1. Introduction

Many local weather events are brought about by characteristic topographic effects. Local fronts formed near the coast of the Kanto Plain, termed the “coastal fronts”, are one of the typical local weather events in Japan. According to Fujibe (1990), a coastal front is defined as a local front between the warm air associated with an onshore wind and colder air inland. It is widely known that such fronts occur near the east coast of North America, especially in New England or the Carolinas (Bosart et al. 1972; Riordan 1990). Many processes associated with the frontogenesis of coastal fronts have been identified: (1) land–sea differential heating, that is, the thermal contrast of the cold land surface with the initiation of warm onshore wind (Ballentine 1980); (2) nighttime cooling over the land (Nielsen 1989); (3) frictional forcing, that is, differ-
ences in friction between the sea and the land (Bosart 1975); and (4) the mountain effect, which involves the upstream blocking of stable (cold) air in a region of warm advection (Garner 1986; Nielsen 1989). Most studies have regarded coastal fronts as fronts associated with extratropical cyclones. In such cases, coastal fronts have been found to coincide with a boundary of rain and frozen precipitation and/or the location of enhanced precipitation (Bosart 1972; Marks and Austin 1979). Moreover, coastal fronts often form the eastern boundary of a cold-air damming (CAD) event, enhancing mountain-parallel cold advection along the eastern slopes of a mountain range (Bell and Bosart 1988; Appel et al. 2005).

As for coastal fronts in the Kanto Plain (Fig. 1), Fujibe (1990) has shown from statistical analyses that (1) coastal fronts are more conspicuous for southerly onshore winds than for northeast winds, from late autumn to winter than in warmer seasons, and in the nighttime rather than in the daytime if precipitation is absent or limited; (2) the orientation of coastal fronts in the Kanto Plain is in the NNE–SSW direction with northeasterly and southeasterly onshore winds, and in the ENE–WSW direction with southwesterly onshore winds in the composite field. Moreover, based on previous research, the coastal fronts in the Kanto Plain can be classified into two types: (a) with southerly onshore winds (S-type coastal front) and (b) with easterly onshore winds (E-type coastal front). Severe air pollution in the Kanto Plain is associated occasionally with the S-type coastal fronts (Mizuno and Kondo 1992); therefore, a series of observations have been performed to determine the three-dimensional structure of the front. These works revealed that (1) warm southerly winds flow over the stagnant cold air mass with a thickness of several hundred meters (Yoshikado et al. 1994; Seino et al. 2003) and (2) the fronts are formed by the effect of the western and northern mountains holding the cold air on the lee side against the southerly wind, along with the cold air being supplied from the downslope wind (Yoshikado 1998; Seino et al. 2003). The E-type coastal fronts are associated occasionally with the winter precipitation form (Yamamoto 1984; Hasemi and Baba 1994) and the location of heavy precipitation (Sakakibara et al. 1985; Hara 2014), showing good correspondence to the New England coastal fronts. In this case, evaporative cooling and condensational heating also contribute to the formation and maintenance of coastal fronts (Hara 2014).

Forecasting the location of coastal fronts accurately is highly desirable in predicting sensible weather, such as air pollution and reduced visibility for the aviation industry (Tokyo Aviation Weather Service Center 2014), and in anticipating the type of precipitation. However, few previous studies have systematically verified forecasts of the “occurrence location” of coastal fronts. In assessing the performance of the Japan Meteorological Agency’s (JMA’s) operational mesoscale Numerical Weather Prediction (NWP) model (MSM) with a horizontal grid spacing of 5 km, Hara (2014) and Kawano et al. (2019) reported that coastal fronts tend to be forecast on the cold air (inland) side of their actual positions. In addressing the model’s underestimation of the cold air mass, Hara (2014) suggested several possible reasons, including (1) overestimation of warm air advection or underestimation of cold air advection, (2) underestimation of evaporative cooling on the cold air side, and (3) excessive heating due to condensation or a turbulent closure scheme. However, Hara did not reach a conclusion regarding the exact cause. Kawano et al. (2019) investigated the coastal fronts by using the Mesoscale Ensemble Prediction System operated by the JMA, which generates the initial and boundary perturbations on the MSM. The illustrative case study revealed that no ensemble member significantly corrected the location error of the coastal front, and suggested that the initial and boundary values have less impact on the systematic error.

The topographic condition of the Kanto Plain is similar to that of the southeastern coast of the United States, both of which are located southeast of a high mountain range. Indeed, the operational U.S. NWP models have been reported to underestimate spatial and temporal scales of the Appalachian CAD, often eroding cold-air domes prematurely (Stanton 2003; Lackmann and Stanton 2004). Such a systematic error remains a challenge to operational forecasters (Lackmann 2011). Even today’s high-resolution mesoscale models show a tendency to erode the cold air too quickly (Grumm 2015). The systematic error has been considered to originate from convective schemes, turbulent closure schemes, and cloud–radiation interactions (Stanton 2003; Lackmann 2011).

This study focused on the MSM forecast error reported by Hara (2014) and Kawano et al. (2019). To date, few studies have dealt directly with the relationship between predicted coastal fronts and a model representation of orography. The Kanto Plain is surrounded by mountains on its western and northern sides (Fig. 1). If the MSM forecast error is due to an underestimation of mountain barriers associated with the model’s use of a mean orography for smoothing, the error may be corrected by introducing an envelope
orography or increasing the model’s resolution. We set out to confirm the systematic MSM error associated with predictions of coastal fronts through statistical validations and to explore their physical mechanisms by conducting sensitivity experiments.

To this end, we selected the coastal fronts observed over the Kanto Plain, while we abandoned the analysis on offshore fronts for lack of data available for verification. Furthermore, given Fujibe’s (1990) observation that coastal fronts are more conspicuous for southerly onshore winds, we selected S-type coastal fronts in this study. For the sensitivity experiments, we used a nonhydrostatic numerical model developed by the JMA (JMA-NHM).

In the next section, we introduce a typical example of the forecast error in the prediction of coastal fronts. In Section 3, we confirm the existence of systematic error through statistical validations of coastal fronts. In Section 4, we explore the systematic error mechanisms by conducting sensitivity experiments, the results of which are presented in Section 5. In Section 6, we consider the mechanism of shifting coastal fronts through our sensitivity experiments. Finally, we summarize our findings in Section 7.

2. Example of MSM error in predicting coastal fronts in the Kanto Plain

We used MSM forecasts of surface temperature and surface wind to validate the coastal fronts. The MSM is the mesoscale operational short-range forecast model used by the JMA to cover Japan and surrounding areas (3,600 × 2,880 km, with a horizontal resolution of 5 km). It provides 39- or 51-hour forecasts every 3 hours. For the period prior to February 2017, the operational MSM was the “JMA-NHM” (Japan Meteorological Agency 2013); this was then replaced by the new “asuca” model (Japan Meteorological Agency 2019). The MSM forecast data were collected and distributed by the Research Institute for Sustainable Humanosphere, Kyoto University (http://database.rish.kyoto-u.ac.jp/index-e.html). The observation data on surface temperature are from the Automated Meteorological Data Acquisition System (AMeDAS) database at the JMA (Fig. 2a) and the Atmospheric Environmental Regional Observation System (AEROS) database at the Ministry of the Environment (Fig. 2b). The verification data on surface wind are from the JMA’s operational hourly atmospheric analysis.

2.1 Comparison of MSM 12-hour forecasts with observations

Figure 3 shows the synoptic field at 03 JST 9 March (= 18 UTC 08 March) 2018. At that time, the Kanto Plain was under the warm sector of a cyclone passing over the Sea of Japan, with a southerly onshore wind from the Pacific Ocean. Figure 4 shows a time series
of surface temperature and surface wind over the
Kanto Plain during the occurrence of the coastal front.
The zone showing numerous isotherms corresponds to
the coastal front, which is the boundary between the
warm southerly flow and the stagnant colder air. The
coastal front here was observed between Yokohama
and Tokyo at 00 JST 9 March 2018 (Fig. 4a), near
Tokyo at 03 JST (Fig. 4b), and near Koshigaya at 06
JST (Fig. 4c). Compared to these observed positions,
the coastal front in the MSM 12-hour forecasts (Figs.
4d–f) was located consistently to the inland (cold air)
side, with northwestward distance errors estimated to
be approximately 30 km.

2.2 Verification of initial condition dependence of
MSM forecast error

Figure 5a shows the MSM analysis (initial condi-
tion) at 03 JST 9 March 2018. Figures 5b–f show the
MSM forecasts at 03 JST 9 March 2018, with a fore-
cast period of 3–15 hours. The coastal front in the ini-
tial condition (Fig. 5a) was placed in nearly the same
position as the observation (Fig. 4b). This is because
the surface observation data and the remote sensing
data were assimilated into the forecast-analysis cycle
of the MSM to create the initial condition. However,
as the forecast period progresses, the location of the coastal front shifts gradually to the inland side of its actual position, conspicuously from the initial condition (Fig. 5a) to the forecast period of 6 hours (Fig. 5c). This may represent a typical error, where the position of a coastal front is shifted inland in the Kanto Plain. We next statistically confirm the robustness of this error.

3. Statistical verification of forecast errors

3.1 Sampling of coastal fronts

We limited our samples to those in which the coastal fronts were observed between Miura and Kuki (Fig. 2a) with southerly onshore winds, during the period between 2015 and 2018. Samples that satisfied the conditions described in Appendix A were selected every hour. We assigned the frontal position to the section with the largest horizontal temperature gradient among four sections, from seaside to inland in the Kanto Plain: Miura–Yokohama, Yokohama–Tokyo, Tokyo–Koshigaya, and Koshigaya–Kuki (Fig. 2a). The distance per section is approximately 25 km. In all, 170 hours of samples were obtained for statistical verification. The cold season (October to March) accounts for 140 of 170 hours of samples. Also, 170 hours of samples account for 14 coastal front cases. The operational MSM forecasts were compared with all the samples selected.
3.2 Confirmation of systematic forecast error

Table 1 shows the frequencies of coastal fronts for each section in the observation and the MSM 10–12-hour forecasts, with a total of 170 hours of samples. The coastal fronts were frequently forecast one or two sections to the northern side of their observed positions. As the distance per section is approximately 25 km, it is inferred that coastal fronts were shifted frequently 25–50 km north along the section line. The underlined orange boxes in Table 1 represent the frequencies with which MSM forecasts placed the coastal fronts on the inland side of their actual sections (“MSM inland”), accounting for 149 of 170 hours. In contrast, the MSM forecast located the fronts in their actual sections (“No Error”) in 21 of the 170 hours. Notably, in none of the forecasts did the MSM locate the fronts seaward of their actual section (“MSM seaward”). Therefore, we concluded that the operational NWP model has a systematic forecast error, shifting coastal fronts inland for the forecast period of 10–12 hours.

Fig. 5. (a) MSM analysis (initial condition) at 03 JST 9 March 2018. (b, c, d, e, f) MSM surface temperature and surface wind forecasts at 03 JST 9 March 2018 with different initial conditions (the forecast period of 3–15 hours). The shading and contour lines are the same as those in Fig. 4.
hours. We also confirmed that this systematic error did not change when the operational model changed in February 2017, and the error appeared in all 14 coastal front cases from 2015 to 2018.

Next, we investigated whether a similar systematic error would emerge for other forecast periods. Here, the frequencies of “MSM inland”, “No error”, and “MSM seaward” were calculated for each forecast period, and the ratio (%) of the frequencies to all samples was determined. Figure 6 shows the relationship between the length of the forecast period and the “MSM inland”, “No error”, and “MSM seaward” ratios. As indicated in the figure, the percentage of “MSM inland” is in the 20% range at the initial condition, but exceeds 80% for a forecast period of more than 5 hours. Thus, the coastal fronts were shifted inland of their actual positions when the forecast period exceeded 5 hours. Moreover, “MSM inland” of 80% or more persists until the 39-hour forecast (maximum forecast range). These results suggest that the forecast error does not come from the initial and boundary conditions, but rather is related to the systematic error of the MSM.

### 3.3 Systematic error verification through composite analyses

We conducted composite analyses for each section of observation, as shown in Table 1. Figure 7 gives the composite fields of the surface temperature and surface wind in the observation (AMeDAS) and MSM 10–12-hour forecasts. In all the sections, we confirmed the northwestward errors of coastal fronts in the MSM 10–12-hour forecasts. As shown in the right panels of Figs. 7c, 7f, 7i, and 7l, the surface temperature differences of the MSM from the observation were near zero on both sides of the frontal zone. Although the MSM forecasts shift coastal fronts to the inland side, they do not have large biases of surface temperatures, either for warm or cold air masses.

We then investigated the relationship between forecast error and precipitation, defining two cases: the “RAINY case” was defined as samples where pre-

<table>
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<th>Observation</th>
<th>MI ~ YO</th>
<th>YO ~ TO</th>
<th>TO ~ KO</th>
<th>KO ~ KU</th>
<th>sum</th>
</tr>
</thead>
<tbody>
<tr>
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<td>0</td>
<td>0</td>
<td>0</td>
</tr>
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<td>0</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Yokohama–Tokyo</td>
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<td>1</td>
<td>0</td>
<td>0</td>
<td>9</td>
</tr>
<tr>
<td>Tokyo–Koshigaya</td>
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<td>18</td>
<td>10</td>
<td>0</td>
<td>46</td>
</tr>
<tr>
<td>Koshigaya–Kuki</td>
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<td>13</td>
<td>33</td>
<td>9</td>
<td>57</td>
</tr>
<tr>
<td>North of Kuki</td>
<td>2</td>
<td>9</td>
<td>17</td>
<td>29</td>
<td>57</td>
</tr>
<tr>
<td>sum</td>
<td>31</td>
<td>41</td>
<td>60</td>
<td>38</td>
<td>170</td>
</tr>
</tbody>
</table>

Fig. 6. Relationship between MSM forecast error ratios (vertical axis, %) and the forecast period (horizontal axis, hours). The solid red line shows MSM forecasting coastal fronts on the inland side of their actual section (“MSM inland”), the broken black line shows MSM forecasting in their actual section (“No error”), and the dotted blue line shows MSM forecasting on the seaward side of their actual section (“MSM seaward”).
Fig. 7. Composite fields of surface temperature (contoured at intervals of 1°C) and surface wind for each section of observation shown in Table 1. The left panels (a, d, g, j) are composites of observation (AMeDAS). The middle panels (b, e, h, k) are composites of MSM 10–12-hour forecasts. The right panels (c, f, i, l) show the surface temperature differences of the MSM from the observation (shaded, contoured at intervals of 1°C).
Precipitation of 0.5 mm h$^{-1}$ or more was observed at more than half of the AMeDAS stations in the Kanto Plain; for the “non-RAINY case”, such an observation occurred at fewer than 10% of the stations. Figure 8 shows composite fields of the RAINY case (14 hours) and non-RAINY case (26 hours), where observed coastal fronts were classified into the section between Tokyo and Koshigaya. As the figure indicates, the MSM forecasts appear to have systematic error regardless of precipitation, although the systematic error is more prominent in the RAINY cases than in the non-RAINY cases. Similar results were obtained in all sections.

4. Numerical sensitivity experiments

We conducted numerical experiments on three typical coastal front cases with systematic prediction error over the Kanto Plain. The coastal fronts were associated with an extratropical cyclone traveling over the Sea of Japan (Fig. 1) and were observed continuously for more than 6 hours with southerly onshore winds. The dates of the numerical experiments and the analysis period used to evaluate the frontal position are given below. The start time of the numerical experiments was prior to the onset of the coastal front. CASE 1 is an example of a RAINY case, as described in Section 2; CASE 2 and CASE 3 are examples of non-RAINY cases:

**CASE 1**: 0900JST 8 March 2018 (initial time) to 0900JST 9 March 2018 (analysis period: 0000JST to 0600JST 9 March, FT15 to FT21)

**CASE 2**: 2100JST 22 February 2017 (initial time) to 1200JST 23 February 2017 (analysis period: 0300JST to 0900JST 23 February, FT06 to FT12)

**CASE 3**: 0900JST 3 February 2019 (initial time) to 0900JST 4 February 2019 (analysis period: 0000JST to 0600JST 4 February, FT15 to FT21)

(FT indicates forecast period (hours))

For the experiments, we used the JMA-NHM (Saito...
et al. 2006) rather than the operational NWP model. Figure 9 shows the computational domain of 1000 km × 1000 km centered over the Kanto Plain. The mesoscale analyses of JMA (Japan Meteorological Agency 2013) were used as the initial and lateral boundary conditions. An experiment with nearly the same settings as the MSM was defined as the control (CTL) experiment. The details of the model configuration of the CTL are as follows. The horizontal grid spacing was set as 5 km. The 50 vertical layers extended from a height of 20 m to 13.2 km. The model included the bulk cloud microphysics by Ikawa et al. (1991) and the Mellor–Yamada–Nakanishi–Niino level-3 turbulence closure scheme by Nakanishi and Niino (2006). No convective parameterization was used. The terrain data were generated by the interpolation of the GTOPO30 (Global 30 arc-second elevation with a horizontal grid spacing of approximately 1 km) dataset. Surface-related parameters were introduced by the National Land Numerical Information dataset provided by the Geospatial Information Authority of Japan. The design of the sensitivity experiments is described below. A summary of the sensitivity experiments is shown in Table 2.

4.1 Horizontal resolution (DX_2km, DX_1km)

If the error is due to an underestimation of mountain barriers associated with a model smoothing using a 5-km grid (CTL orography, shown in Fig. 10a), higher-resolution models could be expected to reduce the error. We conducted experiments by increasing the horizontal resolution from 5 km to 2 km (DX_2km experiment) and 1 km (DX_1km experiment), respectively. The model terrains used in the DX_2km and DX_1km are shown in Figs. 10d and 10g, respectively.

4.2 Envelope orography (EO_5km, EO_2km, EO_1km)

Normally, the elevation at each grid point is given by the mean value of all the GTOPO30 elevation data included in the grid. Here, we introduced an envelope orography as the model terrain, created by using the maximum value of the GTOPO 30 elevation data at each grid point. Hereinafter, experiments using the envelope orography are referred to as EO_5km (Fig. 10b), EO_2km (Fig. 10e), or EO_1km (Fig. 10h), depending on the horizontal resolution in the model. Figures 10c, 10f, and 10i show the differences in the elevation of the envelope minus the mean orography for each resolution. In the 5-km grid, the envelope orography increased the model terrain by ~500 m in the mountainous region, while it caused little change in the Kanto Plain. If the resolution is

<table>
<thead>
<tr>
<th>Experimental name</th>
<th>Grid spacing (m)</th>
<th>Orography</th>
<th>Surface roughness</th>
<th>Evaporative cooling</th>
</tr>
</thead>
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<td>Mean_5km</td>
<td>Original</td>
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</tr>
<tr>
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<td>Mean_2km</td>
<td>Original</td>
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</tr>
<tr>
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<td>Mean_1km</td>
<td>Original</td>
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</tr>
<tr>
<td>EO_5km</td>
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</tr>
<tr>
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</tr>
<tr>
<td>Rough_SEA</td>
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<td>On</td>
</tr>
<tr>
<td>NOEVAP</td>
<td>5,000</td>
<td>Mean_5km</td>
<td>Original</td>
<td>Off</td>
</tr>
</tbody>
</table>
Fig. 10. The left panels (a, d, g) are the terrain created from the grid mean value of GTOPO30 (Mean_xkm). The middle panels (b, e, h) are the terrain created from the grid maximum value of GTOPO30 (EO_xkm); (a, b, c) are 5-km, (d, e, f) are 2-km, and (g, h, i) are 1-km resolution terrains. The right panels (c, f, i) show differences in the elevation of the EO_xkm minus the Mean_xkm terrain.
increased from 5 km to 2 km and 1 km, the envelope orography had less impact on the model terrain (Figs. 10f, i).

4.3 Surface friction (Rough_URBAN, Rough_SEA)

We also considered the possibility of changes in the coastal fronts due to surface friction. Figure 11 shows the surface roughness used in the CTL. In urban and mountain areas, the roughness is relatively high (3.0 m), while over the sea, it is low (0.001 m). In our experiments, the surface roughness in the Kanto Plain was set at 3.0 m for Rough_URBAN and at 0.001 m for Rough_SEA.

4.4 Evaporative cooling of precipitation (NOEVAP)

In RAINY cases, the evaporative cooling of precipitation is expected to reduce the temperature near the ground and affect coastal fronts. An experiment was conducted without the evaporative cooling of precipitation (NOEVAP).

5. Results of numerical sensitivity experiments

5.1 Quantitative evaluation of the reduction rate of the forecast error

The CTL predictions of the JMA-NHM showed the typical error of coastal fronts shifting inland, similar to the operational MSM predictions. In each sensitivity experiment, we quantitatively evaluated the reduction rate of the forecast error. As indicated below, the error reduction rate was defined as the southeastward deviation of the coastal front predicted by the modified model (“Deviation [Sensitivity – CTL]”) normalized by the northwestward displacement error of the CTL model. (A detailed description of the measurement methodology is provided in Appendix B.)

\[
\text{(Error reduction rate)} = \frac{\text{Deviation (Sensitivity – CTL)}}{\text{Error (CTL)}} \times 100\%.
\]

Following the algorithm described in Appendix B, the Error (CTL) values, representing the deviation of the CTL simulation result from the actual position of the coastal front, were 27.6 km for CASE 1, 15.9 km for CASE 2, and 17.8 km for CASE 3. To illustrate, in CASE 1, if the deviation produced by the modified model is 20 km southeastward, the error reduction rate was calculated as +20 km/27.3 km ≈ +73 %. A reduction rate greater than 100 % indicates an overcorrection of the forecast error; a negative rate indicates an increase in the forecast error.

Table 3 shows the deviations of the predicted coastal fronts from the CTL (value in parentheses in each box, km) results as well as the error reduction rates (upper value in each box, %) for the various sensitivity experiments. The CASE-averaged error reduction rates in the rightmost column give the mean values of the error reduction rates for the three cases. As shown, the CASE-averaged error reduction rates were 27 % in DX_2km, 37 % in DX_1km, 111 % in EO_5km, 101 % in EO_2km, and 87 % in EO_1km. It would appear, then, that the error can be reduced, at least to some extent, by increasing the horizontal resolution, and that it can be all but eliminated by introducing the envelope orography.

In contrast, the CASE-averaged error reduction rates for the Rough_URBAN and Rough_SEA experiments were −2 % and −34 %, respectively, indicating that a change in surface friction did not reduce the forecast error. Indeed, the error increased substantially when surface friction was reduced to that of rough sea. Regarding NOEVAP, the error reduction rates of CASE 2 and CASE 3 (non-RAINY cases) were close to 0, whereas in CASE 1 (RAINY case) it was −72 %.

5.2 Results of EO_5km forecasts

Figure 12 shows the CTL and EO_5km forecasts of the surface temperature and surface wind over the
Kanto Plain for the three cases. In all the EO_5km cases, the coastal fronts shifted seaward in comparison to the CTL result. Figures 12c, 12f, and 12i show the temperature differences between EO_5km and CTL (i.e., EO_5km minus CTL), and suggest that the differences were near zero on both sides of the frontal zone. Thus, the EO_5km experiments indicate that the fronts shifted seaward even though the surface temperature gap across the front was virtually unchanged.

Figure 13 shows a vertical cross-sectional view of the potential temperature predicted by the CTL and EO_5km along the line labeled “Line AB” in Fig. 12. The left-hand (a, d, g) and middle (b, e, h) panels of Fig. 13 show the potential temperature forecast of the CTL and EO_5km, respectively; the right-hand (c, f, i) panels show the potential temperature difference, EO_5km minus CTL. We found that the coastal fronts mark the edge of the cold air “trapped” on the southeastern slope of the mountains. Moreover, EO_5km appears to make the trapped cold air thicker and extend farther southeastward.

5.3 Results of NOEVAP forecast

Figure 14 shows a vertical cross-sectional view of the potential temperature predicted by NOEVAP along Line AB at 03 JST 9 March 2018 in CASE 1. Figure 14a shows the potential temperature of NOEVAP, and Fig. 14b shows the potential temperature difference of NOEVAP minus CTL. Figure 14 shows that the trapped cold air became warmer in the NOEVAP than in the CTL and that the front shifted inland. In other words, the evaporative cooling of precipitation cooled the cold air mass and caused the front to shift seaward.

6. Discussion

To further explore the mechanism behind the shifting of coastal fronts in the sensitivity experiments, we used the Margules equation. According to the American Meteorological Society (2020), this equation accounts for the equilibrium inclination of a frontal surface separating two homogeneous air masses in a steady geostrophic motion parallel to the interface. The equation is given as

$$\tan \alpha = \frac{f}{g} \frac{T_2 v_2 - T_1 v_1}{\Delta T} \approx \frac{f}{g} \frac{\Delta \rho}{\Delta T} (v_2 - v_1),$$

where $\alpha$ is the elevation angle of the frontal surface to the horizontal, $f$ is the Coriolis parameter, $g$ is the gravitational acceleration, and $T_1$ and $T_2$ are the absolute temperatures of the colder and warmer air masses, respectively, with wind speeds of $v_1$ and $v_2$ parallel to the front; $\rho$ and $\Delta \rho$ are the average and difference in the density of the two air layers, respectively. This equilibrium condition has been used to calculate the slope of atmospheric frontal surfaces and explains how the elevation angle of the frontal surface is roughly controlled dynamically.

We applied the Margules equation to coastal fronts where the frontal surface is considered to extend from the mountain ridgeline (Fig. 15a). In the cases discussed in the previous section, the cold air mass trapped on the mountain slope is isolated and is almost stationary ($v_1 \approx 0 \text{ m s}^{-1}$) along the coastal front. On the other hand, the warm air has a strong wind velocity with a large parallel component ($v_2 > 0 \text{ m s}^{-1}$) to the coastal front. To illustrate, consider a mountain barrier with a ridge height of 1000 m and oriented from the northeast to the southwest. If we assume that $f = 10^{-4} \text{ s}^{-1}$, $T_1 = 294 \text{ K}$, $T_2 = 300 \text{ K}$ (i.e., $\rho/\Delta \rho \approx 50$), $v_1 \approx 0 \text{ m s}^{-1}$, and $v_2 \approx 20 \text{ m s}^{-1}$, the $\tan \alpha$ will be approximately 0.01 and the trapped cold air will extend 100 km southeastward from the mountain range. It can be inferred, then, that the fronts shift to the southeast...
Fig. 12. Model-simulated surface temperature (shaded, contoured) and surface wind field of the CTL, EO_5km, and temperature difference of EO_5km minus the CTL in each CASE. The upper panels (a, b, c) are CASE 1 (03 JST 9 March 2018, FT18), the middle panels (d, e, f) are CASE 2 (06 JST 23 February 2017, FT9), and the lower panels (g, h, i) are CASE 3 (03 JST 04 February 2019, FT18). The area shown in gray is at an altitude of 200 m or more above sea level. Line AB shows the location of the vertical cross-section used in Figs. 13 and 14.
Fig. 13. Vertical cross-sectional view of the potential temperature (shaded, contoured at intervals of 1 K) predicted by the CTL (a, d, g) and EO_5km (b, e, h) along Line AB shown in Fig. 12. The right panels (c, f, i) show the potential temperature difference of EO_5km minus the CTL. The topography is marked in black.

Fig. 14. As in Fig. 13, but for the result in the NOEVAP run in CASE 1 at 03 JST 9 March 2018.
(seaward) when the ridgelines of the mountains are higher (Fig. 15b). This would explain why, in experiments EO_5km, EO_2km, and EO_1km, the northwestward displacement error was nearly eliminated, even though the elevation angle of the frontal surface, which is determined by the temperature gap across the front, was unchanged.

Increasing the horizontal resolution (DX_2km, DX_1km) also eliminated the northwestward error, at least to some extent. In fact, changing the horizontal resolution affects not only the representation of the terrain but also the representations of other physical processes (e.g., the turbulent closure and micro-cloud physics schemes). As shown in Appendix C, we confirm that when the 1-km resolution model was given the 5-km resolution terrain, the northwestward error was almost the same as that in the 5-km resolution model. Therefore, we concluded that the reduction of forecast errors that emerged when the horizontal resolution was increased was due to higher mountain barriers.

In the NOEVAP experiment (CASE 1), the coastal fronts shifted to the inland side because of the increase in the elevation angle when the trapped cold air became warmer (Fig. 15c). This implies that any forecast error of physical processes (evaporation, radiation, surface heat exchange, etc.) may increase the systematic location error of coastal fronts by the underestimation of the temperature difference between the two air masses. However, from the statistical validations discussed in Section 3, these processes do not seem sufficient to explain the cause of the systematic error, given the fact that the operational NWP shifts fronts to the inland side without a large bias in the surface temperature prediction of warm and cold air masses.

The validity of the Margules equation was examined in Figs. 13a and 13b. By using the potential temperature of the frontal surface in contact with the ground as a guide (289 K), the slope of the frontal surface (isothermal surface) can be estimated to be about 0.008 (tan \( \alpha \)). If we consider that the mountain barrier has risen by 300 m by installing the envelope orography, the front would be calculated to shift 36
km seaward according to the equation. This theoretical shifting is almost the same as the actual deviation in EO_5km. Of course, the Margules equation is too simple to fully explain the complicated dynamical balance in the actual atmosphere. In reality, the wind is under the influence of the ageostrophic component due to surface friction, turbulent vertical mixing, transient motions, etc. In fact, we confirmed the contribution of surface friction in the Rough_URBAN and Rough_SEA sensitivity experiments. The previous observational analyses (Yoshikado et al. 1994; Seino et al. 2003) revealed that the slope of the frontal surface is steep near the frontal head, while it slopes slightly at the back (inland side) of the front. Besides, Yoshikado (1998) and Seino et al. (2003) indicated that the actual front surface has a transition layer due to turbulent mixing, in which the potential temperature is lower than that of the synoptic inflow, but the wind direction is the same as that of the inflow. Nevertheless, the Margules equation can reasonably explain the results of our sensitivity experiments.

In this study’s sensitivity analyses, we assumed that the trapped cold air of the coastal fronts is stagnant ($v_1 \approx 0$). In fact, the Margules equation can also be applied to CAD events ($v_1 \neq 0$). Appendix D shows a numerical experiment involving a coastal front with typical CAD during a heavy snowfall event over the Kanto Plain in February 2014. As indicated, the experiment with the envelope orography reproduced the position of the coastal front and surface temperature in the frontal zone better than that with the mean orography.

7. Conclusion

We investigated the MSM error in forecasting coastal fronts with southerly onshore winds in the Kanto Plain. The operational NWP is shown to produce a systematic error, consistently shifting coastal fronts to the inland side of their actual positions in forecasts exceeding a forecast period of 5 hours, regardless of precipitation. The numerical experiments conducted in the study suggest that the systematic error associated with coastal fronts can be attributed primarily to the underestimation of the mountain barrier surrounding the Kanto Plain in the model. Most of the coastal fronts were found to take the form of the cold air trapped on the southeastern slope of mountains, where the elevation angle of the frontal surface is roughly controlled dynamically (Margules equation). Thus, the expansion of the trapped cold air toward the Kanto Plain depends on the angle of the frontal surface and the height of the mountain barrier. The NWP with a horizontal grid spacing of 5 km considerably underestimates the mountain barrier because of the smoothing of the mean orography and predicts coastal fronts on the inland side of their actual positions. Even though the frontal positions of the MSM are corrected mostly in the initial condition by assimilating the observation data, they shift to the inland side of their actual positions, reflecting the dynamical balance as the forecast period progresses.

The northwestward distance error of coastal front predictions, averaged over three cases, can be reduced by 27% and 37% by increasing the horizontal resolution from 5 km to 2 km and 1 km, respectively. This is mainly because the mountain barriers become higher with increases in the horizontal resolution. We expect that the forecast error can be further improved by using a finer model grid and finer terrain data.

Using an envelope orography where the elevation at each grid point was defined by the maximum value of GTOPO30 elevation data for that grid nearly eliminated the northwestward error in coastal front predictions. Indeed, this method (i.e., using the maximum value for the model grid) has the disadvantage of making the elevation throughout the entire domain higher than it actually is. Nevertheless, it can be concluded that the envelope orography proposed here is a suitable model terrain to predict coastal fronts in the Kanto Plain more accurately.

The evaporative cooling of precipitation shifts coastal fronts seaward. The mechanism involved in this shift can be simply described: the evaporative cooling decreases the temperature of the cold air mass, reduces the elevation angle of the frontal surface, and shifts the coastal front. This result suggests that any forecast error of physical processes—evaporation, radiation, surface heat exchange, and so on—may contribute to the location error of coastal fronts, depending on the initial and boundary values. However, such processes do not seem sufficient to fully explain the systematic error of coastal front predictions as far as the current mesoscale NWP system is concerned.

In the sensitivity experiments in this study, we were concerned mainly with the trapped cold air of coastal fronts whose wind velocity is very weak. However, the Margules equation can be readily applied to CAD events where the low-level cold air has strong wind velocity. The present study suggests that the use of an envelope orography may improve forecasts of high-impact events, such as the boundary between rain and frozen precipitation and the location of heavy frontal precipitation. The mountain heights surrounding the Kanto Plain are considered to have a greater
impact on the weather system over the Kanto Plain by cold air mass trapping/damming.

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Appendix A: Sampling coastal fronts for statistical verification

A procedure to detect coastal fronts in the Kanto Plain was proposed by Fujibe (1990). The procedure was utilized in the present study, with slight modifications. The algorithm for detection is described from (I) to (III) below. We used the surface temperature and surface wind data from five AMeDAS stations (Miura, Yokohama, Tokyo, Koshigaya, and Kuki; see Fig. 2a). Here, the range of wind direction is indicated clockwise, so “SE–SW” or “between SE and SW” means the range from SE through S to SW.

(I) The wind direction at Miura is between SSE and SSW, with wind speeds of 5 m s\(^{-1}\) or more.

(II) The temperature gap of Miura minus Kuki is +5°C or more.

(III) For the samples detected in (I) and (II), we classified the frontal positions into the section with the largest horizontal temperature gradient (seaward minus landward) among the four sections from seaside to inland (Section 1: Miura–Yokohama; Section 2: Yokohama–Tokyo; Section 3: Tokyo–Koshigaya; and Section 4: Koshigaya–Kuki). We also considered that the temperature-only-based classification might include the false detection of coastal fronts. Thus, samples for which a southerly wind was not detected at the seaward-side station of the classified section was excluded from statistical verification.

As mentioned in Fujibe (1990), the sampling method described above can include the passing of a synoptic-scale front. However, to avoid complicating the experimental setting, this study did not set new conditions to avoid the synoptic-scale front. In addition, the altitude of each AMeDAS point is less than 100 m; thus, the temperature is not corrected by altitude.

The MSM forecast was compared with the samples selected above. We classified the frontal positions of the MSM forecast into the section with the largest temperature gradient among the four sections (Miura–Yokohama–Tokyo–Koshigaya–Kuki), after adding the two conditions described below. As the AMeDAS stations are not necessarily on the MSM grid points, the MSM forecast values of the four grid points closest to the AMeDAS stations were interpolated and taken as the MSM forecast values at the AMeDAS point. The two conditions added were as follows:

1. The wind direction at Miura was not southerly (W–E); coastal fronts are classified on the south side (offshore) of Miura.
2. The wind direction at Kuki was southerly (ESE–WSW), and the wind speed was 3 m s\(^{-1}\) or more; coastal fronts are classified on the north (inland) side of Kuki.

Appendix B: Measuring a southeastward deviation of the coastal front

In each sensitivity experiment, a southeastward deviation of the coastal front from the CTL was measured with the following method:

1. The surface temperature data for all experiments were interpolated into a grid of 2 km.
2. A survey line was plotted in the southeast–northwest direction within the judged area shown in Fig. B1a. The grid point with the largest temperature gradient on the survey line was determined as the frontal position. To prevent false detections associated with the elevation, grids with an altitude of 200 m or more were not judged as frontal positions.
3. The average southeastward deviation from the CTL was measured from all survey lines (Fig. B1b).
4. The deviation of (3) was calculated every hour within the analysis period of each CASE; the time average value was defined as the “deviation (Sensitivity-CTL)”.

Appendix C: Impact study on terrain resolution

An experiment in which only the terrain resolution was changed to 5 km (same as Fig. 10a) in the DX_1km experiment was performed in CASE 1.
Fig. B1. (a) Example of the position of coastal fronts determined by the surface temperature (the CTL is shown by ■, EO_5km is shown by ● (red), and correspond to Figs. 12a (CTL) and 12b (EO_5km), respectively). The determination of the frontal positions was performed inside the judged area shown in the figure. (b) Schematic illustration to measure the deviation of the front. The deviation shown in the figure is calculated as +38.5 km southeastward.

Fig. C1. Model-simulated surface temperature (shaded, contoured) and surface wind field of (a) DX_1km and (b) DX_1km_topo5km at 03 JST 9 March 2018 (FT18). (c) The temperature difference of (b) minus (a).

(Referred to as DX_1km_topo5km experiment). Figure C1 shows the 18-hour forecasts of (a) DX_1km (b) DX_1km_topo5km, and (c) the temperature difference between the two experiments. The coastal front shifted inland in the DX_1km_topo5km compared to DX_1km, and we confirmed that the forecast error of DX_1km_topo5km was nearly the same as that of the CTL.

**Appendix D: Sensitivity experiment to coastal fronts accompanied by CAD**

We conducted a sensitivity experiment for a coastal front accompanied by a typical CAD event in February 2014. As indicated in Fig. D1c (analysis), the coastal front was formed between warm southeasterly winds and the northerly colder flow along the eastern
slope of a mountain barrier. In the numerical experiments (12-hour forecast), EO_5km (Fig. D1b) showed that the northwestward error of the coastal fronts that existed in the CTL (Fig. D1a) was nearly eliminated. Figures D1d and D1e show the surface temperature differences from the analysis. The EO_5km experiment reproduced the surface temperature in the frontal zone better than the CTL experiment.

Fig. D1. Twelve-hour forecasts of the coastal front accompanied by CAD. Upper panels show the surface temperature (shaded, contoured) and surface wind field of (a) the CTL, (b) EO_5km, and (c) analysis (MSM initial condition) at 03 JST 15 February 2014. (d), (e) Surface temperature differences (shaded) of (d) the CTL minus analysis; (e) EO_5km minus analysis. (f) Surface pressure map at 03 JST 15 February 2014. The contour interval is 4 hPa.

References


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