Chapter 19

100 Years of Progress in Mesoscale Planetary Boundary Layer Meteorological Research

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ABSTRACT

This chapter outlines the development of our understanding of several examples of mesoscale atmospheric circulations that are tied directly to surface forcings, starting from thermally driven variations over the ocean and progressing inland to man-made variations in temperature and roughness, and ending with forced boundary layer circulations. Examples include atmospheric responses to 1) overocean temperature variations, 2) coastlines (sea breezes), 3) mesoscale regions of inland water (lake-effect storms), and 4) variations in land-based surface usage (urban land cover). This chapter provides brief summaries of the historical evolution of, and tools for, understanding such mesoscale atmospheric circulations and their importance to the field, as well as physical processes responsible for initiating and determining their evolution. Some avenues of future research we see as critical are provided. The American Meteorological Society (AMS) has played a direct and important role in fostering the development of understanding mesoscale surface-forced circulations. The significance of AMS journal publications and conferences on this and interrelated atmospheric, oceanic, and hydrological fields, as well as those by sister scientific organizations, are demonstrated through extensive relevant citations.

1. Introduction

Experience tells us that the surface has a large impact on the weather conditions we experience at the ground. We take walks in the cooler shaded areas provided by forests. We build walls or lines of tall trees to locally slow the wind and decrease wind erosion from the surface. We await sea breezes to cool us on hot summer days on the beach. It is natural, then, to expect that as our understanding of the weather and climate became more quantitative, much research would be devoted to

understanding the shallow, mesoscale weather systems that are directly linked to Earth's surface.

This chapter explores the mesoscale planetary atmospheric boundary layer (ABL) features that are directly tied to variations in the surface of Earth. These features are generally formed through spatial (and temporal) changes in the features on the surface, both natural and man-made. However, look at any region, especially over land, and it becomes very clear that mesoscale variations abound. Different types of crops are planted over regions tens to hundreds of kilometers across, irrigation alters the surface moisture conditions sometimes in large swaths, and the forested areas give way to prairies.

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To give a full accounting of the research on all boundary layer features driven by mesoscale land surface changes would be beyond the space allocated for this chapter. Therefore, we provide information on research on common examples of atmospheric responses to the surface, starting from thermally driven variations over the ocean and progressing inland (coastlines, lakes) to man-made variations in temperature and roughness (urban), and ending with forced boundary layer circulations (wind farms): 1) overocean temperature variations (responses to oceanic boundaries), 2) coastlines (sea breezes), 3) mesoscale regions of inland water (lake-effect storms), 4) variations in land-based surface usage (urban weather and climate), and 5) forced boundary layer motions (wind farms). This chapter provides brief summaries of the historical development of, and tools for, understanding such mesoscale atmospheric circulations and their importance to the field, as well as physical processes responsible for initiating and determining their evolution. In this way, we hope to explore examples of atmospheric responses to those dominated by changes in surface roughness/forcing, to primarily changes in thermal forcing, and to combinations of both of these forcings.

The focus of this chapter is on developments in the field of mesoscale surface-based weather and climate over the time span of the American Meteorological Society—100 years. Knowledge of some of these weather and climate phenomena spans back far longer (such as sea-breeze circulations), while some are more recent developments in the field (such as responses to wind farms). Commonalities and differences between the evolution of these example phenomena provide a useful overview of the drivers of research and the way in which rapidly developing observational, analytical, and numerical modeling techniques largely determine the speed of progress.

2. Responses to oceanic boundaries

While the existence of warm currents along the western boundaries of oceans and cool currents along the eastern boundaries has been known since well before the twentieth century, it was not until the advent of infrared (IR) imagery from polar-orbiting satellites that their mesoscale structure gained widespread appreciation. The existence of a sharp sea surface temperature (SST) front along the northern edge of the Gulf Stream along with mesoscale meanders in that oceanic front were graphically apparent in the *Improved TIROS Operational Satellite (ITOS-I)* IR satellite imagery (Rao et al. 1971). The cold currents along the eastern boundaries of oceans can also exhibit strong mesoscale

temperature gradients, in their case due to Ekman-pumped upwelling of cooler water from below during periods of equatorward alongshore surface wind (Holladay and O'Brien 1975).

Figure 19-1, a three-day IR composite from a later generation of polar-orbiting satellite, clearly shows these features northeast of Cape Hatteras, North Carolina. Long-term satellite observation of the western boundary current meanders revealed both long-period standing waves and shorter, mobile waves with periods around six weeks (Halliwell and Mooers 1983). The meanders in Fig. 19-1 are typical examples of these more rapidly evolving mesoscale waves in the Gulf Stream. Standing waves have also been observed in the Agulhas Current Extension off the southern tip of Africa (Liu et al. 2007).

With both types of ocean currents exhibiting temperature differences and spatial scales similar to those of the Great Lakes, it is not surprising that a similar array of surface-flux-driven mesoscale atmospheric responses arise. Simple mixed-layer models of this process revealed that cold-air advection across the sharp oceanic temperature front of a western boundary current results in enhanced surface fluxes of heat and moisture and the consequent downstream warming, moistening, and deepening of the atmospheric boundary layer. Similarly, mesoscale modeling of cold-air outbreaks (Sun and Hsu 1988) revealed that ABL convection, already enhanced over offshore waters, increased sharply on crossing the Kuroshio's SST front. Further improvement in resolution to a cloud-resolving large-eddy-simulation model (Skylingstad and Edson 2009) showed that western boundary currents result in stronger turbulence and precipitation, as the convective mass flux ties directly to surface forcing and ABL depth, both of which are enhanced by the greater air-sea fluxes.

Observational studies of air-sea flux enhancement near western boundary currents have begun more recently with buoy observations (Konda et al. 2010) revealing that during the cold-air outbreaks latent and sensible heat fluxes were large over the Kuroshio, while during warm monsoonal flow only the latent heat flux was enhanced. These results are consistent with bulk aerodynamic theory given the differing characteristics of continental polar and maritime tropical air masses. Simultaneous rawinsonde launches across the Kuroshio's SST front (Nishikawa et al. 2016) revealed that the Kuroshio warmed the ABL sufficiently to create pressure anomalies extending up to 800 m. In short, the presence of markedly warmer water in the western boundary of the Kuroshio has a profound impact on the heat, moisture, and pressure fields in the ABL.

FIG. 19-1. AVHRR-derived SST composite for the three-day period ending 0107 UTC 27 May 2008. (Source: <http://fermi.jhuapl.edu/avhrr/sst.htm>; copyright 2008 by the Ocean Remote Sensing Group, The Johns Hopkins University Applied Physics Laboratory.)

These atmospheric responses acquire additional mesoscale structure because the SST front bends and closes off in the presence of current meanders and rings. Warm SST anomalies resulted in downward turbulent momentum flux anomalies near the ABL top (i.e., enhanced entrainment) in a high-resolution numerical model (Koseki and Watanabe 2010), while cold SST anomalies reduced entrainment. This link between surface buoyancy flux, ABL growth, and entrainment provides the first of several mechanisms by which western boundary currents, and their associated meanders and rings (Sugimoto and Hanawa 2011), can affect the ABL winds. The greater surface buoyancy flux also results in more momentum mixing in the surface layer, resulting in stronger 10 m wind speeds (Liu et al. 2007). Indeed, this enhanced mixing can reduce vertical wind shear over the entire depth of the ABL (Hashizume et al. 2002).

Western boundary currents, meanders, and eddies can also impact wind fields through their enhanced fluxes'

power to create mesoscale temperature structures spanning the ABL. In the presence of along current winds, trajectories remain in the region of enhanced fluxes for an extended period, leading to an atmospheric heat and moisture front overlaying that in SST (Rouault et al. 2000). If, instead, the wind blows at a significant angle to the SST front the atmospheric front can be displaced downwind of the SST front and its intensity impacted, as revealed by idealized model case studies for the Genesis of Atlantic Lows Experiment (Huang and Raman 1990). Likewise, shipborne in situ observations from the High-Resolution Remote Sensing Experiment revealed that under weak synoptic forcing (i.e., large-scale wind component across the SST front weaker than the atmosphere's potential solenoidal response to the surface flux gradient) a sea-breeze-like atmospheric flow formed in response to the SST-induced atmospheric front (Sublette and Young 1996). Mesoscale modeling revealed that this solenoidal response was primarily a function of

air–sea temperature difference, the synoptic wind component perpendicular to the SST front and the depth of the ABL.

The high horizontal resolution of recent global reanalyses have permitted the large-scale mapping of these mesoscale atmospheric responses to western boundary current SST fronts. Locally enhanced temperature and moisture fluxes led to locally strengthened surface-wind convergence, and thus local maxima in cloudiness and precipitation, in the vicinity of the SST front (Masunaga et al. 2015). These SST fronts created atmospheric frontogenesis not just directly, through their impact on surface fluxes (causing a horizontal gradient in the heating due to vertical flux convergence), but also indirectly through this heating's creation of a hydrostatic trough and thus convergence over the warmer water.

This combination of convergence and warm, moist air over a western boundary current can also enhance deep convection. The sharp atmospheric front observed across the Kuroshio Extension's SST front leads to one such band of deeper clouds, observed in satellite imagery (Tokinaga et al. 2009). Such deeper convection often reaches cumulonimbus stage as revealed by the World Wide Lightning Location Network (Virts et al. 2015). Such thunderstorms over the Gulf Stream exhibit both seasonal (summer maximum) and diurnal cycles. Along the southeastern coast of the United States, the Gulf Stream SST front is close enough to the shore for the land–sea solenoid (i.e., sea-breeze circulation) to impact flow direction and thus the Gulf Stream atmospheric frontal intensity (Sublette and Young 1996) as well as near-surface convergence (Virts et al. 2015), to suppress lightning around 1800 local time. Lightning in association with the Gulf Stream atmospheric front was found to reach a maximum in the late evening through noon of the following day when land–sea wind fields were more favorable for atmospheric front formation over the SST front.

Warm western boundary currents can also impact overland convection in adjacent regions by supplying additional moisture to the ABL. Given onshore flow, such as that across the Agulhas Current, the resulting warm, moist air can be advected overland, enhancing the convective potential of the sea-breeze front (Jury et al. 1993; Rouault et al. 2002).

Meanders in western boundary currents focus the associated atmospheric impact into mesoscale regions over the warm, anticyclonic bends. For the Agulhas Return Current, the Advanced Microwave Scanning Radiometer on the Earth Observing System *Aqua* satellite revealed that SST perturbation-induced differences in ABL depth averaged 300 m (O'Neill et al. 2005). Over the Agulhas Current Extension this enhanced mixing led to faster ABL

wind speeds over the SST maxima and thus convergence and rising motion over their downwind side (Liu et al. 2007). Additional aspects of this behavior have been explored via Weather Research and Forecasting (WRF) Model simulations of Kuroshio meanders (O'Neill et al. 2010). These simulations reveal that mesoscale SST perturbations modify the wind direction within the ABL via SST-induced perturbations to the crosswind components of the pressure gradient, turbulent stress divergence, and the Coriolis force. The pressure gradient responds to the horizontal gradient of vertical flux convergence within the ABL, which is in turn caused by the horizontal gradient in surface heat flux resulting from the SST gradient. The turbulent stress divergence results from the cross-front difference in turbulent mixing associated with the SST-induced difference in atmospheric surface-layer stability. Turbulent momentum flux and SST-induced pressure gradient act together to accelerate the surface flow toward warmer water and decelerate the flow toward cooler water.

Sensitivity studies with a regional atmospheric model reveal that the sharpness of the SST gradient impacts the intensity of the atmospheric response to mesoscale current features (Tanimoto et al. 2011), with simulation of both the pressure trough and ageostrophic winds requiring adequate resolution of the SST front in the model boundary conditions. Again, the turbulent mixing of heat and momentum and the hydrostatic response to the former are implicated.

Just as the SST front of a western boundary current can produce mesoscale solenoidal circulations, meanders in that front can focus those circulations into bands when the synoptic winds align with the meander axis (Young and Sikora 2003). Thus, during cold-air outbreaks over the Gulf Stream mesoscale cloud bands of unusual width and area can form over and downwind of warm-core meanders as shown in Fig. 19-2. The resulting cloud bands are similar to, but of larger scale than, those extending from Chesapeake and Delaware Bays during similar conditions, a result in keeping with the scale difference between the meanders and bays. For the Gulf Stream, these meander-driven cloud bands occur only during that portion of spring when the land–sea temperature difference in New England and the mid-Atlantic coast is small, rendering the Gulf Stream SST front the effective thermal coast of North America.

Western boundary currents are not the only sources of SST forcing for mesoscale atmospheric circulations. Coastal atmospheric circulations, discussed in more detail in section 3 of this chapter, can also be altered by SST variations. Cool currents along the eastern boundaries of major ocean basins typically underlie equatorward winds and thus experience coastal upwelling as

FIG. 19-2. AVHRR channel-4 infrared image from 0713 UTC 8 Apr 2002. [From [Young and Sikora \(2003\)](#); copyright 2002 by the Ocean Remote Sensing Group, The Johns Hopkins University Applied Physics Laboratory.]

seen in the Coastal Upwelling Experiment 2 ([Hawkins 1977](#)). This vertical advection of cooler water from below enhances the diurnal land–sea temperature difference and thus intensifies forcing for sea-breeze circulations. Simulation of these effects with a simple coupled ocean atmosphere numerical model was soon undertaken ([Clancy et al. 1979](#)). These simulations suggested the possibility of a two-way feedback between SST decreases due to coastal upwelling and the sea-breeze circulation.

For appropriate longshore synoptic winds and coastal configuration, the mesoscale diurnal sea breeze can contribute to the longshore stress and thus to upwelling. This in turn enhances the land–sea temperature contrast and the sea-breeze potential. The fact that these processes happen at scales below the Rossby radius of deformation, however, limits their impact on the mean pressure and wind fields, limiting the impact of the feedback. Results of a 10-yr observational study of SST and surface wind at Cabo Frio, Brazil, show that seasonal variations in the land–sea temperature difference due to cold water upwelling near the coast impacts the intensity of the sea breeze ([Franchito et al. 2008](#)). When the synoptic winds are aligned longshore, they enhance upwelling and thus the land–sea temperature difference and sea-breeze intensity. This process has been simulated for Monterey Bay, California, with the Integrated Regional Model System ([Tseng et al. 2012](#)). This model

couples WRF, the Community Land Model, and a variant of the nonhydrostatic Monterey Bay Area Regional Ocean Model. The results confirmed the existence of feedbacks between coastal upwelling and the ABL. Other, less studied, sources of mesoscale SST variability exist and should be expected to exhibit equally complex atmospheric responses.

3. Coastal circulations—Sea breezes

a. The first 50 years of AMS

Knowledge of sea breezes was gained early in human naval history. The timing, changes, and strength of sea breezes played a large role in efficient and safe landings and departures of ships from dangerous coastlines (e.g., [Neumann 1973](#)). Because many early settlements were near large water bodies, sea breezes were a daily part of life. Therefore, much experience had been gained on these circulation systems. However, in the first century of AMS, understanding of sea-breeze characteristics, their formation mechanism, and impacts on the regional environment and society grew considerably. Of course, almost the same can be said of land breezes, but much less is currently known. The primary focus in this section is on illustrating the increase in knowledge of coastal flows with the much more heavily studied sea breeze.

FIG. 19-3. Schematic diagrams of theories for sea-breeze formation processes near 1914. (Credit: David A. R. Kristovich.)

Sea breezes are mesoscale wind features driven by atmospheric thermal differences in the vicinity of coastlines. These differences arise from physical differences in water and land, which control the amount of heat available to the overlying air. Sea breezes are the responses of the lower atmosphere to the resulting temperature differences between air above the land and water surfaces.

A review article by Ward (1914) provides a thorough overview of the state of knowledge of sea breezes near the founding of AMS. Ward clearly shows that much was already known about sea breezes: Sea and land breezes are most likely to occur during clear-sky and light-wind conditions and are driven by differences in heating rates between the ocean and land surfaces; onshore flows of sea breezes are shallow (hundreds of meters), but often exhibit large horizontal extents (tens to hundreds of kilometers); sea breezes often are topped with offshore flows; such offshore flows may precede the near-surface sea breeze; sea breezes usually initiate well offshore (tens of kilometers); wind directions tend to veer with time during the day due to the influence of Earth's rotation; and sea- and land-breeze circulations can be

augmented or decreased by local and large-scale wind patterns.

Ward (1914) pointed out that despite the knowledge of sea breezes at that time, there was considerable dispute about the process by which sea breezes formed. Commonly held theories at that time are illustrated in Fig. 19-3. The convection theory (Fig. 19-3a) holds that during the day, solar radiation heats the coastal land areas much more quickly than the ocean, causing low pressure and ascending air over land. This rising air, in turn, is replaced by denser air from the ocean driving inland in the form of a sea breeze. Others argued that increasing pressure over the cool ocean drove this sea-breeze process (Ward 1914). A modification of the convection theory (Fig. 19-3b) attempted to explain why the sea breeze becomes evident at the shore several hours after the time the land temperature increased above the ocean temperature. This second theory argued that an expanding bubble of air over the warming land would have an outward force in all directions, thus delaying the inland movement of the cool marine air. A third contemporary theory (Fig. 19-3c)

pointed out that heating over land causes the isobaric surfaces to rise considerably with time, thus producing a seaward pressure gradient force (PGF) above the surface (Ward 1914). Surface pressure over the ocean would increase as air slid down the isobaric surfaces toward the ocean, resulting in a landward PGF and a sea breeze.

Over the next few decades, it became generally accepted that the sea breeze begins at higher altitudes (Fig. 19-3c), followed by the surface winds (although the convection theory is occasionally cited even today). Early observations that the surface sea-breeze winds start tens of kilometers offshore (such as indicated by the ocean surface exhibiting a “fine, black ‘curl’ upon the water”; Ward 1914) could not be explained by convection theory (Fig. 19-3a). Observations obtained by a predecessor to AMS, the New England Meteorological Society, added important evidence to support the higher-altitude initiation theory (Fig. 19-3c) by 1914.

A few key issues were identified in the 1920s–40s, which shaped the discourse on sea breezes. Jeffreys (1922) envisioned the sea breeze as a density current, such that its movement was governed by the balance between PGF and surface friction. McAuliffe (1922), meanwhile, added the concept that the sea-breeze intensity and movement were greatly modified from density current dynamics by large-scale winds. Theoretical models allowed for exploration of individual factors beyond purely density current physics, such as the effects of greater surface friction over land (Beers 1947), evolving inland temperature fields influenced by changing surface heat fluxes with time (Schmidt 1947; Showalter 1949), and Coriolis force (Schmidt 1947).

Technological innovations explored during World War II and increased funding availability for atmospheric studies led to new observational techniques that provided important insights. For example, Atlas (1960) established that radar can be used to track sea-breeze fronts remotely over wide areas. Fisher (1960) and Angell and Pack (1965) reported on field experiments dedicated to observing both the evolution of the sea breeze and the return flow above.

These observational efforts supported development of theoretical models incorporating linear and nonlinear factors and development of numerical modeling techniques aided by the advent of the “electronic computer” (Fisher 1961). Particular attention during this time period was on advancing quantification of the individual components of sea-breeze (SB) forcing as well as the nonlinear effects of how the forcing components affected each other, and thus, sea breezes. An important outcome of such studies was the quantification of factors controlling both the intensity (as measured, e.g., by

temperature differences across the front) and movement of the sea-breeze front, especially the role of larger-scale gradient wind fields (e.g., Estoque 1962; Frizzola and Fisher 1963; Angell and Pack 1965; Fosberg and Schroeder 1966). Briefly, sea-breeze fronts tend to be most intense with light opposing surface winds, which cause minimal inland movement. The studies by Estoque (1962) and other contemporaries, have been particularly recognized as forerunners of advanced sea-breeze simulations and were cited in scientific journal articles for many years thereafter.

Other details of sea-breeze structure and evolution were slowly added to the base of knowledge through the 1960s, but they were quite limited in scope both spatially and in the phase space of influencing factors. However, it was not yet possible to explore how important these findings were for coastlines worldwide. These efforts identified the important role that coastal shape and topography play in certain locations (e.g., Leopold 1949; Staley 1957) as well as causes and influences of gravity waves on sea breezes (Donn et al. 1956; Geisler and Bretherton 1969). Importantly, this time period heralded the beginning of applied research focused on how sea breezes affected regional climate and environmental conditions affected by climate. This included the distribution of precipitation (Byers and Rodebush 1948) and atmospheric dispersion of pollutants (Angell and Pack 1965). We will see in the following sections that such applied studies became increasingly numerous and the topic continued to grow in interest to the present time.

b. Rapid advancements in understanding and simulating sea breezes

The 1970s brought in a large increase in the number of articles in the scientific literature on the sea breeze, due to increases in the use of numerical modeling techniques and radar observations. Early in the 1970s, the increase in use of 2D models continued (e.g., Walsh 1974; Neumann 1973; Anthes 1978), with advancements allowing 3D models (Pielke 1974a,b) and ultimately simple fully coupled air–coastal ocean models (e.g., Clancy et al. 1979). At the same time, observational equipment and project design techniques continued to improve, allowing for more information on the climatological 3D characteristics of the SB and information on the influence of simple air–topography–water relationships (e.g., Johnson and O’Brien 1973; O’Brien and Pillsbury 1974; Lhermitte and Gilet 1975; Hawkins 1977; Barbato 1978; Burpee 1979).

Such advancements led to a much deeper understanding of the sea-breeze phenomena, such as 1) recognition and expansion of information on the influences of nonuniform coastlines and coastal and inland

mountains (McPherson 1970; Pielke 1974a; Hawkins 1977; Barbato 1978; Mahrer and Pielke 1977), 2) understanding of reasons for variations from the typical clockwise turning of the sea-breeze winds throughout the day (Neumann 1977; O'Brien and Pillsbury 1974; Lhermitte and Gilet 1975; Burk and Staley 1979; Reed 1979), 3) influence of large-scale conditions on SB intensity and evolution (Johnson and O'Brien 1973; Pearson 1973; Pielke 1974a; Burpee 1979), and 4) feedback mechanisms due to SB air motions affecting coastal upwelling that, in turn, causes density increases of the marine air and thereby affects SB intensity and movement (O'Brien and Pillsbury 1974; Hawkins 1977; Clancy et al. 1979).

Multiple investigations on the importance of the Coriolis force, and thus latitude, continued into the 1980s. Dalu and Pielke (1989), for example, explained the interchange between friction, which slows the wind and thereby decreases the influences of Coriolis turning, and inland penetration of the sea-breeze front. Neumann pointed out a fundamental error in Jeffreys' (1922) formulation that indicated that Coriolis was not important. The error was due to Jeffreys' overestimated wind speed and underestimated inland penetration distance.

Research in the 1980s further refined our understanding of environmental and surface factors controlling the evolution of sea-breeze dynamics, while there was also an increase in the application of mesoscale and chemical models to understand air pollution. Among the physical processes explored, it was found that 1) the inland movement of the sea-breeze front (SBF) fundamentally depends on Coriolis turning that, in turn, is related to frictionally slowed inflow speed and latitude (e.g., Kozo 1982a,b; Rotunno 1983; Dalu and Pielke 1989); 2) increasing surface friction tends to deepen and weaken the SB inflow and slow the SBF inland propagation (Kozo 1982b; Briere 1987; Dalu and Pielke 1989); 3) nonuniformities in the coastline shape results in the possibility of colliding SB fronts from different coasts, complicating pollution trajectories and providing locations for enhanced convection (Abbs 1986; Cautenet and Rosset 1989); and 4) boundary layer wind flows from nearshore topographic features can locally alter the SB evolution by increasing or decreasing the SBF propagation (Abbs 1986; Kitada et al. 1986; Orlic et al. 1988).

Understanding, and the ability to simulate, sea-breeze flows had improved to the point where applications of models to other processes increased. For example, tropical cloud bands lasting hours to days could initiate near SBs developing in the vicinity of complex shorelines (Sun and Orlanski 1981a,b; Burpee and Lahiffi

FIG. 19-4. Dual-Doppler perturbation wind field at 600 m at 1824 UTC with the mean wind of $U = 1.5 \text{ m s}^{-1}$, $V = 0.5 \text{ m s}^{-1}$ subtracted. The sea-breeze front is shown by the 15-dBZ reflectivity contour (light gray line). The surface position of the trailing convergence line is indicated by the dark gray line. The locations of two vortices within the 600-m winds are shown along the trailing convergence line. [From Laird et al. (1995).]

1984). Multiple studies, some using combined dynamic and chemical models, showed the importance of sea breezes on the pollution concentrations at the surface. Horizontal and vertical advection of pollutants within the SB circulation (e.g., Lyons 1972; Lyons and Olsson 1972; Young and Winchester 1980; Bornstein and Thompson 1981; Ueda et al. 1988), as well as mixing of the SB air with its environment (e.g., Mitsumoto et al. 1983), play key roles that had various impacts of pollution dynamics in different locations.

Research in the 1990s continued to increase our understanding of the complexities of physical processes involved in lake breeze evolution, with particular emphasis on smaller-scale features, such as strengthening areas along sea-breeze fronts as they approached the more stagnant, warmer areas in the urban heat island (UHI, discussed in section 5; e.g., Yoshikado 1990, 1992), internal wind and convection variations due to nonstraight coastlines (e.g., Fig. 19-4; Laird et al. 1995) and inland variations in surface heating, resulting in the development of multiple interactions between near-shore river breezes and larger-scale sea breezes (e.g., Zhong and Takle 1992; Laird et al. 1995). Banta et al. (1993) and Banta (1995) introduced the concept of two interacting scales of sea-breeze circulations along the West Coast of the United States. They documented an initial shallow, relatively weak SB circulation that developed in the morning hours due to local surface heating differences. This initial SB was eventually overlain with a deeper, more intense sea-breeze circulation that evolved

FIG. 19-5. Schematic representation of the interaction between the sea-breeze front (SBF) and horizontal convective rolls (HCRs). The SBF is propagating inland, from left to right, into the environment in which HCRs are developing. [From Dailey and Fovell (1999).]

in response to valleys heating farther inland and the cooler regional ocean. Nighttime sea breezes, during which sea breezes propagate into a stably stratified environment over land, were found to develop long-lived borelike disturbances that moved inland for many hours (Sha et al. 1993; Buckley and Kurzeja 1997a), while advances in field techniques allowed for observation of regions of turbulent mixing in the sea-breeze flows (i.e., behind the head of the SB front; Sha et al. 1991; Chiba et al. 1999). External mesoscale flows, such as horizontal convective rolls (Wakimoto and Atkins 1994; Kingsmill 1995; Atkins et al. 1995; Dailey and Fovell 1999) and thunderstorm outflow boundaries (Nicholls et al. 1991; Laird et al. 1995), were identified as having important impacts on the horizontal variations in sea-breeze structure, pollution

mixing, and further thunderstorm development (Fig. 19-5).

It was recognized that the sea-breeze return flow could add a critical complication on pollution dispersion and therefore needed to be carefully studied. Despite the particular difficulty in distinguishing this weak wind system from often stronger large-scale winds, Tijm et al. (1999) made initial horizontal mass flux observations across the coast while Atkins and Wakimoto (1997) provided finescale dual-Doppler 2D wind fields. Understanding of these and other sea-breeze features proved useful in understanding improvements in both air pollution distribution studies (Fig. 19-6; e.g., Steyn and Segal 1990; Moorthy et al. 1993; Lu and Turco 1994; Buckley and Kurzeja 1997b; Kolev et al. 1998) and in convective initiation efforts

FIG. 19-6. (left) Observed and (right) CMAQ-simulated maximum surface 8-h-average ozone on 11 Jul 2011. CMAQ results are from the 1.33-km-horizontal-resolution domain. [From Loughner et al. (2014).]

(e.g., Atkins et al. 1995; Kingsmill 1995; Laird et al. 1995).

c. Recent advancements in applied sea-breeze studies

As the understanding of the complexities of sea breezes, as well as the capability to simulate their evolution in high-resolution numerical models, increased, the most recent few decades exhibited a surge in studies applying understanding of sea breezes to other atmospheric and societal phenomena. We list a few examples here.

Sea breezes, and associated phenomena such as lake and land breezes, play critical roles in the air quality of coastal communities. For example, Lyons and others used multiple observational and numerical techniques to examine the influence of the Lake Michigan lake- and land-breeze circulations near its southern shores (e.g., Lyons and Cole 1976; Keen and Lyons 1978). Positions and heights of coastal emission sources were important to the transport of atmospheric chemical constituents by these coastal circulations at such locations as Tampa Bay (Lyons 1972; Lyons and Olsson 1972; Young and Winchester 1980), New York City (Bornstein and Thompson 1981), Japan cities (Ueda et al. 1988; Yoshikado 1990), Athens, Greece (Clappier et al. 2000; Grossi et al. 2000), the Chesapeake Bay region (e.g., Stauffer et al. 2014), Salt Lake City (e.g., Blaylock et al. 2017), and many others. It was frequently found that multiday processes involving diurnally alternating land and sea breezes were often related to extreme pollution episodes and that slowing of the sea breeze over urban areas could be particularly important in nonuniform air quality across the region.

Coastal atmospheric circulations are often complicated by nearshore nonuniform surface features, such as urbanized areas (e.g., Finardi et al. 2018), nearshore and inland mountainous regions and associated convection (e.g., Kitada et al. 1986; Lu and Turco 1994; Bastin et al. 2006; Chen et al. 2014), nonuniform coastlines, and inland water bodies (e.g., Azorin-Molina et al. 2009; Lombardo et al. 2016; Hughes and Veron 2018). This continues to be a highly active area of research, complicated by unique topographic and coastal environments in areas of interest, the need for very high-resolution simulation techniques, and interactions with local chemical and physical phenomena.

Of course, sea breezes and lake breezes are known to play important roles in both easing heat waves and modulating where severe weather occurs. For example, Soderholm et al. (2016) examined multiple years of radar data to indicate regions where hailstorms are commonly reported in eastern Australia and King et al. (2003) documented regions near the Great Lakes where colliding lake breezes lead to local increases in tornado frequency. Therefore, improving the capability of forecasting such breezes, and particularly their variations in intensity and speed of movement, remain important forecast challenges.

4. Lake-effect snow systems

a. The first 50 years

Large lakes, especially those in midlatitudes exposed to seasonally extreme temperatures, alter their regional

FIG. 19-7. Heavy lake-effect snow in February 2007. (Credit: Sylwia Kapuscinski/The New York Times/Redux; www.nytimes.com/2007/02/12/nyregion/12snow.html.)

climate in ways that frequently produce severe weather conditions. A noted example is the Laurentian Great Lakes, which are responsible for such extremes as corridors of tornadoes (see, e.g., [King et al. 2003](#)) as well as some of the most extreme winter weather seen in North America (e.g., [Kristovich et al. 2017](#)). This section considers the development of understanding of cold-season lake-effect systems, which can produce damaging winds, flooding rains, ice storms, and very deep snowfalls (Fig. 19-7).

By the early twentieth century, increasingly detailed scientific investigations were driven by the topics with the greatest impacts on society in the Great Lakes region at that time: water transportation and agriculture. Great Lakes weather summaries appeared in *Monthly Weather Review* before and in the first decades after the formation of AMS. Such summaries regularly focused on impediments to shipping, such as strong winds (e.g., [Conger 1908, 1910](#)) and lake ice cover (e.g., [Ten Broeck 1900](#); [Conger 1908, 1910](#)) and its variability between lakes, seasons, and years. A review of these early articles also revealed considerable interest in local temperature and winds near both small (e.g., Lake Mendota, Wisconsin; [Bartlett 1905](#); [Wing 1943](#)) and large lakes. Of particular note is work of [Day \(1926\)](#), who developed one of the first regional investigations of the effects of large lakes on precipitation through compilation of climate data for every station in the region near Lake Michigan. A few years later, [Odell \(1931\)](#) sought to examine the influence of Lake Michigan on a wider range of climatic variables, driven by an interest in fruit agriculture. Odell documented the greatly increased chance of killing temperatures for peaches to the west, as compared to the east, where the lake decreased the

temperature range. In addition, winds were greater over and downwind of the lake and cloudiness and precipitation distributions were greatly influenced, as noted earlier by [Day \(1926\)](#). Subsequently, numerous detailed climatological studies were carried out to provide insight into larger regions of the Great Lakes (e.g., [Eichenlaub 1970, 1979](#); [Scott and Huff 1996](#)).

Perhaps of particular interest to communities near the lake shores is the increase in snowfall due to the lakes, called “lake snow” by [Williams \(1963\)](#). Williams identified the primary reasons for lake snow, atmospheric destabilization by the relatively warm lake surface and enhanced water vapor made available to the atmosphere. Concepts of the importance of wind velocity and near-surface convergent enhancement of lake snows were also pointed out by [Williams \(1963\)](#). Such concepts were further quantified by such authors as [Wiggin \(1950\)](#), [Changnon \(1968\)](#), and [Brunk \(1962\)](#).

While considerable spatiotemporal variation in snowfall had become well known, understanding of the dynamics of mesoscale lake-effect snowbands was fairly limited [with such notable exceptions as [Johnson and Mook \(1953\)](#) and [Peace and Sykes \(1966\)](#)] until the increased use of remote sensing for meteorological purposes. For example, [Williams \(1963\)](#) showed evidence by surface-based radar observations that snowbands were present well before reaching the shore. Satellite observations further showed the wide range of convective structures present on the lakes in lake-effect conditions (e.g., [Ferguson 1971](#); [Holroyd 1971](#)).

The 1970s saw a large increase in studies devoted to understanding processes giving rise to lake-effect storms, reflecting improvements in understanding through rapidly improving capabilities in observational techniques

and advancing 3D mesoscale modeling capabilities. Observations using aircraft-borne instrumentation allowed for documentation of how the lake-effect boundary layer grew across a lake (Lenschow 1973; Braham and Kelly 1982). Mesoscale modeling efforts, such as Rao (1971) and Lavoie et al. (1970) examined processes ranging from snow particle evolution to development of several types of lake-effect snowbands. Magaziner (1973) even considered the possible changes artificial cloud seeding could make on lake-effect systems. Of particular significance is the work of Lavoie (1972), who built a numerical model on the concept of lake-effect systems as layered atmospheric boundary layers. As will be seen, this study helped spark work in simulating lake-effect systems for many subsequent years.

b. Climatic variability of lake-effect snows

Because of the large impacts that snowfall has on natural and societal features in mid and high-latitude lake regions, there is much interest in whether there are long-term trends in such snowfall and what controls interannual variability; issues that remain not fully understood. One of the major difficulties is that snowfall near lakes can be due to non-lake-effect systems (such as frontal systems and cyclones), lake-effect-only cases, whereby no snow would be expected if the lakes were not present, and lake modification of precipitation from non-lake-effect systems (lake enhanced).

Several methods have been used to examine long-term trends. Most studies compared trends in snowfall in typical lake-effect snowbelt regions (say, within 50 km of the lake shores) to those in nearby areas that are thought to have little influence from the lakes. Using this method, Eichenlaub (1970) found more than doubling of seasonal snow at lake-effect sites from the early 1900s–10s to the late 1950s. Braham and Dungey (1984) found that seasonal snowfall in the Lake Michigan snowbelt increased from the 1930s to the 1970s but remained nearly constant in other nearby areas. Similarly, Norton and Bolsenga (1993) found that in the central region of the Great Lakes and areas near Lake Ontario, snowfall increased between 1951 and 1980, while little trend was seen west of the Great Lakes. Ellis and Johnson (2004) and Burnett et al. (2003) noted similar increases in snowfall in the Great Lakes snowbelts, which the latter related to decreasing ice cover and warming surface temperatures of the lakes. However, later studies found that such trends were not persistent. Kunkel et al. (2009) also reported increases of snowfall in the Lake Michigan–Huron snowbelts, but their figures showed a slowing or reversal of those trends by the late 1990s.

More recent studies indicate the upward trend may be changing. By subtracting snowfall amounts outside of

the Lake Michigan snowbelt from those west and east of the snowbelt region, Bard and Kristovich (2012) found two estimates of the lake contribution to snowfall in three latitude bands. This analysis revealed an upward trend until the 1970s–early 1980s followed by a strong reversal to the early 2000s. Such reversals in snowfall were also reported for other regions of the Great Lakes (Hartnett et al. 2014; Clark et al. 2016; Suriano and Leathers 2017) often, but not always, through contributions to increased lake-effect rains (e.g., Notaro et al. 2015; Bajinath-Rodino et al. 2018). Other common methods including determining the frequency that synoptic patterns or other weather conditions found to be conducive for lake-effect snows changed over time (e.g., Suriano and Leathers 2017; Bajinath-Rodino et al. 2018).

Climate simulations for latter parts of the twenty-first century generally predict decreases in lake-effect snowfall, as the region continues to warm (e.g., Kunkel et al. 2002; Notaro et al. 2014), possibly after shorter-term increases as ice cover on the lakes diminish (e.g., Suriano and Leathers 2016).

It is noted that quite a few important studies have examined the interannual variability of lake-effect snows and/or precipitation and their causes. These are not summarized in this chapter, however.

c. The lake-effect boundary layer

A considerable difficulty in studying detailed characteristics of lake-effect events is that much of the evolution of the storm takes place over large water bodies, far from operationally available observation sites. Radar (e.g., Steenburgh et al. 2000; Alcott et al. 2012; Yeager et al. 2013; Veals and Steenburgh 2015), satellite (Kelly 1986; Kristovich and Steve 1995; Rodriguez et al. 2007; Ackerman et al. 2013; Laird et al. 2016), and other operational remote sensing techniques (e.g., Steiger et al. 2009) provide the opportunity to examine large numbers of lake-effect events with increasingly precise data, but are limited in the number of variables that are measured. A series of short-term field experiments obtaining data from multiple platforms simultaneously have been conducted to obtain detailed information on the lake-effect boundary layer and many other features that could not otherwise be observed: The University of Chicago Lake Snow Project (1970s and 1980s; Braham and Kelly 1982), Lake Ontario Winter Storms (LOWS 1990; Reinking et al. 1993), Lake-Induced Convection Experiment (Lake-ICE 1997/98; Kristovich et al. 2000), Ontario Winter Lake-effect Systems (OWLeS 2013/14; Kristovich et al. 2017), and numerous other more focused experiments (e.g., Gerbush et al. 2008; Steiger et al. 2013). These projects, in combination with rapidly

FIG. 19-8. Rapid development of a lake-effect boundary layer. (From bradhoc at <https://www.flickr.com/photos/bradhoc/11799104264> under the creative commons license at <https://creativecommons.org/licenses/by/2.0/>. Some cropping of the image was done to save space.)

advancing numerical modeling techniques, have led to advancements in understanding many aspects of lake-effect storms.

As noted earlier, it was recognized early that lake-effect snowstorms are manifestations of internal boundary layers that develop over relatively warm waters of large lakes during the cold season (e.g., Lavoie 1972), causing rapid convective growth (Fig. 19-8). Since that time, multiple studies have examined lake-effect boundary layers in order to develop a better understanding of convective boundary layer thermodynamic properties. Such fundamental issues examined in lake-effect systems include surface flux magnitudes (e.g., Gerbush et al. 2008; Fujisaki-Manome et al. 2017), mass overturning processes and rates in the mixed layer (e.g., Agee and Hart 1990; Braham and Kristovich 1996; Rao and Agee 1996), internal boundary layer growth and heat/moisture budgets (e.g., Chang and Braham 1991; Kristovich and Braham 1998; Kristovich et al. 2003), and entrainment indirectly driven by surface fluxes (Young et al. 2000). A particularly interesting finding from Zurn-Birkhimer et al. (2005) was the apparent presence of a thermal-internal boundary layer and a deeper moisture-internal boundary layer. Such layers may be quite important to understanding convective boundary layer growth and ice particle growth near the entrainment layer.

While it has been known that the area of convergence near the downwind shore, due to the surface winds moving from water to a rougher land surface area,

contributed to locally increased boundary layer growth, rather little work has quantified the influences of divergence near the upwind shore (e.g., Kristovich and Laird 1998; Zurn-Birkhimer et al. 2005). Kristovich et al. (2018) recently provided evidence of the importance of upwind boundary layer conditions (in this case, a modified boundary layer from Lake Erie) on the lake-effect boundary layer over Lake Ontario and developed a conceptual model of how the upwind shore divergence area may influence this process (Fig. 19-9). Moreover, developing lake-effect systems have the additional near-shore influences of heat transfer and particle loading through cloud and precipitation processes and associated radiative property variations (e.g., Chang and Braham 1991), topics of continued significant uncertainty.

d. Mesoscale features in lake-effect systems

Of the various size and time scales involved in the lake-effect storm process, perhaps the feature most directly associated with determining which communities receive the heaviest snow and for how long is its mesoscale structure. The mesoscale structure and evolution of lake-effect precipitation bands are influenced by many factors, including lake–air temperature differences, wind direction relative to the lake orientation, nearshore topography, convective intensity, static stability, wind changes with height (wind shear), and so on. However, as long as conditions are appropriate for lake-effect development, the local factors most influencing the overall structure of lake-effect

FIG. 19-9. Conceptual model of turbulent eddies as they moved from the boundary layer upwind of the Lake Ontario coastline (left), across the more shallow boundary layer near the land-water transition, and then to the lake-effect boundary layer over Lake Ontario (right). The red line indicates the approximate top of the PBL, as well as its relative variability. The schematic is stretched vertically to make the changes more evident. Relative thickness of the turbulent eddies (ovals with arrows) indicates the relative magnitude of the vertical motions. [From Kristovich et al. (2018), caption revised.]

storms have generally been identified as the wind velocity, intensity of surface heat fluxes, and lake shape relative to the wind direction (e.g., Fig. 19-10 from Hjelmfelt 1990; Niziol et al. 1995; Laird et al. 2003; Laird and Kristovich 2004, and many others).

Most classification schemes of mesoscale lake-effect convective structures over individual lakes in the literature tend to fall into three categories: 1) long-lake-axis parallel bands (LLAP, Fig. 19-10, columns II and III, also called lake-parallel bands and type I bands), 2) short-lake-axis parallel bands (SLAP; Fig. 19-10, column I, also called broad-area coverage cells and cloud streets, roll convection, wind-parallel, type II bands, etc.), 3) and vortices (e.g., Fig. 19-10, column IV; Hjelmfelt 1990; Niziol et al. 1995; Laird et al. 2003; Laird and Kristovich 2004). In addition to these fundamental types, numerous subcategories have also been identified, such as lake-to-lake (L2L) bands (multiple-lake bands, type III bands, etc.) and land-breeze shore-parallel bands (type IV bands). Analyses of the daily frequency with which these types of lake-effect bands are present over the Great Lakes region generally show that SLAP bands are most common, followed by LLAP bands and vortices (Kelly 1986; Kristovich and Steve 1995; Rodriguez et al. 2007; Ackerman et al. 2013; Laird et al. 2017). Combinations of, or evolution between, different types of bands also occur very frequently, as do lake-effect events without clear mesoscale organization (Kelly 1986; Kristovich and Steve 1995; Rodriguez et al. 2007; Ackerman et al. 2013; Laird et al. 2016).

In the most recent few years, observations of finescale details of lake-effect storms taken during the OWLeS

experiment, particularly LLAP bands, has provided insights into mesoscale features of lake-effect storms (e.g., Steiger et al. 2013), circulations within and along lake-effect snowbands (e.g., Bergmaier and Geerts 2016; Welsh et al. 2016; Bergmaier et al. 2017; Mulholland et al. 2017; Steenburgh and Campbell 2017), and interactions of lake-modified air with surrounding land characteristics and topography (e.g., Veals and Steenburgh 2015; Minder et al. 2015; Campbell et al. 2016; Campbell and Steenburgh 2017; Veals et al. 2018; Eipper et al. 2018; Kristovich et al. 2018).

e. Forecasting of lake-effect systems, synoptic influence, small lakes

As forecasting techniques improved in the 1970s, multiple studies were conducted to best identify large-scale atmospheric conditions conducive to the most intense storms. Dewey (1975) asserted that thresholds of surface heat fluxes could be of particular benefit to forecasting lake-effect storms. However, since his formulation took into account both surface heating processes and the influences of stability on limiting lake-effect storm depth and intensity, a threshold value of lake surface to 850 hPa temperatures was adopted more readily by forecasters (e.g., Rothrock 1969; Holroyd 1971). As operational numerical simulations and understanding of lake-effect storms improved, Niziol (1987) and Niziol et al. (1995) provided detailed overviews of forecasting techniques employed at those times. Many regional and local forecasters employ increasingly sophisticated numerical modeling systems (e.g., Saslo and Greybush 2017) and techniques

FIG. 19-10. Morphological types of lake-effect snowstorms: (i) broad area; (ii) shoreline band, (iii) midlake band, and (iv) vortex. (a) Satellite pictures of examples of each type (photos courtesy of R. R. Braham, Jr.), (b) schematic depiction, and (c) numerical simulation examples. Grid spacing is 8 km and results are shown for 20 h simulated time except for (iii) at 9 h. Vectors are scaled to $1 \text{ km} = 1 \text{ m s}^{-1}$, shading is cloud cover at 1500 m height. The contour shows vertical velocity $>20 \text{ cm s}^{-1}$ at 1 km. Heavy dark contour is Lake Michigan shoreline. [From [Hjelmfelt \(1990\)](#).]

designed specifically for their areas of interest (e.g., lake snow parameter; [Smith and Boris 2008](#)). Such efforts have led to significant increases in the ability to predict where and with what intensity lake-effect storms will occur (e.g., [Ballentine 1982](#); [Ballentine et al. 1998](#)), but the need for continued sharing of experiences in forecasting these systems remain (e.g., [Kristovich et al. 2017](#)).

Influences from more than one lake, when they are sufficiently close together, can combine to alter lake-effect precipitation locations, modify larger-scale flow fields, and intensify snowfalls. Fundamental fluid dynamics theory identified that areas of warm surface

temperatures would have an important influence on distortion of the wind fields. [Pettersen \(1956\)](#), noting the expected relationship between surface heating and cyclogenesis, found that “all major inland open water surfaces are active cyclogenetic areas during the cold season” ([Rao 1971](#)). In a paper that ultimately proved to be quite important to understanding wintertime modification of the regional weather by the Great Lakes, [Pettersen and Calabrese \(1959\)](#) showed that the Great Lakes as a whole tended to enhance low pressure over the region. Understanding of the feedback of developing cyclonic vorticity over the lakes to changes in snowfall throughout the region improved significantly in the next

two decades (e.g., Rao 1971; Danard and Rao 1972; Danard and McMillan 1974). These were followed by a series of studies by Sousounis and colleagues that added much insight into the combined effects of the Great Lakes and how these modifications, in turn, affect local lake-effect precipitation characteristics (e.g., Sousounis and Shirer 1992; Sousounis and Mann 2000; Mann et al. 2002).

On a smaller scale, lake-effect snowbands stretching from one lake to another could lead to greatly enhanced snowfalls over and downwind of the second lake. Cloud and snowbands stretching from one lake to another are frequently seen over the eastern Great Lakes region (e.g., Rodriguez et al. 2007; Laird et al. 2017) and play an important role in forecasting (e.g., Niziol 1987; Byrd et al. 1991; Niziol et al. 1995). Conditions associated with lake-to-lake lake-effect influences include rapid changes in the lake-effect boundary layer depth and ice particles carried by wind over the downwind lake (e.g., Kristovich et al. 2018).

Much knowledge has been accumulated on lake-effect processes based on studies of large lakes (such as the North American Laurentian Great Lakes), but far less has been learned about atmospheric responses to the much more common smaller lakes. Mesoscale lake-effect convective structures on smaller lakes have received much less attention but have been identified for midlatitude lakes as small as a few tens of kilometers. Climatological studies of lake-effect convection over the Great Salt Lake in Utah (e.g., Steenburgh et al. 2000), Lake Champlain on the Vermont–New York border (e.g., Laird et al. 2009), Lake Tahoe on the border of Nevada and California (e.g., Laird et al. 2016), and the New York Finger Lakes (e.g., Laird et al. 2009, 2010) indicate that these lakes can produce snow multiple times each year, usually under more constrained conditions than needed for large lakes (e.g., Carpenter 1993). Compared to lake-effect storms over the Great Lakes, small lake storms tend to occur less frequently, over a lesser range of environmental conditions (especially lake–air temperature differences and low-level wind direction), and have fewer mesoscale band types. Such studies have provided valuable information on how lake-modified boundary layers interact with topography (e.g., Alcott and Steenburgh 2013) and nonuniform water characteristics (e.g., Onton and Steenburgh 2001), as well as their hydrologic impacts (e.g., Yeager et al. 2013).

5. Urban weather and climate

It has been known since before the founding of AMS that urban areas have large impacts on the local weather

and climate. For example, constructed buildings and other infrastructure act as resistance to the low-level airflow and heat sources in urban areas modify the boundary layer structure. These modifications are due to changes in the type of vegetation and land cover, the presence of streets, buildings, and parking lots, surface albedo, and anthropogenic heat (e.g., waste heat from air conditioners, power plants, transportation, and manufacturing). However, detailed, quantitative investigations of urban modifications of lower-level atmospheric flows, turbulence, temperature, precipitation, and humidity, as well as their impact on transport and dispersion of pollutants in the boundary layer, have only been explored in detail over the last few decades. Moreover, the increasing pace of urbanization has warranted the need for improvements in forecast model predictions, to better prepare for short- and long-term impacts of climate change on health and well-being, as urban areas become larger, hotter and more polluted.

a. A conceptual framework for the urban boundary layer

Before discussing the historical evolution of our understanding of this system, it is useful to provide an overall summary of current understanding of the urban boundary layer (UBL). The UBL is the portion of the ABL whose characteristics are modified by the presence of a city at the surface. It is the most complex atmospheric layer as it responds to many morphological elements, such as buildings, streets, parks, trees, transportation, humans, and green infrastructure. Under quiescent conditions, when there is no synoptic flow and the background flow is weak/absent, the influence of the city is restricted to a self-contained region of the boundary layer called the urban heat dome (Fig. 19-11a; Findlay and Hirt 1969; Li et al. 2015; Fan et al. 2017). The urban heat dome circulation patterns are induced by horizontal pressure gradients between urban and rural areas, and inhomogeneous convective plumes impinging on the surrounding atmosphere. Thus, the urban dome flow is characterized by flow convergence close to the surface of urban areas, updrafts within the UBL and flow divergence near the UBL/free atmosphere interface. Under synoptic conditions with background atmospheric flow, the internal UBL grows upward from the rural–urban interface up to the entrainment zone at the mixed-layer top and spreads downwind (Fig. 19-11b; Oke 1988). The UBL in such ambient conditions can form an urban plume (Clarke 1969), which contains thermodynamics effects of urban areas and can transport them away from the urban area along the synoptic flow.

Conventionally, the internal structure of the UBL is divided among several layers over a city or urban center

FIG. 19-11. Schematic showing different types of UBL at mesoscales: (a) urban dome with no ambient wind and (b) typical urban internal boundary layer with an ambient flow causing a formation of an urban plume. Note, here the boundary layer structures are idealized to illustrate concepts. [Figure is modified from [Oke et al. \(2017\)](#). Copyright T. R. Oke, G. Mills, A. Christen, and J. A. Voogt 2017, published by Cambridge University Press. Reproduced with permission of the Licensor through PLSclear.]

(Fig. 19-12; e.g., [Oke 1988](#)). Figure 19-12a illustrates the daytime progression of the internal UBL structure. It includes an urban canopy layer (UCL), which extends from the ground to about the building roof-level heights. The UCL is strongly dominated by local site characteristics and near-field dispersion of atmospheric constituents along with turbulent kinetic energy (TKE) production and transport processes. The roughness sublayer ranges from above the UCL to a height where TKE production and dissipation balance each other; typically, of thickness 2–5 times average building height. The remainder of the UBL is made up of an inertial sublayer and a mixed layer. The inertial sublayer (also called the constant flux layer) is a horizontally homogeneous region located beyond the height where building wakes are observed. Last, the mixed layer tops the UBL and is the layer directly in contact with the free atmosphere above. The mixed layer is maintained by entrainment of free atmospheric air parcels and is often a location of TKE production.

At night, vigorous buoyant and mechanical turbulent mixing present during the day is diminished due to

regional and local stabilization of the ABL. Thus, the depth of the nocturnal UBL is lower and varies based on the strength of the UHI and urban roughness effects (Fig. 19-12b). However, local sources of heating and TKE production as near-surface winds interact with the rough surface typically maintain a shallow well-mixed layer that varies with urban land use. For sparsely populated urban areas, the cooling during the night restricts the UBL height, while densely built-up urban areas show elevated inversion bases.

The UBL determines the heat and moisture exchange in urban areas and hence impacts the local micrometeorology and air quality ([Miao et al. 2009](#); [Zhang et al. 2011](#); [Sharma et al. 2016, 2018](#)). Therefore, the simulation and analysis of urban temperature and moisture must be accompanied by careful treatment of the UBL ([Fernando et al. 2001](#); [Fernando 2010](#)). As we will see in the following sections, the complex UBL structure is relatively hard to capture with observations due to measurement and logistical difficulties ([Frehlich et al. 2006](#); [Barlow 2014](#)), and hence many recent studies have attempted to use complex urban models ([Fan and Sailor](#)

FIG. 19-12. Schematic of detailed characteristics of an ambient urban boundary layer at mesoscales during (a) daytime and (b) nighttime. [Figure is modified from Oke et al. (2017). Copyright T. R. Oke, G. Mills, A. Christen, and J. A. Voogt 2017, published by Cambridge University Press. Reproduced with permission of the Licensor through PLSclear.]

2005; Sharma et al. 2014, 2015, 2016, 2017; Conry et al. 2014, 2015).

Ultimately, the UBL develops through vertical and horizontal transfers of heat, moisture, and momentum, on microscales. Considerable effort has been made on flows specific to urban environments, such as urban canyon flows (winds down rows of buildings, for example) and individual heat sources (e.g., Oke 1997; Belcher 2005).

b. The inception and rise of urban atmospheric science

Luke Howard (1833), a pioneer of urban climatic studies, recognized the influence of urban areas on the ABL, including their ability to retain more heat relative to surrounding rural areas (Chandler 1965). Excerpts from Howard's (1833) book *The Climate of London* illustrates the understanding of urban climate in London at that time:

Showery summers must be by far the most comfortable, on the whole, to the inhabitants of a great city, provided the rainfall in considerable quantities at a time. Not to mention the annoyances of heat and dust in the streets, the very atmosphere of cities must need this cooling and

washing, being sensibly purified by great rains; so that the sulfurous and other effluvia are no longer so perceptible. I was much struck with the differences in the air of London, in these respects, in the wet weather of the present summer.

Howard developed an early understanding of the UBL based on observations that were quite limited (e.g., meteorological sensors, photographs of urban profiles and clouds from land and sea). Even though Howard never took simultaneous measurements at different sites in London and surrounding rural environments, by taking observations in each environment over different times, he correctly deduced an urban influence. He showed that the mean temperature in London, United Kingdom, was 1.57°C warmer than the countryside over a 10-yr period from 1807 to 1816 (Fig. 19-13). For the first time, he identified UHI phenomena as a difference between the measurements of the Royal Society in the center of London (T_u) and his meteorological station maintained outside the city of London (T_r), which is still fundamentally how UHIs are determined today. Since Howard's work in the year 1833, there has been a substantial evolution in taking diverse approaches for UBL effect studies as described below.

In the early decades of the nineteenth century, scientists described urban effects based on observations collected using conventional meteorological equipment as highlighted by Kratzer (1956). By the mid-1950–60s, statistical methods were predominant, and results were generalized across a wide range of regions (Duckworth and Sandberg 1954; Sheppard 1958; DeMarrais 1961; McCormick and Baulch 1962; Munn and Stewart 1967; Davidson 1967; Bornstein 1968). These analyses, as well as increases in understanding of boundary layer processes in nonurban environments [such as in Kansas (Haugen et al. 1971; Kaimal et al. 1971; Readings and Butler 1972) and in Minnesota (Kaimal et al. 1976)] led to a focus on determining why UBLs behaved differently.

Shortly after the Kansas and Minnesota ABL studies, a field campaign was designed to study New York urban air pollution dynamics and examine transport and diffusion processes with an anemometer and the upper-level wind network (Davidson 1967). This effort was aided by the development and deployment of measurements from airborne platforms (a helicopter and an airplane), which included pressure, temperature, and SO₂ pollutant concentrations. The measurements showed a high frequency of weak elevated inversion layers, above UHI over urban areas as high as 500 m (Fig. 19-14; Bornstein 1968). He also suggested that a heat reservoir over urban areas lead to vertical motions and compensating low-level convergence into the urban

FIG. 19-13. Time series of London mean temperature for 10 years (1807–16). The solid line is for London and dotted line for surrounding country area. [From Howard (1833).]

area. Such flows were also observed when geostrophic winds were weak and there was a quiescent state of the atmosphere (Okita 1960; Pooler 1963). During the same time, Munn and Stewart (1967) compared the vertical distribution of air temperatures in rural and urban locations in Ontario. They found minimal daytime differences between these profiles. However, more frequent

inversion-free nights were observed over the urban areas, which they attributed to nocturnal UHI effects.

c. METROMEX and growth of urban atmospheric models

A large field research project in the Saint Louis, Missouri, area, employing large numbers of observational

FIG. 19-14. Vertical and horizontal temperature distribution over a rural-urban cross section of New York area on 16 Jul 1965 showing explicit urban heat island formation. The figure also shows inversion layers in shaded contours. [From Davidson (1967); reprinted by permission of the Air & Waste Management Association, www.awma.org. Reprinted by permission of the publisher (Taylor & Francis Ltd, <http://www.tandfonline.com>).]

instrumentation, is often cited as having greatly enhanced the growing field of urban atmospheric studies. In the 1970s, urban meteorological theories were developed to scale urban processes using an energy budget framework. In this regard, the Metropolitan Meteorological Experiment (METROMEX; [Changnon et al. 1971](#); [Changnon 1981](#)) in Saint Louis was a major multiyear U.S. campaign to study how a large metropolitan area in a humid continental climate affects the local climate at multiple spatial and temporal scales. They investigated radiation components, temperature, humidity, wind fields, and aerosol concentrations in the boundary layer over an urban area. Following the findings of [Changnon \(1969\)](#) that summertime thunderstorms enhanced precipitation amounts downwind of urban areas, METROMEX also studied local climate modifications on precipitation due to urban–industrial effects. In particular, anthropogenically induced moisture convergence over urban areas using multiple rain gauges, sophisticated rain radars, aircraft flights, and other surface-based instruments provided information on the processes of cloud and precipitation formation, the chemistry of aerosols and rainwater, the urban heat budget, the UBL diurnal evolution, and the three-dimensional patterns of precipitation in an atmospheric column. Recent reviews from [Shepherd \(2005\)](#) highlight the scientific importance of METROMEX in advancing our understanding of thunderstorm precipitation for urban areas with storms normally initiated during strong UHI periods and bifurcated during weak UHI periods. This has been validated from recent UHI observational studies in Beijing ([Dou et al. 2015](#)).

As a follow-up to METROMEX, the Regional Air Pollution Study (RAPS) in Saint Louis evaluated the interrelated processes in regional air quality, meteorology, and pollutant emissions ([Schiermeier 1978](#)). This time, they focused on measurements of surface and mixed-layer turbulent transport and diffusion processes for different land-use types.

Meanwhile, for the first time, analytical and numerical solutions for the nighttime UHI and boundary layer using idealized boundary conditions were developed ([Delage and Taylor 1970](#)). They successfully simulated the two-dimensional wind flow for a vertical cross section of the heat island, using prescribed surface temperatures and no external winds. This work was extended for a nonsteady two-dimensional model that successfully reproduced the daytime flow for a neutral atmosphere over a rough city, and the nighttime flow for a stable atmosphere over a rough, warm city ([Bornstein 1975](#)). Excellent resources like [Yoshino \(1975\)](#) and [Landsberg \(1981\)](#) provide a timely state of the knowledge on urban meteorology during this period.

Studies in the 1970s–90s could be grouped into three primary areas of interest: improved documentation and modeling of turbulent and radiative heat transfer processes, urban influences on air quality, and growth in the understanding of UHI influences on larger-scale atmospheric features. Advancements in field capabilities (e.g., mobile tethered sonde system—[Tapper 1990](#); sodar—[Melling and List 1980](#); lidar—[Kunkel et al. 1977](#); [Cooper and Eichinger 1994](#); [Casadio et al. 1996](#)) led to direct measurements of fluxes and atmospheric flows that were, in turn, used to test hypotheses using scaled urban models. For example, [Auer \(1981\)](#) documented how urban and rural areas differed in energy and radiation balances. [Oke \(1988\)](#) and [Ching et al. \(1983\)](#) showed that the peak values of surface sensible heat flux occur over urban areas and the flux remains positive through all or part of the diurnal cycle. Meanwhile, observations in Sapporo, Japan, revealed nighttime strong and elevated UBL inversion formation during winters with inversion base height approximately twice the average building height ([Uno et al. 1988, 1989](#)). They also found, through analysis of the TKE budget, the important role of the turbulent transport and shear generation from the urban canopy in the nocturnal UBL development. Numerical modeling advancements built on these observations to assess how turbulence transfer models developed in Kansas and Minnesota could be modified for urban systems. Pioneering two-dimensional simulations of the UBL with the urban meteorology (URBMET) model ([Bornstein 1975](#)), [Oke et al. \(1989\)](#) used a one-dimensional idealized constant flux layer approach (e.g., [Stull 1988](#)) to link surface turbulent fluxes and profiles in urban areas. In the early 1990s, [Grimmond and Oke \(1991\)](#) developed a nonlinear model to predict the storage heat flux in urban areas by parameterizing changes in heat storage as a function of net radiation.

Urban air quality and related pollution transport were also investigated extensively in the 1970s–90s. The Southern Oxidants Study in Nashville, Tennessee, used an airborne differential absorption lidar (DIAL) system to profile daytime O₃ buildup in the UBL for a strong, synoptic-scale stagnation period, a phenomenon when an air mass remains over an area for an extended period ([Banta et al. 1998](#)). The Etude de la Couche Limite dans l'Agglomération Parisienne (ECLAP) experiment used sodars, lidars, and surface measurements to study the vertical structure, diurnal evolution of the ABL, UHI effect, and pollution over Paris ([Dupont et al. 1999](#)).

Finally, interactions between urban areas and larger-scale phenomena were identified. [Loose and Bornstein \(1977\)](#) and [Bornstein and Thompson \(1981\)](#) documented that frontal movement during non-UHI periods

was retarded significantly over the entire central urban area due to the increased surface frictional drag exerted on the front by the increased surface roughness of the city as compared to that of its surrounding areas. However, during UHI periods there was a retardation in frontal speed over the upwind half of the city, followed by a significant acceleration of the front over its downwind half. In addition, relationships and feedbacks between the boundary layer and the regional climate began to be addressed efficiently with computer models. With the increase in computational capability, non-hydrostatic assumptions-based models with spatial resolutions <10 km scales were developed using the Met Office model (Tapp and White 1976; Carpenter 1979) and the fifth-generation Pennsylvania State University–NCAR Mesoscale Model (MM5) (Grell 1993; Grell et al. 1994).

d. Use of integrated observations and models since the year 2000

The rapid progress in urban and, in general, ABL modeling and observations are primarily attributed to advancements in affordable parallel processing, and an urgent need to address air pollution and urban heat for increasing populations. To find solutions for these pressing urban issues, researchers began to assess the model results rigorously, and identify deficiencies in models and field studies. The knowledge gaps in urban model development and lessons learned from previous field studies on boundary layers over urban and rural environments sparked new and extensive urban field campaigns over the last 20 years. During this time, there have been many UBL field campaigns, similar to METROMEX, across the globe. A few notable urban field campaigns are the Basel Urban Boundary Layer Experiment (BUBBLE) in Switzerland (Rotach et al. 2005); Expérience sur Sites pour Contraindre les Modèles de Pollution Atmosphérique et de Transport d'Emission (ESCOMPTE) in Marseille, France (Cros et al. 2004; Mestayer et al. 2005); and Investigación sobre Materia Particulada y Deterioro Atmosférico-Aerosol and Visibility Research (IMADA-AVER) Boundary layer experiment in Mexico City, Mexico (Doran et al. 1998). Many field campaigns were also organized in the United States during this time, for example, Vertical Transport and Mixing (VTMX)–Urban 2000 in Salt Lake City (Allwine et al. 2002; Doran et al. 2002); 2001 Phoenix Sunrise Experiment in Phoenix, Arizona (Doran et al. 2003); and Joint Urban 2003 in Oklahoma City, Oklahoma (Allwine et al. 2004). These campaigns were designed to study different urban configurations and spatial scales (ranging from indoor/building or canyon, neighborhood, city, regional) and

used a suite of instruments (station observations, towers, lidars, sodars, radars, aircraft, and satellites).

The 2000s also saw an increase in the number of long-term experiments aimed to provide information on the UBL in a wider range of environmental conditions. For example, since 2005, a long-term mesoscale weather observational network in southern Finland called the Helsinki Testbed is operational and contributing to advance the understanding of mesoscale weather phenomena (1–10 km), urban and regional modeling, and applications in a high-latitude coastal environment (Dabberdt et al. 2005; Koskinen et al. 2011). Similarly, a Winter Fog Experiment (WIFEX) over the Indo-Gangetic Plains is aiming to develop better nowcasting and forecasting of winter fog on various temporal and spatial scales for urban and rural areas (Ghude et al. 2017; Pithani et al. 2018). There is also an impetus for long-term instrumentation of cities (such as CCPN 2018) and creating a cyberinfrastructure of urban measurements. In this regard, a flexible urban-scale network of instruments is deployed in Chicago, called the Array of Things (AoT), to create an “instrumented city” with a platform that is driven by urban science needs (Catlett et al. 2017) (Fig. 19-15).

Parallel processing and increased-efficiency supercomputers have also allowed for substantial improvement in urban parameterizations for models including a larger number of grid points within the UBL (Martilli 2007). Such models can account for the heterogeneous and complex urban surfaces that add to the urban complexity (e.g., buildings, trees, streets, roads, pavements, and elsewhere; Fig. 19-16). Thus, multiple urban schemes of varying physical complexity (based on the application and need) have been developed. These suites of different urban physics formulations were recently compared in an international urban model comparison project to evaluate the strengths and weaknesses of the current urban models (Grimmond et al. 2010; Lindberg and Grimmond 2011). The project identified that urban schemes could capture the net all-wave radiation but not the latent heat flux, and a poor choice of urban parameters can adversely affect the model performance.

Currently, mesoscale models are widely used to provide detailed spatiotemporal information on urban systems related to operational forecasting, future long-term climate projections and evaluating urban mitigation strategies. The mesoscale models broadly follow two approaches to treat urban processes in regional climate simulations (Taha 1999; Kusaka et al. 2001). The first approach is to use a slab model to modify surface heat/energy balance through improved surface characteristics (e.g., heat capacity, thermal conductivity, surface

FIG. 19-15. Array of Things (AoT), a networked urban sensor project to understand cities and urban life showing (a) locations with coverage of interactive, modular sensor boxes that are installed around Chicago to collect real-time data on the city's environment, infrastructure, and activity for research and public use; (b) wireless connected sensor network; and (c) a schematic of mounted AoT sensor with node components and its connection to server database for data dissemination. [Credit: Charles Catlett, Luc Anselin, Julia Koschinsky, Marynia A. Kolak, Anaïs Ladoy, <https://spatial.uchicago.edu/>.]

albedo, and roughness height) to better represent urban areas (Liu et al. 2006; Miao et al. 2007; Chen et al. 2011; Giannaros et al. 2013; Sharma et al. 2017). A slab model treats urban geometries as flat surfaces with large roughness lengths and small albedo, that is, it assumes urban surfaces such as buildings, roads and concrete parking spaces have the same temperature. This is a common and computationally inexpensive approach that provides a reasonable estimate of overall UBL dynamics.

A disadvantage of the slab approach is its inability to represent the heterogeneities present in the urban areas due to the variability of urban morphology at fine spatial scales. The second approach is to allow finescale urban surface morphology to impact UBL fluxes directly, thus providing more flexibility and details about the processes within the boundary layer. One such approach is to use a single-layer urban canopy model (SLUCM) (Kusaka et al. 2001; Kusaka and Kimura 2004; Lee et al. 2011). This approach uses an idealized 2D street canyon to explicitly parameterize canyon radiative transfer,

turbulence, momentum, and heat fluxes. It accounts for different urban surfaces: roof, wall, and road. The SLUCM also can assess the impact of anthropogenic heat (AH) fluxes due to air conditioning, transportation, and human metabolism. This explicit coupling allows the representation of urban land surface heterogeneity by dividing the land use into different classes, based on different morphological and thermal properties, to better understand the spatial distribution of urban heat and moisture within the boundary layer. SLUCM, generally, accurately produces enhanced UHI effects, but it does not reproduce the observed slowing of urban wind speeds by the building barrier effects. Alternatively, more sophisticated urban schemes with multilayer urban canopy models (Martilli et al. 2002; Kondo et al. 2005) allow explicit inclusion of 3D building effects within the boundary layer, like anthropogenic heat and the slowing of UBL winds. For example, building effect parameterization (BEP) explicitly treats urban surfaces three-dimensionally, and considers the heat in the buildings and the effects of vertical and horizontal

FIG. 19-16. Infrared images for (a) Phoenix and (b) Chicago showing different urban structures contributing toward urban heating and the need to account for each heterogeneous urban surfaces accurately in urban models. The figure also shows the temperature within the UBL height. (Credit: H.J.S. Fernando.)

surfaces on temperature, momentum, and turbulent energy (Martilli et al. 2002). Recently, BEP was coupled with a multilayer building energy model (BEP + BEM) to account for energy exchanges happening between the interior of a building and the UCL (Salamanca and Martilli 2010). These approaches represent the most recent and sophisticated updates to mesoscale urban parameterizations.

Apart from these urban model developments, TERRA-URB, an urban parameterization (Wouters et al. 2015, 2016), is an attempt to find a unique solution to bridge the gap between previous bulk schemes, for example, urban climate model (UrbClim; De Ridder et al. 2015), Noah/Urban (Kusaka et al. 2001) or integrated urban land model (IUM; Meng 2015), and explicit-canyon schemes [e.g., the Urban Parameterization for the Community Land Model (CLM-U; Demuzere 2014), BEP (Martilli et al. 2002), Town Energy Budget (TEB; Masson 2000), or SLUCM (Kusaka et al. 2001)]. TERRA-URB provides corrections of the surface parameters (e.g., roughness length, albedo, emissivity, and heat capacity) using the semiempirical urban-canopy (SURY) parameterization, which bridges the gap between bulk urban land surface schemes and explicit-canyon schemes. Instead of explicitly calculating the urban physical processes (e.g., the reduced-albedo effect from multiple reflections inside a street canyon), they are implicitly taken into account in the land surface parameters in accordance with detailed observational studies, modeling experiments and available parameter inventories. The model translates urban canopy parameters into urban bulk parameters by accounting for ground-heat transport, surface-radiation exchanges, and surface-layer turbulent transport of momentum, heat, and moisture.

Notable improvements have also been made in recent years on resolving building-block-level urban phenomena and incorporating all available urban complexities (e.g., urban street canyons) in microscale models and evaluating them with extensive field campaigns. Microscale models, in general, include conventional computational fluid dynamics (CFD) frameworks (Coirier et al. 2005; Gowardhan et al. 2011), and non-CFD based models include ecosystem processes such as vegetation, anthropogenic effects, solar forcing (Matzarakis et al. 2007; Lindberg et al. 2008), and thermophysiological effects for human beings (Matzarakis et al. 2007). Other microscale approaches include hybrids of the above approaches to simulate surface–plant–air interactions inside and outside urban structures (Bruse and Fleer 1998; Ma et al. 2012). There have also been developments to account for the energy balance of individual buildings (Crawley et al. 2008) and human comfort for indoor

FIG. 19-17. Daytime near-surface temperature (2 m) contours overlaid by near-surface winds (10 m) showing a shift in Chicago hot spot inland. [From Sharma et al. (2017); copyright 2016 Ashish Sharma, Harindra J.S. Fernando, Alan F. Hamlet, Jessica J. Hellmann, Michael Barlage, and Fei Chen. International Journal of Climatology published by John Wiley & Sons Ltd on behalf of the Royal Meteorological Society. <https://creativecommons.org/licenses/by/4.0/>.]

(Fanger 1970; Mayer and Höppe 1987) and outdoor (Jendritzky 1990; Ma et al. 2012; Conry et al. 2015) environments.

e. Special topics: Coastal UBLs and UBL mitigation

Coastal urban areas are particularly vulnerable to complex environmental conditions (Pullen et al. 2008) and atmospheric circulations leading to extreme weather, air pollution problems and other climate change effects. The recent field experiment ESCOMPTE (Cros et al. 2004; Mestayer et al. 2005) in Marseille, France, observed that the sea breeze tends to weaken UHI development (e.g., De Tomasi et al. 2011; Melecio-Vázquez et al. 2018). Recently, a numerical study by Sharma et al. (2017) showed that the Chicago's coastal urban environments shift the daytime *hot spot* inland by 5–10 km depending on the strength of the lake breeze (Fig. 19-17). Thus, the confounding effects of land surface processes determined by land use and land cover, varying building heights and morphologies, and different surface materials as well as the presence of water bodies (e.g., lake and sea breeze) add to the complexity of UBL over coastal urban environments (Masson 2006; Han et al. 2015; Sharma et al. 2017).

Many urban areas experience ocean/lake breezes that alter the urban energy balance, impact the urban heat and moisture profiles, and contribute to transport and dispersion of airborne contaminants to surrounding

FIG. 19-18. An illustration of vertical cross section of ozone distribution over rural–coastal urban–lake–rural land cover. Ozone (ppmv) in filled colored contours with potential temperature (K) and vertical winds (m s^{-1}) over Chicago region. (Credit: Ashish Sharma.)

regions (see also [section 3c](#)). For example, the changes in surface sensible and latent heat fluxes modify the UBL dynamics with redistribution of meteorological states and pollutant concentrations in the lower atmosphere. [Figure 19-18](#) illustrates the impact of rural–urban–lake land-use classification on the boundary layer. The boundary layer height is higher over urban areas and reduced over the lake and rural area. A vertical cross section from the surface to 3-km height for different meteorological variables (potential temperature and vertical winds) and pollutant concentration (ozone) along a transect covering inland rural and coastal urban areas around the Chicago region shows high ozone concentration over urban areas within the UBL. [Figure 19-18](#) provides an example coastal–urban circulation with a high concentration of ozone within the suppressed lake boundary layer, illustrating the potential complexity of boundary layer flow dynamics in such regions. While a water body can act as a cooling mechanism for a city, it may also deteriorate the air quality over urban areas in some cases. A combined understanding of synoptic, mesoscale, and small-scale processes is crucial in the modeling of urban air quality as well as climatic effects on the atmosphere for coastal urban regions ([IPCC 2007](#)).

Meteorological urban modeling has shown that green infrastructure can help reduce the surface energy emissions and provide relief from excess urban heating ([Georgescu et al. 2014](#); [Li et al. 2014](#); [Song et al. 2018](#)). However, green infrastructure can also suppress the daytime UBL height and adversely impact the regional air quality. [Sharma et al. \(2016\)](#) showed that daytime convective structures within the UBL were weaker for green/cool roofs cases with thinner updrafts, reduced vertical velocity of air currents, and thicker downdrafts. Thus, it was hypothesized that implementation of green/cool roofs (a proxy for green infrastructure) could lead to stagnation of air close to the surface, with longer retention times of anthropogenic pollutants released in the UBL. This, in turn, would lead to increased exposure rates for urban residents. Thus, an improved understanding of UBL dynamics in response to modifications of the surface energy balance is critical for determining the pollutant dispersion as well as lake and ocean breezes.

Progress over the last 100 years in observational, computational, and laboratory UBL studies have, without doubt, been phenomenal. With advancements in measurements and modeling techniques, we are now qualitatively able to reproduce meteorological processes over urban areas. However, we still lack in

capturing the intensity and magnitude of urban effects (Martilli 2007). The future UBL research would require detailed experiments covering fieldwork, laboratory, and numerical modeling woven together in an integrated fashion to gain insights over a wide range of scales, and enhance parameterizations for representing urban processes at microscale, regional scale, and global scales. At the same time, efforts are needed to improve CFD models at microscales that can run efficiently using high-performance computing (HPC) resources. In addition to model improvements, there is a need to enhance the development of an ultra-high-resolution land-use database for the extremely heterogeneous urban environment across the world [World Urban Database and Access Portal Tools (WUDAPT); Mills et al. 2015]. This approach is similar to the National Urban Database and Access Portal Tool (NUDAPT) resource (Ching et al. 2009) that provides gridded high-resolution data for urban canopy parameters (e.g., buildings, vegetation, and land use) used in model physics to improve urban simulations. Thus, going forward, the challenge in UBL research would be to set a balance between complexity and simplicity in accounting for urban processes based on the needs and computational capability. Such standards would need to be continuously revised with time, to account for increasing urban populations with new urban challenges related to urban security and climate change.

6. Mesoscale responses to wind farms

The history of scientific research on wind farm interaction with the atmosphere is quite short, despite the fact that single wind machines have been used for thousands of years for grinding grain and pumping water (Solaripedia 2018). Collections of wind turbines, positioned in close enough proximity to each other to create an aggregate atmospheric effect, were installed in California in the early 1980s under government incentives, largely in response to oil shortages and interest in alternative energy sources (U.S. Energy Information Administration 2018). The early California wind farms are located east of San Francisco (Altamont Pass), south of Bakersfield (Tehachapi Pass), and near Palm Springs (San Geronio), all in areas having limited prior intensive human land use. The Gansu wind farm in China, consisting of more than 7000 turbines arranged in rows on the edge of the Gobi Desert and currently considered the world's largest wind farm, also does not share the landscape with other intensive human activities. By contrast, more recent development of large wind farms in Texas and Iowa are located on landscapes having prior and current use for grazing and intensive agriculture,

which generates more interest in how these structures might modify local microclimates.

Conceptual models from analogs in natural or human-constructed landscapes provide guidance for thinking about the ways that wind farms would modify microclimate. Wind tunnel simulations of individual turbine wakes and collections of wakes produced by turbines aligned in regular arrays give insight on the changes in aerodynamic (e.g., wind speed, turbulence) characteristics of wind farms. Analysis of field measurements taken within wind farms and leeward of agricultural shelterbelts have identified thermodynamic (e.g., heating/cooling, evaporation/condensation, solar and infrared radiation) characteristics needed to construct an “informed conceptual model” of how turbines influence fluxes and profiles of mean properties. Under unique vertical distributions of moisture and temperature in the wind farm boundary layer, the additions of turbine-generated turbulence have been known to create regions of condensation in the flow field that can persist for minutes to hours in the downwind wake of the wind farm. Occasionally these are captured by radar or photography. Satellite measurements of surface radiating conditions provide spatial maps of surface skin temperature for regions occupied by wind farms for temporal periods of satellite overpasses. By representing the aerodynamic effects of wind farms as changes in surface roughness, global climate modelers have explored how large continental wind farms may create nonlocal climate effects, such as small changes in precipitation over oceans.

Because scientific research on wind farm impacts on the atmospheric boundary layer has only taken place over the most recent couple decades, focus in this section will be on outlining our current understanding, rather than providing a detailed history. In the following sections we use the combined results of in situ measurements, numerical models, wind tunnel studies, satellite observations, and photographic and radar images to provide a summary of how the aggregated effect of wind turbines alters natural conditions on subregions of the atmospheric boundary layer, mesoscale, and global scale.

a. Atmospheric boundary layer

Surface fluxes of momentum, water vapor, heat, and trace gases (primarily CO₂) are fundamental descriptors of how the atmosphere interacts with oceans and terrestrial surfaces. Over managed surfaces (crops, forests, pastures, and rangelands), any changes to natural fluxes caused by wind farms are of scientific interest and candidates for manipulation. Wind machines have been used for decades to

change surface fluxes of heat over vineyards and orchards to prevent frost formation. Agricultural shelterbelts (Wang et al. 2001) were planted throughout the Great Plains after the Dust Bowl to reduce upward fluxes of soil particles that created vast dust storms.

Most recently, operating wind farms throughout the Great Plains and U.S. Midwest are being collocated with range lands or agricultural fields used for growing commodity crops such as corn, soybeans, wheat, and cotton. Crop improvement programs have created cultivars that are more finely tuned to local soils and climate conditions, including the timing and magnitudes of fluxes previously identified. Changes to these basic ambient conditions due to wind farms are then of interest for their potential impact on crop growth and yield. Recent measurements in a large utility wind farm have reported the aggregate changes, or “bulk effects,” of multiple turbines that give insight on the basic characteristics of the wind farm boundary layer (Takle et al. 2019).

When the atmosphere is thermodynamically unstable the ambient turbulence breaks down the organized structure of turbine wakes, and vertical profiles of wind speed downwind of a wind farm recover their upwind characteristics more rapidly than for neutral or stable stratification. By contrast, under stable conditions wakes persist for many kilometers (Hirth and Schroeder 2013). Under neutral (unstable, stable) stratification, the bulk winds in the layer below the rotors deep in a wind farm are increased by about 2%–6% (5%–12%, 5%–12%) of the 120-m winds (Takle et al. 2019). However, winds in the rotor-swept layer are decreased under all stratifications, more so in stable conditions, typically 5%–10% under strong stratification.

A wind turbine converts a fraction, typically a third, of mean kinetic energy of air to rotational energy of the rotor shaft and ultimately electrical energy. Another third is converted to turbulence kinetic energy, with the final third remaining as mean kinetic energy of the downwind wake. The wake region, containing a high but dissipating amount of turbulence, expands outward away from the turbine axis line, ultimately reaching the surface downwind from the turbine. The wake reaches the surface nearer the turbine in daytime conditions compared to during the night due to the greater thermal stratification of the nocturnal boundary. In fact, under strongly stable conditions the wake turbulence may dissipate before reaching the ground. The physical scales of introduced turbulence range from approximately the rotor diameter to less than centimeter scales, which creates a fluid flow field within the wind farm that differs substantially from the natural boundary upwind of the wind farm. This leads

FIG. 19-19. Conceptual model of a turbine wake under (a) unstable daytime conditions and (b) stable nighttime conditions. (Credit: Eugene Takle.)

to changes in surface fluxes, temperature profiles, radiating surface temperatures, and surface shortwave and longwave radiation as described in the following paragraphs. Turbulence intensity (σ_u/u , where u is wind speed and σ_u is standard deviation of u) is decreased by 20%–50% in the layer below the rotors and increased from 10% to 30% in the rotor-swept layer deep in the wind farm. A conceptual model of the conditions created downwind of turbines is shown in Fig. 19-19.

In situ measurements suggest that turbines have little influence of daytime temperatures in wind farms but enhance nighttime surface temperatures by 0.2–1.0 K (Baidya Roy and Traiteur 2010; Henschen et al. 2011; Rajewski et al. 2013, 2014, 2016; Takle et al. 2014). Satellite measurements of radiating surface temperatures over the central United States (Zhou et al. 2012, 2013; Harris et al. 2014; Slawsky et al. 2015) report values of 0.3–1.0 K higher at nighttime satellite flyover times deep within and downwind of utility-scale wind farms compared to cropland outside wind farms. Tang et al. (2017), using Moderate Resolution Imaging Spectroradiometer (MODIS) data, also observed a modest nighttime warming of 0.15–0.18 K by large wind farms in the semiarid Bashang area of northern China. However, contrary to in situ and previous satellite measurements, their research also finds a warming effect of 0.45–0.65 during the daytime.

The aggregate effect of turbines on the rotor-swept layer deep in a wind farm is to increase temperatures by 0.1°–0.3°C under neutral conditions, by 0.2°–0.3°C under stable conditions, and decrease temperatures by 0.2°–0.4°C under unstable conditions (Takle et al. 2019).

Below the rotor-swept layer, the temperatures are increased by 0.1° – 0.3°C under neutral conditions and by 0.0° – 0.3°C under stable conditions, and they are decreased by 0.2° – 0.5°C under unstable conditions.

Turbines tend to dry the surface slightly during daytime (Armstrong et al. 2016; Takle et al. 2019), but in humid climates where heavy dew formation is a frequent overnight occurrence, the additional turbulence suppresses the rate of dew formation, thereby leaving more water vapor in the nighttime boundary layer air (see Fig. 19-19b). Elevated levels of water vapor in the nighttime atmosphere lead to more absorption of upward longwave radiation, with enhanced downward radiation and therefore suppression of the nighttime cooling of near-surface air and of upper surfaces of plants. In semiarid regions having low atmospheric humidity turbulence enhancement from wind turbines may have a strong desiccating effect on vegetation (Tang et al. 2017).

Flux measurements behind isolated turbines (Rajewski et al. 2014) showed an increase in upward latent heat flux and downward CO_2 flux over the crop during the daytime. Obstacles to the flow create a pressure-pumping action on CO_2 fluxes from unsaturated soils (Takle et al. 2004), but this effect, while likely present, has not yet been documented in wind farms. Flux measurement representing bulk effects of wind farms are not available.

Fragments of turbine wakes sometimes can be observed and photographed through their influence on the moisture field. The most famous is the Horns Rev wind farm shown in Fig. 19-20 and discussed in Hasager et al. (2017). Remnants of the wake become visible because the downward mixing of warm moist air under stable stratification into cooler air from below that causes localized condensation. The turbulence that created this localized vertical motion has long since been dissipated but the visible signature becomes “frozen” into the flow field until the droplets reevaporate. Figure 19-21a shows an example where blade rotation inserts periodic fragments of condensation regions in the flow field, and Fig. 19-21b shows a clear trail of a single turbine tracked by radar over a distance of about 65 km. It is emphasized, however, the total impact on surface radiation resulting from turbine-produced cloudiness likely is quite small.

Turbulence generated by turbines has other influences on solar and longwave radiation interactions near Earth’s surface. Baldocchi et al. (1981) reported measurements in an alfalfa field showing that higher levels of natural turbulence reduce boundary layer resistance to downward CO_2 transfer into the plant canopy during daytime. Furthermore, increased plant movement allows light to penetrate to light-unsaturated canopy leaves, thereby increasing overall plant photosynthesis and reducing the net shortwave reflected away

FIG. 19-20. Photograph of the Horns Rev 2 offshore wind farm at 1245 UTC 25 Jan 2016 seen from SSW direction. [From Hasager et al. (2017); copyright 2017 Charlotte Bay Hasager, Nicolai Gayle Nygaard, Patrick J. H. Volker, Ioanna Karagali, Søren Juhl Andersen, and Jake Badger. Licensee MDPI, Basel, Switzerland. <https://creativecommons.org/licenses/by/4.0/>.]

from the crop surface. This may be a contributing factor in corn also due to enhanced daytime turbine-generated turbulence and downward CO_2 flux reported by Rajewski et al. (2013) from concurrent measurements upwind and downwind of a line of turbines.

No studies are currently available that report rigorous in situ evaluations of the impact of turbines on plant or crop growing conditions. However, an in situ report of differences in downward carbon fluxes into identical crop canopies upwind and downwind of turbines suggests an enhanced carbon uptake leeward of wind turbines during daytime and enhanced respiration at night, with the daily average giving an enhanced uptake (Rajewski et al. 2013). Tang et al. (2017), used MODIS data to assess changes in summer growth and productivity from 2003 to 2014 caused by wind farms in the semiarid Bashang area of northern China. Their results show that wind farms inhibit local vegetation growth and productivity, decreasing leaf-area index (LAI), enhanced vegetation index (EVI), and normalized difference vegetation index (NDVI) by approximately 14.5%, 14.8%, and 8.9%, respectively, from 2003 to 2014. This reduces gross primary production (GPP) by 8.9% in summer and net primary production (NPP) 4% annually in wind farms. Potential reasons include decreased photosynthesis because of increased water stress and autotrophic respiration through increasing daytime and nighttime temperatures. Unlike other remote sensing studies (including dry climate regions of Texas) this semiarid site shows that mid-daytime surface temperatures are elevated in the wind farm region.

FIG. 19-21. (a) Fragment regions of condensed water vapor created by turbines, as in Fig. 19-20, in southwestern Minnesota (credit: Eugene Takle), and (b) radar image of a turbine wake having created a trail of condensation in western Illinois, as in Fig. 19-20 (credit: Ray Wolf, NOAA/DVN).

b. Mesoscale influences

Fiedler and Bukovsky (2011) used a regional-scale model to determine that a hypothetical wind farm covering the U.S. Great Plains would enhance warm-season rainfall by about 1% in the near vicinity and several tens of kilometers downwind of such a wind farm. Using a continental-scale climate model for Europe, Vautard et al. (2014) simulated a realistic wind energy development scenario where country commitments to wind energy production were doubled between 2012 and 2020 to a total installed capacity of 200 GW. Overall, they concluded that the aggregate impact of turbines on

Europe would be less than natural variability. Locally, differences in temperature could reach 0.3°C , with increases being largest in winter. Seasonal total precipitation changes due to turbines were calculated to change by a few percent and were significant only in small, isolated regions. Impacts would not be significant during the growing season.

Takle et al. (2014) use both a wind farm model and field observations to explore whether turbine-layer (lowest 120 m) wind speed reduction of wind farms might create horizontal flow convergence, a supporting ingredient for the development of cloudiness and precipitation. Field observations suggested that flow oblique to a line of turbines created flow directional change toward normal to the turbine line. The accompanying numerical simulations revealed a low pressure region that deepened with distance into the wind farm if turbines were sufficiently close together. Together, these results suggest that flow convergence and enhancement of vertical velocity over a wind farm should be examined as a mechanism for mesoscale modification due to wind farms.

c. Global influences

The first attempt to evaluate global impact of wind farms was provided by global model simulations of Keith et al. (2004). They considered massive wind energy deployments in North America, Europe, and China to be represented by increases in the surface drag on atmospheric flow in these regions. They found that wind farms created nonlocal seasonal temperature changes outside the wind farms due to regional changes in global-scale temperature gradients and heat transport. They estimated that a maximum regional and seasonal temperature increase of 0.5°C would be possible if 10% of electrical energy use came from wind by 2100. However, this wind farm contribution to global rise in temperature would be far below natural variability and other human contributions.

Wang and Prinn (2010) used a global climate model with an interactive ocean and enhanced surface drag to examine potential global climate impacts of wind farms. They found that large-scale wind plant deployment could induce changes in temperature and surface heat fluxes even outside the footprint of the wind farm, which could lead to modified global distribution of cloudiness and precipitation. A later study suggested the effect has been overstated by early modeling studies (Fitch 2015).

d. Discussion

Expansion in the number of wind farms in the United States raises the prospect that wind farms also should be

considered for their unique microclimates and even their influence on regional and global scales. While still in its infancy, recent boundary layer research on wind farms has demonstrated small but measurable changes from natural conditions. Because wind farms might be considered an “add-on” land-use type, their effect on the companion land use (e.g., intensively managed agriculture, grazing land, natural ecosystems) deserves research attention to quantify the effects on these primary land use.

7. Summary and future

This chapter provides several examples of mesoscale atmospheric circulations that are tied directly to surface variations. A complete review of all such types of circulations would be beyond the scope of this chapter. Therefore, we provide research on common examples of atmospheric responses to the surface, from relatively simple surface temperature variations over the ocean, inland across coastlines and lakes, to urban temperature and roughness variations, and ending with forced boundary layer circulations: 1) overocean temperature variations (responses to oceanic boundaries), 2) coastlines (sea breezes), 3) mesoscale regions of inland water (lake-effect storms), 4) variations in land-based surface usage (urban weather and climate), and 5) forced boundary layer motions (wind farms).

For each mesoscale atmospheric response, we examine the historical growth in understanding of these phenomena, focusing on how it has evolved since the American Meteorological Society was founded in 1919. The evolution is similar in some overall themes: starting off with observations that either identify a new phenomenon (often with new technological capabilities), sometimes analytical studies, followed by increasingly detailed numerical simulations of the phenomena.

It is interesting to note the range of time devoted to each of these phenomena and its relationship to human impacts. For example, sea breezes have been in the scientific literature since well before AMS was founded due, in part, to the importance of coastlines to societal activities of shipping and transportation. There now appears to be a good understanding of the basic physics behind sea breezes, and much of the recent literature has turned to applied aspects of sea breezes—such as their influence on air pollution and precipitation distributions. Urban meteorology similarly affects a growing percentage of the population and there continues to be rapid development in the various spatiotemporal scales involved as well as applications to mitigation of climate impacts (e.g., extreme heat, air pollution, etc.). Relatively new fields, boundary layer responses to forced

circulations in wind farms and responses to oceanic fronts have built on observational and numerical modeling techniques from other studies and are actively being pursued from both basic understanding of processes, to applicability to other phenomena, to applied components.

Each section sought to not only provide views on the history of understanding the phenomena under consideration, but to point the way toward areas that need future attention. For topics explored here:

- Analyses of atmospheric responses to SST variations, that provide a laboratory for air–surface interaction studies without the complexities of large surface topographic inhomogeneities, should be expanded to examine responses to the wide range of mesoscale variations observed.
- Active research on complex atmospheric motion fields in regions of nonuniform surface features (e.g., nearshore urban, mountainous, nearshore land/water patches) should be expanded to improve models applied to solving societal needs (e.g., air pollution, nearshore ecosystems, transportation economy, severe weather, heat-wave mitigation). Additional work is needed to incorporate the full diurnal, annual, and interannual variations, with emphasis on land breezes.
- Outstanding issues in the study of lake-effect snow systems include a dearth of observations and associated models of microphysical processes (see, e.g., [Barthold and Kristovich 2011](#); including impacts of seeding of LE clouds from higher-altitude clouds and from blowing snow from land) and air–lake–land exchange rates in nonuniform surfaces (including pack ice cover).
- Urban areas are of particular interest due to the potential for high societal impacts, and require the improved ability to link high-resolution, and long-term observations and models on such areas as human health and behavior, economics, energy, water, atmospheric chemistry (including air quality, pollution), and the surrounding environment (rural and natural areas, coastal ecosystems, etc.).
- The impacts of wind turbines and farms, as well as other renewable energy sources, on the atmosphere locally, regionally, and globally, on the atmosphere, society, and the environment need to be expanded, with particular attention on farming and natural ecosystem needs.

Some general themes that may be applicable to the multitude of other surface-forced mesoscale atmospheric phenomena can be made. There is a growing need to understand linkages between atmospheric processes on a range of size/time scales from both an up-scale and downscale perspective—from turbulence-scale

and microphysical processes to global circulations. As our understanding improves, it becomes increasingly important to gain information through observations and theoretical approaches and to develop models capable of fully coupling atmospheric models to those in other fields (i.e., coupling air chemistry models to urban atmospheric models, coupling large-eddy simulation models of wind turbines to finescale agriculture models, etc.). In this way, we will enhance our ability to understand and predict the mutual influences of mesoscale variations in Earth's surface cover, the atmospheric responses, ecosystems, and society.

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