Process-Oriented Analysis of Environmental Conditions Associated with Precipitation Fog Events in the New York City Region

ROBERT TARDIF
Research Applications Laboratory, National Center for Atmospheric Research, and Department of Atmospheric and Oceanic Sciences, University of Colorado, Boulder, Colorado

ROY M. RASMUSSEN
Research Applications Laboratory, National Center for Atmospheric Research, Boulder, Colorado

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ABSTRACT

An analysis of the environmental conditions associated with precipitation fog events is presented using 20 yr of historical observations taken in a region centered on New York, New York. The objective is to determine the preferred weather scenarios and identify physical processes influencing the formation of fog during precipitation. Salient synoptic-scale features are identified using NCEP–NCAR reanalyses. Local environmental parameters, such as wind speed and direction, temperature, and humidity, are analyzed using surface observations, while the vertical structure of the lower atmosphere is examined using available rawinsonde data. The analysis reveals that precipitation fog mostly occurs as a result of the gradual lowering of cloud bases as continuous light rain or light drizzle is observed. Such scenarios occur under various synoptic weather patterns in areas characterized by large-scale uplift, differential temperature advection, and positive moisture advection. Precipitation fog onset typically occurs with winds from the northeast at inland locations and onshore flow at coastal locations, with flows from the south to southwest aloft. A majority of the cases showed the presence of a sharp low-level temperature inversion resulting from differential temperature advection or through the interaction of warm air flowing over a cold surface in onshore flow conditions. This suggests a common scenario of fog formation under moistening conditions resulting from precipitation evaporating into colder air near the surface. A smaller number of events formed with cooling of the near-saturated or saturated air. Evidence is also presented of the possible role of shear-induced turbulent mixing in the production of supersaturation and fog formation during precipitation.

1. Introduction

The present study further develops the characterization of fog events affecting the coastal urban region of New York, New York, in the northeastern United States presented by Tardif and Rasmussen (2007, hereinafter TR07). In TR07, a description of the overall character of fog was presented, focusing on the identification of the various fog types observed in a region centered on New York City (NYC; Fig. 1). Their study concluded that fog in this region is composed of several types, each with distinct features in their spatial and temporal variability. The most frequent type was shown to be fog formation during precipitation (e.g., precipitation fog).

The emphasis of the present study is to perform an analysis of environmental conditions aimed at identifying the processes influencing the occurrence of precipitation fog in the NYC region. Despite the common occurrence of this fog type, relatively few studies describe the conditions leading to its formation. Previous investigations have focused on relatively few fog regimes, namely, radiation fog (e.g., Taylor 1917; Roach et al. 1976; Choularton et al. 1981; Mason 1982; Turton and Brown 1987; Fitzjarrald and Lala 1989; Bergot and Guédalia 1994) and sea (advection) fog (e.g., Taylor 1917; Findlater et al. 1989; Cho et al. 2000; Lewis et al. 2004). From a geographical standpoint, past studies on fog in the United States have focused primarily on the coastal areas of California (Palmer 1917; Petterssen...
1938; Goodman 1977; Noonkester 1979; Pilié et al. 1979; Leipper 1994; Koračin et al. 2001; Lewis et al. 2003), as well as on California’s central valley (Holets and Swanson 1981; Collett et al. 1999; Hoag et al. 1999; Underwood et al. 2004). Other studies characterized fog occurrence in the U.S. Midwest (Friedlein 2004; Westcott 2007) and along the coast of the Gulf of Mexico (Croft et al. 1997).

Coastal regions of the eastern part of the country have not been studied as extensively as the West Coast. Taylor (1917) described conditions under which sea fog formation occurs well offshore of the North American eastern seaboard. Woodcock (1978, 1984) and Woodcock et al. (1981) described conditions associated with fog formation at three coastal locations in southeastern Massachusetts and Rhode Island, while Pilié et al. (1975), Fitzjarrald and Lala (1989), and Meyer and Lala (1990) studied the life cycle of valley fogs at two locations in upper New York State.

Among the most relevant work to the present study, George (1940) found that fog forming in precipitation in association with warm fronts represents a common scenario for locations in the eastern and southern United States. Byers (1959) stated that warm front precipitation fog in the northeastern United States is among the most frequent in the world, because of temperature contrasts in the lower troposphere associated with the presence of polar air masses in combination with the nearby tropical maritime air over the Gulf Stream. However, detailed investigations of the mechanisms leading to the onset of precipitation fog were not presented by these authors.

The objective of the present study is to characterize the phenomenon of precipitation fog in the NYC region. A significant challenge resides in having to consider processes that are difficult to observe. Our approach consists of inferring the role of possible fog-inducing processes by their association with environmental conditions typically linked with each process. The frequency at which conditions were observed prior to and at fog onset are examined to assess the role of various processes. This is performed using a 20-yr historical dataset comprised of surface observations, upper-air soundings, and model reanalyses. Key climatological parameters associated with the formation of precipitation fog in the NYC region are identified.
2. Physical processes

Precipitation fog is generally described as a phenomenon taking place in areas either ahead of warm fronts (George 1940; Byers 1959; Petterssen 1969), behind cold fronts, in association with stationary fronts (UCAR/COMET 2007), or in regions of extratropical cyclones characterized by a transition in precipitation type (Stewart 1992; Stewart and Yiu 1993; Stewart et al. 1995). Goldman (1951) listed factors thought to be linked to the lowering of cloud ceilings in continuous precipitation and proposed empirical rules to forecast the rate of the lowering of cloud bases. The factors cited were the rate of moistening of the subcloud layer by the evaporation of raindrops, mechanical turbulence, and the advection of temperature and moisture. However, no specific investigation of the role of each factor was provided.

Physical processes leading to supersaturation and the formation of fog during precipitation possibly include the following:

- evaporation of hydrometeors that are warmer or much colder than the saturated ambient air [hydrometeors may be melting or freezing particles at 0°C, and thus possibly becoming much colder or warmer than the environment, respectively (Donaldson and Stewart 1993); fog formation from the raindrops that remain warmer than the air when falling in the colder layers of inversions is discussed by Petterssen (1969)],
- adiabatic cooling of saturated near-surface air parcels by orographic lifting,
- diabatic cooling associated with the advection of saturated (from precipitation evaporation) warm air over a cooler surface [warm, saturated air may be flowing against a gradient in the sea surface temperature (SST) in coastal areas, or over a snow-covered surface],
- local-scale turbulent mixing of near-saturated or saturated air parcels with contrasting temperatures (Rodhe 1962; Gerber 1991; Korolev and Isaac 2000), and
- radiational cooling of the near-surface air over a wet land surface (significant radiational cooling is not typically observed under deeper precipitating clouds, but may take place when skies rapidly clear after a precipitation event).

The mechanisms contributing to fog formation are thus varied. Some involve moistening of the lower atmosphere, while others involve cooling. While varied, contributing factors may be classified into distinct categories, namely, synoptic-scale influences (large-scale uplift), interactions between the low-level flow and the surface (adiabatic cooling from upslope flows, flow over colder surface, surface layer turbulent mixing), and microphysical mechanisms (evaporation of melting–freezing particles or warm raindrops into colder air). The work presented herein seeks to identify the most common processes contributing to the formation of fog with respect to liquid vapor pressures because fog is mostly composed of water droplets in the region (TR07).

3. Datasets and methodology

a. Datasets

The characterization of the local conditions leading to precipitation fog formation is performed with the dataset used by TR07. Hourly surface observations of visibility, temperature, dewpoint temperature, wind speed and direction, ceiling height, cloud cover, coded obstruction to vision, and precipitation type and intensity were taken from the National Oceanic and Atmospheric Administration’s (NOAA’s) Techniques Development Laboratory surface hourly observations archived at the National Center for Atmospheric Research (2007). Seventeen stations in a region centered on New York City were chosen (Fig. 1) on the basis of their location with respect to important physiographic features and data availability (see TR07). The period from 1977 to 1996 inclusively is considered to ensure homogeneity in present weather-reporting practices (human-made observations, before the widespread introduction of automated weather stations) and maximize the number of reporting stations. A well-known deficiency of the early generation Automated Surface Observing System (ASOS) was its inability to detect very light precipitation as described by Sears-Collins et al. (2006). Human-made present weather observations, although subjective in nature, are believed to provide better sampling of conditions in the vicinity of weather stations.

Hourly surface observations from a buoy located in the coastal waters south of Long Island, New York, are used to characterize the near-surface offshore conditions during onshore flow. Observations from the Ambrose Light (ALSN6) buoy are used. ALSN6 is located 22 km south of the John F. Kennedy International Airport (JFK) and 15 km east of Sandy Hook, New Jersey (Fig. 1).

The identification of synoptic weather patterns associated with fog formation is performed using analyses of mean sea level pressure (MSLP) from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalyses (Kalnay et al. 1996; Kistler et al. 2001). These were
performed using a global spectral model using 28 vertical levels and a horizontal resolution of approximately 210 km. Therefore, only the large-scale features are highlighted.

In addition, atmospheric profiles are used to characterize the vertical structure of the lower atmosphere within 3 h of fog onset (time of fog formation). Issues of temporal versus vertical resolutions were considered in choosing a dataset. Balloon soundings provide observations with a higher vertical resolution (provided the dataset includes mandatory and significant levels), but with a low temporal resolution (twice daily). NCEP–NCAR reanalyses are available 4 times per day, but lack sufficient vertical resolution to represent shallow temperature inversions. Given the importance of boundary layer stratification in the context of the present study, vertical resolution is privileged over temporal resolution. Historical soundings from the NOAA Integrated Global Radiosonde Archive (Durre et al. 2006) and from the NOAA/Forecast Systems Laboratory (now the Earth System Research Laboratory/Global Systems Division) North American radiosonde data archive (Schwartz and Govett 1992) are used. Soundings in the region were performed at different locations over the time period considered. Soundings from Fort Totten (JFK-FT; 1977–80), and Upton (OKX; from 1994 onward), New York, are paired with fog events at LaGuardia Airport (LGA), JFK, and ISP. The sounding at Fort Totten was located 11 km northeast of LGA, 15 km to the north of JFK, and 56 km west of ISP. The OKX sounding site is located 22 km away from ISP and 81 km to the west of JFK on Long Island. Soundings performed at Atlantic City (ACY), New Jersey, from 1980 to 1994, are paired with fog events at ACY, Millville (MIV), New Jersey, and McGuire Air Force Base (WRI), New Jersey. MIV and WRI are located 44 km to the southwest and 64 km to the north of ACY, respectively. The locations of the sounding sites are shown in Fig. 1. Because the soundings were launched from Atlantic City for 14 yr during the period considered, results are biased toward the conditions observed in the coastal plain of New Jersey. However, noticeable differences in salient features defining the vertical structure at fog onset between the ACY soundings and those at JFK and OKX were not found, suggesting the absence of influence of microclimate variations on the results.

Finally, the satellite-derived Northern Hemisphere Equal-Area Scalable Earth (EASE)-Grid weekly snow cover and sea ice extent dataset from the National Snow and Ice Data Center (Armstrong and Brodzik 2005) is used to assess surface snow conditions. Binary values indicating the presence–absence of snow with a 25-km resolution are used to establish whether a relationship exists between the presence of snow on the ground and the occurrence of fog. This dataset was found to be the most comprehensive source of information on surface conditions. The sparseness of in situ measurements of surface and ground conditions prohibits any conclusive assessment of the relationship between surface snow and fog occurrences in the region.

b. Methodology

The analysis of conditions characterizing precipitation fog relies on the identification of events. Fog events are identified using the concept of “M of N” constructs (Setiono et al. 2005), where M represents the number of hours with fog within a number of N consecutive hourly observations. Values of M = 3 and N = 5 are used (see TR07 for more details). Taking into account the concerns of the aviation community, fog is herein defined as saturated conditions with horizontal visibilities below 1.6 km (1 statute mile), corresponding to the threshold defining the low-instrument flight rules used in the United States (NWS Aviation Services Branch 2004). A precipitation fog event is identified when any type of precipitation was reported at the time of fog onset or the hour prior. Using observations taken 1 h prior to the detected onset enhances the identification of precipitation fog formation during the period between reporting times. However, events forming as a result of rapid cloud clearing and associated radiative cooling of the moistened air may be counted as a precipitation fog event. A careful review of the data revealed that a small fraction of precipitation fog events (9%) identified initially was characterized by precipitation ending at fog onset, cooling conditions, and weak wind. Because the main interest of this investigation is on fog formation during precipitation from influences other than the well-know radiative cooling mechanism, those were removed from the subsequent analysis.

4. Morphology of precipitation fog formation

a. Precipitation versus fog

To characterize the relationship between precipitation and fog, precipitation events are identified using the same methodology as for fog (e.g., presence of any type of precipitation on three out of five consecutive hourly observations), and the proportion characterized by the presence of foggy conditions is compiled. The presence of fog is diagnosed when at least one hourly observation reports fog while visibility is below 1 km (dense fog) in light to moderate liquid and/or freezing precipitation or below thresholds proposed by Rasmus-
sen et al. (1999) in snow. The threshold used in heavy rain and freezing precipitation cases is 800 m (1/2 mile). The various thresholds are used to take into account the varying contributions of various precipitation types and intensities to visibility reduction. The proportion of precipitation events leading to foggy conditions ranges from 11% to 39%, depending on the location (see Table 1), with an average of 18%. Also, 54% of the precipitation events that did not lead to fog had a maximum relative humidity during precipitation (RH_{max}) below 95%, and 46% were characterized by a RH_{max} greater than 95%. Therefore, a relatively limited fraction of precipitation events lead to foggy conditions. Conditions were simply too dry for nearly half of the nonfoggy events because precipitation evaporation was not sufficient to create saturation. The remaining nonfoggy precipitation events were characterized by conditions near or at saturation. This suggests that a high level of relative humidity is a poor indicator of the potential for fog formation as a result of precipitation evaporation. Specific conditions are required in order for supersaturation and fog to appear.

The total number of precipitation fog events identified in the dataset is shown in Fig. 2 for each station. The locations with the highest overall number of events are White Plains (HPN; 321 events), New York, ISP (202 events), MIV (195 events), and JFK (154 events). A smaller number of events occurred at locations closest to the New York City urban center such as Newark Liberty International Airport (EWR; 55 events), New Jersey, Teterboro (TEB; 63 events), New Jersey, and LGA (65 events). Noticeable contrasts exist in the likelihood of precipitation fog in the area, suggesting the role of local influences in enhancing or mitigating precipitation fog formation.

The morphology of precipitation fog formation is described next by examining the evolution of visibility, cloud ceiling height, temperature and humidity tendencies, and precipitation during the few hours prior to fog formation.

b. Visibility and cloud ceiling

A characteristic length of time required for fog to appear after the start of precipitation has been determined to be of the order of a few hours. The median of the distribution (not shown) is 6 h. The distributions of observed visibility in the 6 h leading to fog onset shows that visibility reduction was already taking place prior to foggy conditions for at least 75% of the events. A reduction in the variance during the transition toward fog onset is clearly identified, with values commonly reported in the 2–5-km range (Fig. 3a). Such values are typical of visibility impairments because of the presence of light to moderate liquid precipitation. A tendency for visibility to gradually decrease over time is also suggested because the clear majority of events were characterized by successive decreasing visibility trends equal to or smaller than 1 km h\(^{-1}\) during the few hours leading to fog (Fig. 4). Such behavior is in contrast with cases of advection fog as described in TR07. The response of hygroscopic aerosols to the increasing ambient humidity resulting from moistening by precipitation is likely a contributing factor in the gradual reduction in visibility.

A large proportion of events were associated with the presence of cloud ceilings below 1000 m in the hours

<table>
<thead>
<tr>
<th>Station identifier</th>
<th>Name and state</th>
<th>Elev (m)</th>
<th>Precipitation events</th>
<th>Events with fog</th>
<th>Fraction of events (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EWR</td>
<td>Newark/Liberty, NJ</td>
<td>7</td>
<td>1847</td>
<td>406</td>
<td>22.0</td>
</tr>
<tr>
<td>JFK</td>
<td>John F. Kennedy, NY</td>
<td>9</td>
<td>1865</td>
<td>368</td>
<td>19.7</td>
</tr>
<tr>
<td>LGA</td>
<td>LaGuardia, NY</td>
<td>11</td>
<td>1774</td>
<td>238</td>
<td>13.4</td>
</tr>
<tr>
<td>ISP</td>
<td>Islip/McArthur, NY</td>
<td>43</td>
<td>1942</td>
<td>220</td>
<td>11.3</td>
</tr>
<tr>
<td>TEB</td>
<td>Teterboro, NJ</td>
<td>7</td>
<td>1812</td>
<td>242</td>
<td>13.4</td>
</tr>
<tr>
<td>HPN</td>
<td>White Plains, NY</td>
<td>121</td>
<td>1773</td>
<td>683</td>
<td>38.5</td>
</tr>
<tr>
<td>POU</td>
<td>Poughkeepsie, NY</td>
<td>46</td>
<td>1707</td>
<td>228</td>
<td>13.6</td>
</tr>
<tr>
<td>BDR</td>
<td>Bridgeport, CT</td>
<td>7</td>
<td>1677</td>
<td>301</td>
<td>17.9</td>
</tr>
<tr>
<td>BDL</td>
<td>Hartford, CT</td>
<td>60</td>
<td>1953</td>
<td>319</td>
<td>16.3</td>
</tr>
<tr>
<td>PVD</td>
<td>Providence, RI</td>
<td>16</td>
<td>2038</td>
<td>378</td>
<td>18.4</td>
</tr>
<tr>
<td>ABE</td>
<td>Allentown, PA</td>
<td>114</td>
<td>1962</td>
<td>341</td>
<td>17.4</td>
</tr>
<tr>
<td>WRI</td>
<td>McGuire Air Force Base, NJ</td>
<td>41</td>
<td>1973</td>
<td>402</td>
<td>20.4</td>
</tr>
<tr>
<td>ACY</td>
<td>Atlantic City, NJ</td>
<td>28</td>
<td>1793</td>
<td>339</td>
<td>18.9</td>
</tr>
<tr>
<td>MIV</td>
<td>Millville, NJ</td>
<td>18</td>
<td>1754</td>
<td>484</td>
<td>27.6</td>
</tr>
<tr>
<td>PHL</td>
<td>Philadelphia, PA</td>
<td>23</td>
<td>1772</td>
<td>230</td>
<td>13.0</td>
</tr>
<tr>
<td>PNE</td>
<td>North Philadelphia, PA</td>
<td>23</td>
<td>1727</td>
<td>309</td>
<td>17.9</td>
</tr>
<tr>
<td>ILG</td>
<td>Wilmington, DE</td>
<td>28</td>
<td>1748</td>
<td>339</td>
<td>19.4</td>
</tr>
</tbody>
</table>
preceding fog (Fig. 3b). A clear tendency for lowering cloud bases during the transition to fog is observed with an associated reduction in the variance of cloud-base heights. Cloud-base lowering during the hour leading to fog onset occurred in more than 90% of the events, while the remaining events were characterized by cloud bases that were already close to the ground (below 100 m; see Fig. 4b). The decrease of cloud-base heights over time was generally gradual, with successive ceiling height variations of the order of $-100 \text{ m h}^{-1}$ (Fig. 4). The majority (75%) of time intervals with lowering ceilings occurred as precipitation was reported concurrently.

c. Temperature and humidity tendencies

Not surprisingly, near-surface relative humidity (RH) conditions during the few hours leading to onset were mostly characterized by high levels of RH, with a gradual increase toward values at or near saturation. One hour prior to onset, 29% of the events had conditions at saturation, and 33% had $95\% \leq \text{RH} < 100\%$. Therefore, the latter stages of the transition toward fog generally took place under subtle increases in RH toward supersaturation while a greater jump in RH was required to reach saturation in fewer instances.

Observed changes in specific humidity and saturation specific humidity during the hour prior to onset are examined. Changes in saturation specific humidity reflect changes in temperature. Events are divided among tendencies of 1) moistening conditions, 2) cooling, 3) moistening and cooling, and 4) constant conditions. Moistening occurs when specific humidity and temperature both increase, with the increase in water vapor equaling or dominating the temperature changes with respect to variations in RH. In this case, the addition of moisture by evaporating precipitation particles in the warming air is the most likely mechanism leading to the appearance of fog. Moistening and cooling define events characterized by a decrease in temperature along with an increase in specific humidity. Events with constant conditions are those for which a perceptible change in near-surface moisture and temperature was not observed, within the uncertainty of the measurements. The reader is referred to Table 2 for more de-
tailed definitions. Moistening occurred for 42% of the events, 25% had cooling, and 23% began under the absence of variations in temperature and moisture (Table 2); only 10% were associated with concurrent moistening and cooling. The events characterized by the absence of variations were those for which the relative humidity was already at saturation the hour prior to fog onset. It is hypothesized that the supersaturation and nucleation of fog droplets occur from the slight cooling or addition of water vapor after saturation had been achieved. Because routine instruments do not have the capability to measure supersaturation, observations indicated a dewpoint temperature remaining equal to the ambient temperature as this process occurred. Because of instrument limitations, observed conditions seem to have remained at saturation.

d. Precipitation character

The frequency of the various precipitation types and intensities (in categories defined as light, moderate, and heavy) associated with the onset of precipitation fog events is compiled from the 17 stations. The clear majority of events are associated with light liquid precipitation, such as light drizzle (L−) and light rain (R−; Fig. 5). Even though light rain showers (Rw−) is the second-most-reported precipitation type overall, it is not associated with a significant number of fog events. This may reflect a lack of time for falling precipitation to create and sustain the foggy conditions resulting from the intermittent nature of rain showers.

The possible relationship between fog formation and a transition in precipitation character (type and intensity) is examined next. Frequencies of variations in precipitation character during the hour leading to fog onset indicate that the majority of cases (55%) are characterized by the absence of any transition in precipitation type and intensity, at least in a categorical sense (Fig. 6). Relatively common scenarios are characterized by fog onset occurring at the very end of a precipitation event or during a break in precipitation (17%), or at the start of precipitation (15%). Of all of the events characterized by the absence of a transition in precipitation character, 62% are also characterized by an absence of transition during the period between 2 and 1 h prior to fog onset. This suggests that the occurrence of continuous liquid precipitation is an enhancing factor in the likelihood of fog formation in the NYC region. The preferred scenario leading to precipitation fog does not involve transitions in precipitation type, as mentioned by Stewart (1992), Stewart and Yiu (1993), and Stewart et al. (1995).

5. Environmental conditions and physical processes

Various environmental parameters are investigated with the objective of identifying the physical processes influencing fog formation.

a. Synoptic-scale influences

The respective roles of local-scale mechanisms are in many ways conditioned by the large-scale environment through flow patterns, temperature, and humidity variability and associated advections. Characterizing the large-scale weather patterns is a first step toward establishing the character of the environmental conditions leading to fog formation during precipitation.

Fig. 3. Distributions of (a) visibility and (b) cloud ceiling height observed during the 6 h preceding the onset of fog and at fog onset. Distributions are illustrated using box plots, defined by the 5th percentile (lower whisker), 25th percentile (lower edge of box), median (horizontal line within box), 75th percentile (upper edge of box), and 95th percentile (upper whisker). Open circles represent outliers.
1) Weather Patterns

The synoptic weather patterns at the onset of fog events were identified through a subjective classification of the MSLP analyses from the NCEP–NCAR re-analyses. The synoptic-scale MSLP patterns corresponding to fog events at the EWR (urban environment), JFK (coastal), Allentown (ABE), Pennsylvania (inland), and MIV (coastal plain of New Jersey) are examined in order to gain a comprehensive view of the

![Graphs showing distributions of observed hourly changes in visibility and cloud ceiling height during the hour leading to fog onset, 2 h before onset, and 3 h before fog onset.](image)

**Table 2. Distribution of precipitation fog events according to tendencies in specific humidity ($\Delta q$) and saturation specific humidity ($\Delta q_{sat}$) observed in the hour leading to fog onset.**

<table>
<thead>
<tr>
<th>Tendencies</th>
<th>Conditions</th>
<th>No.</th>
<th>Frequency (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moistening</td>
<td>$\Delta q &gt; 0$, $\Delta q \geq \Delta q_{sat}$</td>
<td>555</td>
<td>41.7</td>
</tr>
<tr>
<td>Moistening and</td>
<td>$\Delta q &gt; 0$, $\Delta q_{sat} &lt; 0$</td>
<td>136</td>
<td>10.2</td>
</tr>
<tr>
<td>Cooling</td>
<td>$\Delta q_{sat} &lt; 0$, $\Delta q \leq \Delta q_{sat}$</td>
<td>335</td>
<td>25.2</td>
</tr>
<tr>
<td>No trends</td>
<td>$\Delta q = \Delta q_{sat} = 0$</td>
<td>305</td>
<td>22.9</td>
</tr>
</tbody>
</table>
FIG. 5. Histogram showing the frequency of various precipitation types and intensity associated with the onset of fog events—L: drizzle, R: rain, Rw: rain showers, Zl: freezing drizzle, Zr: freezing rain, Mx: combination of liquid and frozen precipitation, IP: ice pellets, S: snow, and Sw: snow showers; − and + denote light and heavy precipitation, respectively. The overall frequency of precipitation per type is also shown as a reference (dots). The data cover the 1977–96 period.

FIG. 6. Histogram showing the frequency of observed transitions in precipitation characteristics (type and intensity) corresponding to the onset of fog conditions for fog events identified at the 17 stations for the period from 1977 to 1996.
synoptic-scale features. This subset of stations was chosen as a compromise between the number of patterns examined and the number of events considered, while maintaining an adequate representation of the main physiographic features of the region.

Surface weather patterns were subjectively classified into categories inspired by the work of Meyer and Lala (1990), Westcott (2004), and Keim et al. (2005). Nine synoptic weather patterns were linked to the onset of precipitation fog events. These patterns are defined through the relative position of surface weather features (centers of high or low pressure systems, well-defined frontal boundaries, or less defined troughs and ridges) with respect to the location of target stations. The patterns are listed in Table 3, along with their associated characteristics. Patterns are also illustrated in Fig. 7 using individual representative cases.

The analyses corresponding to the time closest to fog onset were examined for all precipitation fog events. The number of events associated with each synoptic weather pattern is shown in Table 3. One of the dominant scenarios is the pre-warm front pattern with a little over 17% of events (81 out of 477). This is consistent with previously published work, which points out that the region ahead of warm fronts is prone to fog formation with warm raindrops falling into the colder air near the surface (George 1940; Byers 1959; Petterssen 1969). The other leading synoptic weather scenarios are the weak trough pattern with 79 events (17%) and the warm sector of a midlatitude low pressure system (74 events, or 16%). Fog formation in precipitation as a cold front is approaching also represents a common scenario (59 events, 12%), as well as when a low pressure system is centered over the region (48 events, 10%). Fogs within a ridge (8%) or col (7%) in the MSLP pattern are less frequent scenarios. A similar finding characterizes the likelihood of fog forming in the northeast sector of a low pressure system. The same can be said for fog forming under return flow conditions from the Atlantic on the western edge of an anticyclone (6%).

2) Vertical motion

The onset of precipitation fog is associated with a wide variety of large-scale patterns. The common feature for most of these patterns is that fog occurred in regions characterized by large-scale uplift throughout the lower troposphere. The values of lower-tropospheric vertical motion ($\omega; \text{Pa s}^{-1}$) extracted from the NCEP–NCAR analyses at the nearest grid point to the surface stations where fog onset occurred reveal distributions that are mostly characterized by negative va-
ues (upward motion) of the order of a few tenths of a pascal per second (Fig. 8). A Student’s t-test has shown that these distributions are different than the corresponding vertical motion climatology for the same grid point at the 95% confidence level. The relatively fewer events characterized by downward motion \((\omega > 0)\) were associated with the ridge and Atlantic return patterns with light drizzle. This suggests that fog onset in such scenarios is associated with drizzling advection fog or other boundary layer clouds below a subsidence inversion, with different support mechanisms related to mesoscale features and interactions.

3) **Horizontal advection**

Large-scale horizontal (isobaric) advections of temperature and specific humidity were calculated from the NCEP–NCAR reanalyses gridded data corresponding to the time closest to fog onset. Values at grid points nearest to surface stations where fog occurred were used. The values in the lower troposphere (results at 925 hPa are shown in Fig. 9) indicate that precipitation fog generally formed in environments characterized by large-scale positive (warm air) advection and positive humidity advection. A greater proportion of positive values compared to the overall advection climatology are evidenced by the positive skewness \(S_k\) of the distributions corresponding to fog \(S_k = 0.57\) for temperature and \(S_k = 0.64\) for moisture) compared to the negative values of the climatological distributions \(S_k = -1.03\) and \(S_k = -0.36\). The local surface flow conditions at fog onset show that the majority of events at coastal locations [ISP, JFK, LGA, Bridgeport (BDR), Connecticut, Providence (PVD), Rhode Island, and ACY] began while an onshore flow occurred (Fig. 10). Evidence of the effect of onshore flow is also seen in the wind distribution at EWR, MIV, and Wilmington (ILG), Delaware, with maxima in the frequency distribution corresponding to low-level winds from Raritan Bay for EWR and from Delaware Bay for MIV and ILG. Inland locations generally experienced precipitation fog formation during northeasterly flows, a flow aligned parallel with nearby regions of higher elevation in some instances. In terms of wind speed, a difference exists between inland and coastal locations, with a greater proportion of events beginning in low wind speeds at inland stations and when wind speeds were in excess of 6 m s\(^{-1}\) at coastal locations (Fig. 11). This reflects the overall climatology, and indicates that advection processes at the surface were less influential inland.

Events with a lack of surface wind, mostly with the weak trough, low center, or col weather patterns, were associated with the absence of temperature advection.
at the surface. Events at inland locations with a north-easterly flow (with the pre-warm front, northeast sector, or ridge weather patterns) were characterized by cold air advection, while southerly or southeasterly flows (with the pre-cold front, warm sector, or Atlantic return flow patterns) led to surface warm air advection (Fig. 12a). With conditions aloft typically characterized by warm air advection (Fig. 9a), cases with no surface wind or cold air advection at the surface were influenced by positive differential temperature advection. In such instances, large-scale flow conditions contribute to increasing the static stability of the lower troposphere. In fact, the distribution of differential advection between the surface and 925 hPa shows a proportion of 71% of positive values (Fig. 12b). In comparison, the climatological distribution is well approximated by a narrow normal distribution (not shown) with a zero mean advective temperature tendency. Such results suggest that scenarios with warm air advection at the surface and aloft were generally characterized by stronger advection aloft. Inversely, advective tendencies were generally strongest at the surface among the few cases with cold air advection both aloft and at the surface, also increasing the static stability.

The large-scale environment associated with precipitation fog is characterized by a wide variety of surface weather patterns, but with the majority of events associated with large-scale uplift along with warm and moist air horizontal advection aloft. The combination of tropospheric upward motion and the presence of moist air

**Fig. 8.** Frequency distribution of vertical motion (in pressure coordinates) in the NCEP-NCAR reanalyses at (a) 925, (b) 850, and (c) 700 hPa at the nearest grid point from the surface stations with fog. Analyses corresponding to time nearest to fog onset were used. The climatology represents the distribution of values extracted in the analyses at the same grid points for the entire dataset (1977–96).
merely reflect the fact that such factors contribute to the formation of clouds and widespread stratiform precipitation in the area. An important factor is the differential temperature advection contributing to the presence of inversions in the lower troposphere. Such conditions are typical in regions ahead of synoptic-scale low pressure systems (Bluestein and Banacos 2002). The presence of inversions can enhance the role played by microphysical processes involved in the evaporation of precipitation and the formation of extreme low cloud ceilings and fog.

b. Microphysical processes

As stated earlier, scenarios involving warm raindrops or freezing hydrometeors evaporating in the colder air of inversions are believed to be conducive to the formation of fog. Their continuous evaporation even in saturated environments may lead to supersaturation and fog formation. The association between the presence of temperature inversion and the occurrence of precipitation fog is investigated by comparing the frequencies of inversions found in soundings with fog and those corresponding to precipitation events that did not lead to fog. This association is not expected to be perfect because the temporal and spatial sampling of conditions associated with fog from limited sites and infrequent soundings is limited. However, more frequent occurrences of fog in association with precipitation falling in inversions in comparison with occurrences of precipitation without fog provide evidence of the influence of microphysical processes involved in the evaporation of hydrometeors in cold, saturated air.

Atmospheric profiles are examined using historical soundings from JFK-FT, OKX, and ACY launched within 3 h of fog onset at nearby surface stations. The number of valid soundings with fog is 153. The dataset is divided among events characterized by moistening, cooling, and those without perceptible trends in the surface temperature and humidity in the hour leading to onset. An examination of the humidity profiles has revealed that the atmosphere is typically saturated over a deep layer at fog onset, with an average depth of 2300 m. Only on few occasions (16%) was a dry atmosphere found above a shallow (~1000 m) saturated layer at the surface, suggesting the possible influence of drying aloft and associated low-level radiative cooling in fog formation.

A low-level inversion is defined as a layer with a temperature increasing with height found in the lowest 1 km of the atmosphere. Inversions were found in 96% of the soundings associated with the moistening cases and in 93% of the soundings in events without trends, while a proportion of 76% was found for events with cooling (Fig. 13). Inversions with lapse rates greater than +10°C km⁻¹ were found in 60% of the soundings representing moistening cases, in 44% of soundings corresponding to events without trends, and in 28% of the cooling cases. In comparison, inversions were found in 80% of precipitation cases without fog, with only 32% of those with strong stratification (lapse rates greater than +10°C km⁻¹). Statistics were compiled on
the height of inversion base and depth of the layer. Data from soundings associated with moistening events are compared with cases of precipitation without fog. Inversions were mostly found in the lowest 300 m in fog cases, with 32% being surface-based inversions and 46% having a base between 100 and 300 m (Fig. 14a). On the other hand, cases without fog had a much smaller proportion of inversions with bases in the 100–300-m range. More than 50% of inversions with fog had depths between 100 and 300 m (Fig. 14b). The distribution corresponding to cases without fog show a greater proportion of very shallow inversions (depths smaller than 50 m) and a smaller proportion of inversions with depths between 100 and 300 m.

The rare occurrence of dry elevated layers above saturated boundary layers (potential for radiative cooling), combined with the higher frequency of strong saturated inversions in association with precipitation fog, compared to cases without fog, suggest the important role played by precipitation falling in an atmo-
Fig. 10. (Continued) (b) As in (a), but for stations located in the general vicinity of New York City.

Fig. 11. Histograms showing the frequency distribution of wind speed at the time of precipitation fog onset for inland and coastal stations located in the northeastern United States.
sphere with vertical temperature contrasts (a cool layer superimposed by a warmer air) in fog formation. The likelihood of fog is enhanced when an inversion is strong enough (lapse rates $> +10^\circ C \text{ km}^{-1}$), low enough (below 300 m), and deep enough (greater than 100 m in depth). The presence of strong and deep inversions maximize the potential for the hydrometeors falling through to remain warmer than the ambient air, and maintain their evaporation even in saturated air leading to supersaturated conditions. Beyond fog onset, a clear correlation could not be established between inversion characteristics and the severity in visibility reduction during the ensuing fog. This is partly due to the limited number of data and to the influence of other factors not considered here, such as the conditions of ambient aerosols and other mechanisms that can influence the subsequent evolution of relative humidity, such as turbulent mixing.

The examination of precipitation type at fog onset has shown that freezing precipitation (freezing rain, freezing drizzle, or ice pellets) is not a major influence for precipitation fog in the NYC region. This is confirmed by the small fraction (6%) of soundings with a structure characterized by a transition from above-freezing temperatures aloft to below-freezing conditions near the surface. The possible contribution of melting precipitation particles is assessed using surface air temperature observations and temperature profiles to determine the height of the 0°C level and the presence of melting layers (isothermal layers with a temperature near 0°C). Fujibe (2001) describes a maximum near 0°C in the surface air temperature frequency distribution during precipitation and argues that it is a consequence of the cooling from the melting of falling snowflakes. The frequency distribution of the near-surface temperature at the onset of precipitation fog shows a clear maximum near 10°C, while temperatures near 0°C were observed on 13% of the events (Fig. 15). Studies have shown that the maximum depth over which melting snow is expected to penetrate in an environment with above-freezing temperatures is typically of the order of 200–600 m, depending on the size of the flakes and the environmental conditions (Lumb 1961, 1963; Mitra et al. 1990). Therefore, in order for melting particles to play a role in fog formation, the vertical structure of the atmosphere should be characterized by a low enough melting level. The height at which a transition from below- to above-freezing temperatures occurred was identified in the soundings corresponding to the onset of precipitation fog. The frequency distribution reveals that most fog events occurred because the melting level was located much higher than 1 km (Fig. 16). These results suggest that the presence of melting particles was unlikely in the lowest atmosphere at the onset of fog. The majority of precipitation fog events in the NYC region were therefore associated with liquid (rain) drops.

c. Surface interactions

As discussed in section 2, processes in addition to precipitation evaporation should be considered. Those involve air–surface interactions, either from the point

![Fig. 12. Frequency distribution of (a) horizontal temperature advection at the surface corresponding to surface flows from the northeast–east and from the south–southeast, and (b) the difference in horizontal temperature advection between the surface and 925 hPa. The climatology in differential temperature advection represents the distribution of values extracted in the analyses for the entire dataset (1977–96).](image-url)
of view of energy fluxes and/or the interaction of the flow with surface elements. Fog formation may occur as a result of warm air, moistened by the evaporation of precipitation, flowing over a colder surface; adiabatic cooling when the air is forced upward by topography; or the occurrence of turbulent mixing of parcels characterized by contrasting temperatures.

1) SURFACE–AIR TEMPERATURE CONTRASTS

As was previously shown, some precipitation fog events occurred at coastal locations with surface southerly flows with warm air advection aloft in association with southerly to southwesterly winds (Fig. 17). Events at JFK are a good representation of this scenario (see Fig. 10). These events were mostly characterized by low-level temperature inversions. However, because flows at the surface and aloft both had a southerly component, the lack of directional wind shear can lead to weaker differential temperature advection because temperature gradients at low levels are typically aligned from north to south. Therefore, differential temperature advection, while still a contributing factor, is not the only mechanism responsible for the presence of low-level inversions. The climatological analysis of sea surface temperature from the Extended Reconstructed Sea Surface Temperature (ERSST) dataset (Smith and Reynolds 2004) (Fig. 18) reveals that air parcels flowing from the south to reach JFK are flowing from warm to cold SSTs over a relatively tight gradient. This is confirmed by observations from the Ambrose Light tower (ALSN6) at times corresponding to fog onset at JFK. The frequency distribution shows that fog appeared at JFK when the upstream air temperature was warmer than the SST by a few degrees (Fig. 19). This scenario is typically associated with the formation of a stably stratified internal boundary layer (Mahrt et al. 2004). Similar conditions have been observed in association with advection fog events in the New York City region (TR07), in other coastal regions of the United States (Croft et al. 1997), and other parts of the world (Findlater et al. 1989; Cho et al. 2000). Therefore, fog formation may have occurred under the combined influence of warm raindrops evaporating while falling in the inversion, and/or through cooling of the near-surface air (as in advection fog events) with an air mass moistened by precipitation evaporation upstream. The com-
Combination of these influences contributes to the greater number of events at coastal locations.

Another scenario with significant temperature contrasts between the surface and overlying air involves the presence of a melting snow layer on the ground. The falling rain on the snowpack may also lead to latent heat release as the water freezes, producing an upward diffusion of water vapor, further enhancing the potential for fog formation. Examination of the snow cover from the Northern Hemisphere EASE-Grid weekly snow cover and sea ice extent dataset corresponding to fog events associated with light rain and light drizzle revealed that only 5% of events were associated with the presence of snow on the ground. Therefore, scenarios involving snow cover are not common in the context of precipitation fog formation in the NYC region.

2) TERRAIN EFFECTS

Variations in terrain elevation are relatively small in the region under study but can still provide some influence on fog formation. The most important effect is perhaps the occurrence of upslope flow. The highest station in the area, HPN, is located 121 m above sea level. Precipitation fog events are most frequent at that location (see TR07 and Fig. 2 in this paper), and onset typically occurs during onshore flow conditions (Fig.

**Fig. 14.** Frequency distribution of (a) height of the base and (b) depth of temperature inversions detected in radiosoundings corresponding to the onset of precipitation fog events. The frequency distributions corresponding to precipitation events without fog are also shown for reference.

**Fig. 15.** Frequency distribution of the observed near-surface air temperature at the onset of precipitation fog events.

**Fig. 16.** Frequency distribution of the height of the freezing level observed in soundings corresponding to the onset of precipitation fog events.
10) with trajectories over the water of the Long Island Sound and the lowlands of Long Island. Upon reaching the shore, the air is forced to rise by the terrain by about 10 m km$^{-1}$ as it flows toward HPN (TR07). Ceiling and visibility observations at upstream locations (e.g., ISP) at the time of fog onset at HPN show low cloud ceilings between 100 and 200 m on numerous occasions, with observed visibilities typically in the 1.6–4-km range (Fig. 20a). Similar results were obtained for BDR and LGA (not shown). The difference in terrain height between HPN and the upstream coastal stations is of the same order as most of the reported ceiling heights at the coastal locations. This suggests that foggy conditions at HPN may, in part, result from the interception of the low clouds by the local terrain. The temperature at fog onset at HPN is also generally colder than at ISP by a few degrees, indicating that adiabatic cooling of the moist air associated with the upslope flow from the coast to the station could have contributed to the appearance and/or maintenance of foggy conditions (Fig. 20b). The stationary forcing from the terrain effect described here is likely a factor in explaining the significant positive anomaly in the number of fog events at HPN compared to other locations in the area.

3) TURBULENT MIXING

Another factor to consider is the role of low-level turbulent mixing in the generation of supersaturation and fog in the saturated air underneath precipitating clouds. The detailed description of such mechanisms is difficult to achieve without the use of extensive field observations. However, an assessment of turbulent mixing as a possible contributing mechanism is presented using radiosonde observations taken close to fog onset. The likelihood of turbulent mixing is assessed through the bulk Richardson number:

$$R_i = \frac{g(\Delta T + \Gamma_m)\Delta z}{T[(\Delta u)^2 + (\Delta v)^2]}.$$  

(1)

Fig. 17. Frequency distribution of the wind direction at the surface at JFK and ISP (combined), vs wind direction at heights corresponding to the top of temperature inversions in the lower troposphere as observed in soundings at OKX corresponding to precipitation fog events.

Fig. 18. Climatology of the SST in the western Atlantic over the period from 1977 to 1996.

Fig. 19. Frequency distribution of the difference in air temperature and SST observed at the ALSN6 buoy at times corresponding to precipitation fog onset at JFK during onshore flow conditions. The climatological distribution is also shown for reference.
where $\Gamma_m$ is the saturated adiabatic lapse rate and $\Delta$ denotes the finite difference in observed quantities (temperature $T$, $u$, and $v$ wind components) between the top and bottom of the lowest layer in soundings (between surface observations and next reported level). The well-known critical Richardson number ($R_{ic}$) value of 0.25 applies only for local gradients and not for finite differences across thick layers. A customary practice consists of using a larger critical Richardson number, which gives reasonable results from smoothed gradients calculated over the deeper layers. McNider and Pielke (1981) proposed a formulation including the layer depth ($D$),

$$R_{ic} = 0.115D^{0.175}. \tag{2}$$

Applying (1) and (2) on the sounding data, dynamically unstable conditions were found in 65% of the soundings. The distributions of wind speed at the surface are shown for turbulent and nonturbulent surface layers in Fig. 21. Turbulent mixing becomes more frequent than nonturbulent conditions for wind speeds in excess of 4 m s$^{-1}$. This threshold also defines the wind speeds that were more frequently observed at coastal locations than at inland locations (Fig. 11). This suggests that the more frequent occurrence of turbulent mixing could be a contributing factor in enhancing fog frequency at coastal locations.

The possible generation of supersaturation ($S$) by mixing is investigated by considering the mixing of two air parcels after being brought adiabatically to the middle of the turbulent layer—one from the surface and the other from the top of the layer. The temperatures of air parcels prior to mixing are calculated by performing saturated adiabatic expansion for the bottom parcel and compression for the top parcel, corresponding to half the depth of the turbulent layer (see inset in Fig. 22). The specific humidity of air parcels is taken as their respective saturation value. The resulting temperature and humidity of the mixture is then estimated by assuming a 0.5 ratio of mixing, corresponding to the maximum attainable supersaturation (Korolev and Isaac 2000). Uncertainties on the resulting supersaturation are estimated by assuming a realistic $0.5^\circ$C error on the temperature measurements. It is found that supersaturation can result from mixing in 80% of the cases (Fig. 22), with values corresponding to the activation thresholds of common micron-sized haze particles (Seinfeld and Pandis 1998). Turbulent layers with supersaturation were characterized by weak static stabilities (e.g., around the isothermal lapse rate), while the wind shear, supported by the land–sea configurations and frictional influences over land, was sufficient to generate turbulence. Mixing is able to maintain supersaturation until a pseudoadiabatic lapse rate develops. Therefore, fog formation from mixing may be a transient phenomenon.

6. Summary and conclusions

A comprehensive study of observed environmental conditions associated with the onset of fog during precipitation was performed using historical surface-based observations, synoptic-scale analyses, and balloonborne sounding data. The evolution of cloud and visibility conditions prior to fog onset suggests that precipitation...
Fog events can be considered as extreme cases of cloud ceiling lowering (all the way to the surface) in the presence of continuous light liquid precipitation.

The analysis of environmental conditions suggests that fog formation during precipitation involves interactions spanning multiple spatial and temporal scales in a variety of large-scale weather patterns. Despite this significant variability, recurring features were identified. For instance, fog formation tends to occur in regions of stratiform precipitation mostly associated with the large-scale uplift from approaching synoptic-scale low pressure systems. The configuration of the large-scale pressure gradients associated with these systems leads to specific flow patterns, which in turn influence the structure of the lower atmosphere at the meso- and local scales. Inversions forming from differential temperature advection, or stably stratified internal boundary layers in the coastal zone associated with onshore flows and coastal gradients in SST, were found to be common features at the onset of precipitation fog. This provides indirect evidence of the role of moistening by evaporating warm precipitation hydrometeors into the colder air adjacent to the surface. Evidence has also been presented in regards to the role of turbulent mixing at the local scale, as well as the role of the interaction of the near-surface flow with topography in explaining the maximum number of precipitation fog occurrences at White Plains.

The importance of large-scale influences, such as upward vertical motion in the lower troposphere and differential temperature advection associated with directional wind shear in the presence of baroclinic zones, explain the weak diurnal signal in the temporal variability of precipitation fog onset as shown in TR07. Factors influencing precipitation fog formation are not strongly tied to the diurnal cycle as for other fog types. On the other hand, factors related to cyclonic activity exhibit a strong seasonal variability, with a maximum during winter (Sinclair 1997; Hirsch et al. 2001; Eckhardt et al. 2004) coinciding with the maximum monthly frequencies of precipitation fog events (TR07).

**FIG. 21.** Frequency distribution of observed wind speed at the surface in the soundings with the lowest layer diagnosed as turbulent and nonturbulent.

**Fig. 22.** (left) Plot of maximum supersaturation that may be created by mixing of air parcels in the turbulent surface layers vs depth of the turbulent layer, and (right) the corresponding frequency distribution of supersaturation. The inset in the left panel illustrates the method used to estimate the supersaturation, with M being the point representing the resulting mixture (see text for explanation).
The results presented herein represent important steps in efforts required to improve the understanding of precipitation fog formation in relation to synoptic-scale forcing and local-scale influences. Our findings support those of George (1940) and Byers’ (1959) statement that the northeastern United States is one the regions where precipitation fog is most common due to the presence of temperature contrasts associated with SST gradients and the baroclinic zones of extratropical cyclones. More in-depth investigations of the microphysical processes associated with warm raindrops evaporating into colder air and turbulent mixing are required for a more complete understanding of precipitation fog formation. This work will be the subject of future studies.

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REFERENCES


———, ———, and K. T. Redmond, 2004: Sea fog research in the
National Center for Atmospheric Research, cited 2007: TDL U.S. and Canada surface hourly observations. DOC/NOAA/NWS/OST/MDL Dataset 472.0. [Available online at http://dss.ucar.edu/datasets/ds472.0/]
NWS Aviation Services Branch, 2004; Terminal aerodrome forecasts. NWS Instruction 10-813, 57 pp.