a great deal of research in the past 20 years that has evaluated the activity of aerosol particles that serve as CCN or IN. Although many types of dust have been identified as IN, much depends on the morphology of the dust and what other chemical compounds might have been deposited on the surfaces during their transport to Arctic regions from distant sources. Hence, given the diversity of aerosol in regions of ice fog formation, it is reasonable to assume that some fraction of them could serve as IN. On the

other hand, many of the aforementioned aerosols have also been identified and measured as hygroscopic and capable of activating as cloud droplets. Clearly more research and measurements are needed to study the actual nuclei found within ice fog crystals before a definitive answer will be forthcoming.

The first documented measurements of the nuclei of ice fog crystals was by Kumai (1966), who studied ice crystals that were captured on a grid coated with a colloidal film. One set of samples were taken in an ice fog at the Fairbanks airport and another set in the region near a coal-fired power plant located near the same city. The residue of 105 evaporated crystals from the airport and 102 from the power plant were examined under an electron microscope with a 10^4 magnification, along with their diffraction patterns. Figure 4-4a from Kumai (1966) illustrates the nuclei from one of the crystals taken near the power plant. All of the nuclei were identified as combustion by-products, and more than 90% of them were in ice crystals approximately spherical in shape. Oliver and Oliver (1949) had

FIG. 4-4. Properties of nuclei from ice crystals captured near the power plant in Alaska. All of the nuclei were identified as combustion by-products and more than 90% of them were from ice crystals (the others were from liquid droplets). (a) The nuclei shown are identified as combustion particles, likely soot; (b) the nuclei number concentration vs supersaturation at the Syowa Station; (c) the monthly mean IN number concentration vs months as a function of temperature (adapted from Kumai 1966).

hypothesized that the ice crystals in fog in the same region as studied by Kumai (1966) had formed as droplets on smoke particles, but they made no direct measurements to support this conclusion. Likewise,

Thuman and Robinson (1954) had reached a similar conclusion based upon the types of ice crystals that they had observed (droxtals) that had evolved from frozen water droplets (see next section). A later study

(Robinson et al. 1957) had also traced the ice fog formation to local air pollution. It is important to note that the nuclei observed by Kumai (1966) were the residue of an ice crystal; however, that ice crystal might have previously been a water droplet, that is, the combustion particle might have been a CCN rather than an IN.

The ground breaking measurements by Kikuchi (1971a,b, 1972) of CCN, IN, and ice crystals in the Antarctic suggested that ice fog is more likely to evolve from frozen water droplets. Figures 4-4b and 4-4c show concentrations of CCN and IN, the former per cubic centimeter as a function of SS and the latter per liter. Kikuchi measured the CCN using a thermal diffusion chamber (Twomey 1963) and IN using a mixing chamber (Maruyama and Kitagawa 1966). Even at the lowest SS of 0.05%, the CCN concentrations are more than four orders of magnitude larger than those of the IN. The CCN and IN in the Antarctic are very likely of different composition than those found in the Arctic, although Kikuchi did not analyze the composition of the nuclei. During that period of time there was speculation that the IN were meteoritic in composition (Maruyama 1961; Maruyama and Kitagawa 1967), but Kikuchi (1971b) found no correlation with the observed meteor showers. The increased IN with decreasing T (Fig. 4-4c) is some of the earliest evidence of the relationship between IN activity of aerosols and temperature that has subsequently been validated by numerous studies (see Kanji et al. 2017, chapter 1).

As early as 1949, Oliver and Oliver (1949) were reported on Alaskan ice fog and the conditions under which they formed. From their observations near Fairbanks they concluded that ice fog was less likely to form below -30°C , more likely to form between -30° and -40°C , and nearly inevitable when T fell below -46°C . Robinson and Bell (1956), also working in Fairbanks, found that ice fog formed in conditions of low temperatures, clear skies, and low wind speeds associated with polar continental air masses. They observed that the moisture tends to be uniformly distributed within the layer, which was usually 50–100 m deep, with a visually, well-defined upper threshold between clear and cloudy air. Robinson et al. (1957) linked ice fog to air pollution that was related to local sources.

Studies continued in the 1960s on the formation of fog due to local air pollution (Benson 1965; Benson and Rogers 1965), and in parallel, research was under way on the source of ice crystal nuclei and their microphysical characteristics (Kumai and O'Brien 1964; Kumai 1964, 1966). The creation of ice fog by power plant emissions was studied by Richardson (1964) and Porteous and Wallop (1965, 1966, 1970), who also proposed methods

to reduce the impact these fogs had on flight operations at the air force base near Fairbanks. Huffman and Ohtake (1971), Otake and Huffman (1969), and Ohtake (1967, 1970a,b) also studied the formation of ice fog that was generated by the cooling water vapor emissions from power plants, as well as from motor vehicles. Bowling et al. (1971) calculated the temperature difference between radiating ice crystals and the surrounding air while Huffman and Ohtake (1971) proposed a mechanism for ice fog formation by the formation of supercooled water droplets that first grow by condensation of water vapor then freeze and continue to grow by vapor deposition.

The studies that were conducted in Alaska during the 1950s, 1960s, and early 1970s focused on ice fog that formed on anthropogenic CCN and IN that were produced from combustion, that is, aircraft, motor vehicles, and power plants. All of the researchers that were involved with these studies concluded that the ice fog was formed via the pathway of supercooled droplet formation followed by freezing.

Sakurai (1968, 1969) reported on a very well-designed study to investigate the transition of supercooled liquid fog to ice fog in the city of Asahikawa, Hokkaido, Japan. Figure 4-5, taken from Sakurai (1968), illustrates one selected time period when the temperature decreased from -23° to -27°C , and the RH_i increased from approximately 105% to over 110% (Fig. 4-5a). During this period, the N_i increased by almost three orders of magnitude from several tens per liter to over a thousand per liter (Fig. 4-5b). As the N_i increased, the mean size decreased (Fig. 4-5c). Samples of the ice crystals were captured on glass slides covered with cedar oil or water blue solution. Examples of these ice crystals are shown in Fig. 4-5c at the times that they were sampled. After the 1967 study, Sakurai (1968) concluded that ice crystals grew more rapidly in the presence of liquid droplets, a clear indication of the Wegener–Bergeron–Findeisen (WBF) process by which ice crystals grow at the expense of water droplets because of the higher supersaturation with respect to ice compared to liquid (Wegener 1911; Bergeron 1935; Findeisen 1938). Sakurai also noted that there was a distinct phase change that occurred when the minimum T was reached. In a follow-up experiment, Sakurai (1969) repeated his measurements in the same location with similar results, concluding that the most important result was the confirmation that the supercooled droplets were changed rapidly to ice owing to the cooling to lower than -20°C and that the amount of liquid water content (LWC) was kept nearly the same in spite of the phase change to ice crystals. Figure 4-6 shows photos of some of the ice fog samples that were

FIG. 4-5. (a) A time period where the temperature decreased from -23° to -27°C and the RH_i increased from approximately 105% to over 110%. (b) Ice crystal number concentration time series in. (c) As the number concentration increases, the mean size decreases (adapted from Sakurai 1968).

taken at the same time, supercooled droplets (Fig. 4-6a) and ice crystals (Fig. 4-6b) (Sakurai 1969). The time series below the photos (Figs. 4-6c,d) illustrates the transition from supercooled droplets, to mixed phase, then to all ice.

Yamashita et al. (1971) studied ice fog in the city of Asahikawa, as well, but took ice crystal samples in five regions instead of just two, like Sakurai (1968, 1969). A different method was employed than was used previously by other investigators, who collected particles on glass plates. Yamashita et al. (1971) used an oil-coated glass slide attached to a 90-cm rod that he swung at an approximately constant rate in the vicinity of the ice fog crystals. Although he concluded that the rod has a 100% collection efficiency for crystals $> 30\text{ }\mu\text{m}$, he gives no estimate of the efficiency for smaller crystals. Hence, it is likely that the samples were biased toward larger crystal sizes given the results of size distributions reported by Ohtake and Huffman (1969) and Kikuchi (1971a,b). Regardless of this uncertainty, a comparison of the ice fog in different areas provided an interesting insight into

the nature of their formation and evolution. Yamashita et al. (1971) concluded that in the city, by the river that flows through it, there was a continually forming supercooled liquid fog. A similar fog is formed in the lee side of a paper mill that emits a heavy cloud of water vapor. However, whereas the city-formed liquid fog transitions to an ice fog, the paper mill fog instead produces snow crystals that were much larger than the ice fog crystals. This suggests that the nature of the cloud nuclei were quite different between locations and thus leading to different formation and evolution pathways for the ice crystals.

2) ICE FOG MICROPHYSICAL PROPERTIES

The first study reported to capture ice crystals in ice fog was reported by Thuman and Robinson (1954), who captured them on oil coated slides for examination with an optical microscope. These are the first reported droxtals in ice fog. Figures 4-7a and 4-7b-7 show pristine crystals and droxtals, respectively (Thuman and Robinson 1954). The latter term was given by these researchers as a combination of “drop” and “crystal” since they concluded that they are formed by freezing of water droplets. Not until 10 years later were there follow-on studies of ice fog crystals, still in the same region. Figure 4-7c (Kumai 1966) shows a photo of some of the ice crystals that were captured on the oil-coated substrates (the size distribution that was derived from these measurements is shown in Fig. 4-9a). The peak size was between 4 and $5\text{ }\mu\text{m}$, the number concentration from 100 to 200 cm^{-3} , and the ice water content (IWC) ranged from 0.02 to 0.1 g m^{-3} . The methodology for calculating the number and mass concentrations was not described in the paper.

In Alaskan ice fog, Ohtake (1970c) collected ice crystals and reported on the unusual shapes of some of them. Figure 4-7c illustrates polyhedral (inset) and “block” crystals, as defined by Ohtake (1970c). Ohtake and Suchanek (1970) also investigated the electrical properties of the ice fog crystals during this time period, concluding that they carried no significant charge.

Recent measurements of ice fog crystals have been captured with more modern methods in Fairbanks and Yellowknife. Figure 4-7d illustrates droxtals collected in Fairbanks with a video ice detector (VID; Schmitt et al. 2013) an instrument described below in section 4c. An assortment of ice crystals with maximum diameters $< 50\text{ }\mu\text{m}$ that were measured in Antarctic diamond dust with a cloud imaging probe (Lawson et al. 2006) are shown in Fig. 4-7e, and sub- $50\text{-}\mu\text{m}$ crystals observed in Yellowknife during the FRAM-IF campaign (Gultepe et al. 2014) are seen in Fig. 4-7f. This latter project is discussed in the following section.

FIG. 4-6. The photos of (a) supercooled droplets and (b) ice crystals. (c), (d) The time series over two time segments below the photos illustrate the transition from supercooled droplets to mixed phase then to all ice (adapted from Sakurai 1969).

A revolutionary result from the Kumai (1966) study was the evidence discovered for the sintering and splintering of ice crystals. Hobbs and Mason (1964) had demonstrated in the laboratory that ice crystals can become attached to one another, accelerating their fall velocity, through a process called sintering. Kumai

(1966) showed this process occurring with natural fog ice crystals (Fig. 4-8a). In this same figure, a frozen water droplet that has been sintered to an ice crystal is shown forming a splinter as it freezes. Kumai (1966) attributes this splintering to the expulsion of CO₂ during the freezing process. Splintering is one of the processes of

FIG. 4-7. (a) Pristine crystals and (b) droxtals (adapted from [Thuman and Robinson 1954](#)). (c) Polyhedral (inset) and “block” crystals ([Ohtake 1970c](#)). (d) Droxtals captured in Fairbanks with a VID (adapted from [Schmitt et al. 2013](#)). (e) The CPI, with a resolution of about $3\ \mu\text{m}$, measured images over the Antarctic region (adapted [Lawson et al. 2001](#)). (f) Ice fog crystal images taken with a microscope during the FRAM project over Yellowknife.

secondary ice production that has been proposed as a possible ice crystal multiplication mechanism ([Hallett and Mossop 1974](#); [Brewer and Palmer 1949](#); [Findeisen and Findeisen 1943](#); [Field et al. 2017](#), chapter 7). [Kikuchi \(1972\)](#) published similar results from measurements made in the Antarctic. An example of his measurements is shown in [Fig. 4-8b](#). The frequency distributions of the ratio of the neck between the frozen droplets and the droplet radius and the neck width are shown in [Figs. 4-8c](#) and [4-8d](#).

Size distributions of ice fog crystals in Fairbanks were measured by [Huffman and Ohtake \(1971\)](#). These were then used to derive an empirical relationship between ice crystal size and visibility ([Ohtake and Huffman 1969](#)). [Figure 4-9a](#) shows size distributions of mass and number concentration for ice fog that occurred in the Alaska region ([Kumai 1966](#)). The size of ice crystals ([Fig. 4-9b](#)), taken over a temperature range from -32° to -40°C had number concentrations that ranged from

FIG. 4-8. (a) A frozen water droplet sintered to an ice crystal forming a splinter as it freezes (adapted from Kumai 1966). (b) Sintering that occurred over the Antarctic (adapted from Kikuchi 1972). The frequency distributions of (c) the ratio of the neck between the frozen droplets and the droplet radius and (d) the neck width.

30 to almost 700 cm^{-3} , average diameters from 3.5 to $8.1\mu\text{m}$ and median volume diameter from 4.3 to $11.4\mu\text{m}$. At the same time that the ice crystals were collected (on oil coated slides) the visibility was measured with a transmissometer and ranged from 260 to 1660m. An empirical relationship was derived that

relates the visibility to the IWC, average diameter, and total number concentration. This function fit the measured visibility with an uncertainty of -5% to 12% about the mean. Kumai and Russell (1969) also conducted studies of the optical properties of ice fog, but looked at the attenuation and backscattering by ice crystals at

FIG. 4-9. (a) The distributions of mass and number concentration for ice fog events that occurred near Fairbanks, and (b) size distributions at four locations (adapted from Kumai 1966). (c) Representative size distributions from the Fairbanks region and (d) the relative frequency of different type of ice crystals as a function of size (adapted from Schmitt et al. 2013). (e) Number density of ice crystals obtained from a CPI for various time periods during an Antarctic project (adapted from Lawson et al. 2006); (f) for an ice fog event, hourly ice crystal spectra obtained using a GCIP during the FRAM project (Gultepe et al. 2016).

infrared wavelengths. Representative size distributions from the Fairbanks region and the relative frequency of different type of ice crystals as a function of size are shown in Figs. 4-9c and 4-9d (Schmitt et al. 2013), respectively. Similarly, the size distributions of ice fog small crystals in the Antarctic using a cloud particle imager (CPI) and at Yellowknife using the grayscale cloud imaging probe (GCIP) are provided in Fig. 4-9e (Lawson et al. 2006) and Fig. 4-9f (Gultepe et al. 2016), respectively.

3) FIELD PROGRAMS TARGETING ICE FOG

Following the activities in Alaska, Antarctica, and Japan, ice fog field programs seemingly ended in the early 1970s; it appears that there was no more research,

published in the open literature, on ice fog microphysics until almost 40 years later. In the interim period there were studies that looked at the conditions leading to deep temperature inversions in Fairbanks (Wendler and Nicpon 1975), terrain-induced ice fog in the Antarctic (Lax and Schwerdtfeger 1976), attenuation by ice fog at multiple wave lengths (Seagraves 1981), and lidar detection of ice fog over Arctic Sea ice (Schnell et al. 1989). Based on an extensive literature search, there does not appear to be any studies of ice fog that include measurements of the ice microphysics until 2005. In that year the FRAM project was initiated (Gultepe et al. 2008, 2009), focused on fog in general. Then the FRAM-IF project was conducted in 2010/11 that specifically targeted ice fog.

The FRAM-IF project took place from 25 November 2010 through 5 February 2011 near the Yellowknife International Airport situated in the Northwest Territories of Canada (Gultepe et al. 2014, 2015). The primary objective of the campaign was to better document ice fog properties, using a modern suite of instruments, and that would provide more extensive information to improve fog forecasting and parameterization of ice fog in climate models. The suite of instruments, the largest ever to study ice fog properties, included in situ and remote sensors to measure aerosol, droplet, and ice crystal size distributions; supercooled liquid water content (SLWC); and vertical profiles of temperature, humidity, and water content. More details on some of these instruments are found in Gultepe et al. (2014, 2015) and below, in section 4c. During this time period 14 fog events were recorded. The size distributions—measured with two single-particle optical spectrometers, a fog monitor (FM), and GCIP—are shown in Fig. 4-9f as a function of hour of the day. As will be discussed in further detail in section 4d, where we discuss forecasting of ice fog and impacts on weather and climate, the measurements made during this project provided information on the microphysical properties of ice fog and their relationship to visibility.

A more recent study of ice fog took place in a nonpolar region but at a location similar to where other ice fog is frequently encountered in many parts of the world. Fernando et al. (2015), Gultepe et al. (2016), and Gultepe and Heymsfield (2016) describe a field campaign designed to study fog processes in complex terrain. The Mountain Terrain Atmospheric Modeling and Observations (MATERHORN) project was conducted in the Wasatch Mountains near the cities of Salt Lake City and Heber, Utah, from 5 January to 15 February 2015. During the fog events, the visibility ranged from 100 m up to 10 km, IWC varied between 0.01 and 0.2 g m^{-3} , the ice crystal number concentrations were as high as 100 cm^{-3} and surface

FIG. 4-9. (Continued)

radiative cooling was 200 W m^{-2} . Some of the remote sensing measurements from this project will be illustrated in the next section on measurement techniques.

3. Measurement techniques

Documenting the properties of ice fog requires measurements of the meteorological state of the atmosphere (T , RH , and pressure), winds (horizontal and vertical), composition and morphology of CCN and IN, and the microphysical and optical properties of the ice fog crystals. These latter properties include the number and mass size distributions, habit, light-scattering and absorption coefficients, and asymmetry factor. From these basic parameters other important metrics like effective radius, single scattering albedo (SSA), and optical depth can be derived. Ideally, the optical properties are needed over a broad range of infrared and solar wavelengths.

Prior to the mid-1970s there were no instruments available that could take continuous measurements of

ice fog crystal properties. The fog water droplets and ice crystals were captured by impaction on a surface coated with a layer of material where they would impinge and retain their original shape until photographed. Methods had already been developed to measure the concentration of CCN (Twomey 1963) and IN (Maruyama and Kitagawa 1966) and were used in the Antarctic to document the aerosol population there (Kikuchi 1971a,b). It would not be until the twenty-first century that more modern techniques were employed to characterize the microphysics of ice fog. In the following discussion, we describe the instruments that have been employed in the FRAM-IF and MATERHORN projects and briefly introduce others that would be useful to deploy in future ice fog projects.

a. In situ sensors

Sensors that are currently available for measuring the properties of ice fog crystals and their CCN, IN, and

FIG. 4-9. (Continued)

supercooled droplet precursors are described in great detail in other chapters of this monograph (Kanjji et al. 2017, chapter 1; Cziczo et al. 2017, chapter 8; Baumgardner et al. 2017, chapter 9; McFarquhar et al. 2017, chapter 11). Here we list the aerosol and cloud properties that should be documented and the sensors that are available to make these measurements. For more specific details of the operating principles, limitations, and uncertainties, the interested reader is directed to the aforementioned monograph chapters and the extensive references contained therein.

1) CCN AND IN PROPERTIES

An understanding of how ice fog forms requires, at a minimum, the measurement of the number concentration of IN active as a function of T and CCN as a function of SS. However, in order to link these types of nuclei to their sources and to accurately parameterize them in process or climate models, it is beneficial to know their composition and microphysical properties. This type of detailed analysis requires either an approach like Kumai

and Russell (1969) who used electron microscopy to measure the residual of evaporated ice crystals or implementation with a combination of an aerosol mass spectrometer situated behind a counterflow virtual impactor (CVI). The CVI separates water droplets and ice crystals from aerosol particles, evaporates them, and delivers them to whatever instrument is connected to it downstream (Ström and Heintzenberg 1994). An aerosol mass spectrometer (Zelenyuk and Imre 2005) can provide the mass fractions of a variety of inorganic and organic ions, depending on the type and sensitivity of the instrument. There are two problems, however, with either of these approaches. The first is that the residual of an ice crystal is not necessarily an IN if the crystal was formed from the homogeneous freezing of a droplet. The second is that the residuals from a melting and evaporating crystal or droplet may also include aerosol particles that were scavenged by the cloud particle. Both of these issues present a challenge for unequivocally linking the composition of a particle to its potential activity as an IN or CCN.

2) CLOUD PARTICLE SIZE DISTRIBUTIONS

From the distribution by size of cloud particle number, mass (liquid or ice), optical and geometric cross section, and surface area, we can derive other bulk parameters needed to evaluate the impact of ice fog on visibility, water budget, and radiative fluxes. The first obstacle to overcome is the definition of “size.” If the particle measured is a supercooled water droplet the size would be a geometric diameter from which the terminal velocity, mass, and optical cross sections can be readily derived. Unless the ice crystal is a frozen droplet that remains spherical, defining its size is more problematic. The aerodynamic size determines ice crystal fall velocity and its optical size, as well as the optical cross section area; however, except for the simplest of shapes, there is no crystal size from which the mass can be directly derived. Given that (i) the lifetime of the ice fog partially depends on the fall velocity of the crystals, and (ii) the moisture budget on the ice mass and the climate impact on the optical cross section, uncertainties in the definition of crystal size pose a major problem that remains unresolved.

The instruments that are currently available to derive ice crystal size are single-particle optical spectrometers that measure how the cloud particle interacts with coherent light produced at a single wavelength by a laser. These instruments are summarized by Baumgardner et al. (2011, 2017, chapter 9) with a discussion on analysis methods found in chapter 11 of this monograph (McFarquhar et al. 2017). In short, for particles in the equivalent optical diameter (EOD) range from

approximately 2 to 50 μm , light scattering is used to derive the size (Knollenberg 1976). Cloud particles are directed through a focused laser beam and part of the light they scatter is collected with optical components that direct the photons to a photodetector where they are converted to a signal whose peak amplitude is related to the EOD using Mie theory. For ice fog research, such as conducted during FRAM-IF and MATERHORN, the instrument used is a commercially available fog monitor (Spiegel 2012). The principle uncertainty is introduced by the nonsphericity of the crystals, since Mie theory requires an assumption of sphericity. As discussed in chapter 9 (Baumgardner et al. 2017), the asphericity can lead to uncertainties in derived size of as much as $\pm 30\%$. The error propagates into the uncertainties when deriving water mass, terminal velocity, and optical cross section.

Optical imaging is used to measure particle size in a number of ways. The video ice particle sampler (VIPS) used by Schmitt et al. (2013) captures cloud particles on a moving film coated with silicone oil (Heymsfield and McFarquhar 1996; Schmitt and Heymsfield 2009) and photographs them with a video camera at a resolution of about 5 μm . The optical array probe (OAP), developed by Knollenberg (1970), measures the image of a cloud particle that is projected onto a linear array of diodes as the particle passes through a collimated laser beam. To date, the minimum resolution of the commercially available OAPS that use a single linear array to capture the image is 10 μm . The CPI has a resolution of about 3 μm (Lawson et al. 2001). Some of the images measured with this instrument in the Antarctic are shown in Fig. 4-7e. In addition to issues related to focusing (Baumgardner et al. 2017, chapter 9), differentiating liquid from ice is a challenge when the image consists of fewer than about 10 pixels. Given that these are two-dimensional images, the density and the mass estimates will be increasingly uncertain for sizes smaller than about 50 μm because of the limitations in resolving the ice crystal structure. Hence, for ice fog, the size accuracy of OAPs will be about the same as the light-scattering probes.

Another technique that holds promise for ice fog is that of holography (Fugal and Shaw 2009). Although the processing of the images produced by this technique is currently time intensive, the potential for evaluating a large volume of cloud to distinguish droplets and ice crystals with a size resolution as small as 5 μm makes this approach appealing for utilization in future projects.

3) DIRECT MEASUREMENT OF WATER CONTENT AND OPTICAL PROPERTIES

Given the uncertainty in deriving bulk quantities like IWC and extinction coefficient from the size distribution,

it is beneficial to measure these quantities directly. The CVI mentioned previously has been used to measure total condensed water content (Twohy et al. 1997; Mertes et al. 2007) by measuring the water vapor mixing ratio of the evaporated cloud particles as they enter the instrument. The CVI cannot distinguish liquid from ice, but for measuring the total water content it is quite sensitive and can measure as little as a few milligrams per cubic meter.

The extinction coefficient is commonly used to determine visibility and is based on attenuation of visible light. Visibility during ice fog (as well as in other types of fog) is obtained using transmissometers and current weather sensors such as the Vaisala present weather detector (PWD) and Sentry (Gultepe et al. 2014) that converts measured extinction to visibility. An extinction probe that was recently developed by Korolev et al. (2014) for airborne research could also easily be implemented for ground-based operations. The cloud extinction probe (CEP) utilizes the transmissometric method of measuring the intensity of light from a known source after it passes through a known volume.

The phase function of ice fog crystals can also be measured as was described by Lawson et al. (2006) using the polar nephelometer (PN). The PN directs the cloud particles through a focused laser beam around which are located an array of individual photodetectors that measured the light scattered from approximately 5° to 169° . The measurement is not of individual particles but of the scattering from the ensemble that falls within the sample volume of approximately 8 cm^3 .

b. Remote sensing platforms

Whereas in situ measurements provide detailed information on the microphysical properties of ice fog, measurements with remote sensors are necessary to quantify the thermodynamic profiles of the atmosphere and the depth and homogeneity of fog as these evolve over time. In addition, satellite measurements are the only way we can evaluate the impact of ice fog on broader regional and global scales.

1) GROUND-BASED PASSIVE REMOTE SENSING PLATFORMS

Microwave radiometers, such as the one that was used during the FRAM projects (Gultepe et al. 2008, 2014) can be useful for measuring the relative changes in the vertical distribution of liquid water path (LWP), T , and RHw, and providing time–height cross sections of ice fog conditions.

Figure 4-10 illustrates retrievals from one type of microwave radiometer, the Profiling Microwave Radiometer (PMWR) that was deployed during FRAM-IF (Gultepe et al. 2014, 2015). These profiles were made on

FIG. 4-10. An ice fog event observed by a microwave radiometer on the 17 Jan 2011 during the FRAM-IF project in Yellowknife; (top) temperature, (middle) RH with respect to ice, and (bottom) liquid water mixing ratio [adapted from Gultepe et al. (2014); used by permission of AMS].

17 January 2011 as the thermal inversion begins to form around 0300 local time (LST), as observed in Fig. 4-10 (top). By 1500 LST, the temperature near the surface decreased to nearly -35°C while the temperature at the top of the inversion increased to nearly -20°C [where RHi indicates saturation with respect to ice; Fig. 4-10 (middle)]. The water mixing ratio (Fig. 4-10c) shows the ice fog beginning to form below 500 m at 0700 LST, then increasing to a depth greater than 500 m by 2000 LST. The maximum derived water mixing ratio is estimated as nearly 0.8 g kg^{-1} . This particular illustration demonstrates the utility of such measurements for tracking temporal changes in the vertical cloud structure; however, the absolute accuracy of these types of radiometers is still under evaluation.

2) GROUND-BASED ACTIVE REMOTE SENSING PLATFORMS

Active remotes sensors like lidar (i.e., light detection and ranging) and radar provide greater detail on the vertical structure of fog and a more clear discrimination between droplets and ice crystals. A lidar transmits a laser beam, either continuous or pulsed, and collects the light that is backscattered from aerosol and cloud particles. The intensity of scattered light is proportional to the power of the transmitted light and the surface area concentration of the atmospheric

particles (Sassen 1991). When the transmitted light is polarized, the scattered light will be depolarized by nonspherical particles like ice crystals, dust, or volcanic ash, and the depolarization ratio that is derived from this type of measurement is then used to determine if cloud layers are liquid water or ice (Sassen and Zhu 2009). The use of lidar is most effective for very thin clouds like ice fog or cirrus since the attenuation by liquid clouds limits the ability of the lidar to penetrate liquid cloud layers. Figure 4-11a shows a vertical profile of the linear depolarization ratio (LDR) through an ice fog layer that was measured during the 2008 FRAM ice fog project at Barrow, Alaska (now known as Utqiagvik), as part of the U.S. Department of Energy (DOE) Indirect and Semi-Direct Aerosol Campaign (ISDAC) project (Lindeman et al. 2011). This instrument was a micropulse lidar with polarization. In the figure, the ice fog has already formed at 0000 UTC and is about 750 m deep, decreasing to 250 m by 0400 UTC with a maximum LDR between -1.5 and -1 indicative of high ice concentrations (N_i) at the top of the cloud layer. For the same time period, a DOE millimeter cloud radar and a Vaisala CL51 ceilometer were also used for ice fog detection, and the measured reflectivity factor and backscatter coefficients are shown in Figs. 4-11b and 4-11c, respectively. All three platforms indicated the ice fog occurrence, location, and depth.

FIG. 4-11. Ice fog images obtained by a Doppler lidar during an ice fog event occurred at the DOE North Slope of Alaska (NSA) site in Barrow on 9 Apr 2008: (a) attenuated backscatter coefficient and (b) LDR. (c) The backscatter cross section obtained for the same time period using CL31 ceilometer. Ice fog event was observed between 0000 and 1000 UTC.

Millimeter-wavelength cloud radar (35 GHz) are sensitive to even relatively thin fog (Matrosov 2010; Gultepe et al. 2015), as shown in Fig. 4-11b.

3) SATELLITE PLATFORMS

Integration of satellite-based retrievals of ice fog microphysical parameters with surface-based sensors like lidar, cloud radar, ceilometer, and microwave radiometer, as well as surface in situ fog sensors, for example, fog and standard visibility sensors, can be used to improve the prediction and monitoring of ice fog micro and macrophysical properties.

Passive, low-earth-orbit and geostationary satellite measurements have been used for monitoring ice fog in the absence of higher-level cloud layers (Pavolonis 2010, Calvert and Pavolonis 2011; Gultepe et al. 2015). Calvert and Pavolonis (2011) describe the physical basis of the fog/low cloud detection algorithm for the Advanced Baseline Imager (ABI), flown on the GOES-R series of NOAA geostationary meteorological satellites. This algorithm is designed to quantitatively identify clouds that produce instrument flight rules (IFR) conditions for aircraft operations. These conditions are defined as

having a cloud ceiling between 150 and 300 m above ground level (AGL). The GOES-R fog product predicts the probability that the cloud ceiling is below 300 m AGL. At night, the algorithm utilizes the 3.9- and 11- μm channels to detect IFR conditions. During the day, fog/low clouds are determined using the 0.65-, 3.9-, and 11- μm channels. Although the GOES-R measurements hold great promise for documenting ice fog, satellite coverage over the Arctic regions is currently limited, and geostationary satellite observations suffer from reduced resolution in these areas (Gultepe et al. 2015).

4. Ice fog modeling

Models of cirrus and ice fog also have similarities and difference related to how physical processes are simulated. Both types of models either parameterize or explicitly represent the microphysical processes of ice crystal formation via homogeneous or heterogeneous nucleation and the subsequent crystal growth/sublimation/breakup. Because the physics of ice nucleation are the same no matter if crystals are forming in cirrus or ice fog, the primary differences are threefold: 1) the source and

composition of the condensation or ice nuclei, 2) the environmental forcing that leads to the supersaturations with respect to liquid or ice, and 3) the dynamics that produce the conditions under which the ice clouds disperse. In the following sections we discuss how ice fog is represented in current forecast and climate models.

a. Ice fog forecasting

A number of models are currently in use for operational forecasting of fog in general, but not for ice fog, specifically, although they could potentially be configured to do so. These include the Canadian High Resolution Deterministic Prediction System (HRDPS), North American Mesoscale Forecast System (NAM), and Weather Research and Forecasting (WRF) Model.

1) HRDPS MODEL

The HRDPS (Kehler et al. 2016) is based on the Canadian Global Environmental Multiscale forecast model (Côté et al. 1998). The HRDPS uses the ice nuclei parameterization of Meyers et al. (1992), a formulation based on a limited number of aircraft observations and it is written as

$$N_i = \exp\{a + b[100(S_i - 1)]\}, \quad (4-1)$$

where $a = -0.639$ and $b = 0.1296$, and S_i is the supersaturation with respect to ice. The original parameterization was valid only over the temperature range -7° to -20°C , ice supersaturations from 2% to 25%, and water supersaturations between -5% to $+4.5\%$, although it has been extrapolated outside these limits. The maximum values of N_i predicted by this equation can reach up to 100 L^{-1} at $T = -20^\circ\text{C}$, which are significantly less than N_i observed in typical ice fog, for example, during the FRAM-IF project when N_i often exceeded 1000 L^{-1} . Therefore, HRDPS simulations may result in an underestimate of N_i in ice fog. The visibility (Vis) prediction is based on a two-moment version of the Milbrandt and Yau (2005a,b) bulk microphysical scheme (MY2). In the MY2 scheme, ice crystals are represented by two categories: 1) ice, representing pristine crystals, and 2) snow, representing larger crystals and/or aggregates. Particle size distributions (PSDs) are represented by complete gamma functions with the two prognostic variables (for each category type x), total N_i , $N_{i,x}$, and the mass mixing ratio (q_x). Some limited simulations of ice fog that formed during the FRAM-IF project, in Yellowknife, were done with the MY2 scheme. These resulted in low values of N_i , compared to the measurements with a fog monitor, leading to an underestimate of simulated Vis compared to direct measurements of

visibility (Gultepe et al. 2015) where the model calculated visibility is obtained using Gultepe et al. (2015) as

$$\text{Vis} = 0.24(\text{IWC} \times N_i)^{0.5147}. \quad (4-2)$$

This suggests that the HRDPS, in its current implementation, may be limited for ice fog predictions without better parameterization of the ice formation processes. In particular, if the formation of ice is modeled only through heterogeneous nucleation, ice fog that forms via the homogeneous freezing of haze droplets will be underrepresented.

2) NORTH AMERICAN MESOSCALE FORECAST SYSTEM

For operational applications, the National Centers for Environmental Prediction's (NCEP) 12-km NAM (Rogers et al. 2009; Ferrier et al. 2002) is used for regular weather guidance over the continental United States, including Alaska. The NAM runs four times per day (0000, 0600, 1200, and 1800 UTC) and provides forecast products. It is not at the moment used for ice fog forecasting but could be adapted with appropriate microphysical parameterizations. The NAM postprocessor calculates Vis based on the work of Stoelinga and Warner (1999). Evaluation of some test simulations run using the NAM during FRAM-IF showed large biases in forecast versus observed Vis (Gultepe et al. 2009). Comparison of the NAM simulations with observations also indicated large uncertainty in forecast RH and IWC parameters. Therefore, two diagnostic methods for identifying ice fog conditions have been proposed. The first method uses boundary layer parameters from the NAM simulations (Zhou and Du 2010) and then predicts the occurrence of ice fog, but does not provide microphysical parameters such as IWC. The second method utilizes model runs that include prognostic microphysical parameters and provides IWC as a function of altitude. Based on the work of Zhou and Ferrier (2008) and Zhou (2011), this second method was used to predict IWC and subsequently the visibility during the FRAM-IF project. The measured and predicted visibilities were sometimes in fair agreement but there were frequent discrepancies in IWC and Vis that underscored the need for a better representation of the ice fog microphysical and optical properties than is currently possible with the NAM.

3) WRF MODEL

The WRF Model has been used for ice fog predictions over mountainous (Pu et al. 2016; Nygaard et al. 2011) and Arctic regions (Gultepe et al. 2015;

Kim et al. 2014). Nygaard et al. (2011) used one rendition of the WRF Model with a horizontal grid spacing of 333 m and tested three different parameterization schemes for the microphysics: the Morrison two-moment scheme (Morrison et al. 2005; Morrison et al. 2009), the Thompson scheme (Thompson et al. 2004, 2008), and the Eta grid-scale cloud and precipitation scheme (E. Rogers et al. 2001). They concluded that supercooled water amount and droplet concentration N_d over a hilltop in northern Finland strongly depends on the model horizontal resolution and microphysical scheme used. For their particular conditions and location, by assuming $N_d = 250 \text{ cm}^{-3}$, the particle size and amount could be predicted accurately when compared to observations. On the other hand, Pu et al. (2016), using the Thompson microphysics scheme (Thompson et al. 2004, 2008), concluded that ice fog prediction accuracy was very low over the valleys of the Wasatch mountains (Utah) using a WRF Model operationally.

b. Climate models

A detailed analysis of ice fog conditions affecting the radiative budget has been carried out by Girard and Blanchet (2001b) using a single-column model with explicit aerosol and cloud microphysics that was developed specifically to evaluate cloud–aerosol interactions in the Arctic. The aerosol and cloud size distributions are represented with 38 bins from 0.10 to 500 μm diameter. Three equations describe the time evolution of the aerosol, cloud droplet, and ice crystal spectra simulating coagulation, sedimentation, nucleation, coalescence, aggregation, condensation, and deposition. The model also accounts for the water–ice–phase interaction through the homogeneous and heterogeneous freezing, ice nuclei, and the WBF effect.

The model was developed to simulate Arctic ice fog that had an observed weekly mean frequency of 11% at the remote region of Alert in the Qikiqtaaluk Region, Nunavut, northern Canada, during 1991–94 (Girard and Blanchet 2001a). The model uses the Aerosol Dynamics Model (MAEROS2) developed by Gelbard et al. (1980) to compute the time evolution of size-segregated particles in the atmosphere, including condensed water, to simulate winter ice fog by the homogeneous freezing of haze droplets at a temperature below 238 K. The hypothesis tested with the model is that ice fog occurs at temperatures below 238 K, resulting from the homogeneous freezing of haze droplets (Bertram et al. 1996; Thuman and Robinson 1954). Furthermore, based upon high IN concentrations that were observed locally a few meters above the Beaufort Sea and downwind of open leads in the sea ice during spring 1998 during the

SHEBA experiment (D. C. Rogers et al. 2001; Gultepe et al. 2003), it was hypothesized that IN originating from phytoplankton in the sea might be the cause of the high concentrations responsible for the ice fog formation at relatively warm temperatures during the spring (Girard 1998; Girard and Blanchet 2001a,b).

In the ice fog simulations of Girard and Blanchet (2001a,b), as the temperature decreases and relative humidity increases and exceeds 100%, haze droplets form that are a mixture of deliquesced, hygroscopic aerosol particles and activated CCN. When the temperature falls below 238 K, the homogeneous freezing temperature of haze droplets is reached and unactivated haze droplets freeze, becoming ice crystals. The largest aerosol particles deliquesce first, activate as supercooled water droplets, freeze, and then grow rapidly as ice crystals. At the same time, ice supersaturation, which was high previously, decreases rapidly because of the sudden IN availability provided by the haze droplet freezing. The ice crystal number concentration (N_i) then increases above 1000 L^{-1} . The ice crystal mean diameter decreases dramatically to between 10 and 30 μm while the ice crystal mass concentration exceeds 0.01 gm^{-3} . These microphysical properties are typical of ice fog (Benson 1970; Bowling 1975; Curry et al. 1990). One of the major features seen in the time evolution of ice fog is weak variability of microphysical parameters. Although a large number of small ice crystals are formed, the ice crystal aggregation efficiency for 10- μm crystals is low and the sedimentation velocity is very small. Hence, ice fog evolves slowly and may persist many hours in the simulation, mimicking what is observed in real ice fog.

The radiative effect of ice fog is significant because of their long lifetimes. In the ice fog simulations of Girard and Blanchet (2001a,b) the mean downward infrared radiation flux at the surface increased by 7.4 W m^{-2} resulting in a reduction of the surface cooling rate of about 0.3 K day^{-1} . Hence, the ice fog formation kept the surface temperature at about 1 K warmer after three days of simulation.

5. Remaining challenges and recommended actions

Considerable progress has been made since the 1950s, when ice fog research was first initiated, particularly in instrumentation to measure the microphysical properties of these clouds and in the modeling their formation and evolution. There is still, however, a long way to go toward a full understanding of ice fog. The simulations of Girard and Blanchet (2001a,b) are encouraging in that they were able to reproduce, to a certain degree, the ice fog formation and lifetime, as well as the general

microphysical features that were observed; however, these were very limited case studies that focused only on a single region over a relatively short time span. Clearly, similar studies are needed over much broader temporal and spatial scales, validated with in situ and remote sensors. The results of modeling by Pu et al. (2016) underscore the difficulty in simulating ice fog in complex terrain. More sophisticated approaches like large-eddy simulations are needed to successfully model ice fog under these types of conditions.

In particular, forecast and climate models have to

- accurately predict the temperature and water vapor profiles;
- once more extensive data are available from field measurements, implement parameterizations of droplet and ice activation based upon the CCN and IN that are appropriate for a given regions;
- employ the correct optical properties of ice fog crystals to accurately forecast extinction coefficients, optical depths, and visibilities;
- correctly forecast ice fog lifetimes based on crystal fall velocities and associated meteorological variables like turbulence and entrainment of dry air;
- incorporate ice fog in forecasts of regional and global climate change.

There are a number of knowledge gaps that remain:

- What are the regional sources of CCN and IN?
- What triggers the transition from supercooled liquid droplets to ice crystals?
- How do surface conditions, including topography and plants, impact the formation and evolution of ice fog?

As has been underscored a number of times in this chapter, there is a paucity of observations in ice fog, largely because of the lack of field projects that target this type of cloud. As a result, the primary recommendations we can make to rectify this dearth of information are the following:

- A detailed white paper should be written by a panel of experts with experience in a wide range of cloud dynamics and microphysics, not only focused on ice fog. This white paper would guide the development of a document that would expand on what has been summarized in this chapter to create a long range plan that would lay out the observational and theoretical studies that are needed to bridge the knowledge gaps in ice fog formation and evolution. In addition, because of the frequent labeling of diamond dust as ice fog, a consensus should be reached, if possible or necessary, as to what differentiates these two types of surface ice clouds.
- An extensive modeling effort, including large-eddy simulation (LES) models, is needed for better prediction

of ice fog formation and evolution with high-resolution simulations. Then, the new hypotheses can be tested with well-designed field programs. This requires dedicated field projects to be launched to study ice fog in the regions of the world where it is most frequently observed.

- The most current, state-of-the-art instruments need to be employed, particularly to measure the size, shape, and optical properties of ice crystals smaller than 50 μm .

Many challenges face us as we strive to better understand the complex processes that lead to ice fog and its role in modulating the balance of water vapor and radiation in the atmosphere. Yet there are many opportunities to study these processes in greater detail while expanding our general understanding of ice in clouds.

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