Marine fog: A review

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\textbf{Abstract}

The objective of this review is to discuss physical processes over a wide range of spatial scales that govern the formation, evolution, and dissipation of marine fog. We consider marine fog as the collective combination of fog over the open sea along with coastal sea fog and coastal land fog. The review includes a history of sea fog research, field programs, forecasting methods, and detection of sea fog via satellite observations where similarity in radiative properties of fog top and the underlying sea induce further complexity. The main thrust of the study is to provide insight into causality of fog including its initiation, maintenance, and destruction. The interplay between the various physical processes behind the several stages of marine fog is among the most challenging aspects of the problem. An effort is made to identify this interplay between processes that include the microphysics of fog formation and maintenance, the influence of large-scale circulations and precipitation/clouds, radiation, turbulence (air–sea interaction), and advection. The environmental impact of marine fog is also addressed. The study concludes with an assessment of our current knowledge of the phenomenon, our principal areas of ignorance, and future lines of research that hold promise for advances in our understanding.

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1. Prologue

While studying atmospheric and oceanic processes, we realize that they are occurring on a wide spectrum of scales intrinsically coupled through complex interactions. In a more general sense, as we view the development of physics in the twentieth century and the early part of the twenty-first century, theorists have marveled at the interplay of the very large and the very small. In a more general sense, as we view the development of physics in the twentieth century and the early part of the twenty-first century, theorists have marveled at the interplay of the very large and the very small. Perhaps everything, because in any fundamental theory of physics, the large and the small cannot be separated. So what does the quantum have to do with the universe? Perhaps everything, because in any fundamental theory of physics, the large and the small cannot be separated.

Quantum mechanics is often described as the theory of the very small. A true statement, as far as it goes. Quantum mechanics is an absolute necessity, and an everyday tool, in explaining how molecules, atoms, photons, electrons, and other particles behave. It is of no consequence in explaining motion of spacecraft, planets, comets, and whole galaxies. So what does the quantum have to do with the universe? Perhaps everything, because in any fundamental theory of existence, the large and the small cannot be separated. ([Wheeler and Ford, 1998, p. 329])

A corollary to Wheeler’s statement is the impossibility of explaining the phases of marine fog (existence, maintenance, and destruction) without accounting for the very small (the microphysics of fog droplet formation) and the very large (hemispheric circulations) as well as the intermediate scales of motion. Since a typical dimension of fog condensation nuclei is 0.1 μm (10⁻⁵ cm) or less and the synoptic-scale processes linked to marine fog are on a scale of 10⁶ cm or more, a conservative estimate of the ratio of length scales for marine fog is about 10¹³. For the universe as a whole, the ratio of one of the smallest structures (the dimension of a proton: 10⁻¹³ cm) to the largest (the distance across the universe: 10²⁶ cm) is 10²⁶ ([Ford, 1968]).

Fig. 1 schematically displays the collection of processes that are central to the phases of marine fog — ranging from the large/synoptic scale (advection, subsidence, cloud) to subsynoptic scale (land–sea breezes, sea surface temperature structure, and oceanic upwelling) to the mesoscale (radiation, surface fluxes) and to the smallest scale (droplets and aerosol within the foggy air mass). And precipitation, always a challenging aspect of any geophysical problem, plays a double role by moistening and cooling of the subcloud layer and generating cloud thickening and lowering of the cloud base to eventual fog formation, but also inducing fog dissipation in some cases. The solution demands a coupled set of governing

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1 In the decades since Kenneth Ford made his estimates of this ratio, there is convincing evidence that distances significantly smaller than the size of a proton exist, yet precise quantitative measurement of these distances is difficult to confirm (Cary and Michael Hwang [http://htwins.net]).
equations that describe the dynamics, moist thermodynamics, radiation, microphysics, and turbulence processes—a formidable mathematical/physical problem. It is no wonder that this geophysical problem attracted some of the most able scientific minds of the early twentieth century: Geoffrey Ingram (“G.I.”) Taylor (1915, 1917), Harold Jeffreys (1918), Anders Ångström (1920), and Ira Bowen (1926).

The difficulties and the challenges of marine fog investigation, however, do not come without esthetic rewards. One only needs to experience the splendor of a sea fog advance along the ocean’s shoreline to be captivated by the phenomenon. Fig. 2 is a picturesque example of the majesty of sea fog as it advances toward the pier at the Scripps Institution of Oceanography in La Jolla, California. The turrets that form atop the fog bank give evidence of the buoyant plume and the leading edge is dynamic.

2. Introduction

Marine fog is a worldwide phenomenon occurring at both coastal and open ocean areas, especially frequent over the northwestern Atlantic and Pacific Oceans (Fig. 3). Here the warm western boundary currents in each of these oceans, the Gulf Stream (Atlantic) and Kuroshio Current (Pacific), abut the cold oceanic flows neighboring Labrador and Kamchatka, respectively. The processes that give rise to these foggy areas are discussed in Section 5.

Fig. 3 also indicates an area of significant marine fog frequency off the U.S. West Coast. In this area, the cold upwelled ocean water along the coast in association with strong warm-season subsidence and surface winds out of the north–northwest favor the formation of marine fog (Leipper, 1994; Filonczuk et al., 1995). These high frequency areas of marine fog give evidence of the differing factors that lead to fog over the world sea.

Although we mention in the further text many of the investigations and associated areas of frequent fog occurrence, here are some of the examples:

![Sea fog approaching Scripps Pier.](www.flickr.com)
U.S. West Coast

Palmer (1917), Byers (1930), Anderson (1931),
Leipper (1948), Oliver et al. (1978), Pilié et al.
(1979), Hudson (1980), Koračin et al. (2001, 2005a,
b), Lewis et al. (2003, 2004), Thompson et al. (2005),
Johnstone and Dawson (2010) and O’Brien et al.
(2013).

South American West Coast (Ecuador–Peru–Chile coast)

Schemenauer and Cereceda (1991) and Garreaud et al.
(2008).

Atlantic Coast (U.S.–Canada)

Taylor (1917), Willett (1928), Gultepe et al. (2006c),
Tardif and Rasmussen (2008), Toth et al. (2010) and Yang
et al. (2010).

Yellow Sea

Gao et al. (2007), Zhang et al. (2009), Heo and Ha
(2010), Kim and Yum (2010, 2012) and Zhou and Du
(2010).

South China Sea

Huang et al. (2011).

Southern Africa

Taljaard and Schumann (1940), Cermak (2012) and
van Schalkwyk and Dyson (2013).

Arabian Peninsula

Bartok et al. (2012).

From a pragmatic viewpoint, knowledge of marine fog cli-
matology and the ability to forecast it is of utmost importance. It
is estimated that 32% of all accidents at sea worldwide and 40% in
the Atlantic occur in the presence of dense fog (Tremant, 1987).
The disruption of trade and endangerment to life are always at
the forefront of its assessment (Johnson and Graschel, 1992;
Croft et al., 1995; Garmon et al., 1996; Kim and Yum, 2010; Leem
et al., 2005). And from an ecological viewpoint, changes in its
frequency have devastating effects on ecology (Johnstone and
Dawson, 2010). Some have even suggested that the economic
and human losses associated with fog and low visibilities are
comparable to the losses from other weather disasters such as
tornadoes and hurricanes (Gultepe et al., 2007b).

In this review, we pay particular attention to the wide range
of small- and large-scale processes that are germane to the
phenomenon of marine fog. A brief history of research is
presented in Section 3 and expanded upon at various junctures
in the paper. Section 4 reviews the worldwide field programs
aimed at analyzing the fog and the centerpiece of the con-
tribution is found in Section 5 — an explanation of the known
causative factors associated with marine fog. The microphysics
of fog formation is discussed in Section 6, and an assessment of
forecast capabilities and detection of fog via satellite observa-
tions are found in Sections 7 and 8, respectively. The Epilog
summarizes our knowledge of this phenomenon and lists those
areas where ignorance remains. Finally, we speculate on lines of
research that hold promise for advancement in our understand-
ing of the phenomenon.

3. History of sea fog research

As in all modern science, collection and analysis of obser-
vations is the starting point for research. Further, in the presence
of the great challenge to predict sea fog — its origin, main-
tenance, and decay — the climatology of sea fog assumes
paramount importance. The first known “archive” of sea fog
comes from the two Icelandic Sagas that include records of sea voyages from Iceland to northeastern North America in the AD 1050s through the early AD 1100s (over nine hundred years ago) — a new world called the “Wineland” by Leif Eriksson and the other Viking explorers (Bergthórsson, 2000). Bergthórsson, a renowned Icelandic meteorologist and protégé of Carl-Gustaf Rossby, pays particular attention to the meteorological and oceanographic events encountered during these voyages. Further, he reconstructs weather charts from information in the Sagas that give evidence of the sea fog that the explorers encountered (Bergthórsson, 2000, Ch. 2). And as shown on the climatic charts of fog over the world sea that came from the extensive logs at Seewarte and the Naval Observatory in late-19th century, the preponderance of sea fog along the track of these voyages from Iceland to Newfoundland exhibits the largest frequency in the world — ≈ 30% frequency over the Grand Banks of Newfoundland during the warm season (Fig. 3). Augmentation of these data resources continues with collection at the Hadley Climate Centre in England in coordination with COADS (Coupled Ocean and Atmospheric Data Set). Excellent local sea fog climatology based on COADS has appeared in publications such as those from the Scripps Institution of Oceanography in La Jolla, California [see, for example, Filonczuk et al. (1995)]. Fig. 4 presents information on the monthly frequency of sea fog off the California coast that has been extracted from data in Filonczuk et al. (1995).

Taylor made the most celebrated study of sea fog during the summer of 1913 over the Grand Banks of Newfoundland.

Fig. 4. Relative monthly frequencies (in percentages) of sea fog hours along the California coast by regions for the period 1949–1991. These data were extracted from Filonczuk et al. (1995). The plot is reproduced from Lewis et al. (2004). For more details, see Koračin et al. (2005b).
The study was made in response to the RMS Titanic disaster when the British Government and some shipping companies equipped an expedition on an old 230-ton wooden whaling ship, the S.S. [steam ship] Scotia (see Lewis et al., 2004 for details). Although the primary purpose of the expedition was to track icebergs from the glaciers of Greenland, the analysis of meteorological conditions that led to sea fog was also a central theme. One of Taylor’s recollections of the expedition follows:

“... of the 806 occasions on which meteorological observations were taken during the voyage of the whaling ship Scotia over the Banks, there was fog 141 times... It would have been impossible to examine every case in detail to see how the fog arose, but in certain cases I succeeded in raising the kite to explore the upper air, and in most of those observations I traced the cause of the fog which prevailed at the time of the ascent. In every one of these cases, it turned out that the fog was due to air blowing off warm water on to the cold water of the Banks.”

(Taylor, 1917, p. 250)

Fig. 5 shows the track of air for a particular case in July 1913 and the associated thermodynamic structure in the presence of a foggy layer over the Banks. In this case, Taylor was able to track the surface air for a period of nine days through the judicious use of observations from merchant vessels. The sea surface temperatures were obtained from weekly charts published by the British Meteorological Office. Analysis of the kite sounding at the terminal point of the trajectory along with knowledge of the sea surface temperature along the trajectory led Taylor to reason that the air column, characterized by a well-mixed adiabatic state immediately above the sea surface on 18 July, appears above 700 m on 25 July. Extension of this profile to the surface gives a temperature of 26 °C (79 °F) that is close to the sea surface temperature measured by a merchant ship at the origin of the trajectory on 18 July. He concluded that after 18 July, the air traveled over progressively colder water and cooled as the result of turbulence (shear turbulence in the presence of a stable stratification). From the measurements of relative humidity and the saturated state in the lower atmosphere, Taylor concluded that the fog extended to 700 ft (210 m ASL). The high frequency of sea fog in the northwestern Pacific Ocean displayed in Fig. 3 stems from the same action — air flowing over the Kuroshio current to a location above the cold waters of the Oyashio Current that emanates from the Bering Sea. In Taylor’s study, the cold water stemmed from the Labrador Current that flows over the Grand Banks from its Arctic Ocean point of origin.

Taylor was an expert in the theory of turbulence (Taylor, 1915) and he clearly understood the action of turbulence at the air–sea interface. It was Petterssen (1936, 1938, 1939) who expanded the view of turbulence as a factor in sea fog generation. He studied sea fog generation off the southern California coastline and realized that it was buoyancy of low-level air that stemmed from contact with a warmer sea surface (air temperature initially colder than sea surface). If the lifting condensation level (level where buoyant air reaches condensation) is below the level of the inversion, then stratus forms. And with longwave cooling at the top of the stratus that mitigates solar warming, negative buoyancy can be created and lead to cooling of the subcloud layer and subsequent sea fog formation (the stratus thickening/lowering phenomenon for fog creation). It was Douglas (1930) that first postulated and measured this negative buoyancy as it led to the haar, sea fog over the North Sea. Others who have made early contributions to understanding the influence of radiation as a contributor to sea fog are Emmons and Montgomery (1947) and Leipper (1948, 1994). Again, expanded discussion of contributions by these investigators is found in Lewis et al. (2004).

As a complement to this brief history that has emphasized contributions from the United Kingdom and the United States, we refer the reader to the book by Wang (1985). Although Wang does not view the development of sea fog research from a historical perspective, he includes a broad-based bibliography and presents a host of interesting analyses of sea fog in the East Asian area. Further valuable introduction to sea fog processes from a pedagogical viewpoint is found in Roach (1995).

4. Marine fog field programs

Field programs designed to establish a climatic base for sea fog effectively began with the international agreement on collection of ship observations in the 1850s. The first surface measurements and kite soundings in fog-prone areas that supported this climatological effort took place in 1913 as part of the International Ice Patrol over the Grand Banks. It was the same program discussed in the previous section in which G.I. Taylor collected meteorological measurements associated with fog. In addition to the measurements discussed earlier, Taylor developed fog diagrams that related surface variables to fog occurrence (discussed further in Section 5.1.1). In the following decade, the German survey vessel Meteor took surface observations and made kite soundings over the Atlantic in the latitude band 20 °N to 60 °S. This monumental field exercise became known as the Meteor Expedition (German: Deutsche Atlantik Expedition) of 1925–1927. As part of this field program, observations of inversion-base height were made in the eastern subtropical zone of the south Atlantic and in the eastern boundary upwelling zones where fog is extensive. In the period 1910–1960, data were collected from lighthouses along the U.S. West Coast in conjunction with aircraft soundings and finally balloon soundings. These data were used to identify mesoscale characteristics important to fog formation. The 1970s saw single-purpose scientific ships along the U.S. West Coast that focused on identifying small-scale atmospheric and oceanic structures in the upwelling zones that occurred in the presence of coastal fog. Again along the U.S. West Coast during the 1980–2000 period, a series of instrumented aircraft flights complemented observations from automated coast and buoy stations to study coastal fog. After 2000, satellite microwave measurements of foggy areas were used to improve microphysical parameterization for numerical forecast models. Highlights from the various field programs follow.

4.1. 1850s, Maury and ship observations

The most basic climatology, including fog for 71% of the Earth, was directed by Matthew Fontaine Maury when he served as superintendent of the United States Naval Observatory between 1842 and 1861. The systematic collection and analysis of surface meteorological and oceanographic observations from ships’ daily weather logs were the basis for this climatology (discussed in Maury’s well-known book, The Physical Geography of the Sea,
Maury organized the first true international meteorological conference in Brussels in 1853 for the world-wide collection of marine data. As part of the collection strategy, the world oceans were divided into $1^\circ \times 1^\circ$ latitude–longitude squares that were used to generate monthly analyses – a practice that continues to this day. Results are found in atlases such as The Atlas of the Climatic Charts of the Oceans (U.S. Dept. of Agriculture, 1861; reviewed in Lewis, 1996).

Fig. 5. Figures from Taylor (1917): (a) path history of air, and (b) temperature and relative humidity profiles obtained from an instrumented kite where the sea surface temperature is indicated by the arrow along the abscissa. From Lewis et al. (2004).
stratus, often closely related to fog or a stage in its cycle, has been a significant atmospheric phenomenon since ancient times. Anderson (1931) and synoptic setting (Willett, 1928; Petterssen, 1913) have been important controlling factors. The International Comprehensive Ocean–atmosphere Data Set (ICOADS) was developed, which includes variables related to fog occurrence and its nature with factors such as weather type, horizontal visibility, ceiling obscuration, humidity, and sea and air temperatures. For most any modern study being initiated on marine fog, the first step is to gather ship observations and associated analyses initially organized by Maury. We have followed this practice by making use of charts from the Atlas of the Climatic Charts of the Oceans (Sections 2 and 5).

4.2. 1913 International ice patrol off Labrador

4.2.1. 1925–1927 Meteor Expedition in eastern Atlantic

As stated above, Germany sponsored the Meteor Expedition in 1925–1927 (Von Ficker, 1936; Spiess, 1985). Accurate measurements were made with balloon wind soundings and box kite soundings with an attached temperature recorder. As a result, the air temperature inversion height was mapped based on the eastern Atlantic and it remains the only ocean-scale analysis based directly on measurements. This inversion base was so low at a point off southwest Africa that the top of the Meteor’s mast was above the air temperature inversion base. As later recognized from measurements along the California coast, a similar strong, air temperature inversion capping a cool, moist marine layer is a critical factor in fog formation and maintenance in that area.

4.3. 1910–1960 Oregon, California and Baja California

In addition to the Grand Banks, a well-recognized marine fog area with different dynamics occurs over the coastal waters off Oregon, California, and Baja California. Due to the surface pressure forces associated with the summertime subtropical anticyclone over the Pacific Ocean and the heat low over the northwest U.S., the winds along the California are from the north–northwest and they serve to drive ocean upwelling that brings cold water to the surface. A result of the contrast between the cold ocean surface and the dry, subsiding air is an atmospheric marine layer with a cold moisture source at the base capped by dry, hot air, which gives rise to conditions conducive to fog.

Reports on the basic fog conditions along the California–Oregon coast began to appear in the 1910s through the 1960s. These are: the eastern side of the northeast Pacific anticyclone dominating the area with flow toward the equator; wind driven upwelling of cold water with the coldest sea surface temperature close to the coast; and a shallow, cold, moist atmospheric marine layer containing fog and stratocumulus clouds capped by a strong, sharp air temperature inversion. The earliest accounts of these conditions were based upon lighthouse observations for climatology (Palmer, 1917), aircraft soundings (Willett, 1928; Anderson, 1931) and synoptic setting (Willett, 1928; Petterssen, 1938). Crucial to fog in this area is the distinctive air temperature inversion layer which is an important controlling factor, and stratus, often closely related to fog or a stage in its cycle, has been the focus of many early field programs reviewed by Kloesel (1992). Clarity of the atmospheric conditions accompanying fog, stratus, and cumulus in the eastern Pacific Ocean came with tropical–subtropical studies by Herbert Riehl and his students at the University of Chicago near the end of WWII (Riehl et al., 1951; reviewed in Lewis et al., 2012). Of particular interest in this marine fog review is the series of upper-air soundings that were obtained from four stationary U.S. Navy ships that operated on the flight path from San Francisco to Honolulu during the period July to October 1945. Observations from these ships defined the height and depth of the summertime inversion over the subtropical latitudes of the eastern Pacific Ocean. Morris Neiburger and his colleagues at UCLA continued the line of research started by Riehl with special emphasis on the stratus cloud in the eastern Pacific (Neiburger et al., 1961). The work of Neiburger and his associates has formed the backbone for subsequent studies involving near-surface winds, marine conditions and marine clouds. A major result is that the air temperature inversion is lowest (below 500 m ASL) and most persistent during the extended summer between Cape Mendocino in northern California and Point Conception in southern California. Going westward, the inversion base height rises quickly from the North and Central California coast, and then more slowly to be above 1200 m ASL at Hawaii.

4.4. 1970s, California Coast, CEWCOM, US Navy

In the 1970s, the Calspan Corporation (a component of the Cornell Aeronautical Laboratory) and the Naval Postgraduate School, Monterey, CA participated in the Cooperative Experiment in West Coast Oceanography and Meteorology (CEWCOM). The Naval Air Systems Command supported the program. As part of this experiment, a series of single-ship cruises took place along the California coast from northern California to San Diego at distances as far as 500 km from the coastline. On seven cruises, 30 fog events were encountered. Especially along the northern California coastline, the fog areas were characterized by strong, low air temperature inversions and wind-driven coastal upwelling that occurred in patches. Using data from this experiment, Pilié et al. (1979) reported five different conditions under which fog develops. One condition indicates that fog is triggered by instability and mixing of colder air over warm water patches, so that the fog layer grows along downwind trajectories. A second condition links fog development within a stratus layer to radiative cooling atop the cloud layer. The cooling leads to negative buoyancy so that the stratus lowers and thickens to create fog. A third condition relates fog formation to low level convergence, growth of the entire layer, and lifting of the inversion base height. The fourth condition favors fog formation over the coastal waters in response to radiative cooling over the adjacent land at night. In wintertime, a nocturnal land breeze advects the cool air out to sea. Finally, another instance is a complex land–coastal interaction that occurs along the Southern California to form fog during the fall–winter–spring in association with a “Santa Ana” — a warm offshore, dry wind regime. Fog in response to the Santa Ana flow regime was originally proposed by Leipper (1948). Focusing on southern California, radar, acoustic sounder, aircraft, satellite, radiosonde, ship and coastal observations were employed to sort out
shallow marine air movement over water with different temperatures (Noonkester, 1979; Pilié et al., 1979). Warm, dry offshore flow lowers the marine inversion and passes over cooler water where fog forms. As stated in earlier discussions, an important factor in fog maintenance and growth is longwave cooling at the fog top.

4.5. 1980–2000s, ONR, other California coastal studies

With knowledge of the large-scale factors influencing fog climatology, smaller scale, difficult to measure aspects of fog events became a focus along the California coast during the last decades of the 20th century. Long aircraft flights west of San Francisco in September 1981 found that cooler air moving over warmer water created instability and vertical mixing that dissipated fog (Telford and Chai, 1993). Ship and coastal automated surface observations illuminated the smaller scale, diurnal variation and higher concentrations of fog occurrence along California (Filonczuk et al., 1995; Lundquist and Bourcy, 1999). A coastaly trapped atmospheric event with a shallow bore pushing poleward along the central California coast created a shallow stratus-coastal-cloud bore that developed downward into fog by pushing warmer air over colder water (Dorman et al., 1998). The changing trajectories of near-surface air during transient weather events were linked to both the formation and dissipation of sea fog (Lewis et al., 2003; Koračin et al., 2001, 2005b).

4.6. Field programs after the 2000s

In the 2000s, fog field experiments tended to focus on surface observations and near-surface observations that were co-located with satellite measurements. The field work was designed with an eye on strengthening operational numerical forecast models — model development that needed guidance on parameterization of the moist physics associated with sea fog.

FRAM-L was the marine phase taking place in the summer of 2006 along the Nova Scotia, Atlantic coast that was part of a larger fog program that included continental fog. Its major strength was the use of optical probes for detailed measurement of droplets and visibility at the surface combined with satellite microwave radiometer data to examine inferred fog properties that will be used to develop microphysical parameterizations that could be incorporated into numerical forecast models (Toth et al., 2010).

5. Causative factors for marine fog

The wide range of processes governing marine fog have been mentioned earlier and displayed in Fig. 1. Earlier studies have tended to focus on a limited number of these known processes whereas recent work has often tended to dismiss significant results from the early studies — in part due to a restrictive viewpoint tied to the goals of funded research projects. Further, many studies have failed to be even-handed in their attention to the phases of marine fog — often emphasizing initiation and duration without consideration of processes that maintain and destruct fog. As is known by those who have investigated sea fog, once formed, it tends to persist. Essentially, generative processes counteract destructive processes. A negative feedback process is at work. In this section that strives to identify most of the causative factors related to marine fog, the dynamical/physical processes are divided into the following categories: cold sea fog, warm sea fog, and elevated and land forcing. And each of these major themes is further subdivided. Although we have made an effort to avoid overlap in our review, by necessity some overlap occurs — especially noticeable with information in Section 4 (Marine fog field programs) and itemization of causative factors found in this section.

5.1. Advection fog — warmer air flowing over a colder sea (cold sea fog or cold fog)

The sea fog investigated by Taylor (1917) has come to be labeled advection fog — a fog that is generated through the action of air movement over a surface with a different temperature. In Taylor’s study, the warm/moist air initially over the Gulf Stream was transported northward and encountered progressively colder sea surface temperatures on its path to Newfoundland. This we label cold sea fog in contrast to the opposite case where cold air is transported over progressively warmer water — the situation studied by Petterssen off the coast of southern California and labeled warm sea fog (discussed in Section 5.2).

5.1.1. Advection and air mass trajectory history

The panels of Fig. 6 are complements to the sea fog frequency chart shown in Fig. 3. Focusing on the high-frequency fog area over the northeastern Atlantic, we note sea surface temperatures (SSTs) as low as 10 °C north of Newfoundland. In the other cardinal compass directions relative to Newfoundland, the SSTs are significantly warmer (the American continent to the west and the Gulf Stream to the south and east). An airflow originating from any of these warmer areas can potentially induce cooling of the near-surface marine air and eventually produce fog (Fig. 6). Taylor concluded that all fluid states (moisture—temperature structure) in these cases must map onto a straight line in the q–Θ plane (q — total water mixing ratio; Θ — potential temperature) which passes through the surface state (q0, Θ0) and has a slope determined by the ratio of the air—sea moisture and heat fluxes. If the turbulence effects are dominant and radiation effects are negligible, Oliver et al. (1978) showed that this condition implies that temperature and moisture fluxes are the same and that the distributions of q–q0 and 0–Θ0, and their turbulence correlations must be functions of z/Lm and z/z0, where Lm is the Monin–Obukhov length, z is the height, and z0 is the roughness length. To achieve this state, Taylor concluded that one of the most important factors was adaptation and modification of an air mass along the long overwater trajectory originating over the warm sea surface and continuing over the cold ocean current. This use of the similarity theory complements Taylor’s analysis.

5.1.2. Advection over upwelled U.S. West Coast waters

Another fog-prone area is the U.S. West Coast, which is characterized by cold upwelled sea surface in the warm season, a shallow atmospheric marine layer with frequent northerly and northwesterly flows and occasional offshore flows, and subsidence that maintains a strong low-level marine inversion. As itemized in the Introduction, there have been numerous studies that have investigated West Coast fog. Byers (1930)
studied summertime fog over this area and hypothesized that the fog forms by saturation of moderately cold air moving over the colder sea surface. Other investigators including Leipper (1948, 1994) have challenged the premise that fog forms simply as a consequence of cooling and moistening of the transported air. According to these studies and based on many observations and data analysis, a primary mechanism for fog formation is as follows: In the warm season, offshore flows (linked to an anticyclone over the northwestern U.S.) descend over the coastal mountains and additionally warm by compression and adiabatic heating from 5 °C to as high as 24 °C (Leipper, 1994) (Fig. 7). This warm offshore flow effectively depresses the marine inversion almost to the surface and fog forms in a very shallow marine layer. Once fog forms, other processes such as the longwave radiation at the fog top induce cooling and turbulent mixing that initiate fog growth. It should be noted that Johnstone and Dawson (2010) also support the scenario with positive correlations between fog occurrence and Oakland upper air temperatures above the marine layer and negative correlations within the cooler than normal marine layer.

So, to summarize this scenario, fog does not form within the warm advected air, but it forms in the extremely shallow marine layer over the cold ocean waters. The warm advected air serves to lower the marine inversion almost to the surface and effectively contain a shallow and moist marine layer. Pilié et al. (1979) used observations and conceptual modeling and also concluded that one of the main mechanisms for the

**Fig. 6.** a. Same as Fig. 3, except for temperature differences (air temp − sea surface temp). b. Same as Fig. 3, except for wind regime.
fog formation in the warm season over the West Coast is that a warm advected offshore flow erodes the marine inversion almost to the surface and allows saturation of this near-surface warm advected offshore flow erodes the marine inversion.

### 5.1.3. Fog during coastally-trapped disturbances

Interestingly, some authors have proposed physical mechanisms different from the classical mechanisms for cold sea fog mentioned above. One example is related to fog over the cold upwelled West Coast water that appears during occasional coastally-trapped disturbances [CTD, also called coastally-trapped wind reversals (Nuss et al., 2000; Thompson et al., 2005)]. The CTD events usually occur several times each year over the West Coast during the warm season synoptic setup (anticyclone in the northern Pacific and low over the U.S. southwest). In these areas, persistent northerly and northwesterly flow regimes are interrupted by a period of opposing southerly flow near the coast where clouds and fog are spread over several hundred kilometers of the coast and rapidly propagate northward (Nuss et al., 2000). These conditions appear to represent a characteristic case of warm and moist air propagating northward over cold upwelled water [documented by Dorman et al. (1998)]. However, Thompson et al. (2005) argued that according to their sensitivity tests from 3D simulations using the COAMPS model (Coupled Ocean– atmosphere Mesoscale Prediction System; Hodur, 1997; Hodur et al., 2002), the main driver for the fog maintenance during the northward propagation was due to wind convergence and associated rising motions at the leading edge of the cloud/fog. The convergence and rising motions change the stability of the marine layer from strongly stable to a shallow and cool mixed layer with relative humidity increasing upward. Future studies including observations and advanced modeling are needed to further investigate these hypotheses based on testing particular model sensitivity to some of the processes such as surface fluxes.

### 5.1.4. Haar off British Isles

The haar is a well-known sea fog that occurs most frequently along the east coast of Scotland. The unusual name comes from the speech of the inhabitants of Lincolnshire County on the east coast of England. This fog has been traced to the cooling of warm air masses by the relatively cold waters of the North Sea (east of Scotland). British meteorologists Hubert Lamb and C.K.M. Douglas are credited with an early investigation of haar that gave evidence of both classic cold sea fog process augmented by radiative cooling atop the fog layer (Douglas, 1930; Lamb, 1943). Lamb’s study made it clear that local land–sea circulations were often active in haar where foggy air at sea was brought to land by a sea breeze. The poor visibility is often extreme with haar (visibility < 25 m). Further insight into haar is shown by Taylor (1987) and Findlater et al. (1989), among others.

### 5.1.5. Merging synoptic and local effects — Yellow Sea fog

The classic concepts of cold sea fog are in evidence from studies of fog over the Yellow Sea and the East and South China Sea. There are a significant number of observational and modeling studies aimed at understanding the active physical processes governing fog in this area. The main backdrop includes synoptic and local effects. In March–April, heating of the air over central-eastern China usually induces a shallow anticyclone over the cool Yellow Sea and the northern East China Sea that includes a prominent inversion at 100 to 350 m altitude. On the west flank of the anticyclone, this setup generates persistent southerlies which advect warm and humid air over the cool waters of the Yellow Sea and East China Sea that leads to fog occurrence (Zhang et al., 2009). In July–August, the East Asian–western Pacific monsoon initiates a strong shift from southerlies to easterlies over those two areas that brings drier air to the region that erodes the marine inversion and leads to disappearance of the fog.

Additionally to mechanisms for fog along the U.S. West Coast due to offshore flow and lowering of the marine inversion (Leipper, 1948, 1994; Koraćin et al., 2005a), Zhang et al. (2009) indicate that the offshore flow from the heated continent (at about 925 hPa) strengthens the marine inversion and traps the moisture in a shallow marine layer. Compared to the central Yellow Sea, the sea surface approaching the Korean coast is even colder due to intense tidal mixing that promotes more fog events (Cho et al., 2000; Zhang et al., 2009).
Numerical simulations of sea fog over the Yellow Sea have been conducted by Gao et al. (2007). Their results using Mesoscale Model 5 (MM5; Grell et al., 1994) showed that formation of a surface-based inversion was a critically important element in the formation of sea fog. The inversion led to development of a shallow, moist marine layer and formation of a thermal internal boundary layer at the base. Sea fog formed in response to gradual cooling and moistening by turbulent mixing. Dissipation of fog was caused by the shift in advection from warm and moist southwesterly to dry and cold northerlies.

5.1.6. Frequency and persistence of fog over the Yellow Sea

Kim and Yum (2010) analyzed the climatologies of sea fog along the Korean coast and open ocean and categorically separated events into coastal and open ocean fog–sea fog. Although there were significantly more cases of coastal fog, the sea fog (cold sea fog being more frequent than warm sea fog) was characterized by longer durations. Based on the synoptic setup discussed above, the occurrence of cold fog during the warm season is in agreement with results from Zhang et al. (2009) while the occurrence of warm sea fog was usually from January to May. Kim and Yum (2010) noted that the dew point preceding the cold fog was greater than the SST. On sea fog days, this dew point was approximately 5 °C greater than on fog-free days. Consequently, fog was formed by cooling of already sufficiently humid air.

Using numerical simulations, Kim and Yum (2012) investigated a case of cold fog over the Yellow Sea which they found to occur more frequently than warm sea fog. They simulated this fog event with both a 1D model (Bott and Trautmann, 2002) and a 3D model — the Weather Research and Forecasting (WRF) model (Skamarock et al., 2008). They indicated that some processes relevant to cold sea fog begin with cooling of the overlaying warmer air by radiative cooling and turbulent mixing which in turn leads to a thermally stable internal boundary layer. The moisture loss due to the downward moisture flux near the surface is balanced by the continuous advection of the moisture — this keeps the dew point quasi-constant and, due to net cooling, the air mass eventually reaches saturation.

5.1.7. Importance of air–sea interaction for fog evolution

Heo and Ha (2010) used coupled and uncoupled versions of the COAMPS model (Hodur, 1997; Hodur et al., 2002) and the ROMS ocean model (Shchepetkin and McWilliams, 2004) to investigate cold sea fog cases and a steam fog case over the Yellow Sea. Regarding the cold sea fog, there was advection of warm and moist air over the cold coastal waters that were additionally cooled compared to offshore regions by tidal mixing (Cho et al., 2000). Simulated trajectories showed that the warm air mass coming from the south was gradually cooled over the SST that decreased northward. Condensation occurred, thus inducing negative latent heat flux and increasing stability near the sea surface causing weakening of the winds and restricting downward mixing of drier air. Decreasing SST along the advection trajectory appears to be a significant pre-cursor for fog formation allowing cooling and moistening of the near-surface air. This also maintained the fog until the southerly advection was reduced. Heo and Ha (2010) noted that uncoupled simulations showed underestimation of stability, and they were unable to represent these processes relevant to the formation of fog.

Huang et al. (2011) studied observations during a cold sea fog event over southern China coastal waters with warm air advection over the cooler waters. They suggested that the marine boundary layer consisted of two major sub-layers characterized by mechanical turbulence near the sea surface and thermal radiation turbulence effects due to longwave cooling from the stratus/fog top. They indicated that a thermal turbulence interface was observed between 180 and 350 m altitude that separates these sub-layers. Fog dissipation or/and lifting to stratus was attributed to reduction of moisture transport, rapid lifting of the fog top and associated entrainment. Results from this study indicate that the thermal turbulence interface persisted even when the fog lifted to form stratus. In short, the thermal turbulence interface has a significant role in maintaining stratus. This could motivate future observational and modeling studies to further understand these observed characteristics.

5.2. Advection fog — colder air flowing over a warmer sea (warm sea fog or warm fog)

Another concept of fog formation stems from advection of colder air over the warm sea where saturation occurs in response to mixing of the cold and sufficiently moist air with warm/moist air. Taylor (1917) also gave some attention to this process of warm fog formation.

5.2.1. Petterssen’s and Pilié’s viewpoints

Lewis et al. (2004) reviewed the work by Petterssen regarding fog formation due to advection of colder air over warm water (Petterssen, 1936, 1938, 1939). Petterssen discussed warm sea fog in the context of mixing — i.e., mixing of two unsaturated air masses, one the incoming cool and unsaturated air and the other the near-surface warm and saturated or near-saturated air. The vertical mixing is a response to buoyancy. Sufficient mixing and radiative cooling then lead to saturation of the air mixture.

Some of the main processes discussed in this type of fog are discussed in detail by Pilié et al. (1979). They collected shipboard measurements in a situation characterized by fog at the site of the ship but fog-free upwind of the ship. There was a surface inversion layer on the upwind side that effectively trapped moist near-saturated air at the air–sea interface. When this layer traversed warm sea, the stability was eroded. Mixing and saturation then took place between the warm and moist air at the sea surface and the advancing cool, moist air (Fig. 8). The temperature decreased in the fog layer due to mixing and radiative cooling and a superadiabatic layer was maintained near the surface. The cold fog layer also represents a sink for moisture that consequently enhances evaporation from the warm sea surface.

Another study of warm fog evolution was shown by Kim and Yum (2010). They analyzed data from sea fog (cold and warm) and also coastal fog over the west coast of Korea. Their measurements indicated that the air temperature slightly dropped when the fog was formed over the warm waters. They concluded that the dew point increase, with the difference between the air temperature and the dew point during the fog mature phase ranging from 0.5 to 1.9 °C, is due
to moisture transport from the surface and is the dominant process for fog formation. They indicated that this is possibly due to longwave cooling of the fog top which overpowers warming from the sea surface. Notably, Pilié et al. (1979) also observed the effect of fog cooling over the warm waters (Fig. 8; their Figs. 2 and 3). Others also demonstrated pre-conditioning cooling and/or fog cooling over warm waters. Koračin et al. (2001) used 1D simulations emulating cooling of a Lagrangian air mass along a long overwater trajectory due to longwave cooling from the cloud/fog that can eventually overpower warming from the surface. With a continuous supply from the surface and cooling-generated instability and turbulence, this process can lead to fog formation. These effects are further elaborated on in Section 5.3.2.

5.2.2. “Steam fog” — advected air much colder than the sea surface

When a stream of cold, dry air (typically originating over land) traverses a much warmer sea surface, a “steam fog” can occur. Such events also occur with regularity over lakes in wintertime. The large heat capacity of water leads to slower cooling of the lake water relative to the land mass (especially notable in late fall and early winter). In both cases, i.e., over lakes or oceans, the latent and sensible heat fluxes are extreme (often reaching values of 100s or more of W m$^{-2}$ for both sensible and latent heat fluxes). The steam fog is frequently observed in the Arctic (Saunders, 1964; Økland and Gotaas, 1995; Gultepe et al., 2003). Further observations and simulations, especially for cases of larger air–sea temperature differences, are needed to reveal the role of SST in steam fog.

5.3. Air mass transformation leading to fog — processes at an elevated level

The earlier discussion of sea fog clearly indicates that much of the attention has focused on the relationship between the surface atmospheric layer and the ocean surface. However, there are certainly other pathways that can bring the surface layer to near saturation over the sea. Some of the other possibilities occur in response to elevated processes above the surface layer. These include convergence and divergence above the surface layer that can change the subsidence and inversion dynamics on the top of the surface layer. Above this surface layer, the presence of clouds, haze, and particulate matter alter the upward and downward radiation. Precipitation can also affect surface layer processes. Slight changes in these factors can tip the balance for fog formation. However, a near-saturation condition can also persist without fog formation where the sum of elevated processes is a significant factor. Once formed, there are tremendous differences between cloudy and cloud-free marine layers with respect to thermal, radiative, and turbulence processes that favor maintenance of a cloudy layer (Tjernström and Koračin, 1995).

5.3.1. Inversion and cloud forcing leading to sea fog

Koračin et al. (2001) studied a transient-season (April) sea fog over the U.S. West Coast in the presence of an along-coast northerly and northwesterly flow. They emphasized a need to consider the sea fog formation and evolution in a Lagrangian framework that allows for understanding air mass modification along a long over-water trajectory. In essence, these researchers followed Taylor’s (1917) lead on considering air mass transformation in a Lagrangian framework. The modification was characterized by the sea being warmer than the air with continuous cooling of the marine air along the trajectory. According to their analysis of observations and a high-resolution 1D model, they concluded that the main mechanism for fog formation was cloud-top cooling and associated turbulent mixing, supplying moisture due to a positive heat flux, and all in the presence of intense subsidence, maintained strong marine inversion, and shrinking of the marine layer along the advection trajectory (Fig. 9). The 1-D model was run in a Lagrangian mode where the advection was emulated by varying the SST over a spatial trajectory estimated from the weather analysis. The downward propagation of cloud cooling can be seen clearly in this time–height cross section of air temperature (Fig. 9). The onset of cooling first occurs at higher elevations and gradually descends to lower levels. For example, the net cooling begins after 12 h of simulation at 200 m, while it takes 2 more hours for the cooling to reach the 100-m level. Significant cooling occurred after 39 h when the elevated air driven by cloud-top cooling merged with the near-surface layer. Cloud-top cooling and
subsidence appear to be dominant factors for fog formation and evolution during cloud lowering along the advection trajectory. Their sensitivity tests showed that there is an optimum strength of the inversion and that moisture above the inversion controls the longwave cooling and entrainment processes.

5.3.2. Warm advection depressing the inversion — triggering mechanisms

A triggering mechanism for saturation within the shallow marine layer is still under investigation and debate. As mentioned earlier, Koračin et al. (2005b) used a high-resolution 1D model to simulate depression of the marine inversion and consequent fog formation. In this case, differential advection is present — warm air advection above the inversion and cold air below the inversion. Their hypothesis is that the increased longwave flux divergence at the top of a shallow marine layer with sufficiently uniform moisture leads to condensation at the top of this shallow layer, which then rapidly propagates to the surface due to further increases in longwave cooling at the fog top. Their simulations show that longwave cooling at the interface of the warm, dry air and the moist, cool air at the very top of the depressed marine layer is the dominant process for fog formation. Sufficient mechanical and radiatively-driven turbulence initiate saturation at the very top of the marine layer, which rapidly propagates downward to the sea surface within a very shallow marine layer. Consequently, longwave cooling at the fog top is the main driver of the fog growth that is limited by the strength of the subsidence and gradual warming and drying while growing and mixing vertically. Their sensitivity study (Fig. 10) clearly indicates that the model setup (vertical resolution and physics options), and initial conditions (lower boundary — SST, humidity and temperature profiles within the marine layer, and properties above the inversion such as humidity, temperature, and subsidence) significantly influence fog initiation and evolution.

These tests clearly indicate a need for sophisticated ensemble modeling providing probabilities for fog occurrence, evolution, and dissipation (further discussion in the Epilog).

5.3.3. Synoptic forcing for fog formation

Lewis et al. (2003) re-analyzed the case studied by Koračin et al. (2001) focusing on the larger-scale influences on fog formation and dissipation. They contrasted synoptic conditions for fog and fog-free events and found that the
Fig. 10. (Left) Evolution of the simulated fog top for the baseline run (solid line), a case with colder sea-surface temperature by 2 °C (○), a case with warmer sea-surface temperature by 2 °C as compared to the baseline run (*), a case with larger near-surface inversion (×), and a case with a drier hot-air layer (+), as compared with the baseline run; (right) time series of the simulated average net heating of the fog layer (°C day$^{-1}$) due to longwave and shortwave radiative heat transfer for the baseline simulation with 180 vertical points (solid), the test run with 90 vertical points (dash-dot), and the test run with 45 vertical points (dashed line) within 1200 m.

From Koračin et al. (2005b).
strength and evolution of the synoptic-scale subsidence, strength, and height of the marine inversion, air–sea temperature difference, cloudiness at the top of the marine layer as well as air-mass transformation along trajectories were critically important factors in sea fog formation. Their back-trajectory analysis showed that trajectory residence time over the land significantly reduced fog formation over the sea. Koračin et al. (2005a) performed 3D simulations of the same case from Koračin et al. (2001). The simulations were able to demonstrate that the intensity of air mass modification during this advection significantly depended on whether there were clouds along the trajectories and whether the trajectories had residence time over the land or ocean. The model was able to simulate overall cooling of the marine layer (due to longwave cooling at the cloud top) in spite of a gradual increase in the SST along the trajectory. A scale analysis of the model results showed that longwave cooling at the cloud and fog top overpowered surface sensible and latent heat fluxes and entrainment in cases of transformation of cloudy marine layer over long over-water trajectories. Fog dissipation was significantly influenced by the development of land-driven circulations as a result of the complex interplay among advection, synoptic evolution, and specifics of local circulations. Displacement and weakening of the horizontal synoptic pressure gradients and the consequent decrease of the marine winds allowed for the development of offshore flows that induced warming and drying and consequent fog dissipation.

5.3.4. Local circulations and advection of aerosols

While categorizing fog by advection properties, an important component is advection of aerosols and fog condensation nuclei that bears importance for fog both over the sea and land. The advection can be due to local circulations or synoptic patterns. Transport of sea aerosols and water vapor onto the land during sea breeze circulations can trigger fog over the land where the condensation nuclei came from the sea (Bartok et al., 2012). Obviously, fog can form over the land and then drift to the sea during a land breeze and vice versa during a sea breeze (Pilié et al., 1979). This further complicates investigation of the origin of fog and physical processes relevant to fog formation. These issues are comprehensively addressed in Section 6 (microphysics of fog).

5.3.5. Fog and precipitation

Some of the early studies indicated that cloud precipitation cools and moistens the subcloud layer and lowers the condensation level. Pilié et al. (1979) describe this process as cloud thickening. It should be noted that in many cases, this process is inaccurately called “cloud lowering”. The term “cloud lowering” should be reserved for the whole entity of cloud lowering (including the cloud top) and this is actually explained in the study by Koračin et al. (2001) where subsidence overpowered marine turbulence and the whole inversion lowered along the trajectory of the air mass.

As mentioned earlier, Tardif and Rasmussen (2008) showed that the effect of precipitation evaporating and moistening the subcloud layer can occur for a wide range of precipitation intensities and can strongly influence fog formation, evolution and dissipation. Although the majority of the cases fall into categories of light precipitation, rain showers and snow as pre-condition forcing can also lead to fog formation. They claim that the moistening is a more dominant process compared to evaporative cooling leading to fog formation. In a subsequent study, Tardif and Rasmussen (2010) used a Lagrangian-frame simulation of a microphysical column model and showed that evaporation from raindrops departing from equilibrium, i.e., the latent heat loss from droplet evaporation does not balance sensible heat flux from the ambient air, facilitates fog formation. Consequently, they indicate that the non-equilibrium state of the falling raindrops needs to be taken into account while considering precipitation fog events.

5.4. Environmental impact on terrestrial systems

Advecting marine fog transports water and materials onto adjacent coastal and terrestrial systems, profoundly altering the environmental conditions of these systems. The transport of fog water droplets and their associated soluble ions, dust particles, microorganisms, and mixtures of organic and inorganic reaction products changes the hydrologic (Bruijnzeel et al., 2005; Dawson, 1998), thermodynamic (Madej et al., 2006; Jacobellis and Cayan, 2013), nutrient (Weathers and Likens, 1996; Ewing et al., 2009; González et al., 2011), and toxicological (Wurl and Obbard, 2004) regimes of ecosystems along the coast. Although the interactions are variable and complex, several prominent ecological mechanisms associated with each type of flux (water, energy, and microscopic particulates both inorganic and organic) have been elucidated.

The strong ecological response to the spatial and temporal variability of marine fog results in discernible impacts on diurnal to seasonal (Fischer et al., 2009; Carbone et al., 2013), inter-annual to decadal (Cereceda et al., 2007; Garraud et al., 2008), and longer scales (Gutiérrez et al., 2008; Williams et al., 2008). The strong biogeographic climate signature of transported marine fog water and minerals has led to terrestrial proxies for paleoclimate change in coastal upwelling. Fog associated species such as coastal redwoods [Sequoia sempervirens (D. Don) Endl.], coastal fog opportunists such as Bishop pines (Pinus muricata (D. Don)), and endemic fog-dependent species such as Torrey pine (Pinus torreyana ssp. insularis (Hall)) and Chilean Tique [Aextoxicon punctatum (Ruiz and Pav)] have provided pollen (Heusser, 1998; Barron and Anderson, 2011) and tree-ring based chronologies (Roden et al., 2009; Williams et al., 2008; Gutiérrez et al., 2008; Díaz et al., 2001) that are used as paleoclimate proxies (Poole and van Bergen, 2006) in the interpretation of atmospheric–ocean linkages associated with major climate events. Johnstone and Dawson (2010) related how, for example, over the last 8 millennia, expansion and contraction of fog associated vegetation is interpreted as a synchronous response to marine fog-conducive upwelling conditions that are an expression of ENSO/PDO variations (Barron and Anderson, 2011). In the context of longer paleoclimatological timescales, fog imbued coastal regions now act as refugia for forest ecosystems that were once distributed across North America but do not tolerate current combinations of arid and freezing conditions such as redwood forests that are now restricted to a narrow 50 km belt along the California coast (Ahuja, 2009; LePage et al., 2005; Johnstone and Dawson, 2010). Reported decline of summer fog along the California coast (Johnstone and Dawson, 2010) and Hokkaido Island, Japan (Sugimoto et al., 2013) could result in a cascade of ecosystem responses should the trend continue.
5.4.1. Marine fog — hydrologic impacts on ecosystem dynamics

One of the earliest records of hydrologic transport from marine fogs into coastal ecosystems is the “sacred” fountain tree, attributed to Pliny the Elder. In 1764, George Glas described it as a laurel [Ocotea foetens (Aiton) (Benth. & Hook.f.)] on the Canary island of El Hierro used by indigenous people until its uprooting during a hurricane in 1610 (Gioda et al., 1995). In recent times, coastal drylands adjacent to eastern ocean basins such as California, Chile, and Namibia have been used as a productive laboratories to understand the mechanisms of hydrologic input from marine fog at scales from microbiological to landscape. In the Mediterranean climate system of California, precipitation from the strong cyclonic activity occurs during the winter with little to no precipitation falling during the anticyclonic dry summer period. By late summer the water deficit experienced by plants as they transpire more water than is available to maintain tissue turgor is at its most extreme, and if unabated will cause a plant to wilt and die.

Two mechanisms effectively transfer liquid water from marine fogs into ecosystems. The first is fog drip, where fog water accumulates on surfaces such as leaves, needles, or stems then drips or is channeled to the soil surface providing moisture to shallow roots (Means, 1927; Oberlander, 1956; Azevedo and Morgan, 1974; Ingraham and Matthews, 1995; Corbin et al., 2005; Simonin et al., 2009; Roth-Nebelsick et al., 2012) or to collection containers for human use (Schemenauer and Cereceda, 1994). The second is foliar uptake where leaf-wetting fog events increase foliar hydration when water moves directly through leaf pores and surfaces into internal tissues along a water potential gradient (Limm et al., 2009; Burgess and Dawson, 2004; Vasey et al., 2012; Eller et al., 2013). Critically important features for this mechanism are the absolute liquid water content of the marine fog event and deposition rate (Lovett, 1984; Slinn, 1982; Joslin et al., 1990; Katata et al., 2010; Hiatt et al., 2012). The first is predominantly a function of cloud condensation nuclei (CCN) composition and path history while the second is influenced by wind speed, wind direction, and dew-point depression. Liquid water inputs based on annual averages from flat screen fog collectors and fog gauges range from 0.2 to 15 l/m²/day (Schemenauer and Cereceda, 1991; Larraín et al., 2002; Juvik et al., 2011; Klemm et al., 2012).

Marine fog also increases the relative humidity, suppressing transpiration and effectively reducing plant water deficit (Ritter et al., 2009; Fischer et al., 2009; Williams et al., 2008). This effect is also notable in regions with a predominant nocturnal marine fog layer that acts to reduce nighttime transpiration rates (Dawson et al., 2007; Alvarado-Barrientos et al., 2013). The increase in available water and reduction of evaporative stress has marked impact on overall ecosystem productivity. At the canopy level, forest photosynthesis and carbon uptake is substantially higher during fog events than on clear, sunny days — a result of diffuse radiation on canopy CO₂ (Carbone et al., 2013). At the soil level, increased microbial activity measured as soil respiration also increases available nutrients leading to an overall increase in ecosystem productivity (Carbone et al., 2011).

5.4.2. Marine fog — chemical and microbiological impacts

Fogs have long been observed to carry more than water resulting in both deleterious and beneficial impacts. The same mechanisms that can lead to fog formation, such as anticyclone mediated temperature inversion, can trap and concentrate urban, industrial, and agricultural aerosols to toxic levels. Despite an extensive literature on nutrient and pollutant deposition flux from fog (Weathers et al., 1986; Collett et al., 2002) much is still poorly understood about the impact of marine fog chemistry and microbiota on humans and ecosystems. Concerns over impacts from deposition nitrogen oxides and anthropogenic sulfur dioxide (which are also found in marine fogs) have resulted in regulations on industrial and urban emissions (Lynch et al., 2000). Other impacts such as asthmatic or allergic reactions to constituents in marine fogs have not been adequately studied or regulated. In part, this results from the complexity of marine chemical composition and the reactions that occur on multiple temporal scales as well as the difficulty of ascribing causal linkage within an epidemiological context.

The marine fog air mass is a complex and reaction rich aqueous environment that reflects its ocean origin and path trajectory. The droplet “core” varies in size depending on chemical composition and can have a variety of surface films (Fig. 11) that are altered over time (Donaldson and Vaida, 2006). The interstitial space between fog droplets can contain inactivated or hydrophobic dry aerosol particles and aerosol precursor gases. These include, in varying quantities and for various durations, ionic ocean salts, anthropogenic and naturally occurring sulfate and organic compounds, atmospheric particles entrained from above such as wind-blown mineral dust and trace metals, and bioaerosols.

Large differences in reported marine fog water chemistry (Gundel et al., 1994, Northern California; Klemm et al., 1994, New England; Watanabe et al., 2001, Japan; Yue et al., 2012, South China Sea) suggest a strong geographic and temporal response of marine fog to ambient conditions. This is also mirrored in the differences between terrestrial and marine fog chemistry, the latter tending to have higher concentrations of inorganic ions, lower pH, and lower primary sulfates (Kimball et al., 1988). In contrast, marine fog of longer exposure has higher concentrations of secondary ammonium sulfate and nitrate (Chow et al., 1996). The bidirectional coastal flux between marine fog and anthropogenic sulfur dioxide (the precursor of sulfuric acid and sulfate aerosols) from copper smelters in Chile is an example of inland-sourced aerosols affecting marine fog formation and evolution. The reactions that occur in the fog air mass as it moves from the ocean to land surfaces and toward a dissipation or deposition location defines the material composition and size and, as a consequence, the type and level of impact on the receiving body, either organism or landscape unit.

![Fig. 11. Atmospheric processing of a 0.2 μm organic marine aerosol.](From Ellison et al. (1999).)
High chloride ion concentrations from sea spray-originating CCN result in a much more corrosive atmospheric environment causing cars to rust faster in seaside cities (Ma et al., 2009). The impacts of other ions including nitrogen and sulfur compounds are beneficial for nutrient limited ecosystems but deleterious when in excess of ecosystem needs. When marine fog scavenges excess ammonium sulfate, ammonium nitrate, and other air pollutants, air quality improves however the subsequent deposition (acid rain) results in deleterious impacts on vegetation and water bodies. Essential plant nutrients transported by deposition (acid rain) results in deleterious impacts on vegetation and water bodies.

The highly curved surface microlayer of bursting ocean bubbles is an important interface of hydration microphysics. Toxicologically significant pollutants such as chlorinated hydrocarbons, organotin compounds, petroleum hydrocarbons, polycyclic aromatic hydrocarbons (PAH), and heavy metals are up to 500 times more concentrated in this sea-surface microlayer (top 1–1000 μm) than in the underlying bulk water column (Wurl and Obbard, 2004). The flux of these materials into marine fog aerosols and their epidemiological impact is not yet well quantified.

Airborne particles of biological origin such as fungal spores, bacteria, algae, viruses, pollen, excreations, and biotic fragments of varying size represent a significant fraction of air particles (Després et al., 2012). Experiments in 1769 by Spallanzani to disprove spontaneous generation initiated the field of aerobiology. Fog temperature, chemical composition, and acidity affect the concentration of bacteria and yeasts but not of molds (Fuzzi et al., 1997). Metabolic activity of aerosol microbes can transform the chemical constituents of fog aerosols affecting the microstructure, wettability, and hydration of the aerosols (Deguillaume et al., 2008). The sources and impacts of microbial aerosols are not yet well understood. Urbano et al. (2011) used both culture dependent and independent techniques such as DNA clone libraries to examine 55 sequences obtained from a coastal pier in southern California. The majority of these were fungi and bacteria taxa commonly found from terrestrial sources suggesting a beach or intertidal source. Dueker et al. (2012) used 16S rRNA sequencing on bacterial cultures from ocean and onshore coastal Maine collections during foggy and clear days and found that cultures from foggy days were dominated by ocean surface microbial aerosols and microbial viability was enhanced when fog was present.

The epidemiology of airborne material has a rich literature (Fernstrom and Goldblatt, 2013) but the specific flux and consequence from marine fog is sparse. Many different atmospheric aerosols that are also constituents of marine fog have been known human health effects such as viruses, cyanobacteria (Genitsaris et al., 2011), dust (Sandstrom and Forsberg, 2008), and heavy metals such as mercury (Tchounwou et al., 2003). Pandemics that can affect large portions of the global population that also have an aerosol/droplet component, such as the 1918–1920 Spanish Flu, are increasingly investigated using an atmospheric aerosol conceptual model (Fuhrmann, 2010). The epidemiological requirement to ascertain unambiguous causal linkage makes the challenge of identifying impacts of marine fog on human health more challenging. The inter-connectedness of the air–ocean–land–anthropic system at global, mesoscale, and local scales increases the potential for an environmental burden associated with marine fogs. Efforts are underway to investigate the environmental burden in marine fog of several categories of toxicologically damaging constituents such as mercury (Weiss Penzias et al., 2012), nitrosamines (Cornell et al., 2003), and other organic compounds (Herckes et al., 2013).

5.4.3. Marine fog – thermodynamic ecosystem impacts

Marine fog-related reductions in daytime temperature and increases in nighttime temperature moderate climatic conditions for coastal ecosystems and reduce extremes in climate variability (Lebassi-Habtezion et al., 2011; Iacobellis and Cayan, 2013). The impact of the thermal moderation has wide ranging effects both for human communities and natural resources. Marine fog reduction of thermal extremes decreases emergency response to extreme heat events and human mortality (Gershunov and Johnston, 2011). Energy consumption is reduced in several sectors such as irrigation pumping and air conditioning thereby lowering energy infrastructure costs (Miller et al., 2008; Lebassi et al., 2010). Impacts on coastal wildlife from marine fog are also widespread. Sessile organisms such as intertidal marine invertebrates experience high mortality during low tide events coinciding with anomalously low summertime fog conditions (Helmut et al., 2006). Marine fog blocks shortwave radiation during California summer months when coastal stream flow is low helping to reduce salmon mortality from thermal stress. Fisheries managers in the eastern Pacific are greatly concerned that the 20th century trend (Johnstone and Dawson, 2010) of 33% declines in coastal fog could continue.

5.4.4. Conclusion

Marine fogs provide many valuable ecosystem services with most not yet fully quantified. Once considered primarily for the hazard it posed to ship, aviation, and other transportation sectors, it is now also recognized for beneficial properties (Bendix et al., 2011). Climatological projections of future marine fog conditions (O’Brien et al., 2013; Haensler et al., 2011; Tseng et al., 2012) are of extreme interest to many social and management sectors including energy (Gonzalez-Cruz et al., 2013), emergency response (Gershunov and Johnston, 2011), coastal fisheries (King et al., 2011), and biodiversity (Ackerly et al., 2012). Interdisciplinary inquiry will help advance our understanding of the environmental impacts of marine fog and quantify the ecosystem services provided by marine fog.

6. Microphysics of fog

Most of what is known about fog microphysics pertains to land fog. Nevertheless, it is suspected that results obtained from land-based fog observations are applicable in principle to sea fogs and especially to fogs near the coastline.

Probably the greatest difference between fog and cloud microphysics is the greater perceived importance of smaller droplets in fog compared to most aloft cloud studies. This is partly due to the greater prevalence of smaller droplets in more polluted environments that are more frequent at the surface where most of the pollution originates. Another reason is the greater importance of the 2nd moment of droplet sizes compared to higher moments that are more important for many cloud studies; i.e., precipitation, radar measurements. This lower moment, which is directly related to the prime motivation of
most fog studies, visibility, attaches more importance to smaller droplets.

One of the earliest investigations of fog microphysics was conducted by Pedersen and Todsen (1960). They were the first to note bimodal drop size distributions, which were also occasionally observed by May (1961) at 1–2 and 15–25 μm diameters. Garland (1971) found bimodality (Fig. 12) in half of a more extensive investigation of 25 rural England fogs, where small droplet concentrations exceeded 500 cm\(^{-3}\).

He attributed the bimodality to the difference between the larger activated droplets and the smaller unactivated droplets. Activated droplet sizes exceeded their critical sizes (at the peak of the Kohler curve of each particle) because the ambient supersaturation (S) exceeded the critical S (S\(_c\)) of the particles upon which these droplets had formed. Unactivated droplets had grown on particles with S\(_c\) higher than ambient S. Garland noted that although the unactivated (haze) droplets did not contribute much to the liquid water content of the fogs, their disproportionate surface area made a greater relative contribution to visibility reduction because in nine of the twenty five fogs these haze droplets contributed 30% of the atmospheric extinction and in five of the fogs they produced 60% of the extinction. Roach et al. (1976) also noted that the activated droplet mode contributed greatly to the liquid water content (LWC) but little to visibility reduction. Elias et al. (2009) also found that small unactivated haze droplets produced the majority of the extinction in polluted Paris fogs. Fig. 13 shows larger fog droplet size distributions in rather unpolluted fogs at Albany, New York (Fuzzi et al., 1984).

Roach et al. (1976) were the first to point out the importance of gravitational collection of droplets at the ground, which considerably reduced LWC from 1–2 g m\(^{-3}\) to <0.3 g m\(^{-3}\).

Gultepe et al. (2007b) noted that most fogs have LWC 0.01–0.4 g m\(^{-3}\). Roach et al. (1976) also placed more importance on microphysics by noting that the droplets themselves contribute to the radiative cooling that sustains the fog. Although Roach (1976) predicted fog S of a few hundredths of a percent, estimates of fog S by matching activated mode fog droplet concentrations (N\(_a\)) with CCN concentrations (N\(_{CCN}\)) measured at various S, inferred 0.8% S in rural England fogs (Roach et al., 1976). This type of inference of fog or cloud S by matching N\(_{CCN}(S)\) (CCN spectra) with nearby measured N\(_a\) is dubbed effective S (S\(_{eff}\)). With CCN measurements at much lower S by using an isothermal haze chamber (IHC), Fitzgerald (1978) inferred S\(_{eff}\) of 0.055–0.79% in sea fogs off Nova Scotia. With similar equipment, Hudson (1980; Fig. 14) found fog S\(_{eff}\) of 0.06–0.11% at four different locations along the U.S. West Coast, including one on a ship at sea (Fig. 14b).

These similar S\(_{eff}\) in spite of very different N\(_{CCN}\) indicated the validity of the proportionality between N\(_{CCN}\) and measured N\(_a\). Another measurement at San Diego, California in more polluted conditions showed S\(_{eff}\ll0.04%\) where the haze droplets alone reduced visibility well below 1 km (Fig. 14). Gultepe et al. (2007b) noted that the distinction between activated and unactivated droplets is much less in more polluted environments. On the other hand, measurements in mountain impacted stratus by Hudson and Rogers (1986) showed higher S\(_{eff}\) in the cleaner San Marcos Pass, California (S\(_{eff}\geq0.1%)\ than at the more polluted Henninger Flats (0.02–0.05% S\(_{eff}\)), which was immediately downwind of Los Angeles. This seemed to confirm predictions of Twomey (1959) that S\(_{eff}\) should be inversely related to N\(_{CCN}\). Similar aircraft measurements in stratus clouds off the US west coast indicated slightly higher S\(_{eff}\) of ~0.2% (Hudson, 1983). This was confirmed by more extensive aircraft measurements...
measurements of stratus off the US west coast by Hudson et al. (2010) and Hudson and Noble (2014) where a wider range of N_{CCN} and N_{c} definitively showed the expected decrease of S_{eff} with N_{CCN} predicted by Twomey (1959), while clean stratus showing S_{eff} > 1%. This was quite different from fog S_{eff} with similarly low N_{CCN} and N_{c} on a ship off the west coast where S_{eff} was 0.06% (Hudson, 1980). Further measurements with the same CCN instruments and similar droplet instruments in radiation fog at Albany New York showed S_{eff} of 0.026–0.20% whereas similar measurements in impacted stratus at Whiteface Mountain, New York showed S_{eff} of 0.14–0.35% (Hudson, 1984). Thus, in all comparisons fog seemed to show a lower S_{eff} than...
stratus that was either detached from the surface or impacted on mountains. This $S_{\text{off}}$ difference between fog and stratus is probably partly due to the lower vertical velocities of fog due to the inhibition of motion due to the proximity of the surface.

Meyer et al. (1980) differentiated haze from fog conditions in that visual range ($V$) which depends strictly on particle concentrations in unsaturated haze conditions but changes abruptly to dependence on particle/droplet size at $V \sim 1–2$ km where activation to fog/cloud conditions occurs (Fig. 15).

In haze conditions, where there are no activated droplets, $V$ depended only on particle concentrations and very little on particle sizes, but in activated fog conditions particle/droplet size first became equal to particle concentration effects and then overtook particle concentration dependence (Fig. 15b). The larger sized activated mode tends to persist more than the smaller sized unactivated mode during fog dissipation stages.

Precipitation often accompanies fog, especially fogs associated with frontal passages. Tardif and Rasmussen (2007) show precipitation fog as one of the four major fog types along with radiation, advection, and cloud-base-lowering fogs. Fig. 16a shows a fog case with precipitation throughout the fog while Fig. 16b shows precipitation only after the fog.

Fig. 16c shows relationships between fog visibility and precipitation, which shows that at high precipitation rates visibility is much lower than estimated from earlier studies and that visibility decreases much more gradually with precipitation rate than the earlier studies. Furthermore, drizzle seemed to have more effect than rain on visibility. The huge scattering of the data shows that more detail of the drop and droplet sizes are needed to better relate to visibility. Haeffelin et al. (2005) reported that light precipitation contributed to the reduction in visibility in three quarters of non-precipitation types of fogs.

Recent research that has revisited Sean Twomey’s interesting conjectures of over 50 years ago (Twomey, 1959) gives evidence of the interaction of vertical motion fluctuations in stratus (turbulence) and cloud microphysics. Though clouds are generally caused by rising air (positive vertical velocity [$W$]), unlike cumulus clouds, stratus clouds cannot have net positive $W$ because they are usually vertically confined. Thus, over sufficient distances mean $W$ is usually zero in stratus. This is why the fluctuations of $W$ rather than mean $W$ are generally considered in stratus (Peng et al., 2005). Hudson and Noble (2014) showed that droplet concentrations ($N$) were proportional to standard deviations of $W$ ($\sigma_w$) in polluted stratus clouds where $N$ was not correlated with CCN concentrations ($N_{\text{CCN}}$) as was the case in cleaner air masses. Hudson et al. (2010) demonstrated the suppression of cloud supersaturation ($S$) by higher $N_{\text{CCN}}$ that had been predicted by Twomey (1959). At these lower values of $S$, cumulative CCN spectra are generally steeper (i.e., higher $k$; i.e., greater $N_{\text{CCN}}$ differences per $S$ difference). Twomey (1959) pointed out that as $k$ increases, $N$ switches from predominant dependence on $N_{\text{CCN}}$ to predominant dependence on $W$. So at the higher $k$ that is more relevant as $S$ is depressed, variations of $W$ rather than variations of $N_{\text{CCN}}$ become more important for determining $N$, and there is a higher correlation between $N$ and $W$ than between $N$ and $N_{\text{CCN}}$ (Hudson and Noble, 2014). In the case of fog, this is $\sigma_w$ rather than $W$. Twomey (1959) said that high $k$ makes $W$ more important than $N_{\text{CCN}}$ for determining $N$, but it is the high $N_{\text{CCN}}$ and high $k$ of the low $S$ portion of the CCN spectrum that makes $W$ or $\sigma_w$ variations more important than $N_{\text{CCN}}$ variations for determining $N$.

The studies described here affirm the contention in the introductory paragraphs of the relative importance of smaller droplets in fogs compared to clouds. Smaller droplets have greater relative significance for visibility and chemistry, which are more important for fogs than for most cloud studies. Thus, even unactivated haze droplets take on much more significance for fogs compared to clouds. Furthermore, this is exacerbated by the greater losses due to fallout and impaction of the larger droplets, which is more prevalent closer to the surface, where there are also obstacles such as vegetation (mainly trees). Fogs can even occur without supersaturation, especially in more polluted environments.

Fig. 15. Visual range versus (a) cumulative aerosol concentrations and (b) rms diameter squared (mean surface diameter squared). Solid vertical lines indicate one standard deviation.
From Meyer et al. (1980).
where supersaturations are suppressed by competition among droplets. The impact of pollution (higher CCN concentrations) seems to be more recognizable in fogs where the smaller droplets take on more importance and the larger droplets also fall out.

7. Marine fog forecasting

Two-way communication is essential for weather forecasts — timely receipt of observations at the operational prediction center and delivery of the forecast product to clients. This process
was initiated by land telegraph that connected stations across continents by 1870; undersea cables linked the world by 1900. At sea, long-distance communications began with early radio, which received a big expansion in 1912 with the sinking of the Titanic in fog, followed by a steady stream of technical improvements that later included satellites and ultimately the World Wide Web.

Weather observations at sea were taken with basic instruments read by eye and transmitted by radio. This evolved to automated instruments on ships that recorded information digitally and transmitted electronically. The era of satellite measurements has made it possible to sample the entire world ocean, beginning with sea surface temperatures, later including estimates of winds based on cloud tracking, and vertical thermal structure from radiance soundings, and finally surface wind based on sea-state structure.

Over the past hundred years, fog forecasts have steadily improved. In the early 20th century, only the ship-based climatology was available. With the development of aviation during these same decades, land forecast centers were established and issued forecasts based on hand-plotted and analyzed maps, fog-prediction diagrams, and subjective experience of the synoptic meteorologist. By the beginning of WWII, weather forecasting was viewed in terms of the synoptic situation or so-called synoptic typing. The evolution of a weather pattern for periods up to a week was based on the analog method — matching current weather with a similar pattern in the past and using the historical pattern as a guide for the current forecast. As mentioned in the history of sea fog research (Lewis et al., 2004), C. K. M. Douglas had an encyclopedic memory of historical weather patterns and the D-Day forecast benefitted greatly from his memory of past weather. After WWII, the advent of the digital computer led to computer-based numerical weather map analyses that fed into the dynamical prediction models — essentially objective analyses and forecasts based on the physics of the atmosphere and the supporting numerical methods required to solve the governing equations. With knowledge that came from numerical prediction experiments/simulations on the global scale in the 1960s and beyond, operational forecast models began to produce products over the oceanic areas by the 1970s. Nevertheless, the coarse resolution of the early global models made it impossible to capture the small-scale processes related to sea fog initiation and maintenance.

The focus of this section is forecasting fog at sea including those situations where the marine influence laps over the coast. Forecasting is intrinsically perishable — it ceases to be useful to the operational forecaster after a given time point. The “nowcast” is a short-period extrapolation of analyses based on current observations and background information such as a forecast from an earlier time and/or climatology. The longer-range forecast relies on an accurate initial condition and a dynamical model. The merit of a forecast is judged on its ability to improve on the climatological background state.

In the following, we supply details on some of the major issues that have faced sea fog forecasting.

7.1. Initial climatology

Basic climatology of marine fog over the world sea was developed through the international collection of weather observations from ships and lighthouses (Fig. 17) starting in the mid-1800s (see Section 4, Marine fog field programs). An example of a California coastal fog climatology based upon ship and coastal stations was shown in Fig. 4.

Over time, the number of stations and data volume transmission expanded — critically important for weather forecasts on the synoptic scale. Significant weather data transmission by telegraph was transcontinental by the later 1800s; transoceanic undersea cables were well established by 1900, and ship radio was expanding by the 1920s.

Early prediction was based upon fog-prediction diagrams (Petterssen, 1956). The main client was a coastal airport where the forecast relied on surface observations such as air temperature and dew point depression along with sounding data that delivered wind shear used to predict the chance of fog, its hour of onset, severity (dense/moderate/light fog), and time of breakup. While Taylor (1917) was among the first to use this technique at sea, forecasts based on these subjective principles remained in use well into late 20th century until numerical fog prediction replaced this approach.

7.2. Surge expansion with aviation

Dramatic expansion of operational weather surface observations so important for the understanding of fog and short-term forecasts occurred with the initial development of
commercial aviation, especially air mail and passenger service over land in the late 1920s and 1930s (George, 1960, http://celebrating200years.noaa.gov/foundations/aviation_weather/#get). Interest in marine fog forecasting tended to be restricted to coastal airports subject to fog that included bays. In these early days, there were a disproportionate number of floatplane operations due to the major travel junctions at coastal bays and the limited number and quality of land runways (Johnson, 2009). The forecasting technique tended to be based on a combination of climatology, fog-prediction diagrams and a subjective estimate of the synoptic weather trend. The latter for some stations was a subjective analysis from a central agency relying on experience, climatology, surface weather observations and soundings, performed by skilled forecasters such as C.K.M. Douglas and H.H. Lamb in Great Britain (Lewis et al., 2004) and the National Weather Service in the U.S. (http://www.noaa.gov/features/protecting_1208/weatherservice.html). Some forecasters maintained a reputation of superiority over machines in forecasting fog long into the era of central, computer-based forecasts, due to the difficulty of such systems in accurately expressing the subsynoptic scale, never mind the mesoscale, in the early stages of their development.

More systematic collection of surface observations and balloon soundings at fixed sea locations began with the development of Ocean Weather Ships (in part the response to a Pan-American aircraft accident over the Pacific Ocean in 1938) and the heightened trans-North Atlantic ship and air traffic with the escalation of WWII. A dozen Ocean Weather Ship stations were in the north Atlantic and three were in the north Pacific starting about 1940 and ending in the 1970s (Adams, 2010).

7.3 After WWII

Following WWII, the weather forecasting industry in both government and private sectors expanded and flourished. Central to these expansions were the world wide communication circuits that morphed from land, undersea cable and radio teletype to the World Wide Web at century’s end. In the 1950s–1960s, most surface observations were made by human observers at airports, encoded onto paper forms and transmitted by teletype. By 2000, most of these observations were digitized by automated weather devices and entered into the World Wide Web. Satellite detection of marine fog lagged, but has improved with use of visual and IR bands as well as satellite-based soundings for enhanced measurement of the synoptic structure over water. By the late 20th century, the expanded coverage that came with numerical weather prediction included all of the world’s temperate latitudes (see Section 7.4). The forecast of low cloud and properties of the marine layer (including fog) still need considerable development.

Although numerical weather prediction’s development was initially slow following WWII, advances in both computing power and theories of atmospheric processes led to improved models and faster execution times by the mid-1960s. By the 1980s, satellite-based measurement of SST, clouds and temperature structure through inversion of radiance measurements led to improved initial conditions for the numerical models — especially noticeable over the data-sparse oceans.

7.4 Growth of operational numerical models

The main approach to marine fog forecasting starts with the larger scale motions. Output from the large-scale models generally provides boundary conditions for the subsynoptic and mesoscale models that cover a coherent geographical area such as a specific coast or a small section of an ocean and the adjacent area. The horizontal resolution of these local models approaches 10 km or better, although upper-air observations cannot resolve structures at this resolution. The benefits of fog forecasts that use output from these small-scale models in concert with climatology have received some attention (Wells, 2007), but the sample size is limited.

Many coastal stations, such as U.S. Gulf Coast stations, forecast fog based upon a conceptual decision-tree process that is a mix of climatology (large scale and station specifics), observations (local surface, local sounding, satellite imagery), centrally produced numerical guidance and analyses, local model diagnostic software, and qualitative assessment of cloud condensation nuclei (Croft et al., 1997). This combination of factors is used to forecast fog initiation, visibility, and breakup of fog.

Bartok et al. (2012) present a recent example of the use of a high resolution, 3-dimensional numerical model (WRF) coupled with a one-dimensional fog model that employs boundary layer processes and parameterized microphysics to forecast the occurrence of fog for road traffic on the north coast of the United Arab Emirates with significant skill. The fogs are formed at night from Persian Gulf marine air advected over land. Comparisons of a forecast with corresponding satellite image are shown in Fig. 18, where a correct forecast is when fog is forecasted and occurs (27 out of 84 cases), a correct negative forecast is when fog is forecasted to not occur and does not (16 out of 84 cases), and a false alarm is when fog is forecasted but does not (19 out of 84 cases). Not shown is the case when fog occurs when not forecasted, which happened only in a small proportion of the cases studied (5 out of 84 cases).

It has been suggested that improvements could be made from better integration of operationally available surface based measurements, remote measurements (satellite, ground based profilers), and numerical model outputs (Ellrod, 1995; Isaac et al., 2006). Another example is Zhou et al. (2007) who propose two different approaches that use an operational numerical model — one improves the probability of fog occurrence while the other also improves liquid water content which is required for visibility forecasts.

A variation of working on a numerical weather prediction system itself is to base fog prediction on multi-mesoscale models related in subgroupings to form systems (Zhou et al., 2007; Zhou and Du, 2010). An advantage is that this allows a relative check on how individual models and various groupings perform on deterministic and probabilistic forecasts. Ensemble-based forecasts are significantly better than those based on individual models.

Another alternative to sea fog forecasting comes with the pattern recognition method (a statistical methodology as opposed to a dynamic prediction methodology). Among these statistical methods is the classification and regression tree (CART) method — a decision tree-building technique (Lewis, 2000). The advantage of this approach is that when the available data sources, including numerical models, are not
strongly related to fog formation, combinations of indirect factors can sometimes improve the forecast. Lewis (2000) applied this technique to fog forecasting at the Kunsan Air Base in Korea. His model input included sea surface temperatures, land-based surface observations, upper-air soundings, and output from a numerical weather prediction model. Forecasting for fog has been also conducted using artificial neural networks (e.g., Fabbian et al., 2007) and for cloud ceiling and visibility using fuzzy logic (e.g., Hansen, 2007) and neural networks (Marzban et al., 2007).

7.5. Satellite development and expansion

The analysis of current observations is the essential component of the “nowcast” — the prediction on the order of several hours. As expected, satellite observations are crucial for the nowcast over the oceanic areas. The visible imagery can generally detect thick cloud, but thin cloudy areas including foggy areas are difficult to detect — essentially transparent during the daylight hours. The infrared radiance measurements at night generally fail to differentiate between the temperature of the fog top and the temperature of the sea surface. Nevertheless, work is underway to differential these temperature based on radiance differences in the shorter- and longer-waves in the infrared spectrum (Eyre et al., 1984; Ellrod, 1995). A filtering process helps identify those image pixels with partial coverage by fog. Operational algorithms for daytime detection of fog and low stratus have been proposed by Bendix et al. (2006) (Terra MODIS) and Cermak and Bendix (2007) (Meteosat SEVIRI).

Satellite-based fog detection over a range of oceanic scales is paramount to improving the fog forecasts at sea (Ellrod and Gultepe, 2007). Bendix et al. (2006) have proposed a fog detection algorithm for the MODIS instrument that includes channels in the near infrared that reportedly can detect fog to 500 m resolution.

7.6. The elusive sub-mesoscale and microscale

A major obstacle to operational numerical prediction of fog is the inability to directly incorporate fog microphysics into the model. In part, this is due to increased computational demands that come with the microphysics parameterization (Müller et al., 2007; Gultepe et al., 2006c; Tardif and Rasmussen, 2010). Further, as presented and discussed in Section 6 (Microphysics of marine fog), equations governing the microphysics of fog droplets are not easily linked with the equations that govern a mesoscale model. That is, the parameterization is not straightforward and generally involves serious assumptions. A lot of uncertainty is also present in unknown initial and boundary conditions of aerosols. Another challenge is to optimally couple the background climatology with the observations. Thus, on both the large- and small-scale, factors critical to fog and its evolution are not easily incorporated into a forecast system (Stoelinga and Warner, 1999; Gultepe et al., 2006a).

There has been considerable work on the relationships between the sub-mesoscale or microscale and fog. To give an impression of the complexities missed by operational models, some of the work on treatment of physical processes and needs for their parameterizations in models follows.

Fog occurs in aerosol-laden surface air with high relative humidity, ranging from undersaturated to slightly supersaturated (Pruppacher and Klett, 1997); it is a mixture of micron-size haze (unactivated) particles and activated particles reaching 10s of microns in size (Pinnick et al., 1978; Hudson, 1980; Gerber, 1981). Fog’s microstructure and life cycle depend on the properties of aerosols (Bott, 1991), as does superstasion (Pilié et al., 1975; Gerber, 1991). Fog droplets are generally

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**Fig. 18.** Comparison pairs of fog forecast (left, WRF model, relative humidity in units of percent, gray scale with white = 100%; arrows are 10-m flow streamlines) with corresponding satellite image (right, EUMET satellite) for north coast of the United Arab Emirates at time and date posted in upper left. White areas denote foggy regions. See text for discussion.

Figure adapted from Bartok et al. (2012).
smaller than cloud droplets; fog liquid water content is generally small, and most fog liquid water content ranges from 0.01 to 0.4 g m$^{-3}$ (Gultepe et al., 2007b).

Liquid water content is related to droplet concentration (Gerber, 1981, 1991; Garcia-Garcia et al., 2002; Fuzzi et al., 1992), gravitational settling of larger drops (Bott et al., 1990), and droplet size (Jiusto, 1981). Liquid water content has been empirically related to extinction coefficients (Eldridge, 1971; Tomasi and Tampieri, 1976; Kunkel, 1984; Gultepe et al., 2006a) while liquid water content times the droplet number concentration has been related to visibility (Gultepe et al., 2006a), an important forecasting characteristic.

The role of small-scale turbulence in fog has been investigated by Zdunkowski and Barr (1972), Turton and Brown (1987), and Musson-Genon (1987). Turbulence exchange coefficients in the nocturnal boundary layer with fog have been explored in Turton and Brown (1987) and with a second-order turbulence closure in Nakanishi and Niino (2004, 2006).

The cooling of moist air by radiative flux divergence has been analyzed by Duynkerke (1991). Another complexity is that clouds above the surface layer increase the downward longwave radiation, reducing the longwave radiation loss at the top of the fog which is important to fog dynamics (Gultepe et al., 2007a).

The role and importance of advection terms and their role in fog formation and evolution have been shown by Guéダlia and Bergot (1994). Related is that large-eddy simulations show distinct flow regimes in different stages of fog layer evolution (Nakanishi, 2000).

Fog has been explored using models of different dimensions. Local surface measurements were assimilated into a 1-dimensional fog model (Bergot et al., 2005). Oliver et al. (1978) used a second-order closure model to investigate turbulence radiation properties in fog. Koračin et al. (2001, 2005b) used a 1D model in a Lagrangian framework and also a 3-dimensional model (Koračin et al., 2005a) to simulate inversion and cloud forcing leading to fog. Bott et al. (1990) employed a 2-dimensional fog model to examine the effects of fog microphysics and radiation processes. Ballard et al. (1991) and Pagowski et al. (2004) used 3-dimensional models to explore fog variations. On the other hand, Bretherton et al. (1999), Duynkerke et al. (1999) and Teixeira (1999) used single-column versions of existing 3-D models that have been used for fog studies. Muller (2006) developed a 1-D variation assimilation scheme for surface observations and coupled two 1-D models with several operational 3-D models to produce an ensemble forecast. Zhou and Du (2010) have developed a multimodel mesoscale ensemble prediction system for fog and showed that the ensemble-based forecasts are in general superior to the single control forecasts.

As noted in the previous paragraph, a major restriction for fog forecasting improvement was that forecasting is done largely through operational numerical models that do not deal directly with the fog physics but employ coefficients and parameterizations. To improve this, the field project FRAM included data from coastal and continental sites (Gultepe et al., 2006a; Gultepe and Milbrandt, 2009; Toth et al., 2010). The marine phase took place in the summer of 2006 along the Nova Scotia Atlantic coast. Extensive measurements were made of the lower atmosphere, including variables known to be important for fog but not operationally available such as aerosol size and concentration, cloud-base height, droplet size, droplet number concentration, liquid water content, liquid water path, radiative fluxes, vapor mixing ratio, precipitation type and intensity, 3D wind speed and turbulence, and wind profiler with RASS while satellite observations were included from GOES, MODIS Terra. The data can be used to develop and improve microphysical parameterizations which will be incorporated into numerical forecast models. An example parameterization is that of visibility versus the inverse of the liquid water content times the cloud water drop number concentration (N_d) (Gultepe et al., 2006b). Another example is the concentration weighted particle terminal velocity times liquid water content versus a function based upon the liquid water content and N_d.

Future improvement of world-wide forecasting of marine fog will be based upon better observations of fog that will most likely be through technical developments of satellite based, remotely sensed systems such as hyperspectral channels to provide a basis to improve detection algorithms (Ellrod and Gultepe, 2007). Better remotely sensed observations will, in turn, provide a basis for comparison with large scale operational numerical models to deal with their lack of sensitivity in the smaller scale, near sea surface conditions to better handle fog forecasting.

It is expected that fog forecasting will become at least somewhat more realistic via algorithm development to make up for unmeasured variables that are essential throughout the life of fog (i.e., drop size distribution, condensation nuclei, sub-grid scale motions and structure) as well as to have a basis to forecast useful but previously unapproachable variables that characterize fog conditions such as visibility and fog depth.

In recent years there has been strong interest in the evolution of open sea and coastal zone fog from a climatological perspective. While global models are still operating with coarse horizontal and vertical resolutions, regional climate models (e.g., O’Brien et al., 2013) are showing some promise in addressing this issue.

8. Remote sensing of marine fog

Satellite imagery is used to identify low clouds and fog both in conjunction with in-situ field studies (Gultepe et al., 2009) and operationally by meteorological services (Molenar et al., 2000). The standard approach to detecting fog relies on detection of passive infrared emission or scattering from the cloud-top. This is achieved by evaluating the difference between the narrow-band brightness temperature observed in the infrared window region of the spectrum (typically a wavelength in the 10–12 μm range) and a brightness temperature observed in the near-infrared (typically in the 3–4 μm range). The emissivity of both increases with cloud depth, but the difference is relatively uniform for clouds thicker than approximately 100 m (Ellrod, 1995). For low, liquid water clouds, the infrared window channel emissivity is greater than the near-infrared emissivity, leading to a positive difference (window channel minus near-infrared channel) in the observed brightness temperatures that increases with cloud thickness (Ellrod, 1995). At night, the difference is small (generally less than 2 °C) for clear-sky conditions and increases to as much as approximately 5 °C for thick stratus or stratostratus clouds. Thin cirrus clouds yield a negative brightness temperature difference that distinguishes them from lower liquid water
stratus clouds. Fig. 19 shows an operational application of the technique applied to GOES 15 geostationary satellite imagery depicting nighttime low clouds and fog intruding on the San Francisco Bay Area. In addition to the presence of fog, depth of the fog based upon the brightness temperature difference (Ellrod, 1995) is also estimated operationally.

During the day, the signal of reflected near-infrared radiation from the sun overwhelms the emission signal. However, the near-infrared reflectance of low liquid water cloud is substantially brighter than the reflectance of surface or cirrus clouds, which generally yields a large negative brightness temperature difference useful for daytime detection of fog and low stratus clouds (Fig. 20).

The infrared brightness temperature difference technique is applicable over the ocean and a wide variety of land surface types. It has been applied in a range of fog studies (Ellrod, 1995; Lee et al., 1997; Bendix, 2002).

One fundamental limitation of this approach is that it does not precisely discriminate cloud near the surface from higher-level stratocumulus or altostratus clouds. Higher-level clouds may either obscure fog below or be mistaken for low-level clouds. Furthermore, even for low clouds, the infrared emission signatures of the cloud measured from above yield little information regarding the proximity of the cloud base to the surface. One solution to this problem is combining satellite data with information about surface conditions. Gultepe et al. (2007a) report that successful fog detection rates using satellite imagery compared against surface monitoring stations in Canada are only between 0.26 and 0.32, largely owing to the presence of mid- or high-level clouds. However, they improve the rate to between 0.55 and 1.00 by using numerical weather prediction model-derived near-surface temperature estimates. They report a false alarm rate of only 0.10. Similarly, Zhang and Yi (2013) use a climatology of sea surface temperature to discriminate fog from low stratus clouds over seas adjacent to China. They find that the difference between the temperature at the fog top and the sea surface temperature differs between cases of fog from cases of low stratus cloud. Hence the monthly-mean climatology of SST is used to determine a dynamic threshold on infrared brightness temperature useful for detecting fog.

Daytime application of the brightness temperature difference technique is also complicated by large variations in near-infrared illumination from the sun through the day. Data from Lee et al. (1997) documents the large daytime variation in the brightness temperature difference through the day, and demonstrate that the daytime detection of fog and low cloud is made more reliable by first estimating the near-infrared reflectance, which is a considerably more stable quantity through the day, and using the high reflectance of low stratus clouds to distinguish fog and low cloud from less reflective cirrus, ocean, or land surface. Nevertheless, further ambiguity in the brightness temperature difference is present in the dawn and dusk hours, when the weak signal in reflected near-infrared radiation renders fog indistinguishable from the surface in the brightness temperature difference. Lee et al. (2011) make use of the cloud-free visible reflectance derived from 15 days of prior satellite imagery to identify scenes that are brighter than the cloud-free case and distinguish brighter fog in the visible band from darker surface. Based on the daytime and nighttime variations.
brightness temperature difference and the additional dawn/dusk criterion, Lee et al. (2011) present a smooth 24-hour fog detection algorithm.

The limited visibility caused by fog is a consequence of the microphysical aspects of the cloud. Detection of the potential presence of fog using satellites is valuable, but further interpretation of the potential impact on visibility relies on retrieval of the microphysical properties of the clouds. The reflectance of near-infrared radiation, which has so far been discussed as a means of detecting daytime fog, is also substantially dependent upon the size of the cloud droplets. A cloud composed of smaller drops enhances the reflectance of near-infrared radiation relative to a cloud with larger drops. Near-infrared reflectance, when combined with a measure of visible reflectance (for example at 0.64 μm), has been used to perform a combined retrieval of cloud optical thickness and cloud droplet effective radius (Nakajima and King, 1990). These data are now produced routinely using imager data such as that from the MODIS instrument (Platnick et al., 2003). Wetzel et al. (1996) compare retrievals of effective radius and optical thickness for a fog case over California against in situ profiles of fog droplet sizes. Based on a favorable agreement, they discuss the potential to estimate the surface visual range in the presence of fog based on some assumptions about the vertical profile of the drop sizes. MODIS retrievals for another fog case are presented in Gultepe et al. (2009). Retrievals of cloud drop number concentration are also made by satellite with some success (Rausch et al., 2010), which could be applied to translating satellite observations to surface visibility estimates. This retrieval, however, also relies on an assumption for the vertical structure — in this case the adiabatic model for the vertical variability of LWC and drop effective radius.

Active remote sensing of fog can eliminate some of the ambiguity of satellite remote sensing, but at present is limited to select surface sites, and hence does not offer the global coverage of satellite data. Visible light is strongly attenuated by clouds, therefore lidar technology deployed from the surface is useful for detecting cloud base, such as with operational ceilometers. However, such technology cannot profile a fog layer. Suborbital aircraft and satellite lidar systems can identify cloud top heights with substantially greater precision than passive infrared imaging techniques, but are similarly unable to determine whether the cloud base reaches the surface.

Radar systems have been used to study the structure and evolution of fog layers. Operational weather radars operating at centimeter wavelengths are designed to detect precipitation sized cloud drops, but the smaller drops typical of fog layers do not effectively scatter radar signals at these wavelengths. Experimental cloud radars, however, have been deployed at 35 GHz and 95 GHz frequencies to study the layer thickness and vertical structure of cloud layers. Hamazu et al. (2003) describe a scanning Doppler radar system at 35 GHz and use it to describe the variability of reflectivity within a sea fog case. Gultepe et al. (2009) and Boers et al. (2013) evaluate the prospects for determining surface visibility from radar reflectivity. While relationships between the two quantities are apparent (Fig. 21), Boers et al. (2013) conclude that the visibility–reflectivity relationship varies as the fog layer evolves.

A 94 GHz satellite cloud radar, CloudSat, is in orbit, offering the prospect of global fog studies. However, weak sensitivity and clutter attributable to the surface return complicates the observation of cloud properties near the surface by space-borne radar.

Operational techniques for satellite remote sensing of sea fog based on passive visible and infrared imagery are now routinely applied to monitoring fog events. These techniques work best at night and are limited in their ability to distinguish fog at the surface from ordinary low stratus clouds. Active remote sensing of fog by radar shows potential for retrieving visibility estimates in fog layers, but is so far limited to experimental cases rather than widespread monitoring.

9. Epilog

Viewed historically, we have been investigating fog at sea for exactly 100 years if we label G.I. Taylor’s monumental work of 1913 as the initial point in time. What do we know about the life cycle of sea fog commencing with this early investigation? First of all, the factors identified by Taylor have stood the test of time — Lagrangian trajectories of initially warm/moist air that
is cooled from below by colder ocean temperatures and mixed vertically by velocity-shear turbulence. Since then, we also know that turbulence in an unstably stratified air mass (warm ocean temperatures relative to the overlying air) very effectively mixes the air upward and entrains air from atop the mixed layer. Further, we more clearly understand the warming and cooling of the fog layer in response to the processes of both long- and short-wave radiation; and the “capping lid” — marine inversion atop the fog layer is generally a response to subsidence associated with the semi-permanent anticyclones over the oceans. These are among the processes that have been thoroughly investigated and we have some confidence that their action is understood.

What do not we know about the cycle of sea fog? Certainly we are ignorant of the complex interactions of the various processes, some of which were mentioned above. And part of the dilemma rests on our inability to observe these processes individually, let alone collectively. And this should come as no surprise since the phenomenon exhibits extreme scales — a range of $\sim 10^{13}$.

Among the issues that are poorly understood are the following:

- Accurate estimates/predictions of subsidence elude us.
- Models cannot maintain a strong inversion.
- Many model parameterizations have been calibrated for different conditions (frequently over the land).
- Models with coarse resolution cannot fully represent coastal local circulations.
- Ocean input to atmospheric models is generally too coarse.
- Over-ocean measurements are sparse and non-existent in many areas.
- Model coupling (atmosphere–ocean) and surface fluxes are frequently approximated.
- Initial state, composition, and history of air masses including fog condensation nuclei are generally unknown.
- Condensation triggering for various subsaturation and supersaturation scenarios is difficult to represent in models with limited condensation parameterizations.
- Detailed microphysics and precipitation are usually simplified in model parameterizations.
- Physical processes relevant to marine fog in Lagrangian and Eulerian frameworks need to be more closely examined.
- Fog is an elusive target for airborne field programs; airborne measurements in fog are limited by the Federal Aviation Authority (FAA) and other agencies.
- Climatology of marine air characteristics is poorly known over many regions of the world sea.

Improvement in our knowledge of sea fog will undoubtedly come from detection of sea fog through satellite observations. While limitations persist in the quantitative information attainable about fog from satellites, the vastly improved spatio-temporal sampling of the oceans afforded by satellites and the demonstrated ability of satellite detection of fog events suggests that many more fog events can be identified and studied with the aid of satellites.

Of course, we will gain much from measurements of sea fog in field programs that incorporate observations from ship and coastal stations. The future trend will emphasize programs that amalgamate observations in the hope of clarifying the interactions of processes. Field measurements will be combined with model output to obtain the most accurate representation of a system state that stretches across the scales from synoptic to microphysical.

From the many modeling studies of sea fog, essentially numerical experiments/simulations/forecasting that started in the immediate post WWII period, it becomes clear that deterministic forecasting of sea fog onset and its duration has generally been unsuccessful. The extreme sensitivity of model output to elements of control (initial conditions, boundary conditions, and forcing (empirical/physical parameterization)], in concert with the chaotic nature of dynamic prediction, is at the heart of prediction inaccuracy. Ensemble prediction is a possibility, but it comes with complications when applied to sea fog. The complications arise because the phenomenon is discontinuous with an impulsive start and an abrupt end. The fundamentals of ensemble prediction applied to discontinuous dynamical systems such as this one are in their infancy. Without doubt, however, this area of investigation is needed and promising.

When will we be able to make accurate forecasts of sea fog onset and its duration? At this time, we are unable to answer this question with any degree of certainty. However, we know from the history of geophysical science that improvements in understanding and forecasting come incrementally with dependence on better observations (both temporal and spatial) that lead to improved four-dimensional analyses which in turn lead to improved dynamical forecasts. And not only from models, but advances will come from the individual scientists (observationalists, analysts, and theoreticians) with their phenomenological viewpoints — viewpoints that lead to conjectures or hypotheses that when followed to their terminal points contribute to the incremental advance.

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