A Numerical Study of an Extreme Cold-Air Outbreak over the Labrador Sea: Sea Ice, Air–Sea Interaction, and Development of Polar Lows

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ABSTRACT

In this paper, the ability of the MM5 mesoscale forecast model to simulate the air–sea interaction, boundary layer development, and mesoscale structure associated with a cold-air outbreak over the Labrador Sea is investigated. The case chosen was one for which research aircraft data and satellite imagery are available for validation. The default surface-layer parameterization included in the model is shown to grossly overestimate the magnitude of the air–sea interaction resulting in forecasts of boundary layer growth and mesoscale development that differ substantially from observations. It is also shown that a representation of the inhomogeneities in sea-ice cover results in a significant improvement in simulations of the air–sea interaction, boundary layer development, and mesoscale structure both within the marginal ice zone and downstream over the open ocean. Finally, the mesoscale cyclones or polar lows observed in the wake of the cold-air outbreak are shown to be coupled to the evolution of an upper-tropospheric potential vorticity anomaly that was advected over the region. The model simulations suggest that MM5, most probably due to inaccuracies in and the limited resolution of the analyzed fields that supply the initial and boundary conditions to the model, was unable to correctly simulate the development and track of this anomaly and this ultimately led to an incorrect forecast of the polar lows’ finite-amplitude behavior.

1. Introduction

The Labrador Sea region in winter is a region of extremes that result in a variety of interesting meteorological and oceanographic phenomena that play a direct role in the dynamics of the global climate system. As one moves from the frozen continent to the marginal ice zone and then to open water, there is a dramatic change in surface temperature, moisture availability, and roughness. The region is strongly influenced by the passage of synoptic-scale cyclones as they transit from Newfoundland to Iceland. Their passage results in sudden and pronounced changes in the air temperature, humidity, and wind speed and direction in the region. In the northwesterly flow that is established after the passage of an archetypical cyclone, the advection of cold and dry arctic air over the warm surface waters of the Labrador Sea results in intense air–sea interaction and a significant transfer of heat and moisture from the ocean to the atmosphere. Evidence for this interaction can be found in the linear and cellular convective clouds observed in satellite imagery to develop in the wake of these cyclones. The region is also home to intense and short-lived mesoscale cyclones or polar lows that often intensify through latent heat release driven by air–sea fluxes. The region is also often transited by ozone-rich upper-tropospheric potential vorticity anomalies. The weak stratification in the region, which is also a result of air–sea interaction, allows these anomalies to strongly couple to and intensify lower-tropospheric circulations such as polar lows.

The northflowing warm and salty West Greenland Current and the southflowing cold and fresh Labrador Current form a cyclonic gyre that, through geostrophic balance, results in a doming up of isopycnals that acts to expose the deep ocean to the atmosphere. The exchange of sensible and latent heat to the atmosphere leads to a deep and weakly stratified oceanic mixed layer that is susceptible to convective overturning. Indeed the Labrador Sea is one of the few locations where deep ocean convection occurs and as such is an important source of abyssal water for the World Ocean.

The Labrador Sea Experiment was established with the objective of understanding the process of deep ocean convection and the role that the atmosphere plays in its forcing (Lab Sea Group 1998). As part of the experiment, meteorological and oceanographic observations

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were made in the region during February and March of 1997. This included in situ sampling of weather systems over the Labrador Sea by research aircraft from the U.S. Air Force (Renfrew et al. 1999) as well as surface and upper-air observations from the R/V *Knorr* (Lab Sea Group 1998). Renfrew and Moore (1999, hereafter RM) describe aircraft observations of the convective roll clouds, boundary layer structure, and the air–sea interaction during a cold-air outbreak that occurred on 8 February 1997.

This paper will concentrate on the verification of a numerical model against the observations of RM with respect to the evolution of the planetary boundary layer (PBL) and the spatial distribution of sensible and latent heat fluxes. Two mesoscale vortices or polar lows developed to the north of the region sampled by RM. We will use satellite imagery as a validation tool for the model's ability to forecast their evolution. As we shall see, the detailed structure of the marginal ice zone plays an important role in the air–sea interaction in the region.

**Fig. 1.** Sea level pressure (mb) over the Labrador Sea region from the ECMWF analysis at (a) 0000 UTC 6 Feb, (b) 0000 UTC 7 Feb, (c) 1200 UTC 7 Feb, (d) 0000 UTC 8 Feb, (e) 1200 UTC 8 Feb and (f) 0000 UTC 9 Feb 1997.
 Accordingly, in this paper we present a new parameterization of the surface inhomogeneities within the marginal ice zone appropriate for a mesoscale forecast model. The layout of the paper is as follows: section 2 describes the synoptic situation that led to the cold-air outbreak and the development of the polar lows. In section 3, the model used in the simulations will be described with an emphasis on the authors’ modifications and extensions to its standard or “stock” configuration. The experimental design is also presented in this section. Sections 4 contains our results with respect to the impact that sea-ice and surface-layer parameterizations have on the evolution of the flow. In section 5, the role that the upper-level potential vorticity anomaly played in the development of the polar lows will be discussed. Our summary and conclusions are contained in section 6.

2. Synoptic overview

Figure 1 illustrates the sequence of events that resulted in the cold-air outbreak into which the research flight was undertaken and in the wake of which the polar lows developed. The maps are based on 2.5° × 2.5° global European Centre for Medium-Range Weather Forecasts (ECMWF) analyses enhanced with objectively analyzed upper-air and surface observations. On 6 February a deepening synoptic-scale low (998 mb at 0000 UTC on 6 Feb, 982 mb 24 h later) traveled eastward from Hudson Bay over the frozen waters of the Hudson Strait (Figs. 1a,b). On 7 and 8 February, the low filled while it drifted slowly eastward over the northern Labrador Sea (Figs. 1c–d). During this same period, another synoptic-scale low was moving from the Great Lakes over Newfoundland toward Greenland (Figs. 1a–d). It underwent an explosive deepening with its central pressure falling from 1001 mb at 0000 UTC on 6 February to 973 mb 24 h later. In its wake, a region of high pressure moved eastward toward Labrador (Figs. 1d–f). The resulting combination of the low over the northern Labrador Sea and the ridge over Labrador swept cold and dry arctic air over the open waters of the Labrador Sea (Figs. 1e,f).

The mesoscale circulation over the Labrador Sea that developed in this synoptic environment is depicted in the satellite imagery displayed in Figs. 2–4. At 2127 UTC on 7 February (Fig. 2), the high cloud associated with the synoptic low over the northern Labrador Sea is clearly visible. The cloud-free sky over Labrador is a signature of the ridge that was developing at this time. The resulting cyclonic flow and the concomitant strong air–sea interaction that took place as the cold polar air flowed out over the relatively warm waters of the Labrador Sea triggered the development of convective roll clouds that can be seen in the image. The cloud streets of shallow cumuli are aligned roughly in the direction of the mean horizontal low-level wind (see, e.g., Atkinson and Zhang 1996). As the separation between cloud streets (or alternatively their wavelength) increases with fetch, the convection becomes cellular.

Sea ice with its distinctive mottled gray shading can be seen along the Labrador coast at this time. Superimposed on Fig. 2 are contours of 50% and 75% sea-ice concentration as deduced from the passive microwave Special Sensor Microwave/Imager (SSMI) data (Comiso et al. 1996). The transitional region between full ice cover and open water is generally referred to as the marginal ice zone (MIZ). From Fig. 2, one can see that there is considerable variability in ice cover within...
the MIZ. Between 55° and 57°N, the ice concentration was between 50% and 75%, while to the south of 55°N and to the north of 57°N it was in excess of 75%. This figure shows that the cloud streets have their origin over the MIZ where the PBL first receives an injection of heat and moisture from polynyas and leads in which the ocean is exposed to the atmosphere. The presence of sea ice decouples the atmosphere from the ocean and it therefore follows that regions of reduced ice concentration should exhibit enhanced convective streamer activity with the opposite occurring in regions of high ice concentration. This can be clearly seen in Fig. 2 if one compares the region between 55° and 57°N with the region to the south of 55° and north of 57°N. It can be argued that the amorphous cloud feature centered at 57°N, 60°W developed due to the intense convection over a coastal polynya, which could have formed in strong offshore wind conditions (P. Guest 1999, personal communication). Also at this time, a weak disturbance in the low-level cloud field can be seen near 58°N, 57°W (labeled A in Fig. 2).

Figure 3 shows an image from 1119 UTC on 8 February. Disturbance A identified in Fig. 2 has now developed into a mesoscale cyclone or polar low with a well-defined center and frontal-like cloud band, C. Evidence in the cloud field for a second and weaker cyclonic circulation, disturbance B, can be seen in the vicinity of 60°N, 58°W. Estimates suggest that the cloud feature seen in Fig. 2 (labeled D in this image) has been advected off the MIZ. The high-level flight track, aligned approximately in the direction of the low-level wind, during which dropsondes were launched is also displayed in Fig. 3. The same track was flown approximately 10 h earlier providing a unique view as to the evolution of the PBL during a high-latitude cold air outbreak (RM).

Unfortunately, no observations were available in the area where disturbances A, B, and C developed. Since satellite imagery provided the only proof that the disturbances occurred we cannot directly assess strength of their circulations. As could be noted in Fig. 1, 2.5° × 2.5° ECMWF analysis did not reproduce the events in the Labrador Sea with sufficient accuracy.

Figure 4 shows an image from 2109 UTC on 8 February. By this time, the two mesoscale cyclones have begun to dissipate. A family of shortwave disturbances can be seen along cloud band D.

As discussed by Allart et al. (1993) and Reader and Moore (1995), total column ozone data from the Total Ozone Mapping Spectrometer (TOMS) instrument provides a useful signature for the existence of upper-level disturbances associated with tropopause folds and potential vorticity (PV) anomalies that play a role in surface cyclogenesis (Hoskins et al. 1985). Previous case studies of mesoscale and synoptic-scale cyclones over the Labrador Sea (Businger and Reed 1989; Reader and Moore 1995; Moore et al. 1996; Rasmussen et al. 1996) have all emphasized the role of upper-level disturbances in the genesis and evolution of these events. TOMS data for 7–9 February displayed in Fig. 5 show that the period of interest was one in which a region of anomalously high total column ozone was advected from Hudson Bay.
to the east of Greenland via the Labrador Sea. Further proof as to the existence of this stratospheric intrusion of high PV air into the troposphere can be seen in Fig. 6, which shows imagery from the Geostationary Operational Environmental Satellite-8 (GOES-8) water vapor channel at 1200 UTC on 8 February. In this image, dark regions correspond to high radiance and low upper-tropospheric moisture content as would be expected in the presence of a tropopause fold associated with a PV anomaly (Appenzeller and Davies 1992). We therefore suggest that the development of the multiple polar lows identified in the Advanced Very High Resolution Radiometer (AVHRR) imagery was closely associated with this upper-level potential vorticity anomaly. We can also note that the family of disturbances (labeled D in Fig. 4) had a similar origin. They developed after polar lows A and B began to dissipate in the area of strong heat fluxes and significant horizontal wind shear across the convective band once a new source of high-level PV, associated with the southern branch of the anomaly traveling from the Gulf of St. Lawrence (see Figs. 5b and 6), became available. Similar small-scale vortices have been observed along other convective bands over the Labrador Sea (Moore et al. 1996) and early stages of such development have been attributed to barotropic instability in a convectively unstable environment (Moore 1985).

Based on this analysis, we believe that this event provides a good case with which to study the complex interaction between sea ice, the ocean, and the atmosphere that occurs in the climatologically sensitive Labrador Sea region during the winter. It also presents an opportunity to assess the skill of a regional numerical model in forecasting high-latitude mesoscale features such as polar lows. Moreover, thanks to the aircraft data collected by RM the model results can be directly compared to observations.

3. Model description and experiment design

a. Model description

The Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) fifth-generation Mesoscale Model (MM5) was used for this case study. The system offers users great flexibility in choosing physics options. It has been used to simulate a broad range of atmospheric phenomena on varying spatial and temporal scales. It has been also previously used to study air–sea interaction (e.g., Kuo and Reed 1988; Albright et al. 1995).

Different global or regional analyses can be employed as an input and subsequently enhanced through assimilation of upper-air and surface observations (Manning and Haagenson 1992). The core of the system is a non-hydrostatic, primitive-equation atmospheric model (release 2.8 in the current simulations). The model employs a terrain-following vertical coordinate:

$$\sigma = \frac{p_0 - p_s}{p_t - p_s},$$

where $p_0$ is the reference-state pressure at a model level, $p_s$ is the reference-state surface pressure, and $p_t$ is the reference-state pressure at the top of the model.

MM5 offers several choices for the parameterization of grid-scale resolvable moist processes (warm rain, simple ice, mixed phase, ice with graupel) and subgrid-scale cumulus convection (Anthes–Kuo, Arakawa–Schubert, Grell, Fritsch–Chappell, Kain–Fritsch, Betts–Miller) as well as parameterization of shallow convection [see Grell et al. (1994) and Dudhia et al. (1997) for details and references].

b. Surface-layer parameterization

Special attention in this section is given to surface-layer and PBL parameterization where some changes to the original schemes were implemented by the authors. Several PBL schemes are available in the model. Comparison of cross sections obtained from the model with those of RM based on aircraft measurements over the Labrador Sea during the cold-air outbreak showed that the PBL parameterization developed by Troen and Mahrt (1986) and implemented in MM5 following Hong and Pan’s (1996) PBL scheme in the National Centers for Environmental Prediction’s Medium-Range Forecast (MRF) model is the most appropriate for the strongly convective conditions over the ocean present in our case. This result is not surprising since the above closure takes into account the nonlocal nature of the vertical turbulent
transport in the convective PBL. The Blackadar parameterization (Blackadar 1976, 1979; Zhang and Anthes 1982) displayed vertical mixing that was too strong. The Burk and Thompson (1989) parameterization, which is based on a 1.5-order local closure of Mellor and Yamada (1982), had an insufficient vertical mixing and resulted in a boundary layer that was very unstable. For these reasons, the MRF PBL scheme became our parameterization of choice and further discussion in this and following sections will concentrate on this scheme. We note that the Blackadar PBL surface layer is parameterized in a nearly identical manner so our comments should apply as well to this scheme.

The calculation of latent heat flux in MM5 is, with slight modifications, based on Carlson and Boland (1978) who developed bulk formulas for urban–rural canopy surfaces:

$$q(z_a) - q(z_0) = \frac{q_u}{k} \left[ \ln \left( \frac{k z_a}{z_l} \right) \right], \quad (2a)$$

$$u_g q_u = C_f E \left[ q(z_a) - q(z_0) \right] = \frac{E}{\rho L_c}, \quad (2b)$$

where $q$ is the specific humidity, $z_a$ is normally the anemometer height or in numerical models lowest vertical level, $z_0$ is the surface roughness, $q_u$ is the turbulent humidity scale, $k = 0.4$ is von Kármán constant, $u_g$ is the friction velocity, $K_u$ is the molecular diffusivity, $z_l$ is the depth of molecular layer, $\Psi_H$ is the similarity function, $L$ is the Monin–Obukhov length, $C_f$ is bulk transfer coefficient for moisture, $E$ is the latent heat flux, $\rho$ is the air density, and $L_c$ is the latent heat of condensation of water vapor. In MM5, the total flux from a grid square is obtained by multiplying $E$ in Eq. (2b) by the moisture availability parameter, which is defined as a ratio of completely wet to completely dry surface areas within a grid cell. Carlson and Boland (1978) used a similar formula to (2a) for the sensible heat flux so with
the appropriate modifications the equation for it should take the form
\[
\theta(z_a) - \theta(z_0) = \frac{\theta_a}{k} \left[ \ln \left( \frac{\bar{k} u_a z_a}{K^*} + \frac{z_a}{z_i} \right) - \Psi_H \left( \frac{z_a}{L} \right) \right].
\] (2c)
\[
u_a \theta_a = C_{hu} [\theta(z_a) - \theta(z_0)] = \frac{H}{\rho c_p},
\] (2d)

where \( \theta \) is the potential temperature, \( \theta_a \) is the turbulent potential temperature scale, \( C_{hu} \) is bulk transfer coefficient for heat, \( H \) is the sensible heat flux, and \( c_p \) is the specific heat of air. In MM5, however, the sensible heat flux is calculated using the bulk formulation after Blackadar:
\[
\theta(z_a) - \theta(z_0) = \frac{\theta_a}{k} \left[ \ln \left( \frac{z_a}{z_t} \right) - \Psi_H \left( \frac{z_a}{L} \right) \right].
\] (3)

This inconsistency in the way that MM5 calculates the latent and sensible heat fluxes leads to unrealistic behavior of the Bowen ratio (Bo = \( H/E \)). In appendix A, we present evidence that over the ocean the sensible heat flux in the stock MM5 parameterization is disproportionately large with respect to the latent heat flux especially at high wind speeds and for polar conditions under which the sensible heat flux dominates. This in turn results in a Bowen ratio that is a strong function of roughness. This behavior of the MM5 surface-layer parameterization is unphysical as the nature of the heat and moisture transport in the interfacial layer is similar as both are associated with molecular diffusion. It should be noted that the discrepancy between the magnitudes of latent and sensible heat fluxes in MM5 will be even larger over the land where surface roughness is typically higher than over the water. Moisture availability over the land modulates this behavior but we believe that the underlying assumptions in the default formulas are still inappropriate. Furthermore, we add that Carlson and Boland (1978) derived their formulas for urban–rural canopy surfaces and its application over the ocean is questionable.

Oncley and Dudhia (1995) compared surface fluxes from MM5 over land to observations. The results were very much dependent on moisture availability and rather insensitive for a broad range of atmospheric conditions. We are, however, unaware of verification of the MM5 model fluxes against measurements over the ocean.

We note that standard bulk formulas (see, e.g., Stull 1988; Garratt 1992) distinguish between mechanical roughness \( (z_a) \) and roughness scales for temperature and humidity \( (z_t \) and \( z_i \), respectively). The latter are usually smaller than \( z_a \) over rough surfaces, often by an order of magnitude. Nevertheless, no distinction between these scales is present in MM5.

Since we believe that the stock MM5 parameterization of surface layer is inappropriate for use over the ocean for the reasons given above, we rewrite the bulk formulas (2a) and (3) in the standard form:
\[
q(z_a) - q(z_0) = \frac{\theta_a}{k} \left[ \ln \left( \frac{z_a}{z_t} \right) - \Psi_H \left( \frac{z_a}{L} \right) \right],
\] (4a)
\[
\theta(z_a) - \theta(z_0) = \frac{\theta_a}{k} \left[ \ln \left( \frac{z_a}{z_i} \right) - \Psi_H \left( \frac{z_a}{L} \right) \right],
\] (4b)

where \( z_a \) and \( z_i \) in Eqs. (4a) and (4b) are calculated after Garratt (1992):
\[
\ln \left( \frac{z_a}{z_t} \right) = \ln \left( \frac{z_a}{z_i} \right) = 2,
\] (5a)

over land and
\[
\ln \left( \frac{z_a}{z_t} \right) = 2.48 \text{ Re}_{u_a}^{0.25} - 2, \quad \text{and}
\] (5b)
\[
\ln \left( \frac{z_a}{z_i} \right) = 2.28 \text{ Re}_{u_a}^{0.25} - 2, \quad (5c)
\]
over ocean. In Eqs. (5b) and (5c), \( \text{Re}_{u_a} = u_a z_a / \nu \) is roughness Reynolds number and \( \nu \) is kinematic molecular viscosity of air.

In short, our modifications of the stock equations for the moisture [Eq. (2a)] and heat transfer [Eq. (3)] in the surface layer consists of replacing \( z_0 \) with new flow-dependent roughness scales for humidity and temperature. In other words bulk transfer coefficients for moisture and heat change from

\[
C_E = \left[ \ln \left( \frac{z_a}{z_0} \right) - \Psi_M \left( \frac{z_a}{z_0} \right) \right] \left[ \ln \left( \frac{\bar{k} u_a z_a}{K^*} + \frac{z_a}{z_i} \right) - \Psi_H \left( \frac{z_a}{L} \right) \right]
\]

and

\[
C_H = \left[ \ln \left( \frac{z_a}{z_0} \right) - \Psi_M \left( \frac{z_a}{z_0} \right) \right] \left[ \ln \left( \frac{z_a}{z_t} \right) - \Psi_H \left( \frac{z_a}{L} \right) \right]
\]
to

\[
C_E = \left[ \ln \left( \frac{z_a}{z_0} \right) - \Psi_M \left( \frac{z_a}{z_0} \right) \right] \left[ \ln \left( \frac{z_t}{z_0} \right) - \Psi_M \left( \frac{z_t}{L} \right) \right]
\]
and

\[
C_H = \left[ \ln \left( \frac{z_a}{z_0} \right) - \Psi_M \left( \frac{z_a}{z_0} \right) \right] \left[ \ln \left( \frac{z_i}{z_0} \right) - \Psi_M \left( \frac{z_i}{L} \right) \right]
\]
respectively.

In MM5, the roughness over water is calculated using a modified Charnock formula (Delsol et al. 1971):
\[
z_0 = \alpha \frac{u^2_a}{g} + 0.0001,
\] (6a)
with the Charnock constant \( \alpha_c \) equal to 0.032. We alternatively used the formula of Smith (1988):

\[
\tau = \frac{2}{\alpha_c} \frac{\mu}{g} + \frac{b}{\nu} \tag{6b}
\]

with \( b \) equal to 0.11. (We note that in MM5 the Burk–Thompson PBL scheme takes \( \alpha_c \) to 0.0144.) The first term in the above equations is due to the roughness induced by wind induced capillary waves while the second term is due to molecular effects over a smooth water surface. The model simulations showed that differences between the two parameterizations (6a) and (6b) are negligible for the conditions under investigation. The use of the Charnock equation is justifiable as long as the sea is mature, that is, the phase velocity of the surface waves approaches the wind velocity (Garratt 1992). When the sea is young, for example, for a short fetch over water or in very strong and gusty wind conditions, the Charnock equation is not reliable as shown by coupled ocean–atmosphere modeling (Doyle 1995; Bao et al. 1998). However, we believe that for the case under consideration where wind velocities are typically lower than 15 m s\(^{-1}\) and for distances greater than 300 m from the ice edge (Smith and Macpherson 1996) use of Eq. (6b) is justifiable. Our estimates based on parameterizations of Fairall et al. (1994) showed that for winds below 20 m s\(^{-1}\) the role of sea spray is not significant. Also, observations (see Smith 1988; Garratt 1992) show that the value of \( \alpha_c \) over ocean is in the range of 0.011–0.016, that is, less than half of that used in the model. Furthermore, Bao et al. (1998) found this value of \( \alpha_c \) too high based on coupled ocean–wave–atmosphere model simulations. As a result, we have adopted \( \alpha_c = 0.0156 \) based on oceanic data of Wu (1969).

We also note that the stock MM5 parameterization of the surface layer was, despite all its inconsistencies, used by numerous authors to investigate the impact of surface fluxes on, for example, marine cyclogenesis, polar lows, and tropical cyclones. It is not clear what impact if any the use of a more physical surface layer parameterization, such as the one that we have proposed, would have on these previous simulations. It is however important to recognize that at a minimum, the sensible and latent heat fluxes in these studies may be suspect. In appendix A we elaborate on effects that modifications to the stock MM5 parameterization have on fluxes and Bowen ratio over ocean for different atmospheric conditions.

c. Parameterization of sea ice

At high latitudes, sea ice plays a very important role in the exchange of heat and moisture between the atmosphere and ocean and as a consequence the weather in these regions is strongly affected by its presence. Sea ice, especially in the MIZ, is highly inhomogeneous and consists of leads and polynyas (with horizontal scales varying from several meters to hundreds of kilometers) interspersed with regions of solid ice cover (see Fig. 11 in Renfrew et al. 1999). Significant differences in skin temperature between ice, which is a good insulator, and open water result in a high spatial variability of heat, moisture, and momentum fluxes into the atmosphere (Overland 1985). Turbulent heat fluxes over leads and polynyas can reach 300–500 W m\(^{-2}\), which is almost two orders of magnitude larger than the heat flux from a homogeneous ice sheet (Maykut 1978). Simple area averaging of skin temperature is not justifiable as the heat flux depends on vertical temperature gradients in the surface layer in a highly nonlinear manner. In addition, the inhomogeneities are randomly distributed and of subgrid scale and thus need to be parameterized.

Our starting point in parameterizing the effects of sea ice is information on ice concentration, which is the ratio of the area covered by ice to the total area. In our case we make use of the global analysis based on passive microwave data (Comiso et al. 1997) that is available on a daily basis from the National Snow and Ice Data Center (NSIDC) at the University of Colorado. These data are provided at a horizontal resolution of 25 km on a polar stereographic grid and are interpolated to our model grid using the Akima interpolation scheme for irregularly spaced data from the IMSL numerical library. This scheme provides for superior performance in cases such as ours where satellite data are often unavailable at all grid points and no ice concentration over land needs to be assigned.

In calculating the heat and moisture fluxes, we use a mosaic method (see, e.g., Vihma 1995); that is, the total flux from a grid square is obtained as the sum of fluxes over sheet ice and open water so that:

\[
\text{FLUX}_{\text{TOT}} = \text{icec} \times \text{FLUX}_{\text{ICE}} + (1 - \text{icec}) \times \text{FLUX}_{\text{WATER}}, \tag{7}
\]

where icec is the ice concentration. The temperature of open water is set to the freezing point of saltwater with a salinity of 40 psu at 271.45 K and remains constant during the simulation. The skin ice temperature is calculated with a five-layer interactive ice model, which uses an energy balance equation at the surface:

\[
\rho C_i \frac{\partial T_i}{\partial t} = (1 - \alpha_i) R_s + L_\text{l} - L_\tau - H - E + G, \tag{8a}
\]

where \( R_s \) is the shortwave solar radiation flux, \( L_\text{l} \) is the longwave radiation flux from the atmosphere to the ice, \( L_\tau \) is the longwave radiation flux from the ice to the atmosphere = \( \sigma T_i^4 \), \( H \) is the sensible heat flux from the ice to the atmosphere, \( E \) is the latent heat flux from the ice to the atmosphere, \( G \) is the substrate flux = \(- \rho C_i \frac{\partial T_i}{\partial z} \), \( \alpha_i \) is the ice albedo (0.60), \( \sigma \) is ice emissivity (0.97), \( \rho_i \) is ice density (920 kg m\(^{-3}\)), \( C_i \) is ice heat capacity (2230 J K\(^{-1}\) kg\(^{-1}\) m\(^{-2}\)), \( \Delta z_i \) is thickness of the top model layer, \( T \) is ice temperature,
and $K$ is ice diffusivity ($0.92 \times 10^{-6}$ m$^2$ s$^{-1}$). In the substrate, a diffusion equation

$$\frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial z^2},$$

(8b)
is used with a boundary condition on the lowest layer $T_i = 271.45$ K. For the MIZ of the Labrador Sea, a constant thickness of 60 cm, which is appropriate for the first-year ice in the region, is assumed.

The ice model uses semi-implicit integration in time to avoid numerical instabilities and time step limitations. The vertical grid is stretched logarithmically with high-resolution at the top of the model. Experiments with this and other models (Guest and Davidson 1994) showed that the top model layer, $\Delta z_i$, should not be thicker than 5 cm so that model results are not affected by the resolution.

The roughness of the sea within the MIZ is calculated using the Charnock formula (6b) while roughness of the sheet ice is a function of ice concentration. Guest and Davidson (1991) give neutral drag coefficient, $C_{dn}$, for young and first-year ice in the range of 2.0–4.2 $\times$ 10$^{-3}$. The following function for the neutral drag coefficient was prescribed:

$$C_{dn} = \begin{cases} 
C_{dn\text{max}} = 2.0 \times 10^{-3} & \text{for } icec \leq 0.4, \\
C_{dn\text{max}} + 5(icec - 0.4)(C_{dn\text{max}} - C_{dn\text{min}}) & \text{for } 0.4 < icec < 0.6, \\
C_{dn\text{min}} = 4.2 \times 10^{-3} & \text{for } icec \geq 0.6.
\end{cases}$$

(9)

From the definition of neutral drag coefficient,

$$C_{dn} = \left[ \frac{k^2}{\ln \frac{\zeta}{z_a}} \right]^2,$$

the surface roughness can be found.

Mason (1988) notes that local fluxes from inhomogeneous surfaces should be calculated at the blending height, $b$, rather than at the anemometer height (10 m). He gives the following expression for $b$:

$$b = 2 \left[ \frac{u_b}{V(b)} \right]^2 L_{H},$$

(10)

where $L_H$ is a horizontal scale of local inhomogeneities and $V(b)$ is wind speed at height $b$. For typical values in the MIZ in the present case $u_b \approx 1$ m s$^{-1}$, $L_H \approx 1000$ m, $V(b) \approx 15$ m s$^{-1}$, and the value of $b$ is therefore of order 10 m. This shows that the flow above this height, or in our MM5 setup with the lowest model level at about 20 m over the ocean, is little affected by the local inhomogeneities so the use of wind velocity, temperature, and humidity at this level in calculating surface fluxes is justifiable.

The effective roughness ($Z_{eff}$) and effective friction velocity ($u_{\text{eff}}$) of the model grid square are calculated as in Taylor (1987):

$$\ln Z_{\text{eff}} = (\ln \zeta_a) + a \sigma_{\text{inc}}^2,$$

(11a)

$$u_{\text{eff}} = (u_b)^2 (1 + a^2 \sigma_{\text{inc}}^2) = (u_b)^2,$$

(11b)

where $(\cdot)$ denotes area averaging and $\sigma_{\text{inc}}$ is the variance of the logarithm of the roughness length and $a = 0.09$. Thus Eq. (11b) is also well approximated by Eq. (7). It should be noted that in calculating the roughness length of the polynyas within the MIZ, the use of the Charnock relation is less justifiable than it is over the open ocean as a result of the limited fetch. It is thus possible that values of sensible and latent heat fluxes from water in the MIZ may be underestimated by our method. However, available data do not allow for strict verification. The value of $Z_{\text{eff}}$ is calculated from linear combination of Eqs. (5a) and (5b):

$$\ln \left( \frac{Z_{\text{eff}}}{Z_{\text{eff}}^*} \right) = 2\text{icec} + (2.48 \text{Re}_b^{2.5} - 2)(1 - \text{icec}).$$

The grid-averaged Monin–Obukhov length is given by

$$L = \frac{u_{\text{eff}}}{k^2 (u_b \theta_b)^{\gamma_{\text{TOT}}}}.$$

The bulk Richardson number required for determining vertical profiles of diffusivities in the PBL is calculated from

$$R_b = \frac{z_a}{L} \left[ \ln \left( \frac{z_a}{Z_{\text{eff}}} \right) - \Psi_H \left( \frac{z_a}{L} \right) \right]^2 - \ln \left( \frac{z_a}{Z_{\text{eff}}} - \Psi_M \left( \frac{z_a}{L} \right) \right)^2.$$

(13)

In closing, we note that this approach to the representation of surface fluxes over the MIZ would be generally applicable for the parameterization of other subgrid-scale surface inhomogeneities.

d. Experiment design

To begin, we investigate the impact that inhomogeneities in the sea ice have on the evolution of the flow by comparing a model simulation with our MIZ parameterization to one in which 100% coverage is assumed up to the ice edge defined as 0% ice concentration isolinewe will validate our solutions against the in situ observations of RM. Finally, we will examine the role of air–sea interaction and upper-level potential vorticity anomalies in the development of the observed polar lows. In this latter work, we will not address the role of latent heat release plays in the formation of polar lows as it has been documented in numerous theoretical (e.g., Craig and Cho 1988) and numerical (e.g., Mailholt et al. 1996; Bresch et al. 1997) studies.
The limited horizontal resolution of our simulations and the inadequacy of available turbulence closure schemes in MM5 does not allow us to investigate the convection that manifests itself as the cloud streets and cells evident in Figs. 2–4. Rather our focus will be on the evolution of the PBL, the spatial variation in the air–sea fluxes, and the mesoscale circulation pattern over the Labrador Sea.

For all simulations, we have used 2.5° × 2.5° ECMWF analyses as the first-guess fields. These were enhanced with available surface and upper-air observations. The area of interest is the model domain shown in Fig. 7. It is worth noting that the density of upper-air and surface stations in sparsely populated northern Canada is several times lower that in the United States. Within this domain there are only 21 upper-air stations and about 60 surface stations north of 50°N and west of 50°W that could be used in the ECMWF data assimilation cycle and for the enhancement of the analyses. There were no ship/buoy data in this area for the time of interest.

The coarse model domain is covered by 76 × 94 grid points with 60-km spacing. Embedded within it was the second domain (20-km grid resolution) that covered the Labrador Sea. Experiments with a third domain (6.6-km grid resolution) over the area where polar lows developed were also performed but except for some rather minor details results for this domain were similar to the results for the coarser second domain and differences were not essential for our conclusions.

For all the simulations, the number of the vertical full σ levels is set to 28. To avoid inaccuracies in calculating vertical derivatives in the discretized equations, if σ is not a continuous function, care has been given to the adequate spacing of the levels by stretching them vertically in a log-linear manner with highest resolution at the surface (1.0, 0.9943, 0.9872, 0.9784, 0.9675, 0.9541, 0.9377, 0.9179, 0.8941, 0.8659, 0.8327, 0.7942, 0.75, 0.7, 0.65, 0.6, 0.55, 0.5, 0.45, 0.4, 0.35, 0.3, 0.25, 0.2, 0.15, 0.1, 0.05, 0.0). The model uses velocity, temperature, and humidity at its lowest level, about 20 m in the current case, to calculate surface fluxes based on similarity theory as implemented in the surface layer described in the beginning of this section. Thus in cases when boundary layer forcing is of concern, it is important that the lowest model level lie within the surface layer. Otherwise, if the lowest model level is placed too high, significant errors may occur when strong surface inversions are present.

The shallow convection scheme that parameterizes forced shallow nonprecipitating convection in the PBL and also midtropospheric shallow convection caused by subgrid-scale effects (see Grell et al. 1994 for details) was included in all the runs. Several convective parameterizations were tested but little difference between the results for particular runs could be noted. We decided to use the Arakawa and Schubert (1974) scheme modified by Grell et al. (1994) to account for convective downdrafts. Grid-scale resolvable precipitation was calculated using a simple ice scheme devised by Dudhia (1989). Again, small differences in results could be noted when more complex mixed-phase schemes (Reisner et al. 1998) were used in simulations. None of the model runs produced precipitation even though snow showers were observed over the Labrador Sea from the R/V Knorr at this time (P. Guest 1999, personal communication). This might be attributed to the fact that parameterizations of resolvable and nonresolvable precipitation in MM5 are not fully applicable to polar regions.

The most reliable results from the model in comparison with satellite images and available analyses were obtained for 36-h simulations beginning at 1200 UTC on 7 February, that is, about 14–24 h before the in situ measurements that were made on 8 February.

All the simulations performed for this case study are listed in Table 1.
4. Model simulations

a. Sensitivity to sea-ice parameterization

To evaluate the impact that fractional ice cover in the MIZ has on the flow, two simulations were performed: the first (MODI.MIZ) in which our parameterization of subgrid inhomogeneities in sea-ice cover are included and the second (MODI.NOMIZ) in which 100% coverage up to the ice edge is assumed. In both simulations, the same parameterizations of physical processes as described above were used. In particular, the surface layer was represented by our modified parameterization. The sea-ice concentration in the Labrador Sea interpolated from NSIDC data to the inner domain is shown in Fig. 8. It is interesting to note that this figure corresponds quite well with Fig. 2. In particular a coastal polynya centered at 56.5°N, 61.5°W is shown clearly in both figures, as are areas of high ice concentration located to the northwest and southeast of the polynya. The thick line in Fig. 8 (from 54.02°N, 54.91°W to 55.35°N, 48.11°W) denotes the high-level dropsonde leg flown twice on 7 February 1997 (see RM for details).

A sounding from 0307 UTC on 8 February 1997 at 54.02°N, 54.91°W was taken over the MIZ where the ice concentration according to Fig. 8 was approximately 75%. In Fig. 9, a comparison of the dropsonde data with the two model simulations at the 15th hour of each simulation, that is, 0300 UTC, is shown. From Fig. 9, we note that the thermal structure, with its neutral stability, observed in the PBL is represented well in the
Fig. 9. A comparison of observed profiles (solid lines), model profiles from the MODL.MIZ simulation (dashed lines), and model profiles from the MODL.NOMIZ simulation (dashed-dotted lines) at 0300 UTC 8 Feb 1997—15th hour of the simulations in the vicinity of 54°N, 55°W, a location over the MIZ. (a) The potential temperature, (b) the specific humidity, (c) the zonal component of the wind, and (d) the meridional component of the wind.

Simulation with our parameterization of the MIZ. In contrast, the simulation with 100% ice cover has a PBL that is too cold and is stably stratified. The height of the boundary layer (defined here as the height of the temperature inversion) is slightly overestimated in the MODL.MIZ simulation though it is difficult to assess whether the sensible heat flux is too large or if the PBL scheme requires adjustment. The humidity sounding shows 100% saturation throughout the atmosphere. We consider this unlikely and attributable to instrument error in the dropsonde. Nevertheless, it is useful to plot, as a reference, the specific humidity of the saturated
atmosphere. There is significant difference in humidity between the two simulations due to the fact that saturation vapor pressure is strongly dependent on temperature and thus much higher in the simulation that includes a representation of MIZ. Wind profiles are in general less affected by surface inhomogeneities in MIZ. Unfortunately wind data below 940 mb were unavailable.

A comparison of the model results for the 25th hour of each simulation, that is, 1300 UTC, with a sounding over the ocean on 8 February at 1303 UTC is shown in Fig. 10. Figure 10 shows that at about 200 km from the 50% ice concentration contour there still exist identifiable differences in the potential temperature and specific humidity between the two model runs (winds were not available for this sonde). It should be noted that our parameterization of the MIZ produces a stratification in the PBL that is closer to observations (though the air is slightly too cold). The MODI.MIZ simulation is too humid while the MODI.NOMIZ simulation compares well with observations in the lower PBL. The situation is reversed in the upper PBL. Both simulations underestimate the PBL height by about 150 m.

In Fig. 11, the sensible and latent heat fluxes for the MODI.MIZ and the difference in the fluxes between the MODI.MIZ and MODI.NOMIZ simulations are shown at 1200 UTC on 8 February (24 h into the simulations). In both simulations, the maximum sensible heat fluxes occur in the vicinity of the ice edge where the cold and dry arctic air first comes into contact with the relatively warm waters of the Labrador Sea. The area of maximum latent heat fluxes is located farther SE where the sea surface temperature is higher and does not coincide with that for sensible heat. This is a result of the nonlinear dependence of saturation vapor pressure on temperature. It should be noted that the differences in fluxes between the two simulations are most significant in the marginal ice zone and are correlated with ice concentration in Fig. 8. Over the ocean, the differences between the two simulations decrease with distance from the MIZ. It should be noted that the fluxes in the MODI.NOMIZ simulation are systematically higher than in the MODI.MIZ simulation. This can be understood as being the result of the fact that the PBL air in the MODI.MIZ simulation had already undergone some modification while passing over the MIZ and hence offshore of the ice edge; it was thus warmer and moister than the air in the MODI.NOMIZ simulation.

To complete this section, we consider the impact that the MIZ has on the evolution of the polar lows that were observed to develop. Since, as noted in section 2, satellite imagery provided the only proof that the polar lows developed, we cannot directly verify accuracy of the model output except for the location of disturbances. Figure 12 compares surface streamlines and wind speed for the simulations at 1200 UTC on 8 February. It is clear from the figure that a representation of the surface inhomogeneities in the MIZ has an impact on the development of the polar lows. The more southerly position of vortex A attained in the MODI.MIZ simulation agrees better with the satellite imagery (see Fig. 3) than its position in the MODI.NOMIZ run. Furthermore, this vortex in the MODI.NOMIZ simulation has a dipole structure that is not present in the imagery. With regard to vortex B, we believe that the MODI.MIZ simulation better reproduces its development and location than the
simulation MODI.NOMIZ, which has only a weak signature of a circulation in the region where it developed. The region of strong horizontal shear to the south of vortex A in the MODI.MIZ simulation corresponds well with the cloud feature D in Fig. 4, along which several small-scale vortices can be seen. Their origin has been discussed in section 2. Further comments on the quality of MODI.MIZ simulation follow in section 5.

Inspection of Fig. 12 shows higher wind speeds over the ocean in the MODI.MIZ simulation as compared to the MODI.NOMIZ simulation. This is especially true and is accompanied by a difference in wind direction...
in the region bounded by $50^\circ$–$55^\circ$N and $50^\circ$–$55^\circ$W. Possibly, the lower roughness of water in the MIZ for simulation MODI.MIZ resulted in the higher wind speeds despite the fact that the surface sensible heat fluxes and mixing are stronger for this simulation. It is also possible that winds for MODI.MIZ are stronger since higher fluxes promote entrainment of momentum from above the PBL and also affect PBL height. One can also notice a discontinuity in surface wind direction in flow from land over ice and from ice over ocean for MODI.NOMIZ, which is absent for MODI.MIZ. In both simulations wind speed attains high values of over $25\text{ m s}^{-1}$ over the marginal ice zone just west of the center of the polar low.

The primary model sensitivity with respect to our parameterization of MIZ appears to be related to the ice concentration and thickness. Uncertainties in these fields are greater than those associated with other parameters such as ice roughness, albedo, emissivity, and water salinity (which affects the freezing point of water). Since model results are quite sensitive to the ice concentration, it is presumed that error induced by inaccuracies in its recovery from the microwave sensor data and interpolation to the model grid might be bigger than that introduced by the inaccuracies associated with the properties of ice/water.

\textit{b. Sensitivity to marine surface-layer parameterization}

In this section, we will consider the impact that the stock parameterization of the surface layer in MM5 has on the evolution of the flow. This will be achieved by comparing a simulation with this parameterization (STOCK.MIZ) to one in which a more conventional surface layer parameterization, as described above, is applied (MODI.MIZ). In both cases, our parameterization of the MIZ was included in the simulations. The results attained will be compared to estimates of the surface sensible and latent heat fluxes and their spatial gradients given in RM. No tuning of the surface-layer parameterizations to the data was attempted.

A sounding from 8 February at 0200 UTC at $55.35^\circ$N, $48.11^\circ$W is compared with the above models simulations (14th hour of the simulations) in Fig. 13. It is apparent that the stock MM5 surface-layer parameterization leads to a PBL that is significantly deeper (by over 500 m) and warmer (by 3 K) than observed. In marked contrast, our modified surface layer treatment results in a PBL that is considerably closer to observations albeit slightly too deep by about 100 m. In both simulations, the potential temperature at higher levels is very close to observations. Unfortunately, the humidity on this sonde was again not reliable and it is only plotted as a reference to the saturated atmosphere. It is interesting to note that the two simulations show that transport of moisture penetrates the lower temperature inversion and reaches several hundreds of meters above where it rapidly ceases at an inflection point in potential temperature. Unfortunately, the accuracy of this model prediction cannot be assessed at this time. For both simulations, the horizontal winds remain close to observations at low levels but diverge away from observations at the higher levels. The wind profile extracted from the sounding shows rather erratic behavior and its reliability can be questioned.

A sounding from the same day at 1303 UTC (25th
hour of model simulations) at 54.84°N, 51.22°W in Fig. 14 (wind not shown as discussed previously) confirms that the default MM5 surface-layer parameterization leads to a warmer, moister, and deeper PBL than is observed.

Figure 15 shows a comparison of surface sensible and latent heat fluxes for the two simulations at 1200 UTC on February 8. It can be noted that the differences identified above are significant throughout the region. For example, the peak sensible heat fluxes in the STOCK.MIZ simulation are larger by about 40% than in the MODI.MIZ simulation, while the peak latent fluxes are higher by about 25%. Moreover, a comparison of PBL heights and surface streamlines (Figs. 12 and
Fig. 14. As in Fig. 11 except for model profiles from the MODI.MIZ simulation (dashed lines) and model profiles from the STOCK.MIZ simulation (dashed-dotted lines) at a location over the open ocean (55°N, 51°W) at 1300 UTC 8 Feb 1997—25th hour of the simulations.

5. High-level potential vorticity and polar lows, or why the model makes a wrong forecast

As discussed in section 2, the development of multiple polar lows over the Labrador Sea was probably associated with the tropopause fold over the area. Thus the ability of any model to correctly reproduce PV structure should be crucial in forecasting such events. In this section we consider this aspect of the models performance.

For brevity, we restrict our attention to the MODI.MIZ simulation as we believe, for the reasons outlined above, that it best captures the physics of high-latitude air–sea interaction in the presence of sea ice. Furthermore, the other simulations (MODI.NOMIZ and STOCK.MIZ) exhibited similar behavior regarding the evolution of the PV field.

In Fig. 17, surface streamlines are displayed at 1800 UTC on 8 February and 0000 UTC on 9 February. A comparison with the satellite image at 2109 UTC on 8 February (Fig. 4) indicates that movement of polar low A was not well forecast. While the model predicted its eastward drift with the mean flow the observations show that it remained close to the ice edge. To explain this discrepancy we will look at the development of the high-level PV structure.

In Fig. 18, analyzed and modeled 295-K isentropic potential vorticity (hereafter IPV) surfaces for 1200 UTC on 8 February and 0000 UTC on 9 February are shown. First, note that at the earlier time analyzed IPV compares well with the TOMS data (Fig. 5b) over northern Labrador and the Labrador Sea. Also the model-
simulated IPV is of relatively good quality over this region though the anomaly maximum over the Labrador Sea is about 0.8 PVU lower than the analyzed. As a result, we suspect that the model-simulated depression in polar low A (Fig. 19) was smaller than in reality and the strength of the circulation associated with it was underpredicted.

We should also note that both analysis and the model failed to reproduce a secondary maximum of PV over the Gulf of St. Lawrence visible in Figs. 5b and 6.

Comparison of the analyzed and modeled IPV at the later time shows significant discrepancies between the two. While the model predicted eastward drift of the primary IPV maximum with the mean flow, the analysis
kept the IPV maximum at the ice edge. We lack any observations related to PV structure at the later time on 8 February that would allow for the verification. However, TOMS data valid at approximately 1600 UTC show that the drift of the primary PV maximum was occurring in reality. This would suggest that the model correctly predicted the movement of the primary PV maximum. However, since the secondary PV maximum over the Gulf of St. Lawrence was not captured at the earlier time the new source of PV over the Labrador Sea was unavailable to sustain polar lows A and B at the ice edge and to allow for disturbance D to develop. More in-depth analysis of the apparent forecast failure would require use of an adjoint model.
A simulation that began on 7 February at 0000 UTC, 12 h earlier than the MODI.MIZ, was also performed. In this simulation, the PV anomaly was weaker than for the MODI.MIZ simulation and dislocated by approximately 5° to the east. Subsequently, a polar low developed at the location several hundred kilometers from the true location.

The above discussion highlights difficulties the modeller faces in simulations of mesoscale lows in the areas of sparse observations where quality of analyses is questionable. We suggest that the incorporation of a variational data assimilation cycle using satellite data is the best solution to this vexing problem.

6. Summary and conclusions

In this paper, we have evaluated the performance of the PSU–NCAR mesoscale forecast model MM5 in simulating a cold-air outbreak over the Labrador Sea. In particular, the impact on air–sea interaction, boundary layer development, and mesoscale structure arising from the parameterization of the marine surface layer and the marginal ice zone have been examined. The case chosen was one in which mesoscale cyclones, or polar lows, developed in the outflow. It was also an event in which two research flights were flown to investigate the response of the atmosphere to the transfer of heat and moisture that occurred during the outbreak. This data combined with available satellite imagery was used to validate the model’s performance.

Our analysis of the stock parameterization of the surface layer in MM5 suggests that it is inappropriate for use over the high-latitude ocean. In particular, it leads to large overestimates of the sensible and latent heat fluxes that results in a marine boundary layer that is too deep, too warm, and too moist. In its place, a modified parameterization was proposed that did not exhibit the seemingly unphysical behavior of the stock parameterization. Furthermore, there were significant errors in the forecast position and structure of the polar lows. Our modified surface-layer parameterization resulted in a boundary layer closer to the observations albeit slightly too cold and shallow. It is important to note that surface fluxes are important input parameters into MRF PBL parameterization and affect model performance in the whole boundary layer. It is conceivable that inaccurate surface fluxes may produce meteorological fields that agree well with a limited set of observations. This result may seem acceptable to the meteorologist but is not satisfactory to an oceanographer concerned with the magnitude of the corresponding changes to the density of the surface waters. This aspect of MRF PBL parameterization is beyond the scope of the current paper but should also be examined.

The modified surface-layer parameterization was also able to correctly forecast the initial development of the polar lows. However, regardless of the choice of surface-layer parameterization, the model was unable to capture the latter stages of the polar lows’ life cycle due to the incorrect structure of the upper-level potential vorticity field. We suggested that in high latitudes quality of the analyses used for model initialization and lateral boundary conditions is responsible for poor model performance.

We also proposed a new parameterization of the surface inhomogeneities within the marginal ice zone and discussed its effect on fluxes and mesoscale flow features. Our results suggest that open water within the
MIZ modifies the temperature, humidity, and depth of the planetary boundary layer to such an extent as to affect the spatial distribution of the surface heat fluxes over the open ocean. Indeed, impacts were observed several hundred kilometers downstream of the ice edge. This result is important in determining the buoyancy forcing that the atmosphere provides to the ocean and that determines the location and intensity of deep ocean convection that occurs in the Labrador Sea (Lab Sea Group 1998). As was demonstrated, it is also crucial to the correct forecast of mesoscale cyclogenesis that took place.

To summarize our discussion of the impact of surface layer and MIZ parameterization on air–sea interaction, we present in Fig. 20 the sensible heat flux from the four model runs (STOCK.NOMIZ, STOCK.MIZ, MODI.NOMIZ, MODI.MIZ) and estimates from RM derived from dropsonde data along the flight track shown in Figs. 3 and 8. This figure confirms our earlier assertion that the stock surface-layer parameterization
in MM5 results in fluxes that are substantially larger than those obtained with our modified parameterization and observations. The impact that a parameterization of the MIZ has on the spatial variability on the sensible heat flux can also be clearly seen. The representation of the air–sea interaction that occurs within the regions of open water within the MIZ results in a substantial transfer of heat to the atmosphere. The airmass modification that takes place results in a significant reduction in the magnitude of the air–sea interaction several hundred kilometers downstream of the MIZ over the open ocean. The figure also shows that the best agreement with the estimates of RM occurred in the MODIMIZ simulation. There is still however a systematic bias with the model predicting fluxes higher than estimated by RM. In appendix B we present yet another independent estimate of the sensible heat flux where a similar bias between model and observations is present. As discussed, the nonlinear response of the atmosphere makes it difficult to isolate the source of the bias. There is still room for improvement in the model’s representation of the surface layer, for example, the use of a coupled wave model to generate roughness (Doyle 1995) or in the parameterization of physical processes within the planetary boundary layer or in the representation of the MIZ. The aircraft used by RM were unable to directly measure the turbulent air–sea fluxes and thus there is some uncertainty in the estimates derived with the bulk method. In addition, the dataset collected, although important as evidenced in this paper, is furthermore limited in its spatial and temporal sampling especially over the MIZ.

As a result it is not of sufficient quality or density to assist in resolving the source of the remaining bias. There is a clear need for additional direct aircraft observations of the turbulent air–sea fluxes over the MIZ and offshore region that can be used to further improve the model’s ability to predict the atmosphere’s response to the large air–sea fluxes that occur in the Labrador Sea and other high-latitude regions.

The results presented in this paper clearly show the need to use appropriate surface-layer parameterizations and a representation of the MIZ in mesoscale forecasts made in high-latitude marginal seas such as the Labrador Sea where intense air–sea interaction and mesoscale cyclogenesis occurs.

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Fig. 19. Forecast at 1200 UTC 8 Feb 1997 of sea level pressure from the MODIMIZ simulation.

Fig. 20. A comparison of observed (estimates based on aircraft data from RM) and model sensible heat fluxes along the flight track shown in Figs. 3 and 8. The dotted line is from the STOCK.NOMIZ simulation. The dashed–dotted line is from the STOCK.MIZ simulation. The dashed line is from the MODI.NOMIZ simulation. The solid line is from the MODI.MIZ simulation. The circles with error bars are the RM estimates based on dropsonde data.
APPENDIX A
Comparison of Surface-Layer Parameterizations

To illustrate our concern about inappropriate parameterization of surface fluxes in the stock MM5 we present results obtained with different surface-layer representations for unstable conditions. Specifically, we compare the default MM5 parameterization (‘STOCK’) with our modified parameterization (MODI) and with the widely used parameterization of Large and Pond (1982), hereafter LP:

\[ C_{m} = \frac{2.7 \times 10^{-3}}{u_{10}} + 1.42 \times 10^{-4} \]
\[ + 7.64 \times 10^{-5} u_{10}, \tag{A1} \]

and \( z_{0} = 10 \exp\left[-k/\sqrt{C_{m}}\right], \)
\( z_{i} = 4.9 \times 10^{-5} \) m, and \( z_{q} = 9.5 \times 10^{-5} \) m.

The fluxes for all the parameterizations were calculated by iterating the corresponding bulk formulas (see section 3) until convergence has been reached.

To illustrate our concern, three different atmospheric conditions were used: polar, midlatitude, and tropical. The assumed oceanic and atmospheric conditions for each case are provided in Table A1. Figure A1 shows our results with respect to the sensible and latent heat fluxes as well the Bowen ratio. It can be noted from Fig. A1 that the sensible heat flux for all atmospheric conditions is significantly overestimated by the STOCK parameterization in convective conditions with respect to both LP and MODI. For strong winds that result in a high value of sea roughness, sensible heat flux is especially unrealistic with values twice as large as for the
The latent heat flux of STOCK does not diverge significantly from LP and MODI. The only significant difference in latent heat fluxes occurs between LP and MODI for tropical conditions in hurricanes for winds on the order of 30 m s\(^{-1}\). Under these conditions and especially in gusting winds the above calculations are less reliable since the modification of latent heat flux due to the presence of sea spray is significant (Fairall et al. 1994) and applicability of the Charnock formula might be limited.

The friction velocity (not shown) was about 10% higher for STOCK compared to LP and MODI under all conditions.

Plots of Bowen ratio illustrate how skewed the magnitude of sensitive heat flux is compared to the latent heat flux. For LP and MODI, the Bowen ratio is independent or only slightly dependent on ocean roughness; for STOCK it increases rapidly with roughness. We believe that the fact that the ratio of the fluxes is so unrealistic might have multiple negative and unpredictable consequences affecting the quality of simulations over a very broad range of atmospheric conditions. Apparently, the effect of inappropriate parameterization of the surface layer in the default MM5 is most strongly felt under the polar conditions where the sensible heat flux over the ocean is much larger than latent heat flux.

**APPENDIX B**

**An Integrated Estimate of the Sensible Heat Flux**

On 8 February, two different aircraft missions were flown to investigate the response of the atmosphere to the intense air–sea interaction that occurred during the cold-air outbreak. On both missions, the same high-level dropsonde leg was flown (the track indicated in Figs. 3 and 8). As a result, a unique view of the temporal response of the atmosphere to the air–sea interaction is available along this track. In this appendix, we will use two two dropsondes launched approximately 12 h apart (the first at 0200 UTC and the second at 1229 UTC on 8 Feb) in the vicinity of 55°N, 48°W (the farthest offshore point on the flight track; Fig. 8) to provide an integrated estimate of the sensible heat flux.

The potential temperature profile from the two soundings is shown in Fig. B1 and it indicates that, apart from a deepening of the PBL, the stratification as well as the vertical potential temperature gradient in the inversion remained almost unchanged. This result is somewhat surprising since scaling of the thermodynamic equation for the current case suggests that horizontal advection should have the same order of magnitude as turbulent mixing. Based on comparison of figures rather than scaling arguments we proceed with an estimate of the sensible heat flux using an equation for the PBL growth (Garratt 1992, p. 155):

\[
\frac{\partial \left( h \frac{\partial h}{\partial t} \right)}{\partial t} = \frac{H}{\gamma_o}, \quad (B1a)
\]

where \( \beta = (\theta'_{w'})_{0}/(\theta_{w'})_{0} \approx 0.2 \) is the ratio of entrainment virtual flux and the surface virtual turbulent flux (slightly higher values of \( \beta \) result for moist air; however, in polar regions moisture content and fluxes have smaller magnitude) and \( \gamma_o = \partial \theta/\partial z \) is the potential temperature gradient in the inversion above the PBL. Upon integration of Eq. (B1a) from \( t_1 \) to \( t_2 \) we arrive at

\[
H = \frac{h_{t_2}^2 - h_{t_1}^2}{2(1 + 2\beta)(t_2 - t_1)} \gamma_o. \quad (B1b)
\]

The above equation assumes that the turbulent heat flux is a dominant factor affecting potential temperature in the PBL and neglects advection and effects of condensation and evaporation. The resulting value of \( H \) represents an average flux during the time of integration. Using the hydrostatic equation,

\[
\frac{\partial \Pi}{\partial z} = -\frac{g}{\theta_c} \approx -\frac{g}{\theta}. \quad (B2a)
\]

where \( \Pi = c_p(p/p_0)^{\gamma/\theta} \) is the Exner function, or
\[ \frac{\Delta \Pi}{\Delta z} \approx -\frac{g}{\theta} \]  

(B2b)

We find by interpolation from the dropsonde data that \( h_1 \approx 1600 \text{ m} \) and \( h_2 \approx 1900 \text{ m} \). Also, based on hydrostatic assumption vertical potential temperature gradient \( \gamma _{\theta} \) can be found from the data to be approximately equal to \( 8 \times 10^{-3} \text{ K m}^{-1} \). By substituting these values into Eq. (B1b) we obtain time-averaged sensible heat flux \( H \approx 300 \text{ W m}^{-2} \). In model simulation MODI.MIZ time-averaged sensible heat flux was equal to 380 \text{ W m}^{-2}.

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