Evaluation of WRF Model Resolution on Simulated Mesoscale Winds and Surface Fluxes near Greenland

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(Manuscript received 28 March 2012, in final form 11 September 2012)

ABSTRACT

Southern Greenland has short-lived but frequently occurring strong mesoscale barrier winds and tip jets that form when synoptic-scale atmospheric features interact with the topography of Greenland. The influence of these mesoscale atmospheric events on the ocean, particularly deep ocean convection, is not yet well understood. Because obtaining observations is difficult in this region, model simulations are essential for understanding the interaction between the atmosphere and ocean during these wind events. This paper presents results from the Weather Research and Forecasting (WRF) Model simulations run at four different resolutions (100, 50, 25, and 10 km) and forced with the ECMWF Re-Analysis Interim (ERA-Interim) product. Case study comparisons between WRF output at different resolutions, observations from the Greenland Flow Distortion Experiment (GFDeX), which provides valuable in situ observations of mesoscale winds, and Quick Scatterometer (QuikSCAT) satellite data highlight the importance of high-resolution simulations for properly capturing the structure and high wind speeds associated with mesoscale wind events and surface fluxes of latent and sensible heat. In addition, the longer-term impact of mesoscale winds on the ocean is investigated by comparison of surface fluxes and winds between model resolutions over a two-month period.

1. Introduction

Greenland’s southeastern coast is one of the windiest locations in the world (Sampe and Xie 2007) because of frequent barrier winds and tip jets that form when synoptic flow interacts with and is deflected by Greenland’s terrain (Renfrew et al. 2008). These powerful winds occur over the ocean in a region important to the downwelling component of the ocean’s meridional overturning circulation (MOC). In the seas around Greenland (Fig. 1), including the Irminger Sea, transfer of heat from warm North Atlantic surface waters to the atmosphere causes ocean water to sink as part of the MOC, an important component of the global heat budget (Bacon et al. 2003; Pickart et al. 2003; Spall and Pickart, 2003; Martin and Moore 2007). Strong winds affect the ocean surface by inducing large sensible and latent heat fluxes from the ocean to the atmosphere, which decreases ocean water buoyancy and, in the right conditions, lead to oceanic convection on horizontal scales of 1–10 km (Gascard and Clarke 1983; Lavender et al. 2002; Pickart et al. 2003; Spall and Pickart, 2003; Martin and Moore 2007).

Understanding the detailed connection between short-lived winds and the longer-term MOC is difficult because of the dearth of in situ atmosphere and ocean observations, and because atmospheric reanalysis and large-scale global climate models do not resolve mesoscale winds well (Kolstad 2008; Sproson et al. 2010). Previous studies have shown that an ocean model driven by large-scale atmospheric model or reanalysis data does not properly represent the ocean mixed layer depth and upper-ocean currents, and that mesoscale atmospheric features, like high-speed tip jets, are important for accurately simulating these ocean features (Våge et al. 2008; Haine et al. 2009). However, because open-ocean convection occurs at so few sites globally, the connection between these winds and the ocean is an important topic of study because of the potential implications for global climate (Bacon et al. 2003; Petersen and Renfrew 2009). Ideally, future model simulations with a high-resolution fully coupled atmosphere and...
ocean would better elucidate the importance of these high-speed mesoscale winds on the MOC.

The three types of high wind speed events that occur near the southeast coast of Greenland are barrier winds (BW), westerly ("forward") tip jets (WTJ), and easterly ("reverse") tip jets (ETJ; Moore 2003). Climatologies of BW and tip jets show that these mesoscale wind events often have surface winds in excess of 25 m s$^{-1}$ and occur frequently during the Northern Hemisphere winter months as synoptic flow from passing cyclones interacts with the high topography of Greenland (Moore and Renfrew 2005; Sproson et al. 2008; Våge et al. 2009; Harden et al. 2011). The BW form when low-level winds run into Greenland’s east coast where steep terrain (Fig. 1) can block the air from flowing over the topography. This blocked air converges along the coast, causing a localized high pressure region (Doyle and Shapiro 1999; Moore and Renfrew 2005). The resulting flow is northerly barrier-parallel winds that are accelerated by the component of the pressure gradient along the jet (Outten et al. 2009; Olafsson and Agustsson 2009).

Tip jets occur over the oceans near Cape Farewell, Greenland, and the location of the cyclone center is a main determinant for which type of tip jet occurs (Moore 2003). An ETJ forms when a cyclone is located south of Cape Farewell, so when barrier flow reaches Cape Farewell the local high pressure decreases and the pressure gradient aligns with the wind direction causing the air to accelerate (Moore and Renfrew 2005; Outten et al. 2009). A WTJ occurs when a cyclone center is located between Iceland and Greenland, and air accelerates as it travels around Cape Farewell (Doyle and Shapiro 1999; Moore and Renfrew 2005; Våge et al. 2009). While these wind events have similarities, they can have differing effects on the ocean because of differences between wind stress curl of the jet, location of maximum winds, and near-surface temperatures (Spall and Pickart 2003; Moore and Renfrew 2005; Sproson et al. 2008; Pickart et al. 2008; Våge et al. 2009; Harden et al. 2011).

Determining the impact of these mesoscale wind events on the ocean is important, yet there are still few in situ observations of BW and tip jets. In the summer of 2004, a surface buoy was sited to measure high-speed winds over the ocean, but it was blown off its mooring before winter when the strongest mesoscale winds occur (Moore et al. 2005). No further standard surface observations are available over the ocean in this region, and land surface stations do not capture mesoscale flows above the ocean (Pickart et al. 2003). Strong winds make scientific ocean cruises difficult and dangerous, and the authors know of only one aircraft campaign studying these mesoscale flows in the region (Renfrew et al. 2008).

As a result of a dearth of observations, modeling studies are an important tool for understanding the connection between the ocean and mesoscale winds.

Many studies have been conducted to understand flow in complex terrain. These studies have demonstrated that high-resolution numerical models are important for capturing topographic effects such as low-level gravity wave breaking and associated turbulence around Greenland, Iceland, and Norway (Olafsson and Agustsson 2009; Agustsson and Olafsson 2007; Barstad and Grønås 2005), increased surface pressure drag by mountains (Smith et al. 2006), and small-scale variability in surface fluxes and wind speeds in Spitsbergen (Kilpeläinen et al. 2011; Reeve and Kolstad 2011). These studies all use horizontal resolutions between 1 and 10 km and clearly demonstrate the importance of high resolution for accurately modeling wind in complex terrain that can be important for forecasting.

However, to better understand the detailed coupling between the ocean and atmosphere during strong winds it is important to force the ocean with realistic winds, yet no large-scale dataset at sub-10-km resolution for forcing ocean models exists to the authors' knowledge. Indeed, because of data constraints with large datasets of high-resolution data, many ocean models still use coarse reanalysis or global atmospheric model data for forcing (Hunke 2010; Kwok et al. 2008; Hunke and Holland 2007; Lisaeter et al. 2007). Previous work has created methods to parameterize mesoscale winds for ocean models driven by reanalysis (Sproson et al. 2010), yet to tease out details in the atmosphere–ocean exchanges, explicit representation of mesoscale winds is ideal. In addition, many tip jet and barrier wind climatologies have used relatively low-resolution reanalysis data from the National Centers for Environmental Prediction (NCEP; Moore 2003; Reeve and Kolstad 2011), the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis Interim (ERA-Interim; http://www.ecmwf.int/research/era/do/get/era-interim; Harden et al. 2011), or the 40-yr ECMWF Re-Analysis (ERA-40; http://www.ecmwf.int/research/era/do/get/era-40; Våge et al. 2009). The Quick Scatterometer (QuikSCAT) climatologies (Moore and Renfrew 2005; Kolstad 2008; Reeve and Kolstad 2011) are problematic in the Arctic where sea ice concentration impacts data availability, and some wind events may not be captured in the satellite data. As such, using high-resolution models might provide a more accurate climatology of high wind events throughout the Arctic, and this paper will address model resolutions between current model scales (~100 km) and very high-resolution studies (~10 km) to better understand what resolution is necessary to capture mesoscale wind events but is also data efficient enough to be used for forcing ocean models or creating multidecadal climatologies of mesoscale winds.
This paper will use several case studies to compare simulations from the Weather Research and Forecasting (WRF) Model during an ETJ and several BW events with observational data from an aircraft campaign and with QuikSCAT surface wind satellite data. These case study comparisons emphasize the importance of high-resolution numerical simulations for accurately representing the horizontal and vertical wind structure and maximum speeds of mesoscale winds. Section 2 describes the numerical modeling methods and observational data used in the analysis. Section 3 illustrates the results from case study comparisons of mesoscale ETJ and BW events between WRF and the satellite and aircraft observations. Section 4 presents a two-month simulation comparison between WRF and QuikSCAT winds as well as longer-term differences between WRF at different resolutions. Finally, section 5 offers conclusions and directions for future study.

2. Data and methods

a. WRF Model simulations

The WRF Model is a regional mesoscale atmospheric model developed for research and forecasting applications (Skamarock et al. 2008). The WRF model has an effective resolution of $7\Delta x$, where $\Delta x$ is the model’s horizontal grid spacing (Skamarock 2004). The focus of this study is to determine what atmospheric resolution best captures mesoscale wind features by running the Advanced Research WRF (ARW) model, V3.2.1, at four different horizontal grid increments: 10, 25, 50, and 100 km (WRF10, WRF25, WRF50, and WRF100, respectively). The 10-km lower limit was chosen because high-resolution features are known to be important for forcing the ocean, but atmospheric resolutions greater than 10 km are still too computationally expensive for multidecadal fully coupled model simulations that would more fully explore the connection between atmosphere and ocean during these events. The 100-km grid increment provides a baseline for WRF performance at the scale of global models or reanalyses. We have not included WRF100 figures in the remaining analysis because WRF’s effective resolution suggests that tip jets, which have a scale of a few hundred kilometers, would need to be modeled with no less than 50-km horizontal grid increment, so we have focused our discussion on the WRF10, WRF25, and WRF50 simulations. Inland, the maximum terrain heights were generally higher for high-resolution simulations in southern Greenland with WRF10 having a maximum terrain height of about 3260 m (WRF50: 3100 m) near the Geikie Plateau and 2785 m (WRF50: 2761 m) near Cape Farewell. Along the coasts, the slope is steeper in the 10-km model configuration than in the 50-km configuration; near the Geikie Plateau the slope for WRF10 was about 1.5 times greater than for WRF50, and near Cape Farewell the WRF10 slope was 1.6 times greater than for WRF50.

The domain for all WRF simulations is shown in Fig. 1, and all WRF simulations had a 10-hPa model top and 40 vertical levels, about 10 of which are in the lowest 1 km of the atmosphere. The ERA-Interim (see section 2b) was used to specify the WRF lateral boundaries every 6 h. Sea ice was specified from the National Snow and Ice Data Center (NSIDC) bootstrap sea ice concentration satellite product, which has a $0.25^\circ \times 0.25^\circ$ resolution and provides sea ice fraction as a WRF lower boundary condition (Comiso 1999). For grid cells that contain sea ice, WRF calculates surface variables over the open ocean and sea ice portion of the grid cell separately and then calculates an area-weighted average over the entire grid cell (Bromwich et al. 2009). We used WRF physics options (Table 1) found to maximize performance in a pan-Arctic domain (Cassano et al. 2011). For the case studies presented here, WRF was initialized two days before the date of aircraft or satellite observations (Table 2), and the model provided hourly output. Additionally, we performed continuous
1 February–31 March 2007 simulations on the same domain for each WRF resolution to evaluate longer-term atmospheric differences due to resolution; these were initialized on 1 February and then run only with lateral boundary forcing until 31 March. No nesting was used in any WRF simulation, and the box in Fig. 1 shows the area of particular interest for this study.

We calculated the relative vorticity from QuikSCAT observations and wind stress curl from WRF 10-m winds to better understand how the different resolutions were able to capture the narrow jet features. Wind stress curl, like relative vorticity, is a measure of the change in horizontal wind speed over a horizontal distance and large values of wind stress curl or relative vorticity indicate strong horizontal wind shear. Furthermore, positive wind stress curl causes ocean upwelling from Ekman pumping and could lead to preconditioning of the water column for convection. In addition, to understand the forcing and dynamical differences between the different resolution simulations, we calculated the pressure gradient (PG), Coriolis, advection, and acceleration terms of the horizontal momentum equations (Holton 2004) with the remaining terms determined as a residual following the procedure of Nigro et al. (2012).

Grid points with terrain above 100 m were excluded because of errors associated with calculating derivatives over the terrain along Greenland’s southeast coast, and the time derivative in the acceleration term was calculated using a forward finite difference with model output from 0 and +1 h relative to the time of interest.

Over the ocean, the surface layer physics scheme calculates sensible and latent heat fluxes, each with units of watts per square meter, that describe energy exchange between the surface and the atmosphere (Skamarock et al. 2008). The sensible heat flux is given by

$$SH = \rho C_p u_\theta \theta_\phi,$$

where \(\rho\) is air density, \(C_p\) is a dimensionless transfer coefficient, \(u_\theta\) is the friction velocity, and \(\theta_\phi\) is a temperature gradient term. The friction velocity is given by

$$u_\theta = \frac{k V_v}{\ln \left( \frac{z_v}{z_0} \right) - \Psi},$$

where \(V_v\) is the wind speed at the lowest model level, \(z_v\) is the height of the lowest eta level, \(z_0\) is the momentum roughness length, and \(\Psi\) is a stability function. The temperature gradient term is given by

$$\theta_\phi = \frac{k \Delta \theta}{\ln \left( \frac{z_v}{z_{0h}} \right) - \Psi},$$

where \(z_v\) and \(\Psi\) are the same as in Eq. (2), \(\Delta \theta\) is the temperature difference between the surface and lowest model level, and \(z_{0h}\) is the heat roughness length (Dudhia 2010). Similarly, the latent heat flux is given by

$$LH = \rho u_\theta q_\phi,$$

where \(\rho\) and \(u_\theta\) are the same as in Eq. (1), and \(q_\phi\) is a moisture gradient term given by

$$q_\phi = \frac{k \Delta q}{\ln \left( \frac{z_0}{z_{0q}} \right) - \Psi},$$

where \(z_v\) and \(\Psi\) are the same as in Eq. (2), \(\Delta q\) is the mixing ratio difference between the surface and lowest model level, and \(z_{0q}\) is the moisture roughness length (Dudhia 2010). However, the MYJ parameterization uses a single value for the momentum, heat, and moisture roughness lengths. Most bulk aerodynamic formulas have a form similar to Eqs. (1) and (4), and differ in the way that they calculate \(\Psi\) or other transfer coefficients.

### Table 1. WRF physics options (Cassano et al. 2011): turbulent kinetic energy (TKE), Rapid Radiative Transfer Model for GCMs (RRTMG), Goddard Cumulus Ensemble (GCE), and the Noah land surface model (LSM).

<table>
<thead>
<tr>
<th>Planetary boundary layer</th>
<th>Surface layer</th>
<th>Radiation</th>
<th>Microphysics</th>
<th>Cumulus</th>
<th>Land surface</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mellor–Yamada–Janjic (MYJ) TKE</td>
<td>Monin–Obukhov (Janjic Eta)</td>
<td>RRTMG (both SW and LW)</td>
<td>GCE</td>
<td>Grell–Devenyi</td>
<td>Noah LSM</td>
</tr>
</tbody>
</table>

### Table 2. Observational data availability.

<table>
<thead>
<tr>
<th>Day</th>
<th>Wind feature</th>
<th>GFDex flight No.</th>
<th>GFDex dropsondes</th>
<th>GFDex surface obs</th>
<th>QuikSCAT satellite obs</th>
</tr>
</thead>
<tbody>
<tr>
<td>21 Feb</td>
<td>Easterly tip jet</td>
<td>B268</td>
<td>1400 UTC</td>
<td>—</td>
<td>0700 UTC, 2200 UTC</td>
</tr>
<tr>
<td>2 Mar</td>
<td>Barrier wind</td>
<td>B274</td>
<td>1200 UTC</td>
<td>1400 UTC</td>
<td>0700 UTC, 2200 UTC</td>
</tr>
<tr>
<td>5 Mar</td>
<td>Barrier wind</td>
<td>B276</td>
<td>—</td>
<td>1400 UTC</td>
<td>0700 UTC, 2200 UTC</td>
</tr>
<tr>
<td>6 Mar</td>
<td>Barrier wind</td>
<td>B277</td>
<td>1500 UTC</td>
<td>1100 UTC</td>
<td>0700 UTC, 2200 UTC</td>
</tr>
<tr>
<td>9 Mar</td>
<td>Barrier wind</td>
<td>B278</td>
<td>—</td>
<td>1200 UTC</td>
<td>0700 UTC, 2200 UTC</td>
</tr>
</tbody>
</table>
b. ERA-Interim

In addition to being used as the WRF initial and lateral boundary conditions the ERA-Interim (or ERA-I) data were also compared with WRF simulations. When compared to the previous generation ERA-40, the ERA-I data have a longer time range, more vertical levels, a higher resolution (~79 km), assimilated scatterometer winds (including QuikSCAT), and different physics parameterizations (Dee et al. 2011). The ERA-I data used in this study were gridded to a 1.5° × 1.5° resolution, 37 vertical pressure levels, about 6 of which are in the lowest 1 km of the atmosphere, and a 0.1-hPa model top (ECMWF). Data are available in 6-hourly increments (0000, 0600, 1200, and 1800 UTC), and the nearest ERA-I output time was chosen to compare with observations (0000, 0600, 1200, and 1800 UTC), and the nearest ERA-I data were also compared with WRF simulations. When compared to the previous generation ERA-40, the ERA-I data have a longer time range, more vertical levels, a higher resolution (~79 km), assimilated scatterometer winds (including QuikSCAT), and different physics parameterizations (Dee et al. 2011). The ERA-I data used in this study were gridded to a 1.5° × 1.5° resolution, 37 vertical pressure levels, about 6 of which are in the lowest 1 km of the atmosphere, and a 0.1-hPa model top (ECMWF). Data are available in 6-hourly increments (0000, 0600, 1200, and 1800 UTC), and the nearest ERA-I output time was chosen to compare with observations (0000, 0600, 1200, and 1800 UTC), and the nearest ERA-I output time was chosen to compare with observations. However, surface flux data are only available every 12 h (0000 and 1200 UTC), and these data were not used because these times often differed by more than 2 h from observations.

c. QuikSCAT satellite

The QuikSCAT satellite provides high-resolution surface wind data over the global oceans during two passes each day with an 1800-km-wide swath (Sampe and Xie 2007; Dunbar and Perry 2001) from 1999 to 2009, when the scatterometer failed (Dee et al. 2011). Microwave radiation emitted from the satellite is backscattered off the ocean surface, and wind stress and wind vectors can be calculated based on the surface roughness. However, QuikSCAT is unable to provide observations in locations with heavy rainfall or over land, sea ice, or in the marginal ice zone because they do not have wind-altered surface roughness, so these areas are masked in the QuikSCAT product. Nevertheless, QuikSCAT is a valuable resource for verifying the spatial details of the simulated wind field. Table 2 lists the average time of the QuikSCAT passes used for comparison to WRF and ERA-I.

Two retrieval algorithms, the National Aeronautics and Space Administration direction interval retrieval with thresholded nudging (NASA-DIRTH) and Remote Sensing Systems (RSS), are available for surface wind speed. Previous studies comparing aircraft and buoy observations with the QuikSCAT products found that both algorithms appear to have positive wind speed biases (Moore et al. 2008; Renfrew et al. 2009a,b). Over all types of observations, the NASA-DIRTH bias ranged from +1.9 to 2.6 m s⁻¹, and this algorithm had a more conservative sea ice mask than RSS. The RSS bias was from +2.3 to 3.3 m s⁻¹, and this algorithm particularly overestimated speeds in high wind speed conditions. This study uses NASA Jet Propulsion Laboratory’s level 3 daily ocean wind vector data that have been gridded to 0.25° (~25 km) spacing and derived with the NASA-DIRTH algorithm (Dunbar and Perry 2001). These data have been flagged for land, sea ice, and heavy rain, so data gathered in these locations is flagged as missing in the dataset and the remaining data are quality controlled and located over the open ocean (Dunbar and Perry 2001).

We calculated the relative vorticity of the QuikSCAT 10-m winds to evaluate the wind speed gradients associated with ETJ and BW and for comparison with the WRF wind stress curl. We linearly interpolated the missing wind values over the open ocean before calculating the relative vorticity. We did not calculate wind stress curl for QuikSCAT because wind stress curl requires use of the drag coefficient. In the following discussion we have compared the QuikSCAT relative vorticity with the WRF wind stress curl. While these two variables are not the same, they are both strongly influenced by horizontal wind gradients and comparing the horizontal scale of QuikSCAT vorticity with WRF wind stress curl gives an indication of the spatial scale of the horizontal wind shear features in the mesoscale winds of interest for this study.

d. Greenland Flow Distortion Experiment

The Greenland Flow Distortion Experiment (GFDex) was the first aircraft campaign to obtain observations of an ETJ and BW near the southeast Greenland coast (Renfrew et al. 2008). GFDex flew a total of 12 missions between 21 February and 10 March 2007, and of these missions, 1 observed an ETJ while 4 others observed a BW (Table 2). We use GFDex dropsonde measurements, which provide vertical information about winds, moisture content, and equivalent potential temperature, and observations of near-surface variables and fluxes (Renfrew et al. 2009a,b; Petersen and Renfrew 2009; Petersen et al. 2009). These observations are important not only because they provide the first in situ data of these mesoscale wind events, but also because they provide variables besides wind speed that are important for ocean–atmosphere interaction. Complete flight paths and observations can be found in Renfrew et al. (2009a) and Petersen et al. (2009), but the dropsonde flight legs used for comparison with WRF in this paper are shown on Fig. 2.

The aircraft and dropsonde observations of jet vertical structure as well as fluxes provide some of the only in situ data for these high-wind events, and these data will be a basis for comparison in the rest of this paper. Vertical cross sections, interpolated from dropsonde observations (Petersen et al. 2009; Renfrew et al. 2009a), are compared with the simulated vertical structure in WRF. GFDex near-surface observations were taken at an average flight altitude of 39 m above mean sea level with a total range

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of 32–51 m. Surface temperature values are derived from a downward-looking Heimann radiometer, while the surface mixing ratio, 2-m mixing ratio, 2-m temperature, and 10-m winds were interpolated from flight altitude using the Coupled Ocean–Atmosphere Response Experiment, version 3.0 (COARE 3.0), bulk flux algorithm (Renfrew et al. 2009b). Following GFDex convention, in situ covariance measurements of turbulent fluxes were not compared with WRF Model data because of large variability in the observed fluxes (Renfrew et al. 2009b; Petersen and Renfrew 2009); instead, latent and sensible heat fluxes were calculated using both the COARE 3.0 (in 2003) and Smith (1988) bulk flux algorithms. Only the COARE value will be shown here since the fluxes calculated using each algorithm match well (Renfrew et al. 2009b). The COARE 3.0 algorithm has been adjusted for use in higher wind speed conditions and uses separate values for momentum, heat, and moisture roughness lengths (Smith 1988; Fairall et al. 2003).

To compare with aircraft observations, WRF output was interpolated to the flight latitudes and longitudes using a distance-weighted average of the four WRF grid points closest to the flight path. For cross sections, the horizontally interpolated WRF output was linearly interpolated to regular vertical levels. Equivalent potential temperature for the WRF cross sections was calculated from the WRF temperature, humidity, and pressure. For flux comparisons, the surface variables, $q_{2m}$, $T_{sfc}$, and $T_{2m}$, were direct model output; surface saturation mixing ratio, $q_{sfc}$, was calculated using the surface temperature and sea level pressure. For the four near-surface flights, an average root-mean-square error (RMSE) and correlation were found between the near-surface variables for each WRF simulation and the GFDex observations to evaluate how well the WRF simulations captured the GFDex observations.

### 3. Case studies

#### a. Overview of case studies

Detailed analysis of the synoptic situation of each of these case studies has been explored in detail in the
literature. The following descriptions are meant to provide a brief description of each case for this study, but more information can be found in Renfrew et al. (2008), Petersen et al. (2009), Petersen and Renfrew (2009), Renfrew et al. (2009a,b), and Outten et al. (2009).

GFDex observed one ETJ in February 2007 and BWs in the Denmark Strait during four days in March 2007. Figure 2 shows the QuikSCAT 10-m winds for each of the five days examined and provides a snapshot of the synoptic situation during the mesoscale wind event present at the particular time when QuikSCAT observations were available. On each of the BW days, the QuikSCAT sea ice mask obscures coastal regions with strong wind speeds, but on some of these days GFDex dropsonde observations over the marginal ice zone provide information about the jet in this region. Dropsonde flights that will be discussed in this paper are marked on Fig. 2; “A” marks the starting point (0 km) for cross sections created from the dropsonde data. Additionally, on the BW days GFDex flights observed surface variables and fluxes.

On 21 February 2007 (Fig. 2a) there was a strong, distinct ETJ present during the 0700 UTC QuikSCAT satellite pass that extended from Cape Farewell over the southeast Labrador Sea. This ETJ was present, though weakening, during the 2200 UTC pass (not shown). GFDex dropsonde observations were available at 1500 UTC along the marked flight path near Cape Farewell, but in situ surface data are not available.

At 2200 UTC 2 March there was a BW to the southwest of the Denmark Strait, along the Greenland Coast (Fig. 2b). A strong BW in the Denmark Strait existed at 2200 UTC 5 March (Fig. 2c) and at both 0700 (not shown) and 2200 UTC 6 March (Fig. 2d). The 5 March QuikSCAT sea ice mask covers half of the Denmark Strait (Fig. 2c), while the 6 March sea ice mask covers the entire Denmark Strait (Fig. 2d), possibly from sea ice advection by the strong winds. By 0700 UTC 9 March (Fig. 2e) the BW was weaker.

The following sections will discuss each flow regime—ETJ and BW—in more detail with particular emphasis on the low-level winds, boundary layer structure, and surface fluxes because these are particularly important for understanding how the atmosphere and ocean interact. Exemplary comparisons have been chosen to highlight relevant differences for the case study days between the different resolution WRF simulations.

b. Easterly tip jet regime

Comparison between QuikSCAT observed winds, ERA-I, and WRF reveals differences in maximum wind speeds and shape of the ETJ due to resolution. The QuikSCAT observations (Fig. 3a) show a sharp distinction between the windward east coast of Greenland with maximum ETJ speeds of 42 m s\(^{-1}\) over the ocean and the leeward west coast that has wind speeds below 10 m s\(^{-1}\). The WRF10 (Fig. 3b) and WRF25 (Fig. 3c) simulations show a discrete ETJ that closely matches QuikSCAT observations in location, shape, and length, but WRF10 has 35 m s\(^{-1}\) maximum wind speeds, while WRF25 has 32 m s\(^{-1}\) maximum speed. As resolution decreases, the ETJ becomes less distinct and the maximum wind speeds over the ocean decrease, and there is a less distinct lee region on the west side of Greenland, particularly in WRF50 (Fig. 3d), WRF100 (not shown), and ERA-I (not shown).

Using the half-width of the tip jet as defined from QuikSCAT, we found the wind speed gradient for each simulation starting at the maximum wind speed of each simulated jet. We found that the wind speed gradient in QuikSCAT was about 1.6 times greater than the gradients in WRF10 and WRF25, 1.7 times greater than the WRF50 gradient, 4 times greater than the WRF100 gradient, and 3.5 times greater than the ERA-I gradient. This means that the low-resolution ERA-I and WRF100 simulations were not able to capture the observed narrow, high-speed jet as well as the higher-resolution WRF simulations. Additionally, the QuikSCAT relative vorticity (Fig. 3e) shows the small spatial scale of strong wind shear around the ETJ. The wind stress curl for WRF10 (Fig. 3f) and WRF25 (Fig. 3g) have a similar spatial scale to the relative vorticity shown by QuikSCAT. In addition, the magnitude of wind stress curl in WRF10 and WRF25 had a more detailed structure and was about 2 times larger than WRF50 (Fig. 3h) and 10 times larger than WRF100 (not shown). The high-resolution WRF simulations better capture the strong, narrow jet and the sharp wind speed drop from the center to the edge of the jet.

The sea level pressure field and calculated PG (not shown) during the ETJ illustrate the importance of correctly representing the terrain slope. Comparison between the WRF10 and WRF100 sea level pressure fields shows that the WRF10 pressure along the coast is higher when compared with WRF100, and the result is that the PG momentum term is larger and nearer the coast in WRF10 than WRF100. While the wind fields are similar between WRF10 and WRF25, the PG fields show that there is more detailed structure in the WRF10 field, particularly near terrain and in the lee of Iceland. Differences in terrain height between the WRF resolutions are most pronounced along the coast, and the lower-resolution simulations lose the sharp elevation gradient near the coast necessary for proper flow blocking. WRF simulations run at 10 km but with 100-km terrain heights imposed (not shown) have reductions in ETJ wind
FIG. 3. ETJ 10-m wind speed (m s\(^{-1}\)) at 0700 UTC 21 Feb 2007 from (a) the QuikSCAT satellite, (b) WRF10, (c) WRF25, and (d) WRF50. (e) The relative vorticity (10\(^{-5}\) s\(^{-1}\)) corresponding to the 10-m winds for the QuikSCAT satellite, and wind stress curl (10\(^{-2}\) N m\(^{-1}\)) for (f) WRF10, (g) WRF25, and (h) WRF50. The QuikSCAT wind field in (a) has been linearly interpolated to replace missing data and this interpolated field was used when calculating (e) the QuikSCAT relative vorticity.
speeds by 6–10 m s⁻¹ and reductions in both sensible and latent heat fluxes by 100–200 W m⁻². The simulations emphasize that having both high resolution to adequately represent atmospheric processes and the correct terrain heights and slopes is important for simulating observed wind speeds. Without the necessary barrier the winds are effectively smeared over land and are not confined to the coast where the localized high pressure causes the high-speed ETJ as seen in the QuikSCAT data.

Cross sections of wind speed and equivalent potential temperature with height, from the GFDex dropsondes, show the vertical structure of the jet. The GFDex-observed winds (Fig. 4a) indicate that the jet core, with speeds of 48 m s⁻¹, is located approximately 800 m above sea level. With increasing distance from the coast the speed of the jet decreases, and aloft the jet core tilts slightly upward and away from the coast. The WRF output shows distinct differences with the simulated jet between the different model resolutions used. The simulated winds from the WRF10 (Fig. 4b), WRF25 (Fig. 4c), and WRF50 (Fig. 4d) simulations have jet cores near 800 m with 46 m s⁻¹ maximum wind speed for WRF10 and WRF25 and 48 m s⁻¹ for WRF50, while the WRF100 jet (not shown) has a higher core and lower core speeds. In addition, the WRF10, WRF25, and WRF50 simulations capture the shape, tilt, extent, and speeds of the jet core with distance from the coast shown from GFDex.

The GFDex equivalent potential temperature observations (Fig. 5a) show an unstable surface layer, topped by an 800-m-deep well-mixed boundary layer near the coast that increases in depth away from the coast. Additionally, above 800 m a strong vertical equivalent potential temperature gradient marks an inversion above the jet. WRF10 (Fig. 5b), WRF25 (Fig. 5c), and WRF50 (Fig. 5d) all capture the unstable surface layer and boundary layer depth, but WRF10 best captures the strong inversion above 800 m as well as the horizontal equivalent potential temperature gradient. One key difference between GFDex and all WRF simulations is that WRF has a cold equivalent potential temperature bias ranging from 6 (WRF10 and WRF25) to 4 K (WRF50).

However, the physical structure of the boundary layer in the WRF simulations match observations, which implies that the simulations are correctly replicating the fluxes.
and associated surface layer temperature gradients necessary to create the unstable surface layer and mixed layer depths in spite of the temperature bias.

Comparisons between simulated surface sensible (Fig. 6) and latent (not shown) heat fluxes reveal differences between resolutions in the vicinity of the ETJ. The WRF10 (Fig. 6a) and WRF25 (Fig. 6b) simulations’ maximum surface sensible heat fluxes are both above 400 W m$^{-2}$ and are directly collocated with the ETJ. In contrast, the WRF50 (Fig. 6c) sensible heat fluxes do not exceed 200 W m$^{-2}$ in the ETJ region and the WRF100 fluxes are the same in the ETJ region as the surrounding the Irminger Sea (not shown), which has strong implications for air–sea exchanges in the ETJ vicinity. In addition, the large fluxes occur where we find large positive wind stress curl, and the resulting upwelling would shoal the thermocline and lead to a shallower surface layer that would need to lose less energy before ocean convection could take place. Similar differences in flux magnitude can be seen just south of Iceland where strong winds to the south of the island drive larger fluxes in higher-resolution WRF simulations. The maximum fluxes for the high-resolution simulations are more physically plausible because bulk flux formulas show that the magnitude of both sensible and latent heat fluxes depends linearly on the low-level wind speed. Thus, all other physical inputs being approximately equal (not shown, although verified), one would expect both latent and sensible fluxes to peak with highest surface winds. Thus, it seems reasonable that the large heat fluxes collocated with high-speed winds in the high-resolution simulations are more realistic, and over a long period of time the increased fluxes would have a greater impact on the ocean than the fluxes in the low-resolution simulations.

c. Barrier wind regime

At 2200 UTC 5 March, QuikSCAT observations identified an area with wind speeds greater than 44 m s$^{-1}$ in the Denmark Strait north of Iceland (Fig. 7a). WRF10 (Fig. 7b) and WRF25 (Fig. 7c) simulate peak BW speeds between 34 and 36 m s$^{-1}$, but the WRF50 (Fig. 7d) and WRF100 (not shown) maximum wind speed is less than 33 m s$^{-1}$ and the strong winds do not extend as far south as those in the high-resolution simulations. In WRF10, there is a wake region south of the plateau that becomes

![Fig. 5. Cross section of equivalent potential temperature (K) from GF Dex flight path at 1400 UTC 21 Feb 2007 from (a) the GF Dex dropsonde observations, (b) WRF10, (c) WRF25, and (d) WRF50. The “A” in Fig. 2a marks 0 km on the cross section and the “B” marks 150 km on the cross section.](image-url)
less pronounced with decreasing resolution. The directions of the winds are also subtly different: WRF10 has an easterly component to the winds that is not as strong for WRF25, WRF50, or WRF100, implying that in WRF10 the air is being forced through the Denmark Strait instead of flowing over the Geikie Plateau (elevation ~2 km) to the north.

Like the ETJ case, the wind stress curl has a greater magnitude and more detailed structure in WRF10 (Fig. 7f) than WRF25 (Fig. 7g), WRF50 (Fig. 7h), or WRF100 (not shown). The high-resolution WRF10 wind stress curl field has features that are similar in size to the QuikSCAT (Fig. 7e) relative vorticity field. This reinforces that the wind gradients and jet locations are better simulated in high-resolution WRF. In addition, the WRF10 PG momentum term (not shown) has strong pressure gradients around the Geikie Plateau that would lead to strong winds, but the structure is less defined and the PG is weaker in the other WRF simulations. Thus, lower-resolution WRF have less distinct and weaker BW because of flow over the Geikie Plateau instead of around the plateau due to blocking similar to the ETJ. While these spatial wind field features are not verifiable with QuikSCAT because of the sea ice mask, the physical explanation of reduced terrain blocking in low-resolution WRF makes sense and would result in weaker PG forces, weaker maximum winds, and an indistinct lee.

In addition to QuikSCAT observations, GFDex dropsonde observations help illustrate how the different resolutions simulate the BW. Comparison between GFDex observations and WRF output show that low-resolution WRF does not distinguish between barrier and downslope flow, two flow regimes forced by different dynamics. On 6 March, observations in the north Denmark Strait reveal a BW with maximum speeds of 35 m s\(^{-1}\) hugging the coast below 1000 m and a second jet over the land (Fig. 8a). WRF10 (Fig. 8b) simulates the primary jet core to be approximately the same height and distance off the coast as observations, though it has maximum wind speeds of 45 m s\(^{-1}\). In WRF10, like in the observations, the secondary jet over land is distinct from the oversea BW, but the other WRF simulations (Figs. 8c,d) merge the two jets and have lower total wind speeds.

Breaking down the total wind into flow parallel and perpendicular to the cross section reveals that these two jets are different flow regimes that WRF is not able to separate in low-resolution simulations. The parallel component shows that WRF10 (Fig. 8e) and WRF25 (not shown) clearly have strong westerly downslope flow tied to Greenland’s topography, but WRF50 (Fig. 8f) shows weaker downslope flow. In WRF10 (Fig. 8g) the
FIG. 7. Barrier wind 10-m wind speed (m s\(^{-1}\)) at 2200 UTC 5 Mar 2007 from (a) the QuikSCAT satellite, (b) WRF10, (c) WRF25, and (d) WRF50. (e) The relative vorticity (10\(^{-5}\) s\(^{-1}\)) corresponding to the 10-m winds for the QuikSCAT satellite, and wind stress curl (10\(^{-2}\) N m\(^{-1}\)) for (f) WRF10, (g) WRF25, and (h) WRF50. The QuikSCAT wind field in (a) has been linearly interpolated to replace missing data and this interpolated field was used when calculating (e) the QuikSCAT relative vorticity.
FIG. 8. Cross section of wind speed (m s\(^{-1}\)) from GFDex flight path at 1500 UTC 6 Mar 2007. Total wind speed from (a) GFDex dropsonde observations, (b) WRF10, (c) WRF25, and (d) WRF50. Cross-sectional parallel flow for (e) WRF10 and (f) WRF50 and cross-sectional perpendicular flow for (g) WRF10 and (h) WRF50. For the cross-sectional parallel figures easterly winds are shown in blue and westerly winds in orange. For the cross-sectional perpendicular figures northerly winds are shown in blue and southerly winds are in orange. The "A" in Fig. 2d marks 0 km on the cross section and the "B" marks 280 km on the cross section.
BW dominates the southerly perpendicular flow over the ocean, while a smaller alongslope jet is present along the Greenland terrain. WRF25 (not shown) and WRF50 (Fig. 8h) combine the barrier and downslope flow perpendicular to the cross section into a weaker jet.

The equivalent potential temperature cross sections in the BW regimes indicate that high-resolution WRF simulates more realistic boundary layer structure than low-resolution WRF. On 2 March, GFDex shows the deepest boundary layer depth as approximately 2 km over the sea, and the boundary layer height decreases toward the Greenland coast (Fig. 9a). WRF10 (Fig. 9b), WRF25 (Fig. 9c), and WRF50 (Fig. 9d) all capture the boundary layer depth along the cross section, but WRF10 best matches observed surface horizontal gradient of equivalent potential temperature and vertical equivalent potential temperature gradient.

A combination of WRF simulated surface fluxes and in situ observations reveal important differences between WRF-simulated fluxes due to resolution. At 2200 UTC 6 March, QuikSCAT observations show a strong BW in the Denmark Strait (Fig. 2d). The surface sensible (Fig. 10) and latent (not shown) heat fluxes on this day reflect the large differences in wind speed across the WRF resolutions. WRF10 (Fig. 10a) produced sensible heat fluxes greater than 500 W m$^{-2}$ along the marginal ice zone, while WRF25 (Fig. 10b), WRF50 (Fig. 10c), and WRF100 (not shown) have fewer grid cells with large sensible heat fluxes. High-resolution WRF was better able to represent the marginal ice zone and thus had more grid cells with open or partially ice-covered ocean. So in WRF10 more open water was exposed to strong winds and large temperature and moisture gradients, and resulted in a larger ocean surface area covered by high fluxes when compared with the other WRF simulations. While the WRF simulations have similar sea ice extent, the high-resolution simulations are better able to capture the gradient from full ice cover to open ocean (not shown). The better simulated wind field and marginal ice zone in WRF10 allows for more grid points with high positive fluxes because where the highest winds are located more of the ocean surface is totally or partially ice free.

GFDex took near-surface observations during the four BW flights. The 2, 6, and 9 March BW flights took place over open ocean, while the 5 March flight was over

![Figure 9](image_url)
open water and the marginal ice zone (Renfrew et al. 2009b). Data from the 2 and 5 March flights and simulations are shown in Fig. 11. On 2 March there is a large difference between the WRF and GFDex surface mixing ratio and temperature (Figs. 11a,d). On 2 March, the 2-m mixing ratio and temperature best match between WRF10 and GFDex (Figs. 11b,e); however, the surface bias results in large positive mixing ratio and temperature vertical gradients for WRF (Figs. 11c,f), with the WRF10 gradients most closely matching the GFDex gradients. Additionally, on 2 March the 10-m simulated wind speed (Fig. 11g) from WRF10 and WRF25 better represent the higher speeds from GFDex. On 5 March the surface and 2-m variables and the wind speeds have less spread between resolutions and generally match the GFDex observations at the surface, 2 m, and 10 m.

The average RMSE and correlation between the model simulations and the GFDex observations are given in Table 3, and bolded RMSE values indicated RMSE values for WRF25, WRF50, and WRF100 that are significantly different than the WRF10 RMSE at the 95% confidence level. The mixing ratio gradient and temperature gradients have significantly different RMSE values, and decrease with increasing WRF resolution. The 10-m wind speed RMSE is significantly different for WRF50 and WRF100 and also decreases with increasing resolution. For the vertical gradients and 10-m wind speed, the correlation with GFDex increases with higher WRF resolution. Surprisingly, the RMSE for the latent and sensible fluxes did not decrease with higher resolution despite the better representation of the physical variables with high-resolution WRF.

As seen in Fig. 11, there are large differences between the simulations and the observations in the latent and sensible heat fluxes. In general the WRF fluxes are larger than the GFDex COARE fluxes, sometimes by more than 200 W m$^{-2}$ (Figs. 11h,i). The discrepancy between these flux values is surprising considering the physical variables for flux calculations—the vertical gradients and wind speed—agree relatively well between GFDex and the WRF simulations. There are two explanations for this difference. The first is related to surface conditions: on 2 March the WRF temperature and mixing ratio gradients between the surface and 2 m have a positive bias when compared with the GFDex gradients. Larger vertical gradients in WRF would lead to larger fluxes, which is what we observe on 2 March. WRF10, in particular, best matches the 2-m mixing ratio and temperature, but has a large positive bias for the surface mixing ratio and temperature; thus, the poor surface initialization in WRF10 could be a large driver for the vertical gradient biases on 2 March. Previous WRF studies have found...
FIG. 11. Surface variables and turbulent fluxes along the GFDex flight track, both taken around 1400 UTC 2 and 5 Mar 2007. WRF10 is the blue line, WRF25 is the green line, WRF50 is the red line, WRF100 is the orange line, GFDex observations are the solid black dots, and GFDex fluxes recalculated with the WRF MYJ parameterization are the unfilled black dots. The mixing ratios associated with the latent heat flux are (a) surface saturation mixing ratio, (b) 2-m mixing ratio, and (c) mixing ratio gradient (surface–2 m). The temperatures associated with the sensible heat flux are (d) surface temperature, (e) 2-m temperature, and (f) temperature gradient (surface–2 m). (g) The 10-m wind speed, (h) latent heat flux, and (i) sensible heat flux.
that the temperature gradient can be more important than wind speed for determining surface fluxes (Kilpelainen et al. 2011), so better initialization of the sea surface to more realistic temperatures could help improve the flux bias we see. Unfortunately, the Arctic is data sparse and we were not able to find satellite data of sufficient spatial and temporal resolution to attempt simulations initialized with observed sea surface temperatures.

The second potential reason the WRF fluxes have a positive bias is because of the WRF flux parameterization; on 5 March the temperature and mixing ratio gradients between the WRF simulations and GFDex are similar, yet all WRF resolutions have positive flux biases. Because the physical gradients and wind speed agree closely between WRF and GFDex, the way WRF calculates the fluxes must cause this bias. We investigated this algorithm bias several ways. First, we ran experiments using the Yonsei University (YSU) surface layer parameterization instead of Mellor–Yamada–Janjić, and the results also indicated a positive WRF flux bias (not shown). The YSU and MYJ parameterizations differ in how they determine friction velocity, stability, and roughness length. When strong winds lead to a large friction velocity (\(u_\tau > 0.7\)), both parameterizations use a single value for momentum, heat, and moisture roughness lengths [see Eqs. (2), (3), and (5)]. Therefore, we experimented with the roughness length calculation by altering the Charnock parameter to 0.015 from the default value of 0.018 because the lower value is more appropriate for open ocean (Charnock 1955), but this did not result in a noticeable change in surface fluxes in either the MYJ or YSU simulations (not shown).

Finally, we used the MYJ parameterization to recalculate the latent and sensible heat fluxes using the GFDex observed sea surface and 2-m mixing ratios and temperatures, and the 10-m wind speed to calculate the friction velocity. The recalculated latent heat flux (Fig. 11h) had a positive bias even greater than the WRF fluxes, and the recalculated sensible heat flux (Fig. 11i) had a positive bias around the same magnitude as the WRF simulation biases. Because the physical mixing ratio and temperature gradients and the 10-m winds going into the flux calculation were the same as the COARE values, we can narrow down which variables may be causing the WRF flux bias. The WRF-calculated friction velocity exceeded the GFDex COARE calculated friction velocity at all open-water points, and a larger friction velocity would cause both larger sensible and latent heat fluxes, as seen in the recalculated fluxes. In addition, because of the large friction velocities, the recalculated fluxes used a single value for momentum, heat, and moisture roughness lengths, while the COARE algorithm uses separate roughness lengths (Fairall et al. 2003). The COARE stability parameterization (Fairall et al. 2003) is also more detailed than the MYJ stability parameterization. Thus, we conclude that the root cause for the bias between WRF fluxes and GFDex-reported fluxes is related to the simplified surface layer parameterization in WRF, specifically the calculation of friction velocity in high wind conditions; the use of a single roughness length value for momentum, heat, and moisture; and the stability calculation that are not appropriate for conditions such as those observed around Greenland. Despite the positive WRF bias concerning fluxes, the spatial patterns we see with regard to where the largest fluxes are located, the total area covered, and the correlation between high winds and high fluxes demonstrates that high-resolution WRF capture the patterns expected for turbulent fluxes, even if the parameterization needs to be modified for this region.

d. Case studies summary

Table 4 provides a summary of the differences between each resolution simulation for each case study over the ocean grid points in the box shown in Fig. 1. The table lists the maximum and average values for 10-m wind speed and surface latent and sensible heat fluxes, all of which are important for understanding the atmosphere–ocean connection during mesoscale wind events. The maximum wind speeds are located in the ETJ or

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**Table 3. RMSE and correlation calculated between WRF and GFDex for near-surface variables shown in Fig. 11. Values correspond to (top) RMSE and (bottom) correlation, respectively. Boldface RMSE values indicate that the WRF10 RMSE is statistically significant compared to WRF25, WRF50, or WRF100 RMSE at the 95% confidence level.**

<table>
<thead>
<tr>
<th>Variable</th>
<th>WRF10</th>
<th>WRF25</th>
<th>WRF50</th>
<th>WRF100</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface mixing ratio (kg kg(^{-1}))</td>
<td>0.64</td>
<td>0.64</td>
<td>0.67</td>
<td>0.67</td>
</tr>
<tr>
<td>Mixing ratio gradient (kg kg(^{-1}))</td>
<td>0.41</td>
<td>0.42</td>
<td>0.48</td>
<td>0.42</td>
</tr>
<tr>
<td>2-m temperature (K)</td>
<td>1.9</td>
<td>2.1</td>
<td>2.3</td>
<td>2.2</td>
</tr>
<tr>
<td>Temperature gradient (K)</td>
<td>1.3</td>
<td>1.8</td>
<td>2.4</td>
<td>2.6</td>
</tr>
<tr>
<td>10-m wind speed (m s(^{-1}))</td>
<td>1.5</td>
<td>1.6</td>
<td>2.6</td>
<td>3.4</td>
</tr>
<tr>
<td>Latent heat flux (W m(^{-2}))</td>
<td>133.7</td>
<td>151.2</td>
<td>128.67</td>
<td>101.1</td>
</tr>
<tr>
<td>Sensible heat flux (W m(^{-2}))</td>
<td>156.0</td>
<td>198.2</td>
<td>174.0</td>
<td>144.2</td>
</tr>
</tbody>
</table>

---
For all case study days WRF10 simulated the greatest maximum latent and sensible heat fluxes. The largest difference in maximum latent heat flux between WRF10 and WRF25 was 20%, though WRF10 was at most 6% higher than WRF25 for average latent heat flux. WRF10 exceeded WRF25 by at most 29% for maximum sensible heat flux, but the average sensible heat flux values between the two were no more than 3% different, with WRF25 sometimes exceeding WRF10. However, the difference in maximum fluxes was larger when comparing WRF10 with WRF50 and WRF100: WRF50 (WRF100) had maximum latent heat flux values 3%–32% (17%–56%) less than WRF10 values and maximum sensible heat flux values 5%–60% (29%–71%) less than WRF10 values. The average values were closer and WRF10 differed from WRF50 (WRF100) by a maximum of 9% (17%) for latent heat flux and 5% (21%) for sensible heat flux.

Because latent and sensible turbulent fluxes depend on wind speed, simulating a realistic wind field including the strongest speeds is essential for capturing

<table>
<thead>
<tr>
<th>Time</th>
<th>Max wind speed (m s(^{-1}))</th>
<th>Avg wind speed (m s(^{-1}))</th>
<th>Max latent heat flux (W m(^{-2}))</th>
<th>Avg latent heat flux (W m(^{-2}))</th>
<th>Max sensible heat flux (W m(^{-2}))</th>
<th>Avg sensible heat flux (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>0700 UTC 21 Feb 2007</td>
<td>34.7</td>
<td>15.1</td>
<td>610.0</td>
<td>69.3</td>
<td>775.0</td>
<td>31.2</td>
</tr>
<tr>
<td>2200 UTC 2 Mar 2007</td>
<td>34.9</td>
<td>13.3</td>
<td>883.2</td>
<td>168.9</td>
<td>1528.7</td>
<td>177.9</td>
</tr>
<tr>
<td>2200 UTC 5 Mar 2007</td>
<td>34.2</td>
<td>12.9</td>
<td>552.8</td>
<td>108.4</td>
<td>789.1</td>
<td>87.8</td>
</tr>
<tr>
<td>2200 UTC 6 Mar 2007</td>
<td>34.5</td>
<td>10.8</td>
<td>343.8</td>
<td>82.3</td>
<td>421.8</td>
<td>56.5</td>
</tr>
<tr>
<td>0700 UTC 9 Mar 2007</td>
<td>34.5</td>
<td>14.5</td>
<td>629.1</td>
<td>167.3</td>
<td>861.5</td>
<td>114.8</td>
</tr>
</tbody>
</table>

This table shows the maximum and average wind speed, latent heat flux, and sensible heat flux for each case study day. These values are calculated from the oceanic portion of the model domain shown by the boxed region in Fig. 1. Listed values correspond to (from top to bottom) WRF10, WRF25, WRF50, WRF100, ERA-I, and QuikSCAT, respectively. Dashes are given for the QuikSCAT and ERA-I flux values. ERA-I data were taken from the daily 0600 UTC output for 21 Feb and 9 Mar, and the 0000 UTC output on the following day for 2, 5, and 6 Mar. Following each value is the percentage of the corresponding WRF10 value given in parentheses.

BW for each case study day, and for each day WRF10 simulated the highest maximum wind speed, and the maximum winds exceeded the WRF25 maximum winds by 5%–11%, the WRF50 winds by 6%–14%, and the WRF100 winds by 24%–32%. Additionally, comparison between WRF10 and QuikSCAT show that QuikSCAT maximum wind speeds generally exceed WRF10 except on 6 March, when the QuikSCAT sea ice mask obscured the Denmark Strait and likely the strongest winds. However, the average wind speed between all resolutions and QuikSCAT on each day was similar; the largest difference between WRF10 and WRF25 was 3% and between WRF10 and WRF100 was 22%. Spatial analysis shows that during the ETJ and BW events the higher-resolution WRF simulations yielded more correct wind fields during the mesoscale wind events because of more detailed and realistic PG forcing.
Table 5. WRF two-month maximum and average wind speed, latent heat flux, and sensible heat flux. These values are calculated from the oceanic portion of the model domain shown by the boxed region in Fig. 1. Listed values correspond to (from top to bottom) WRF10, WRF25, WRF50, and WRF100, respectively. Following each value is the percentage of the corresponding WRF10 value given in parentheses.

<table>
<thead>
<tr>
<th></th>
<th>Max wind speed (m s(^{-1}))</th>
<th>Avg wind speed (m s(^{-1}))</th>
<th>Max latent heat flux (W m(^{-2}))</th>
<th>Avg latent heat flux (W m(^{-2}))</th>
<th>Max sensible heat flux (W m(^{-2}))</th>
<th>Avg sensible heat flux (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean (Feb–Mar 2007)</td>
<td>17.3 (90)</td>
<td>11.7 (100)</td>
<td>222.9 (86)</td>
<td>100.8 (100)</td>
<td>284.4 (100)</td>
<td>75.9 (100)</td>
</tr>
<tr>
<td>95% values (Feb–Mar 2007)</td>
<td>35.9 (84)</td>
<td>20.1 (97)</td>
<td>524.9 (84)</td>
<td>279.5 (93)</td>
<td>840.3 (93)</td>
<td>236.0 (93)</td>
</tr>
</tbody>
</table>

The strongest fluxes. As seen in Figs. 5 and 10, the high-resolution WRF has concentrated maximum sensible heat fluxes collocated with the strongest winds over open ocean. WRF has a positive flux bias that appears to be related to both incorrectly initialized surface variables due to lack of observational data with sufficiently high temporal and spatial resolution and a parameterization problem that is not related to resolution. However, the differences we see between the flux fields are caused by the wind fields, which are better simulated by high-resolution WRF.

4. Two-month simulations

A simulation from 1 February to 31 March 2007 was run for each WRF resolution to examine longer-term implications of model resolution since the previous case studies are snapshots of short-lived mesoscale wind events. This analysis focuses on the surface winds and surface latent and sensible heat fluxes because of their impact on the ocean and is restricted to the boxed area in Fig. 1. The two-month average values of these variables will be analyzed to assess differences due to resolution. However, the largest wind speeds and fluxes occur during BW and tip jets, so to better understand how the extreme conditions vary because of model resolution, the standard deviation and 95th percentile values of the variables are discussed as well.

The BW and both ETJ and WTJ dominate spatial plots of both the mean and the highest 95th percentile wind speeds (not shown) over this two-month period. As with the case studies, the mesoscale wind features are more distinct in high-resolution WRF; the difference between 95th percentile WRF10 and both WRF50 and WRF100 wind speeds shows WRF10 with winds 4–6 m s\(^{-1}\) greater in the ETJ and BW regions, while both WRF50 and WRF100 have winds 6–8 m s\(^{-1}\) greater in the wake regions of Cape Farewell and the Geikie Plateau. The mean wind speed for the two months is similar for all WRF simulations, but the maximum values from the two-month mean and two-month 95th percentile wind fields are from WRF10 and decrease with resolution (Table 5). Both the average and 95th percentile winds reinforce that the high-resolution simulations show stronger winds along the southeast Greenland coast and better-defined BW and tip jet features. The standard deviation in wind speed, particularly in the locations where barrier winds and tip jets occur, is also larger for high-resolution simulations, which means that these high-resolution simulations are able to capture extreme outlier events that low resolution is not able to capture.

The average and 95th percentile turbulent surface fluxes in this region are both large, reflecting the frequent presence of cold air over warm ocean waters and the influence of strong winds. The two-month mean fluxes for WRF10 and WRF25 have three areas of maximum oceanic heat loss that correspond to locations associated with BW or tip jets. The two-month mean WRF10 (Fig. 12a) and WRF25 (Fig. 12b) sensible heat flux exceeds 100 W m\(^{-2}\) along most of the southeast Greenland coast, with many points exceeding 150 W m\(^{-2}\). WRF50 (Fig. 12c) has fewer grid cells with fluxes exceeding 100 W m\(^{-2}\) in locations with strong surface winds, possibly because WRF50 simulates lower wind speeds and less distinct mesoscale wind features. All simulations show a sharp division between the open ocean, with high mean fluxes, and sea ice, where there are low or negative fluxes. In general, the latent heat fluxes (not shown) follow a similar pattern to the sensible heat flux field. Again, Table 5 illustrates that the maximum mean fluxes are from the WRF10 simulation and decrease in magnitude with decreasing resolution.

The high-resolution WRF10 (Fig. 12d) and WRF25 (Fig. 12e) have the largest standard deviations for sensible heat flux, and because there are high standard
FIG. 12. Surface sensible heat flux (W m$^{-2}$; positive upward) from the 1 Feb to 31 Mar 2007 simulations. The mean, standard deviation, and 95th percentile highest sensible heat flux values are shown for (a),(d),(g) WRF10, (b),(e),(h) WRF25, and (c),(f),(i) WRF50, respectively.
deviations collocated with tip jets or barrier winds this means that these high-resolution simulations are able to capture the strongest fluxes from the ocean to the atmosphere that occur in this region. Along the Greenland coast in all WRF simulations, the highest 95th percentile latent and sensible heat fluxes exceed 600 W m\(^{-2}\). For WRF10 (Fig. 12g) and WRF25 (Fig. 12h) there are many grid cells with sensible heat fluxes greater than 600 W m\(^{-2}\); for WRF50 (Fig. 12i) there are fewer grid cells with fluxes of this magnitude. The same pattern is true of latent heat flux values (not shown). In all cases, the highest of the 95th percentile sensible and latent heat fluxes are collocated with the regions where high-speed tip jet and barrier winds occur over open ocean. The maximum and average turbulent fluxes from the 95th percentile fields tend to have the highest magnitude in WRF10 simulations and decrease with decreasing resolution. In addition, the mean 95th percentile flux values and standard deviations shown in Table 5 also have the highest magnitude for WRF10 and decrease with decreasing resolution, and this reinforces that using high-resolution WRF results in more events with a larger net oceanic energy loss and would therefore be more likely to induce ocean convection when compared with WRF50 or WRF100 simulations. The total energy loss decreases with decreasing resolution, so the lower the WRF resolution the less likely the ocean would lose enough energy to cause convection.

5. Conclusions

Comparison between WRF simulations and available observational data reveal the importance of high resolution in accurately simulating the mesoscale BW and tip jets that occur around the southeast Greenland coast. In general, higher-resolution WRF simulations had similar and more physically realistic winds and PG fields when compared with lower-resolution WRF. While all model resolutions tended to simulate the placement of synoptic features similarly, the higher-resolution simulations were better able to capture local modifications to the synoptic pressure field and the resulting intense winds. The low-resolution WRF simulations used terrain that was smoothed out—too high over the ocean and too low inland—so physically realistic pressure gradients did not build up and force strong mesoscale winds in the correct locations.

Comparisons between GFDex dropsondes and WRF indicate that WRF10 simulated the vertical structure of the ETJ and BW best when compared with WRF25 and WRF50. We have confidence that WRF10 physically models the boundary layer well because the WRF10 simulations generally capture the observed boundary layer winds and mixed layer depth, despite the near-surface equivalent potential temperature bias evident in the simulations. Comparisons between the ETJ and BW sensible heat flux fields show that the surface area covered by high fluxes is larger for higher-resolution WRF and for the BW cases, likely due to slightly stronger winds and a larger near-surface temperature gradient. In addition, high-resolution WRF simulated stronger maximum fluxes in the case studies and two-month simulations. However, the main concern was that WRF had a positive flux bias that is likely due to surface conditions and model parameterizations, neither of which can be improved by increasing resolution. Investigations into improved surface initialization did not return any data with enough spatial and temporal resolution to be included in the simulations.

This study shows that using WRF at 10-km resolution is necessary for properly simulating the mesoscale wind field and boundary layers associated with BW and tip jets. Because turbulent fluxes depend on the wind field, a high-resolution atmosphere is necessary for simulating localized and intense energy removal from the ocean. Tables 3 and 4 illustrate that for both case studies and the two-month simulations the mean energy removal across all WRF resolutions is similar, but that high-resolution WRF produces the maximum fluxes. Examination of spatial flux fields reveals that in high-resolution WRF the energy loss is concentrated in small areas instead of spread across the region, while low-resolution WRF has a smaller spatial area with the maximum fluxes. Without localized and intense fluxes, captured by high-resolution WRF, small-scale (~10 km) ocean convection might not occur because energy is not being lost in the necessary locations.

Ultimately we plan to use a fully coupled regional model, the Regional Arctic System Model (RASM), to study connections at the atmosphere–ocean interface, and better quantify how important mesoscale atmospheric features are for driving ocean convection. RASM uses WRF as its atmospheric component, and this study has shown the importance of using a high-resolution atmospheric model to best simulate tip jets and barrier winds. By using RASM to examine the atmosphere–ocean connection using a high-resolution atmosphere we can be sure that we are driving the ocean behavior with realistic and reasonable high-speed wind events. The use of the high-resolution atmosphere will be important for understanding how precisely tip jets and barrier winds impact the ocean over which they take place.

Acknowledgments. This research was supported by the U.S. Department of Energy (DOE) Grants DE-FG02-07ER64462 and DE-SC0006178, as well as a National
Oceanic and Atmospheric Administration (NOAA)/Earth System Research Laboratory (ESRL)/Cooperative Institute for Research in Environmental Sciences (CIRES) graduate fellowship. We thank two anonymous reviewers and the editor for their comments, which have improved this paper. Thanks also to Melissa Nigro and Mimi Hughes for guidance with the momentum budget analysis.

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