Toward Improved Forecasts of Sea-Breeze Horizontal Convective Rolls at Super High Resolutions. Part I: Configuration and Verification of a Down-Scaling Simulation System (DS³)

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(Manuscript received 23 June 2014, in final form 26 January 2015)

ABSTRACT

Horizontal convective rolls (HCRs) that develop in sea breezes greatly influence local weather in coastal areas. In this study, the authors present a realistic simulation of sea-breeze HCRs over an urban-scale area at a resolution of a few meters. An advanced Down-Scaling Simulation System (DS³) is built to derive the analyzed data using a nonhydrostatic model and data assimilation scheme that drive a building-resolving computational fluid dynamics (CFD) model. The mesoscale-analyzed data well capture the inland penetration of the sea breeze in northeastern Japan. The CFD model reproduces the HCRs over Sendai Airport in terms of their coastal initiation, inland growth, streamwise orientation, specific locations, roll wavelength, secondary flows, and regional differences due to complex surfaces. The simulated HCRs agree fairly well with those observed by dual-Doppler lidar and heliborne sensors. Both the simulation and observation analyses suggest that roll updrafts typically originate in the narrow bands of low-speed streaks and warm air near the ground. The HCRs are primarily driven and sustained by a combination of wind shear and buoyancy forces within the slightly unstable sea-breeze layer. In contrast, the experiment without data assimilation exhibits a higher deficiency in the reproduction of roll characteristics. The findings highlight that CFD modeling, given reliable mesoscale weather and surface conditions, aids in high-precision forecasting of HCRs at unprecedented high resolutions, which may help determine the roll structure, dynamics, and impacts on local weather.

1. Introduction

Horizontal convective rolls (HCRs) are a common feature of mesoscale shallow convection (Etling and Brown 1993; Atkinson and Zhang 1996; Young et al. 2002). HCRs typically form in the convective boundary layer (CBL) with moderately unstable stratification (Kuo 1963; Asai 1970; LeMone 1973; Weckwerth et al. 1997). The rolls are aligned in long parallel lines that highly influence the local distributions of wind and temperature. Roll convection also exhibits a coherent updraft/downdraft, which provides an efficient means of vertically transporting momentum, heat, humidity, and air pollutants in the boundary layer (e.g., LeMone 1976; Wyngaard and Moeng 1992; Weckwerth et al. 1996; Inagaki and Kanda 2010). A unique type of roll develops within the sea-breeze layer in coastal areas, where cool marine air penetrates inland (Iwai et al. 2008; Oda et al. 2010). As sea breezes occur almost daily during summer and a majority of the population lives in coastal areas, further understanding and improved modeling of sea-breeze HCRs are valuable for

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DOI: 10.1175/MWR-D-14-00212.1

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forecasting local weather and for determining related societal implications.

A general difficulty with numerical modeling of HCRs has been resolving their finescale structure and mesoscale evolution. High-resolution calculations, such as large-eddy simulation (LES), are widely applied for explicit resolutions of roll convection at the scale of the CBL depth (Moeng 1986; Moeng and Sullivan 1994; Khanna and Brasseur 1998; Liu and Sang 2011). Sea-breeze HCRs have a particularly small wavelength of several hundred meters (Iwai et al. 2008; Oda et al. 2010). Thus, numerical models must have a grid spacing of tens of meters or smaller. At such a high resolution, the complicated surface conditions (buildings and steep topography) need to be described explicitly, for example, by a computational fluid dynamics (CFD) model (Kanda et al. 2004; Ashie and Kono 2011; Castillo et al. 2011; Inagaki et al. 2012; Park and Baik 2013). Moreover, to elucidate the evolution of HCRs, numerical models not only need a spatial resolution fine enough to capture individual rolls, but also need to possess a domain size large enough to capture their mesoscale features (Atkinson and Zhang 1996). Such a high-resolution calculation in a mesoscale domain requires a huge computational resource, which has only become available in recent years. These calculations allow for the simulation of roll clouds during cold air outbreaks over the ocean (Liu et al. 2004; Gyschka and Raasch 2005; Eito et al. 2010) and of roll-like temperature patterns over urban areas (Ashie and Kono 2011). However, realistic urban-scale simulations of sea-breeze HCRs via building-resolving models are sparse. The roll structure, evolution, and dynamics over coastal lands are not fully understood.

Another difficulty in simulating HCRs involves the representation of roll formation environment. Roll characteristics vary considerably for different weather conditions and underlying surfaces. For instance, the roll aspect ratio (a ratio of the roll wavelength to the CBL depth) tends to increase with CBL instability, while the roll orientation may change with CBL wind direction (e.g., Asai 1970; Weckwerth et al. 1997). The structures of secondary flows are also responsive to surface heating and the ambient wind speed and direction (Raasch and Harbusch 2001). As rolls evolve in a mesoscale area, they are likely subject to the impacts of changes in the atmospheric and surface conditions. Many previous studies performed simulations using idealized conditions or typical profiles on uniform surfaces; thus, these studies were limited to examining the overall statistics of roll behaviors (e.g., Moeng and Sullivan 1994; Khanna and Brasseur 1998). Recent studies suggested that including inhomogeneous surfaces in models is important for the realistic simulation of roll structures and their mesoscale evolution (Prabha et al. 2007; Maronga and Raasch 2013). To describe appropriate weather conditions, an option is to couple the CFD model to a mesoscale model (Baik et al. 2009; Ashie and Kono 2011), potentially through advanced nesting strategies (Mirocha et al. 2014; Muñoz-Esparza et al. 2014). In coastal areas, the variations in winds and temperature can be significant because of the land–sea contrast (Miller et al. 2003). Therefore, to achieve realistic modeling of sea-breeze HCRs, coastal weather and surface conditions should be precisely represented.

Considering the above-mentioned difficulties, forecasts of sea-breeze HCRs are a challenge. In this study, we propose a solution with a numerical prediction system based on advanced techniques. The system is built to carry out two major tasks: represent the mesoscale weather associated with roll formation and implement a building-resolving CFD simulation to describe individual rolls over an urban-scale area. The goal is to capture a specific sea-breeze event at a very high resolution and to reproduce HCRs more realistically than those reported previously. In Part I of this study, we evaluate the system performance by comparing numerical results with intensive observations; in Part II (Chen et al. 2015, hereafter Part II), we examine the detailed impacts of land use and buildings on roll characteristics. Part I is organized as follows: Section 2 describes the configuration of the numerical prediction system, the experiment design, and the verification data. Section 3 estimates the ability of the mesoscale prediction data to represent the inland progress of a sea breeze. Section 4 extensively verifies the characteristics and structure of the CFD-simulated rolls. Section 5 presents a generation budget of the roll kinetic energy to verify the reproduction of roll-forcing mechanisms. Then, the conclusions are presented.

### 2. Numerical system, experiment design, and observational data for verification

This section introduces the Down-Scaling Simulation System (DS³), which is designed for a high-precision forecast of mesoscale weather at a super high resolution. The system features one-way nesting with a parallelized CFD model coupled to a mesoscale model. In contrast to other simple nesting systems, DS³ derives analyzed data to describe the local weather using a convective-scale data assimilation scheme in a mesoscale model, thereby potentially improving the accuracy of multiple downscaling processes. The system applies high-resolution analysis data to obtain a short-range forecast of weather variables at even higher resolutions. Then, the system employs a building-resolving CFD model to explicitly describe complex surfaces and to simulate individual
HCRs over a suburban area near Sendai Airport. These major components of DS³ are detailed in the following subsections.

a. Meteorological model for mesoscale weather

DS³ employs the Japan Meteorological Agency Nonhydrostatic Model (JMA-NHM) to perform the mesoscale calculations for the data assimilation and extended forecast. The JMA-NHM is a community model designed for operational forecasts and research studies on mesoscale weather (Saito et al. 2006). The model is based on the fully compressible equations for hybrid terrain-following coordinates. Many physical processes, such as cloud microphysics, surface processes, boundary layer scheme, convective parameterization, and atmospheric radiation, are included in the model. The operational runs have provided numerical weather predictions for Japan since September 2004 (Saito et al. 2007). The model outperforms the preceding hydrostatic model in many aspects of mesoscale weather prediction. For further details on the operational and research applications of the model see Saito (2012).

b. Data assimilation scheme

The forecast accuracy of the mesoscale models is quite sensitive to the initial conditions. One important method for improving the quality of initial conditions is to link observations with the numerical models via data assimilation. Miyoshi and Aranami (2006) established a data assimilation scheme for JMA-NHM using the local ensemble transform Kalman filter (LETKF). Seko et al. (2011) modified the scheme to assimilate dense observations, such as radar radial winds and precipitable water vapor, to derive more accurate data. Seko et al. (2013) recently extended the LETKF to a two-way nested system to obtain analyzed data at the convective scale. The analyzed fields were used to improve the forecast of local intense rainfall. This nested system is incorporated into DS³ to obtain high-resolution analyzed data. The system’s ability to represent mesoscale weather in northeastern Japan is evaluated in section 3 as one part of our verification effort. These mesoscale-analyzed data are then used to initiate a short-range weather forecast at a higher resolution.

c. CFD model for local weather

For downscaling to a 10-m resolution, DS³ employs a local meteorological model based on a LES (Sha 2002). The CFD model specifications are summarized in Table 1 and are detailed as follows. The model is established in three-dimensional Cartesian coordinates to resolve steep topography and buildings, rather than in terrain-following coordinates that are subject to severe truncation errors at large slope angles. The fully compressible Navier–Stokes equations are solved by an algorithm of the Semi-Implicit Method for Pressure-Linked Equation Revised code (SIMPLERgo; Patankar 1980; Sha et al. 1991). A full time-implicit scheme is applied; variable fields are corrected iteratively until arriving at converged solutions at each time step. A high-order upwind advection scheme Quadratic Upstream Interpolation for Convective Kinematics (QUICK; Leonard 1979; Hayase et al. 1992) is employed. The discretization approach is a finite-volume method on a staggered grid system. The block-off technique is used to handle the complex geometries such as buildings and topography (Patankar 1980; Sha 2002). For more detailed treatments of real land use and buildings see Part II. Regarding turbulence, a classic Lilly–Smagorinsky LES is equipped to explicitly resolve large eddies and to parameterize small eddies (Lilly 1962; Smagorinsky 1963). Moisture and radiation processes are excluded. Using this model, Sha (2008) conducted a series of idealized experiments to illustrate the impacts of steep orography, building arrays, and street blocks on local wind and temperature. The results from small-area calculations are encouraging for realistic applications on an urban-scale domain in this study.

d. Design of the numerical experiments for the sea-breeze event

The numerical experiments are set up to examine the sea-breeze event along the Pacific coast of northeastern Japan and the HCRs over Sendai Airport (Fig. 1). Downscale simulations are run on five domains centered at Sendai Airport (38.135°N, 140.94°E), with horizontal resolutions of 15 km, 2 km, 400 m, 100 m, and 10 m, respectively. The calculations for domains 1–4 are implemented using the JMA-NHM with topography from GTOPO30. For domain 5, the CFD modeling is performed over realistic surfaces with high-resolution topography and GIS building information retrieved from the Zenrin Map Company.

The DS³ experiments with data assimilation (DS³_DA) are conducted over domains 1–2 using the nested

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FIG. 1. Domain setup of the DS$^3$ experiments. (a) Outer domain with a 15-km mesh, (b) inner domain with a 2-km mesh, (c) downscale domain with a 400-m mesh, (d) downscale domain with a 100-m mesh, and (e) CFD model domain with a 10-m mesh. In (a)–(d), the dashed rectangle denotes the next nesting domain. In (a),(b), the shaded area denotes topographic altitudes. In (c),(d), the shaded area denotes land-use categories. In (e), the yellow blocks mark buildings.
LETKF system (Figs. 1a,b). Both domains have 80 × 80 horizontal grid points and 12 ensemble members. The member number is limited here considering the compromise between local computational resources and real-time runs at a reasonable efficiency (Seko et al. 2011, 2013). The forecast-analysis cycles of all members start at 0300 Japan standard time (JST; JST = UTC + 9 h) 18 June 2007 (∼1.5 days ahead the validation time of the HCRs). The operational pressure-level analysis data (5-km mesh over Japan) within 3 days before the experiments are used for the initial conditions of 12 members in domain 1. Surface parameters, such as albedo and wetness, are set using 100-m mesh land-use data from the Geospatial Information Authority of Japan. The unperturbed boundary condition is applied to domain 1, and a covariance inflation of 1.2 is used in the LETKF system. Data assimilation is conducted in a 6-h cycle using conventional hourly observations, such as sounding profiles and surface pressure. The ensemble forecasts of domain 1 provide initial/boundary conditions for domain 2. The data assimilation in domain 2 is conducted in 1-h cycle using 10-min surface records of pressure and winds in northeastern Japan, as well as 30-min wind profiles derived from the velocity–azimuth display (VAD) analysis of the National Institute of Information and Communications Technology (NICT) lidar over Sendai Airport (Ishii et al. 2007). To highlight the effect of the analyzed data in DS3_DA, we conduct a parallel run without data assimilation (DS3_noDA). All parameters in DS3_noDA are the same as those in DS3_DA described above, except that no observational data are input to the LETKF system in domains 1–2. Note that DS3_noDA can somewhat present the weather conditions but not as accurately as DS3_DA because the initial and boundary conditions of domain 1 are given by operational analysis data.

The latest analysis data from domain 2 can be used to initiate an extended short-range forecast in domains 3–4 (Figs. 1c,d). Domain 3 (4) has 80 × 80 (120 × 120) horizontal grids and a forecast time of 1 h. The output from domain 4 offers initial and boundary conditions for the CFD modeling in domain 5 for a horizontal size of 10 km and a vertical extent of 500 m at a uniform 10-m mesh (Fig. 1e). In the CFD model, both the lateral and upper boundary conditions are handled by an improved Orlanski radiation condition that is free of reflection from gravity wave propagation (Orlanski 1976), while the bottom boundary condition (ground temperature) is fixed because it minimally changes over a short period. The building wall temperature is set to the prescribed ground temperature from domain 4. The time step is 1 s. The forecast time is 600 s, which is approximately 4–5 large-eddy turnover times.

e. Intensive observations during the study period

We focus on a typical sea-breeze event along the Pacific coast of northeastern Japan. The event occurred in the daytime on 19 June 2007, when thermally induced low pressure was established over the Japan Islands under fair weather (Fig. 2a). By 1300 JST, the southeasterly sea breeze had penetrated most land surfaces along Sendai Bay (Fig. 2b). The strong cooling of the sea breeze led to a low temperature of 23°–25°C, which was approximately 5°C lower than that on the unaffected inlands. Within the surge of the cool marine air over the heated land, the HCRs formed during the afternoon hours (Iwai et al. 2008; Oda et al. 2010).

In June 2007, an observation campaign on sea breezes was conducted over Sendai Airport. In addition to conventional surface observations, intensive observations from ground-based NICT and Electronic Navigation Research Institute (ENRI) Doppler lidars (Komatsubara and Kaku 2005) and Japan Aerospace Exploration Agency (JAXA) helicopter flights (Matayoshi et al. 2005) were obtained. The lidar locations, helicopter flights, and measuring equipment were detailed in an early report by Iwai et al. (2008). The NICT and ENRI lidars have range resolutions of 90 and 30 m, respectively. Using these intensive observation data, we attempt to evaluate the mesoscale-analyzed data’s representation of the sea-breeze intrusion during the daytime on 19 June. For the observed surface conditions, we use 10-min records of winds and temperature from the JMA Automated Meteorological Data Acquisition System (AMeDAS). For the vertical structure of the sea breeze, we use the 30-min wind profile from the VAD analysis of the NICT lidar. We also verify the CFD-simulated HCRs over Sendai Airport at 1305 JST. To observe the structure of the sea-breeze HCRs, we use the radial velocities of 1-min plane position indicator (PPI) scans from the ENRI lidar and the retrieved three-dimensional winds from the dual-Doppler lidar. These observational data are compared with the CFD simulations to verify the roll characteristics and structure. We also use high-resolution (∼1.16 m) winds and temperature from helicopter measurements to verify the detailed dynamic and thermodynamic properties of the HCRs, as well as the associated roll-forcing mechanisms.

3. Verification of the sea-breeze event over northeastern Japan

a. Inland penetration of the sea breeze

We compare the hourly data from the DS3 experiments with the AMeDAS records to verify the system’s representation of mesoscale weather. Figure 3 shows the
FIG. 2. A sea-breeze event over northeastern Japan on 19 Jun 2007. (a) Analysis chart of the mean sea level pressure at 1200 JST; and (b) the surface temperature and winds at 1300 JST as observed by AMeDAS. A full (half) wind barb is 2 (1) m s$^{-1}$. In (b), the topography is shaded in gray.
FIG. 3. Hourly variations in the surface wind and temperature trend. (a) AMeDAS observation, (b) analyzed data in DS* DA, and (c) forecast in DS* noDA. The temperature trend is estimated as a temperature difference between the valid hour and the following hour. In (a), the sea-breeze front at each valid hour is marked by the dashed line based on its passage over the AMeDAS sites. In (b) and (c), a contour of 0.2°C h\(^{-1}\) outlines the approximate sea-breeze front, where the cooling due to sea breeze is comparable to the surface heating due to solar radiation. A full (half) wind barb is 2 (1) m s\(^{-1}\).
progress of the sea breeze on 19 June 2007. At 0900 JST, the onset of the sea breeze is recorded at Sendai Airport, where a southeasterly wind of 4 m s\(^{-1}\) is followed by a temperature drop of \(-0.5^\circ\text{C h}^{-1}\) (Fig. 3a). The temperature at the other AMeDAS sites increases by approximately \(1^\circ\text{C h}^{-1}\). The numerical experiments of both DS\(^3\)_DA and DS\(^3\)_noDA capture a narrow zone of cooling along the coastline and extensive warming over land (Figs. 3b, c). At 1000 JST, the sea breeze forms over most of the coastal areas and begins to penetrate inland (Fig. 3a). The progress is relatively fast to the west of Sendai Airport, with an inland extent of \(\sim 16\) km from the coastline. A cooling rate of up to \(-1^\circ\text{C h}^{-1}\) occurs along the coastal areas. At this time, DS\(^3\)_DA captures the sea-breeze progress in terms of its penetration distance and cooling rate near the airport (Fig. 3b). In contrast, DS\(^3\)_noDA presents a relatively slow progress of the sea breeze and a weak cooling rate (Fig. 3c).

At 1100–1200 JST, the sea breeze continues to penetrate inland perpendicular to the coastline (Fig. 3a). The progress is relatively fast to the northwest of Sendai Airport, and the sea breeze gradually arrives at the foothills of the Ou Mountains. An enhanced easterly or southeasterly wind follows the passage of the sea-breeze front, with the strongest cooling of approximately \(-2^\circ\text{C h}^{-1}\) at the leading edge. At coastal areas near Sendai Airport, the sea breeze increases to 4–6 m s\(^{-1}\) and induces a gentle but continuous cooling. Such features are well captured by the analyzed data in DS\(^3\)_DA (Fig. 3b). In contrast, DS\(^3\)_noDA presents a delayed cooling at the leading edge of the sea-breeze front (Fig. 3c). Over coastal areas to the south of the airport, both numerical experiments simulate a sea-breeze progress that is slower than the observed progress, probably due to the orographic effects of coastal mountains.

For 3 h (1000–1200 JST), the sea breeze penetrates 30–35 km inland in both the AMeDAS observations and DS\(^3\) experiments; this movement indicates a speed of 10–12 km h\(^{-1}\). As suggested by Simpson (1969) and Sha et al. (1991), the sea breeze is a typical gravity current, whose speed can be given by \(U = 0.62 (g D \Delta T/T)^{1/2}\), where \(g\) is the gravitational acceleration, \(D\) is the depth of cool marine air, \(T\) is the temperature of cool marine air, and \(\Delta T\) is the temperature difference between marine and inland air. Given \(D = 220\) m from the lidar analysis (Iwai et al. 2008; Oda et al. 2010) and \(\Delta T = 3–4\) K from the AMeDAS records at 1000–1200 JST, the theoretical \(U\) is estimated as 10–12 km h\(^{-1}\). Therefore, the observed and simulated progress speeds in Fig. 3 are consistent with the theoretical analysis.

At 1300 JST, the mature phase of the sea breeze is characterized by a strong surge of southeasterly flow from the Pacific coast (Fig. 3a). Extensive cooling is experienced over most of the lowlands on the lee side (east) of the Ou Mountains. Both DS\(^3\) experiments produce a cooling pattern (Figs. 3b, c) very similar to the surface observations (Fig. 3a), except that DS\(^3\)_noDA simulates a weaker cooling rate to the west of Sendai Airport. From the overview in Fig. 3, it seems clear that the downscale experiment (DS\(^3\)_noDA) can offer a reasonably good forecast, while the experiment with data assimilation (DS\(^3\)_DA) better captures the sea-breeze event in terms of its timing and cooling rate.

b. Atmospheric conditions over Sendai Airport

The local weather over Sendai Airport is further evaluated by comparing AMeDAS records with analyzed data. Figure 4 shows that the observed temperature increases rapidly after sunrise and reaches a peak of 23.2°C at 0900 JST. Even under fair weather, the temperature decreases in the daytime because of the sea breeze, except for a small rebound at 1200 JST. Both DS\(^3\) experiments reproduce the rapid temperature increase after sunrise at 0600 JST but at a smaller amplitude than the observations. DS\(^3\)_DA produces both the temperature maximum at 0900 JST and the secondary minimum at 1100 JST, which are consistent with the observations. DS\(^3\)_noDA, however, produces a delayed temperature peak at 1100 JST. Note that the wind speed from DS\(^3\)_DA is \(\sim 1\) m s\(^{-1}\) stronger than that from DS\(^3\)_noDA and becomes closer to the observations. The well-simulated temperature and winds indicate that DS\(^3\)_DA produces a superior representation of the sea breeze strength.

For accessing the vertical structure of sea breeze, the profiles of horizontal winds by the DS\(^3\) experiments are compared with those obtained using the 20° elevation VAD scan from the NICT lidar over Sendai Airport. Figure 5a shows that in the early morning before 0700 JST, the observed wind vectors are northerly above 1000 m AGL (above ground level) and northwesterly in the lower layer. In the morning after 1000 JST, the southeasterly wind becomes dominant in the layer below 500 m AGL, while the westerly wind prevails above 1000 m AGL as the return flow of the sea-breeze circulation. In the afternoon, the depth of the low-level southeasterly wind continues to increase over time, while the maximum wind speed remains at \(\sim 200\) m AGL and the wind direction below only slightly changes, as also indicated by Iwai et al. (2008). Such a wind pattern, with speed shear and little directional shear in the CBL, is favorable for roll formation and sustenance (e.g., Weckwerth et al. 1997).

Figure 5b shows that DS\(^3\)_DA well reproduces the establishment of the sea-breeze circulation in the daytime. The analyzed data also display a gradual

\[
g = \frac{D \Delta T}{T} \frac{1}{2}\]
deepening of low-level southeasterly wind in the afternoon, consistent with the lidar observations. Figure 5c shows that DS3_noDA presents a persistent northwest-erly wind above 1000 m AGL. The run reproduces a southeasterly sea breeze near the surface at 0900–1100 JST but with a weaker wind speed than that of the observations and DS3_DA. These differences between DS3_DA and DS3_noDA suggest that data assimilation improves the representation of temperature and wind variations associated with the sea-breeze event over Sendai Airport.

4. Verification of the sea-breeze HCRs over Sendai Airport

a. General features of the CFD-simulated HCRs

For an overview of the roll characteristics, we examine the spatial pattern and temporal evolution of the CFD-simulated HCRs. Figure 6a shows a full-domain snapshot of the simulated vertical motion by DS3_DA. It is clear that vertical motion exhibits a streaky structure; parallel convective bands are oriented from southeast to northwest and are spaced several hundreds of meters apart. All of the HCRs form over land at a distance of 1 km or more from the coastline. The major rolls continue to grow downstream and stretch several kilometers inland. The rolls are more evident downstream of the major buildup areas (A1, A2, and A4); some of them initiate early in the coastal villages (A6 and A7). In contrast, the rolls are relatively suppressed or absent from the rice paddy fields (A3 and A5).

Figure 6b shows that the simulated HCRs, based on the forecast by DS3_noDA, also form at coastal areas and develop over land. The rolls are aligned from southeast to northwest but with a more northward component than those in Fig. 6a. The specific roll locations over the airport and inland regions in Fig. 6b seem to differ from those in Fig. 6a. For instance, the rolls over area A4 are shifted northward and are located near the rolls over area A2. The rolls over areas A2/A4 and downstream are more widely spaced than those in Fig. 6a. This finding is somewhat expected because different initial/boundary conditions are applied in the two CFD experiments. To estimate forecast accuracy, the simulated rolls are compared with the observations. As indicated, we focus our verification on the roll orientation, location, and wavelength, and the associated regional differences over Sendai Airport.

Figure 7a shows the temporal evolution of the vertical velocity at X = 3 km in DS3_DA. At the initial time step, the condition corresponds to a forecast of the mesoscale model at 100-m mesh. The updrafts have a magnitude within ±0.6 m s⁻¹ and behave like rolls with nearly regular spacing. The updrafts are much weaker than the observed updrafts as shown later; thus, the rolls are partly resolved in the mesoscale model. A possible cause
is that mesoscale model, with a resolution ranging from \( O(1) \) km to \( O(100) \) m, may fall within the “gray zone” or “terra incognita,” in which neither the 1D vertical diffusion in boundary layer scheme nor the subgrid-scale LES closure are appropriate (Wyngaard 2004; Dorrestijn et al. 2013; Shin and Hong 2013; Beare

Fig. 5. Hourly variations in the vertical profile of the horizontal winds over Sendai Airport during 0400–1500 JST 19 Jun 2007. (a) VAD analysis of the NICT lidar from 20° elevation scans, (b) analyzed data in \( Ds^3 \) DA, and (c) forecast in \( Ds^3 \) noDA.

Fig. 6. Vertical velocity at 100 m AGL over a full domain of the CFD model at a forecast time of 300 s (valid at 1305 JST 19 Jun 2007). (a) Simulation based on 1-h forecast from the analyzed data of \( Ds^3 \) DA. (b) As in (a), but based on the forecast from \( Ds^3 \) noDA. Buildings are plotted in the gray contour. The major distributions of land use in the vicinity of Sendai Airport are labeled as follows: airport terminal (A1), industry compounds (A2), rice paddy fields (A3 and A5), residential village (A4), and coastal villages (A6 and A7).

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The following series in the CFD model show that the updrafts become strong and exhibit regional features. From the initiation of weak and uniform perturbations, the CFD model gradually reproduces the strengthened updrafts, particularly downstream of the village and airport runway ($Y = 3.8$ and $5.5–8$ km). Some updrafts are displaced from their initial positions, resulting in an uneven spacing between the rolls. An analysis of the turbulence spectra also suggests that the rolls become well developed after a forecast time of ~300 s, and the CFD simulation reaches a quasi–steady state (Fig. 7b). The time scale is substantially shorter than that in other studies on a typical CBL (1–2 km deep), because the sea breeze layer is very shallow (~200 m) in this particular case. Similar growth of the rolls and turbulent energy occur in DS$_3$ noDA (figure omitted). In sections 4b–d, the instantaneous fields at that time step are discussed.

**Fig. 7.** Temporal variations in (a) the vertical velocity at $X = 3$ km and (b) the turbulent energy spectrum averaged over an area of $X = 3–5$ km and $Y = 0.5–9.5$ km. The estimate is made for a south–north cross section at 100 m AGL from the DS$_3$ DA experiment as shown in Fig. 6a. In (b), the wavenumber has been multiplied by the $Y$ distance along which the spectrum is estimated. The dashed line denotes the $K^{-5/3}$ inertial range slope.
b. Roll orientation, location, wavelength, and regional differences

Figure 8 shows the horizontal analysis of the wind fields over Sendai Airport. The lidar-observed HCRs are identified as the bands (CC’–JJ’0) of coherent updrafts (Figs. 8a,b). The bands extend from southeast to northwest, parallel to the mean wind direction. The CFD-simulated HCRs in the two CFD experiments are also aligned streamwise (Figs. 8c–f). In particular, the rolls from DS3_DA exhibit an orientation almost identical to the observed orientation, while those from DS3_noDA are oriented more northward. The difference in roll orientations corresponds well to the average wind direction, which has an azimuthal angle of \( \sim 135^\circ \) (lidar observations), \( \sim 133^\circ \) (DS3_DA), and \( \sim 139^\circ \) (DS3_noDA). Figure 8 also shows that the roll orientation change with height is small from 50 to 100 m AGL in both the observations and simulations. Such a wind-parallel roll orientation is expected when the CBL directional shear is small and when the speed shear is directed downwind (e.g., Asai 1970; LeMone 1973; Weckwerth et al. 1997). A comparison with the next volume of PPI scans 12 min later (figure omitted) indicates that these observed rolls are persistent and nearly steady, except for a slight displacement toward the northeast (Iwai et al. 2008) likely due to a slow veering wind direction (Figs. 4–5).

In addition to the roll orientation, the lidar measurements allow us to verify the specific locations of individual rolls. In the southern zone of the dual-Doppler lidar analysis domain, there are two elongated rolls observed near the lidar sites, as marked by EE’ and FF’ (Figs. 8a,b). The roll locations are reproduced well by the CFD simulation from DS3_DA (Figs. 8c,d). The CFD simulation from DS3_noDA also captures the location of roll FF’ (Figs. 8e,f). Although the roll EE’ is reproduced on the upstream side, it moves northward and merges with its northern neighbor. As a result, the roll location does not match the observed roll EE’ on the downstream side. A full-domain view of Figs. 6a and 6b shows that the rolls EE’ and FF’ originate from the coastal buildup area A6. The corresponding newly formed rolls share similar locations in the two CFD simulations, possibly because they are initially disturbed by the same underlying surfaces. As the rolls evolve inland with slightly different orientations in the two simulations, their specific locations gradually differ in the Sendai Airport area (Figs. 8c–f).

In the southernmost corner of the dual-Doppler lidar analysis domain, there is a signature of two other rolls (CC’ and DD’), with stronger updrafts observed in the southern roll CC’ (Figs. 8a,b). The CFD modeling from DS3_DA produces two rolls shifted slightly to the south (Figs. 8c,d), while that from DS3_noDA yields two rolls shifted northward with respect to the observations (Figs. 8e,f). The shifted locations correspond well to the deviated wind direction and roll orientation in the two CFD simulations. These rolls can be traced back to coastal village A7 (Figs. 6a,b). During its inland evolution, the southern roll mainly passes village A4, while the northern roll extends over rice paddy field A3. Both simulations reproduce a relatively strong updraft for the southern roll that corresponds to the observed roll CC’ (Figs. 8c–f). The warm surfaces of buildup areas may be conducive for roll growth and sustenance (Raasch and Harbusch 2001; Miao and Chen 2008; Ashie and Kono 2011). The underlying physics are further investigated in Part II.

In the northern zone of the dual-Doppler lidar analysis domain, there are four rolls GG’–JJ’ observed in Figs. 8a and 8b. The simulation from DS3_DA well reproduces roll GG’. Although the simulated rolls HH’–II’ match the observations at the upstream locations, they tend to shift westward on the downstream side, likely because the mean flow is more westward than observed. Roll JJ’ is weak in the observations and is barely distinguishable in DS3_DA. DS3_noDA also has a deficiency in simulating the locations of rolls GG’–JJ’ (Figs. 8e,f). The simulated rolls generally have a northward shift following the mean wind direction. On the other hand, these rolls observed in the northern zone are somewhat weaker than those in the southern zone (Fig. 8b). Such a gradient of vertical velocity is not simulated immediately at the corresponding zones in the CFD model, but it becomes clear in the downstream areas near the NICT site (Figs. 8d,f). The simulated rolls that are well defined near the NICT site appear to be stronger than those detected by lidar, likely because of the effectively lower resolution of the lidar retrieval.

Figure 8 also shows that the rolls are spaced at regular intervals, which enables us to estimate the extent to which the CFD simulations reproduce the roll wavelength. In the northern zone of the dual-Doppler analysis, the observed rolls display a mean spacing of \( \sim 450 \) m (Figs. 8a,b). In the southern zone, the mean spacing between two major rolls (EE’ and FF’) is \( \sim 515 \) m. The wavelength and regional difference are consistent with those estimated by power spectral analysis (Iwai et al. 2008). Figures 8c and 8d show that the simulated rolls from DS3_DA have a wavelength of \( \sim 415 \) m in the northern zone and of \( \sim 475 \) m in the southern zone; these values are slightly smaller than the observed values. The difference in roll wavelength between the two zones is \( \sim 60 \) m, which is nearly the same as the lidar observations. Figures 8e and 8f show that, in the simulation from
FIG. 8. Vertical velocity and horizontal winds over Sendai Airport at 1305 JST 19 Jun 2007. (a),(b) Dual-Doppler lidar retrieval; (c),(d) CFD simulation from DS³_DA; and (e),(f) CFD simulation from DS³_noDA. The lines CC’-JJ’ mark the axis of the observed roll updrafts. In (b), the dashed line marks the helicopter flight in Fig. 11. In (c), the dashed line AB marks the cross-roll section in Fig. 10.
DS$^3_{\text{noDA}}$, the rolls have a wavelength of $\sim 400$ m in the northern zone. However, the rolls exhibit a rather large wavelength of $\sim 625$ m in the southern zone. The roll wavelength difference between the northern and southern zones is $\sim 225$ m, which is much larger than that in the lidar observations and in the simulation from DS$^3_{\text{DA}}$.

c. Near-surface streaks of wind speed and their relation to the HCRs

Because the HCRs develop in a shallow sea-breeze layer, their relationship with the near-surface winds is examined and verified in this section. Figure 9a shows the radial velocity of the PPI scan from the ENRI lidar at a low elevation angle. Clearly, the streaky wind perturbations are embedded in the mean flow, with the axis oriented in the downwind direction. The velocity difference between the streaks of fast-moving and slow-moving flows is $\sim 2$ m s$^{-1}$. Comparing Figs. 9a and 8a, the velocity streaks roughly represent the perturbations of the major streamwise flow, given that the cross streamflow is secondary and slightly displaces the streak locations (Oda et al. 2010). Figure 9a shows that low-speed streaks are well collocated with roll updrafts, while high-speed streaks are located at the interval between updrafts. Such a good relation between streaks and rolls is due to the wind perturbations that distinctly manifest the convergence–divergence pattern as revealed by Iwai et al. (2008). On the other hand, Fig. 9a presents a much wider scan than the dual-Doppler lidar analysis domain of Fig. 8a. Specifically, the streaks are weak and less organized in the coastal areas in the southeast. The streaks become well defined at Sendai Airport and over inland areas in the northwest. Given the close link to updrafts, these near-surface streaks indicate coastal initiation and inland growth of sea-breeze HCRs.

Considering the streak-roll relation, the simulation of near-surface winds is a key issue for realistically modeling HCRs. Figure 9b shows the radial velocity derived from the DS$^3_{\text{DA}}$ simulation. Low-speed streaks are distinctly embedded in the sea breeze. These streaks experience inland growth from the coast and become evident over Sendai Airport. These features agree with those observed by the PPI scan from the ENRI lidar (cf. Figs. 9a and 9b). Figure 9b clearly shows that the low-speed (high-speed) streaks are well collocated with the roll updrafts (downdrafts). The DS$^3_{\text{DA}}$ simulation seems to reproduce the near-surface streaks associated with roll-scale convection. For comparison, Fig. 9c shows that the simulated streaks notably change from those in Figs. 9a and 9b. The streaks tend to stretch much farther to the north, following a change in the wind direction in DS$^3_{\text{noDA}}$. The spacing between the streaks...
in the southwestern quadrant becomes large (~625 m). These features correspond to the deficiency of the roll orientation, location, and wavelength in DS³_noDA (Figs. 8e,f). However, the simulated velocity fields in Figs. 9b and 9c are smoother than the observations in Fig. 9a, indicating that less-developed small-scale turbulence occurs in the CFD model.

d. Vertical structure of the HCRs and associated secondary circulation

Figure 10 portrays the vertical structure of HCRs in a cross-roll plane at 1305 JST. The HCRs are indicated by the presence of several counter-rotating circulations in the internal boundary layer (IBL). The updrafts are strongest at ~80 m AGL, with a magnitude of ~1.5 m s⁻¹. The updrafts are closely associated with the convergent (divergent) cross-roll flows in the lower (upper) half of the IBL. Such a secondary circulation simulated by the CFD model is similar to that observed by the dual-Doppler lidar analysis [Fig. 3d in Iwai et al. (2008)], except the maximum updrafts are at a slightly lower level. In addition to dynamic features, Figs. 10a and 10b show that the convective updrafts grow in thermal plumes, with a temperature ~1 K higher than the ambient air. Compensation downdrafts develop in the spaces between updrafts, where cool air mass can nearly reach the surface and absorb sensible heat. As the heated surface flows converge toward convective zones, they act to refuel the updrafts. Such a manner, in which the thermally driven secondary circulation creates a feedback to support roll sustenance, is consistent with what has been reported for other types of roll convection (Etling and Brown 1993; Atkinson and Zhang 1996; Young et al. 2002; Liu et al. 2004; Miao and Chen 2008).

Figure 10 shows that the CFD-simulated rolls extend throughout the entire IBL. The inversion of the air
temperature, along with near-surface temperature, suggests an IBL depth of 140–200 m in Fig. 10a and 150–240 m in Fig. 10b over Sendai Airport. These values are comparable to or slightly lower than the lidar observation of ~220 m (Iwai et al. 2008; Oda et al. 2010). Figure 10a also shows that the temperature contour exhibits a wavelike pattern at the IBL top and implies the possible existence of gravity waves. A gravity wave may be triggered by an overshooting updraft, while its persistence helps modulate roll vortices (Kuettner et al. 1987; Liu and Sang 2011). The gravity wave is most evident over the airport runway, and it corresponds to the regularly spaced rolls. Figure 10b portrays a slightly warmer IBL and a weaker IBL-top inversion compared with that in Fig. 10a. This difference may be due to the more northward wind direction in DS3_noDA. The marine air mass is expected to travel farther over the heated land surface before arriving at Sendai Airport. The well-mixed IBL seems to allow updrafts to develop higher in Fig. 10b. In contrast to the rolls near the NICT lidar site, the rolls exhibit relatively low heights in the northeastern region, where the IBL is shallow near the shore in the CFD simulations. Therefore, a good relation between the roll height and IBL depth, as reported by previous studies, is also observed in the sea-breeze HCRs.

Combining the roll wavelength (~515 m) and IBL depth (~220 m) near the NICT lidar site, the aspect ratio of the observed rolls is estimated to be 2.34, as also indicated in early analyses (Iwai et al. 2008; Oda et al. 2010). Correspondingly, the aspect ratios of the CFD-simulated rolls are estimated to be 2.38–3.39 in DS3_DA and 2.60–4.17 in DS3_noDA. These values are slightly larger than the lidar observations but are rather small compared with the typical values (ranging from 2 to 20) of roll family (Atkinson and Zhang 1996; Young et al. 2002). Many previous studies that use satellite cloud images have only counted the well-developed rolls with clouds to estimate the roll wavelength (Rao et al. 1999). In this particular case, the roll updrafts do not induce cloud formation because they occur in a shallow sea-breeze layer with relatively dry land conditions. Thus, sea-breeze HCRs, while likely common in coastal areas in summer, are not well detected by conventional observations because of their small wavelengths, shallow structures, and lack of clouds.

To illustrate the detailed structure of HCRs, Fig. 11a portrays the high-resolution variables measured by a helicopter flight over Sendai Airport. The west-east-oriented flight path is 5450 m long, with a spatial resolution of 1.16 m. The flight height of ~95 m AGL is nearly in the middle of the sea-breeze layer, where the strongest updrafts occur. Figure 11a shows that updrafts are active
over the concrete surface of the airport runway, while they are less pronounced over the rice paddy fields and shore regions. The major updrafts are generally spaced over several hundreds of meters; thus, the IBL convection has a roll-scale mode. The updrafts are overlapped by small-scale turbulence that spans tens of meters (the thin line in Fig. 11a). For a roll-scale mode (bold line), the small-scale turbulence is filtered out by a running window of 100 m. Figure 11a shows that roll updrafts (downdrafts) coincide with the perturbations of low speed (high speed) and high temperature (low temperature), with a correlation coefficient of $-0.37$ and $0.69$, respectively. Based on the eddy correlation method, the roll-scale momentum and heat fluxes are estimated as $-0.148 \text{ m}^2 \text{s}^{-2}$ and $0.134 \text{ K m s}^{-1}$, respectively. These fluxes account for two-thirds of the total fluxes estimated from the perturbation quantities (thin lines in Fig. 11a when removing the background value of linear regression). Therefore, roll convection may be effective in transporting momentum and heat vertically within the IBL.

Figure 11b shows that the CFD simulation from DS$_3$-DA appears to capture the major mode of the rolls. The simulated updrafts are evident over the airport runway but less pronounced over the rice paddy fields, which are consistent with helicopter measurements. The updrafts/downdrafts also concur with the fluctuations of the wind speed and temperature. The resolved vertical transport of momentum and heat are estimated as $-0.085 \text{ m}^2 \text{s}^{-2}$ and $0.077 \text{ K m s}^{-1}$, respectively. These values are somewhat smaller than the fluxes estimated from the helicopter data when filtered to roll mode. We also note that the CFD simulation reproduces smooth variables and tends to underestimate small-scale turbulence. An analysis of the turbulence spectra (Fig. 12) confirms that the CFD model captures roll-scale energy peak but falters in the reproduction of turbulence at an inertial range of $\sim 100 \text{ m}$ or smaller, despite some improvement compared to mesoscale model. A possible reason is that the transition from smooth mesoscale to resolved turbulence in a LES model does not guarantee rapid growth of turbulence in the nested LES area (Mirocha et al. 2014; Muñoz-Esparza et al. 2014). The relatively stable inflow may also hinder high-frequency turbulence over the upstream ocean surface in the LES domain. Other causes involve the numerical dissipation sources from the subgrid model in LES and the diffusion in the advection scheme (Lilly 1962; Smagorinsky 1963; Beare 2014). Although this study focuses on roll-scale convection, it calls for improved nesting strategies in the future to reproduce small-scale turbulent flows under actual weather conditions.

Figure 11c shows that the cross section from DS$_3$-noDA shares many common features with that from DS$_3$-DA. DS$_3$-noDA generally reproduces the roll-scale perturbations, the relation between the updrafts and warm/low-speed air, and the regional differences over inhomogeneous surfaces. We note in Fig. 11c that the amplitudes of the major updrafts, wind speed, and temperature fluctuations are slightly larger than those in
Fig. 11b. The spacing between the updrafts over the airport runway is also larger than that in Fig. 11b. These features manifest stronger rolls that develop in a well-mixed IBL in DS3_noDA. The resolved vertical transport of momentum and sensible heat are \( -0.095 \text{ m}^2 \text{s}^{-2} \) and \( 0.092 \text{ K m s}^{-1} \), respectively. These values are slightly larger than those in DS2_DA and are more similar to the observed roll-scale fluxes.

On the other hand, the helicopter measurements indicate that most of the major updrafts have a maximum speed of more than \( 2 \text{ m s}^{-1} \), while the downdrafts are approximately \( -1 \text{ m s}^{-1} \) (Fig. 11a). The updrafts are confined to local sharp peaks, while the weak downdrafts are more extensive; the updrafts and downdrafts are thus asymmetric (Deardorff et al. 1969; Raasch and Harbusch 2001). The updraft peaks are clearly coincident with the narrow bands of low wind speed and high temperature perturbations. Note that these features are not identified in the lidar-retrieval data at an effectively lower resolution (Figs. 8a,b). Figures 11b and 11c show that the simulated updrafts are more intense and narrower than the downdrafts; this finding is consistent with the helicopter measurements. Such asymmetric features are also observed in the other CFD simulations at a high resolution (Liu et al. 2004). These features are closely related to the narrow convergent zones, as indicated by the low-speed streaks (Figs. 9b,c). The convective updrafts develop in the narrow convergent zones of the thermally unstable air, while the downdrafts behave like passive compensation flows (Fig. 10). The CFD model appears to reproduce the intense/asymmetric roll-scale updrafts, despite the originally weak/uniform rolls in the absence of resolved smaller turbulence.

5. Forcing mechanisms of the sea-breeze HCRs

To illustrate roll-forcing mechanisms, we estimate the production terms of the resolved roll-scale turbulent kinetic energy (TKE) budget. Following LeMone (1973) and Moeng and Sullivan (1994), the terms of the roll-scale TKE budget are given by

\[
\frac{\partial E'}{\partial t} = \frac{g}{\rho_0} \bar{\theta}'_w \bar{w}' - \left( \bar{u}' \bar{w}' \frac{\partial \bar{U}}{\partial z} + \bar{v}' \bar{w}' \frac{\partial \bar{V}}{\partial z} \right) - \frac{\partial \bar{w}' E'}{\partial z} - \frac{1}{\rho_0} \frac{\partial w' \bar{p}'}{\partial z} - D, \tag{1}
\]

where the primes denote roll-scale perturbations and the overbars denote regional means. The roll-scale perturbations, as a kind of low-frequency turbulence, can be defined as the deviations from the spatially averaged means (Weckwerth et al. 1997). The term \( \bar{\theta}'_w \bar{w}' \) is the kinetic buoyancy flux; and \( \bar{u}' \bar{w}' \) and \( \bar{v}' \bar{w}' \) are the along- and cross-roll kinetic momentum fluxes, respectively. The first term on the right is the TKE production by buoyancy, and it represents the thermal forcing on rolls. The second term is the TKE production by wind shear, and it expresses the shear forcing on rolls. The third (fourth) term is the TKE change induced by turbulence (pressure) transport. These two transport terms only redistribute TKE vertically and do not contribute to the TKE source; \( D \) is the TKE loss by viscous dissipation. The TKE production terms are discussed here as we focus on roll formation mechanisms (Weckwerth et al. 1997), whereas the effects of all of the production and transport terms on the vertical growth of the rolls are investigated in Part II of this study.

To estimate the observed TKE budget terms, the momentum and heat fluxes are derived using either lidar-retrieved winds or helicopter measurements; the IBL wind shear is given by the lidar-retrieved wind profile. The shear production term derived from the lidar data is valid from 25 to 175 m AGL. The shear and buoyancy production terms are also derived from three helicopter flights along a west–east route over the airport runway at 45, 95, and 145 m AGL. The filtered helicopter data are used in this TKE analysis to consider the roll-scale mode of the perturbations. Because the time lag between flights is \( \sim 6 \text{ min} \), we assumed that the change in mesoscale environment is small during the helicopter measurements.

Figure 13a shows that, in the observation analysis, the roll-energy production by shear is visible throughout the IBL, with a maximum near the surface. The roll production by buoyancy is clearly observed in the middle of the IBL. Figures 13b and 13c show that the budget profiles derived from the two CFD runs exhibit magnitudes and vertical structures that are similar to those in Fig. 13a, except that the level of maximum buoyancy is relatively lower. This similarity indicates that the CFD model well reproduces the roll dynamics. Figures 13a–c illustrate that shear and buoyancy effects are comparable in producing roll kinetic energy near the surface, while buoyancy dominates most of the IBL above 40 m AGL. Such a shear effect in roll convection is distinct from the unorganized or cellular convection in which buoyancy dominates both near the surface and throughout the CBL (Moeng and Sullivan 1994; Weckwerth et al. 1997). Here the evidence from the observations and simulations consistently suggest that a combination of dynamic and thermal effects acts to drive and sustain roll convection in the sea breeze.

Figures 13b and 13c show that an IBL integration of the TKE production by buoyancy is approximately 2–3 times greater than that produced by shear. Thus, buoyancy serves a major energy source of convection: it manifests an evident thermal effect induced by the cool air intrusion over the heated surface to drive CBL
convection. The maximum buoyancy is located at \( \sim 60 \) m AGL, where roll updrafts significantly strengthen; the buoyancy decreases by half at \( \sim 100 \) m AGL. With the diminishing buoyancy aloft, the roll updrafts weaken with increasing height, as shown in Fig. 10a. The buoyancy reduces to near zero at the IBL top and above, where the updrafts dissipate. A comparison of Figs. 13b and 13c shows that the maximum buoyancy in DS\(^3\)_noDA is 10% larger than that in DS\(^3\)_DA. These values correspond to slightly stronger updrafts in DS\(^3\)_noDA, as shown in Figs. 10b and 11c. Another feature of Fig. 13c is that buoyancy is still observable at the levels of 150–230 m AGL and is distinct from a near-zero value in Fig. 13b. This extra buoyancy might explain the higher extent of the roll updrafts in DS\(^3\)_noDA. Buoyancy due to thermal instability greatly determines whether any convection will occur (Etling and Brown 1993; Atkinson and Zhang 1996). Here, thermal instability also has a strong influence on the intensity and vertical extent of roll updrafts.

Figures 13a–c show that the shear production of roll energy is most evident near the surface. The maximum production is mainly due to the near-surface shear (likely related to surface roughness) that is several times larger than the IBL shear (Figs. 14a–c). The profile is

![Figure 13](image13.png)

**FIG. 13.** Vertical profiles of the TKE production by buoyancy and shear over Sendai Airport. (a) Lidar (square with lines) and helicopter (square only) measurements during 1300–1318 JST 19 Jun 2007, (b) CFD simulation from DS\(^3\)_DA, and (c) CFD simulation from DS\(^3\)_noDA. The triangles mark the mean IBL height, which is estimated either from the NICT lidar observation of the signal-to-noise ratio gradient (Iwai et al. 2008) or from the CFD-simulated inversion near the NICT lidar site (Fig. 10).

![Figure 14](image14.png)

**FIG. 14.** As in Fig. 13, but for the wind speed, momentum fluxes, and heat fluxes.
similar to that reported by other analyses, suggesting that the low-level shear is more important than the CBL shear in producing roll energy (Pennell and LeMone 1974; Kristovich 1993; Weckwerth et al. 1997). Wind shear also serves to organize convection into along-shear sheets and to suppress modes that are transverse to it (e.g., Kuo 1963; Asai 1964, 1970). In the case of the sea breeze, the role of wind shear in maintaining roll structures is most efficient near the surface, as shown in Figs. 13a–c. This effect is thought to accumulate warm fluid into low-speed streaks at the lowest level, localizing buoyancy forces to sustain convective bands (Lin et al. 1997; Khanna and Brasseur 1998). Therefore, the dynamic effect due to near-surface shear plays a key role in organizing and maintaining the onshore HCRs within the sea breeze. Such shear and buoyancy forcings, while experiencing a gradual inland evolution, also generally hold for inland rolls, which are further discussed in Part II of this study.

Figures 13a–c show that the shear production of roll energy occurs above 40 m AGL and helps sustain the roll structure throughout the IBL. Note that the shear production term consists of wind shear and momentum fluxes. This production of roll energy in the IBL mostly arises from momentum flux, as the IBL shear is as low as \( -4 \times 10^{-3} \) s\(^{-1}\) in the sea breeze (Figs. 14b,c). As revealed previously, the interaction between rolls and streaks is primarily responsible for the strong momentum flux. The roll–streak interaction thus contributes to the production of roll energy, even under a low-shear sea breeze. These findings support an early analysis of Weckwerth et al. (1997): although wind shear is necessary for roll convection, rolls may occur in a very low CBL shear condition. We note that coherent downdrafts persisting in the IBL are crucial for transporting high momentum from the IBL top to the surface (Fig. 14). Because the downward momentum transport tends to increase wind speed in the downdraft zone, it provides a continuous kinetic source for sustaining the streaky flow patterns near the surface (i.e., Foster et al. 2006). In this respect, the IBL momentum fluxes and near-surface streaks may work together to maintain roll-type convection.

Just as in many theoretical and observational works, roll formation involves the combined effect of dynamic and thermal instabilities. The CBL instability parameter \(-z_i/L\), where \(z_i\) is the CBL depth and \(L\) is the Obukhov length, is often used to measure the roles of dynamic and thermal instabilities in determining roll shape. Here \(L\) is calculated from the surface fluxes as in Stull (1988) and Weckwerth et al. (1997), which has a value of \(-107\) m in DS\(^3\)_DA and \(-103\) m in DS\(^3\)_noDA. With \(z_i\) within 140–200 m in DS\(^3\)_DA (150–240 m in DS\(^3\)_noDA), the parameter \(-z_i/L\) is estimated as 1.31–1.87 (1.46–2.38) over Sendai Airport. The values are comparable to that derived from lidar and helicopter measurements (1.79) (Iwai et al. 2008). Both the observation and simulation analyses manifest a moderately unstable CBL that favors the existence of roll convection (Deardorff 1972). A relatively large instability in DS\(^3\)_DA is consistent with the increased buoyancy-to-shear ratio in the roll-energy production (Fig. 13c). This enhancement corresponds to the roll aspect ratio that increases from 2.38–3.39 in DS\(^3\)_DA to 2.60–4.17 in DS\(^3\)_noDA. The results agree with previous studies that report rolls with larger aspect ratios developing with larger \(-z_i/L\) parameters (e.g., Kuo 1963; Kelly 1984; Weckwerth et al. 1997). We also note that DS\(^3\)_DA, compared with DS\(^3\)_noDA, reproduces a more realistic roll aspect ratio observed over the airport (2.34). Therefore, capturing the CBL structure may be related to roll-forcing mechanisms, which contribute to realistic simulations of roll features.

6. Conclusions and future perspectives

Forecasting sea-breeze HCRs over coastal cities is a challenge. We addressed this issue using an advanced numerical prediction system that integrates a mesoscale model with convective-scale data assimilation and a building-resolving CFD model. We performed experiments on a typical sea-breeze event in northeastern Japan, and we validated the numerical results using intensive observations around Sendai Airport. The results indicate that a realistic simulation of HCRs can be achieved by combining reliable mesoscale weather forecasts with explicitly resolved individual rolls. The findings are summarized as follows:

1) A nested LETKF system is applied to perform data assimilation experiments. The mesoscale-analyzed data adequately capture the sea-breeze progress during the daytime on 19 June 2007. Consistent with surface records, the sea breeze exhibits a southeasterly surge of 4–6 m s\(^{-1}\), it penetrates inland with a speed of 10–12 km h\(^{-1}\), and it induces extensive cooling of \(-1^\circ\)C h\(^{-1}\). Compared with a simple downscale forecast without data assimilation, the analyzed data more accurately capture the passage of sea-breeze front and consequent cooling at most observational sites. In particular, the analyzed data well represent the temporal variations in the surface temperature and low-level winds over Sendai Airport. The data assimilation scheme thus provides an improved representation of the mesoscale weather associated with the sea-breeze event in this particular case study.
2) The characteristics and structure of the HCRs simulated by the nested CFD model are verified using dual-Doppler lidar and helicopter measurements over Sendai Airport. The simulated rolls form along coastal areas and develop over land. The rolls become active over buildup areas but are less pronounced over rice paddy fields; thus, they clearly respond to inhomogeneous surfaces. The rolls are oriented from southeast to northwest, parallel to the CBL wind direction and shear vectors. The rolls exhibit a wavelength of 400–500 m and extend through the entire IBL with a depth of ~200 m. The intense updrafts are typically rooted in the narrow bands of low-speed streaks and high-temperature perturbations near the ground, while the downdrafts are collocated with the high-speed streaks and low-temperature perturbations. The updrafts (downdrafts) are associated with cross-roll secondary flows that are convergent (divergent) in the low levels of the IBL. These simulated HCRs better agree with lidar and helicopter observations than those in the experiment without the data assimilation. Even the specific locations of individual rolls can be reproduced with reasonably good precision. To our knowledge, this research produces the first realistic simulation of observed sea-breeze HCRs with a high accuracy and super high resolution over an urban-scale domain.

3) The TKE production is estimated to elucidate the roll-forcing mechanisms. The simulated budget profiles are quite similar to those derived from the lidar and helicopter measurements; thus, the roll dynamics are well reproduced in the CFD model. Similarly to many theories of roll formation, wind shear and buoyancy forces work together to drive and sustain roll convection in the sea breeze with moderate instability and wind speed shear. Specifically, buoyancy provides a major energy source for convection occurrence and influences the strength and depth of updrafts. Buoyancy reflects the intrusion of cool marine air over warm land that produces thermal instability driving convection. The dynamic effects of wind shear and momentum fluxes also produce a large amount of roll-scale kinetic energy, particularly near the surface. This process greatly regulates the streaky pattern of the winds and temperature at the lowest level; thus, it plays a key role in maintaining roll-type convection. Considering that the IBL structure and surface conditions closely relate to roll-forcing physics, their adequate representation in the CFD model explains the improved forecast of the sea-breeze HCRs. The application of high-precision downscaling may lead to the more realistic HCRs simulations in this study, in contrast to previous idealized studies. In this study, HCRs are strongly dependent on complex surfaces, as also noted in Iwai et al. (2008). The rolls over Sendai Airport thus present a good opportunity to study the influence of land use/buildings on roll structure and evolution. Other studies also reported that roll convection tends to occur over urban areas, where the surface conditions are distinct from the surrounding areas (Kropfi and Kohn 1978; Newsom et al. 2008; Miao and Chen 2008; Inagaki and Kanda 2010; Oda et al. 2012). Complex surfaces are known to involve thermal/dynamic effects that greatly regulate turbulent structures and local circulations (Kanda et al. 2004; Prabha et al. 2007; Ashie and Kono 2011; Castillo et al. 2011; Kang and Bryan 2011; Inagaki et al. 2012; Park and Baik 2013; Maronga and Raasch 2013). Additional effort is required to quantify the detailed impacts of complex surfaces on roll behaviors. Using explicit resolutions of individual buildings, the CFD model can serve as a useful tool to describe these effects. In Part II of this study, we will perform additional experiments to illustrate how the sea-breeze HCRs respond to land-use heterogeneity and building morphology under actual weather conditions (Part II).

Because the mesoscale model output is fed into the CFD domain, the forecast error may introduce uncertainty. This study produces an improved forecast through a downscaling framework with convective-scale data assimilation. It is also noted that a further improvement in the framework may involve two other issues: the “gray zone” and strategies to bridge mesoscale and microscale models. To improve the mesoscale modeling at gray-zone resolutions, the parameterization of the turbulent fluxes can be optimized using several methods (Wyngaard 2004; Dorrestijn et al. 2013; Shin and Hong 2013). Beare (2014) recently proposed a length scale to quantify the effect of numerical dissipation at gray-zone resolutions. Ching et al. (2014) suggested options for handling the convectively induced secondary circulations that are poorly resolved in mesoscale models. On the other hand, downscaling from smooth mesoscale flows does not necessarily lead to rapid growth of turbulent flows in the nested LES domain. Mirocha et al. (2014) noted that nesting of a finer LES within a coarse LES helps accelerate turbulence growth. Muñoz-Esparza et al. (2014) developed a new method based upon the use of potential temperature perturbations near the inflow boundaries in the LES domain that substantially accelerates the development of turbulence at a low computational cost. Their idealized experiments also indicate that the turbulence growth in a nested LES depends on boundary layer stability and surface geometry. These results motivate us to evaluate and improve the
downscaling framework for a wide range of realistic weather and surface conditions.

This study is one part of the Strategic Programs for Innovative Research (SPIRE) of Japan that aims to improve high-precision mesoscale forecasts (Saito et al. 2013). Here we evaluated the performance of a downscaling system for one typical weather case. Because the sea breeze is a common phenomenon in summer, verification should be conducted for more cases to estimate the extent to which we can improve local weather forecasts by better modeling HCRs. In addition to sea-breeze events, the nested system of data assimilation is a powerful tool to improve forecast accuracy for a variety of mesoscale weather, including extreme events such as heavy rainfalls and typhoons (Seko et al. 2011, 2013). Here a combination of reliable mesoscale forecasts and the CFD model is feasible for the super-high-resolution modeling of regional weather over complex surfaces. One of the most valuable applications of this concept is to simulate mesoscale weather in large cities with many tall buildings. Because DS3 can explicitly resolve buildings, it is potentially applicable for realistic modeling of finescale winds and temperature within a city. Further verification of DS3 or similar downscaling systems is warranted to improve the forecast skill regarding urban weather and to establish operational numerical prediction at a resolution of a few meters in the future.

Acknowledgments. The authors thank the three anonymous reviewers for their helpful comments and constructive suggestions. This study was supported by the Strategic Programs for Innovative Research (SPIRE) funded by the Japan Ministry of Education, Culture, Sports, Science and Technology (MEXT). The numerical calculations were performed using the K computer at RIKEN Advanced Institute for Computational Science (Proposals hp120282, hp130012, and hp140220) and the supercomputing resource SX-9 at the CyberScience Center, Tohoku University. The observational data were collected in a field experiment funded by the MEXT (PI: Prof. T. Iwasaki, 19204046) through the collaboration of NICT, ENRI, JAXA, and Tohoku University.

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