Case Studies of High Wind Events in Barrow, Alaska: Climatological Context and Development Processes

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ABSTRACT

The Beaufort–Chukchi cyclones of October 1963 and August 2000 produced the highest winds ever recorded in Barrow, Alaska. In both cases, winds of 25 m s\(^{-1}\) were observed with gusts unofficially reported at 33 m s\(^{-1}\). The October 1963 storm caused significant flooding, contaminated drinking water, and interrupted power supplies. The August 2000 storm caused the wreck of a $6 million dredge, and removed roofs from 40 buildings. Both storms were unusual in that they tracked eastward from the East Siberian Sea into the Chukchi and Beaufort Seas, rather than following a more typical northward track into the Arctic Ocean.

This paper addresses, through modeling and analysis, the development processes of these two storms. The October 1963 system was a long-lived, warm core, zonally elongated cyclone that traversed around the Arctic basin through the Canadian Archipelago. The August 2000 system was an open-wave cyclone that dissipated rapidly into a weak, cold core eddy in the Alaskan sector of the Beaufort Sea. Approximating the contributions to development using terms in a quasigeostrophic omega equation, it was found that both storms were characterized by the increasing importance of the convergence of the \(Q\) vector (representing differential vorticity advection and thermal advection) in the midtroposphere, at the expense of forcing by the turbulent fluxes of heat, moisture, and momentum in the boundary layer. However, the influence of surface turbulent fluxes in the early stages of development was important, particularly for the August 2000 cyclone, which passed over an extensive coastal lead in the East Siberian Sea. This study concludes that the observed retreat in western Arctic ice cover is unlikely to be an important contributor to increasing cyclonic activity in the region, but that ice retreat north of Eurasia could have an impact.

1. Introduction

Barrow, Alaska, is located at the northernmost point of Alaska, on a broad sloping coastal plain that extends from the Brooks Range to the south to the Arctic Ocean (Fig. 1). The coastline is heavily indented with shallow bays and lagoons, and the continental shelf is relatively narrow. The coastal region is predominantly low-lying wetland tundra, dotted by numerous thaw lakes. Sand and gravel barrier islands, island relics of earlier coastal retreat processes, and the unprotected coastline are all subject to considerable erosion by wave action. Retreating sea ice in the arctic seas (Gloersen et al. 1999) combined with rising sea level and the coastal and human geography may contribute to increased impacts of meteorological events on coastal areas, including damages from high winds, storm surge, flooding, and shoreline erosion. Based on discussions with elders, students, local officials, and the general public, Barrow residents are concerned that increasing open water may also provide a significant source of energy for the intensification of cyclones. These meteorological events can also spawn secondary threats such as hazardous materials spills, which are particularly damaging to the coastal environment at high latitudes.

In this paper we examine the climatological context and development processes of two severe storms that caused high winds (around 25 m s\(^{-1}\)) and flooding in Barrow. These storms were mesoscale frontal cyclones, rather than “polar lows” (as will be shown later), which traversed the Chukchi and Beaufort Seas. Polar lows are violent subsynoptic cyclones that form poleward of the polar front, with largely convective cloud systems (Businger and Reed 1989). It has been noted that small vortices with extreme winds (>30 m s\(^{-1}\)) can be embedded within a synoptic system (Rasmussen 1985)—such a “multitype system” may be considered a polar low if...
Fig. 1. Storm tracks for the 1963 and 2000 cyclones. (a) Center of the 1963 cyclone (solid line) based on the NCEP–NCAR reanalysis; center of the 2000 cyclone (dashed line) based on the ECMWF operational analysis; and interpolated storm frequencies per year from the storm track center’s dataset generated from the six-hourly NCEP–NCAR reanalysis (Serreze 1995) for 1958–96. Each storm track center grid point was assigned a frequency value, which was then interpolated using splining. The NCEP–NCAR reanalysis data was not used for the 2000 storm due to errors in the reanalysis from 1997 forward at the time of calculating. (b) A more detailed storm track for the 1963 storm, showing dates and intensities, based on the NCEP–NCAR reanalysis (dashed line) and the Polar MM5 simulation (solid line). (c) A more detailed storm track for the 2000 storm, showing dates and intensities, based on the ECMWF analysis (dashed line) and the Polar MM5 simulation (solid line).
surface heat and moisture are crucial for development. True polar lows are considered rare in the Beaufort-Chukchi sector (Parker 1989).

In fact, the sector encompassing the eastern Chukchi and western Beaufort Seas, where these two cyclones reached their maximum intensity, is a region of low cyclone frequency (Fig. 1a) and intensity compared to other regions of the Arctic. Keegan (1958) conducted one of the earliest synoptic climatologies for the Arctic, focusing on the 15 winter months of 1952–57. Keegan noted a pronounced minimum of cyclones in this sector, suggesting that orography forms a barrier to the passage of cyclones from the northern Pacific, and cyclones originating over eastern Siberia do not generally track as far to the east. LeDrew analyzed a dynamic climatology for the region for the winters (LeDrew 1985) and summers (LeDrew 1983) of 1975 and 1976, and noted the winter storm track extending from the Norwegian Sea along the Eurasian coast to the Kara, Laptev, and East Siberian Seas [the latter area does not appear in the 1952–57 Keegan climatology but does appear in the 1957–61 Gaigerov climatology (LeDrew 1985)]. In the summer analysis, LeDrew (1983) noted regular occurrences of low pressure in the Laptev Sea with high pressure over the Beaufort Sea. Depressions off the North Slope coast of Alaska appear to be very rare in these climatologies.

Lynch et al. (2002, manuscript submitted to Bull. Amer. Meteor. Soc., hereafter LCBM), Walsh et al. (1996), and Maslanik et al. (1996) use different measures but all conclude that the late 1960s through to the mid-1980s had a very different storm climatology compared to later records. This is a period of much lower storm frequency in general, and in particular in the Beaufort–Chukchi region, compared to the period from the mid-1980s to the present. LCBM focus on Barrow and extend the record back to 1945, showing that this change is not a linear trend of increasing storms—the 1940s through to the early 1960s showed more high wind events than the next two decades. Proshutinsky and Johnson (1997) suggest a cyclic behavior with a period of 10–14 yr—in this scheme the Keegan climatology encompasses only the cyclonic regime, whereas the Gaigerov and LeDrew climatologies encompass anticyclonic regimes. Hence the differences between the LeDrew and Keegan climatologies are not unexpected.

Serreze et al. (1993, 1995) examined the characteristics of arctic synoptic activity from the early 1950s to the present. Using data from the National Meteorological Center (NMC), Serreze et al. (1993) confirmed that cyclone motion along the Eurasian coast, with the tendency for the Laptev, East Siberian, and Chukchi Seas to be common zones of migration into the Arctic Ocean, are more typical than motion farther eastward into the eastern Chukchi or Beaufort Seas (Fig. 1a). Cyclogenesis often occurs in preferred regions along the Arctic front—such preferred regions in Siberia are evident in summer and persist into the autumn (Serreze et al. 2001). Using the same dataset to focus on the Beaufort–Chukchi sector, and extending the record to 2001, LCBM note that while cyclone frequency is highly variable over long timescales, there is no significant trend in cyclone frequency for any season over the period of record, but that there is a significant increase in the intensity of summer cyclones. Thus, from the early work to the more modern climatologies, the Beaufort–Chukchi region is not the home of frequently occurring intense cyclones.

A further issue is the influence of these synoptic patterns on the evolution of sea ice cover, which in turn has a strong influence on the harshness of the storm impacts on the coastal environment. Rogers (1978) associates light ice years with more cyclonic activity in the East Siberian Sea. Maslanik et al. (1996) noted that increased cyclone activity north of Siberia since the late 1980s places the East Siberian Sea in the warm sector of cyclones, leading to more frequent warm southerly winds and hence ice retreat. The extension of the storm track through the Chukchi and Beaufort Seas can be associated with strong ice retreat in this region [e.g., Maslanik et al. (1999) for the 1998 Surface Heat Budget of the Arctic Ocean (SHEBA) year]. Hence, coastal impacts are likely to be large in the summer and autumn due to the combined effects of intense cyclonic activity and associated ice retreat.

In section 2 we describe the data and models used for the analysis of the two storms on which we focus in this paper. Section 3 places these storms in the climatological context described earlier and section 4 presents an analysis of the development processes leading to the formation and intensification of these storms, based on model simulations. Section 5 presents some conclusions regarding Beaufort–Chukchi region storms.

2. Methodology

a. Polar MM5

The Polar MM5 nonhydrostatic mesoscale atmospheric model, version 3.4, is used to simulate the storm events (Cassano et al. 2001). The Polar MM5 is modified from the standard version of the fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) (Dudhia 1993; Grell et al. 1994) to better represent atmospheric processes occurring in the polar atmosphere, as described in this section. In the Polar MM5, the grid-scale cloud and precipitation processes are represented by the Reisner explicit microphysics parameterization (Reisner et al. 1998). This parameterization predicts the mixing ratio of cloud water and ice crystals as well as the rain and snow water mixing ratios. Excessive cloud cover was found to be a problem over the Antarctic in sensitivity simulations using an older version of MM5 (called MM4; Hines et al. 1997a,b), similar to results found by Manning and Davis (1997) for cold, high
clouds over the continental United States. Replacement of the Fletcher (1962) equation for ice nuclei concentration with that of Meyers et al. (1992) in the explicit microphysics parameterization, as suggested by Manning and Davis (1997), has helped to eliminate this cloudy bias in polar simulations.

The Polar MM5 uses a modified version of the National Center for Atmospheric Research (NCAR) Community Climate Model, version 2, (CCM2) radiative transfer parameterization (Hack et al. 1993) for shortwave and longwave radiation. Since the parameterization of cloud cover in the original MM5 radiation code tended to significantly overestimate the cloud liquid water path (Hines et al. 1997a,b), the predicted cloud water and ice mixing ratios from the explicit microphysics parameterization are used for determination of cloud radiative properties in the Polar MM5. This allows for a consistent treatment of the radiative and microphysical properties of the clouds and for the separate treatment of the radiative properties of liquid and ice-phase cloud particles. It should be noted that this method treats cloud particles as equivalent spheres and has been shown to be inaccurate for nonspherical ice particles (Grenfell and Warren 1999).

Turbulent fluxes are parameterized using the 1.5-order turbulence closure parameterization used in the National Centers for Environmental Prediction (NCEP, formerly NMC) Eta Model (Janjic 1994). Other improvements in the Polar MM5 include modifications of the thermal properties used in the soil model for snow and ice surface types (following Yen 1981), an increase in the number of soil levels, and the addition of a sea ice surface type (Hines et al. 1997b). The sea ice surface type allows for the specification of fractional sea ice cover in the model initial conditions for any oceanic grid point, and this sea ice distribution does not evolve during the model simulation. The surface fluxes for the sea ice grid points are calculated separately for the open water and sea ice portions of the grid point, and then averaged. The sea ice thickness varies from 0.2 to 0.95 m and is dependent on the hemisphere and sea ice fraction at the grid point.

The model domain (e.g., Fig. 2) is centered at 68°N latitude and 172°W longitude and has a horizontal extent of 2820 km × 2370 km, with a horizontal grid spacing of 30 km. A total of 23 vertical levels are used, of which four are located within the lowest 450 m of the atmosphere. The lowest sigma level is located at a nominal
height of 38 m above ground level. The model top is set at a constant pressure of 100 hPa. The initial and boundary conditions for the model atmosphere are initialized using European Centre for Medium-Range Weather Forecasts (ECMWF) Tropical Ocean Global Atmosphere (TOGA) analyses for the August 2000 case, and the NCEP–NCAR reanalysis for the October 1963 case. Boundary forcing is achieved using linear relaxation, and fields without boundary values are specified to be zero on inflow boundaries and to have zero gradient on outflow boundaries. The sea ice concentrations for both cases are provided by the National Snow and Ice Data Center (NSIDC). For August 2000, the ice concentrations are derived from the Defense Meteorological Satellite Program (DMSP) Special Sensor Microwave Imager (SSM/I), generated using the bootstrap algorithm (Comiso 2002). For the October 1963 case, the sea ice concentration was assembled from the Arctic and Southern Ocean Sea Ice Concentrations (Walsh and Chapman 2001).

b. Omega equation analysis

An important issue in the analysis of the development of these systems is the relative contributions of differential absolute vorticity advection [term A, Eq. (1a)], thermal advection [term B, Eq. (1a)], and diabatic processes associated with the atmospheric boundary layer and from latent heat release within the system [term C, Eq. (1a)]. Such a diagnosis can be approximated by using a diabatic form of the quasigeostrophic omega (vertical velocity) equation (e.g., LeDrew 1988; Roebber 1989; Deveson et al. 2002):

$$\nabla^2(\omega) + \frac{f_o^2}{\rho} \frac{\partial^2 \omega}{\partial p^2} = f_o \frac{\partial}{\partial p} \left[ \nabla \cdot \left( \frac{1}{f_o} \nabla^2 \Phi + f \right) \right] + \nabla \cdot \left[ \nabla \left( -\frac{\partial \Phi}{\partial p} \right) - \frac{R}{\rho c_p} \nabla^2 H \right]$$

$$= -2 \nabla \cdot Q - \frac{R}{\rho c_p} \nabla^2 H, \quad (1a)$$

where all symbols are defined in the appendix. The right-hand side of Eq. (1b) represents all forcing of the vertical motion in a quasigeostrophic system on an $f$-plane. In such a system, with no diabatic forcing, vertical velocity is forced solely by the divergence of $Q$ (Hoskins et al. 1978). This form of the equation avoids the inaccuracies associated with the tendency for the differential vorticity advection term and the thermal advection term [terms A and B in Eq. (1a)] to be of similar magnitude and opposite sign. The diabatic heating rate $H$ is made up of contributions from the surface sensible heat flux ($H_{sh}$), the latent heat release ($H_{in}$), and convective processes ($H_{conv}$):

$$H_{sh} = \frac{F_r}{\rho \Delta Z}$$

$$H_{in} = \frac{L \Delta C}{\Delta t}$$

$$H_{conv} = \frac{L \Delta q_{conv}}{\Delta t}$$

(4)

The equation can be solved as a Poisson equation by specifying the Polar MM5 vertical velocity at the top and sides of the grid cell of interest, and a friction-induced component of vertical velocity at the bottom of the grid cell $\omega_{friction} = (\nabla \times \tau_0) \cdot k/\rho f$. The diabatic heating rate due to latent heat release is further assumed to be partitioned according to the proportion of water vapor in the grid cell that is horizontally advected compared to that which is vertically transported by latent heat fluxes from the surface.

In our application, we focus on the 850-hPa layer to remain above the boundary layer throughout the simulation. The zone of maximum uplift associated with the cyclone is identified, and for this location, the contributions to omega from each term in Eq. (1b) are calculated for the 850-hPa grid cell at this location (with differencing calculated between the 800- and 900-hPa layers). In this way, it is possible to identify the critical development processes at each stage of the cyclone’s life cycle.

3. Case study descriptions

It is often reported that the most severe storm along the Beaufort Sea coast occurred from 3 to 5 October 1963. This storm caused extensive damage at Barrow: the damage estimate was $3.25 million (about $19 million in 2001 dollars). Highest observed winds were 25 m s$^{-1}$, with gusts unofficially reported at 33 m s$^{-1}$ (Hume and Schalk 1967). Water moved inland 122 m from the shore and large chunks of sea ice were washed inland about 4.6 m. Waves were estimated to be 3-m high with a storm surge of 3 m. Coastal areas in the vicinity of the Naval Arctic Research Laboratory (around 5 km northeast of Barrow airport) and shorelines backed by lakes were severely flooded, contaminating local drinking water. Damage included the destruction of 19 buildings, destruction of power lines, erosion of bluffs southwest of Barrow by 3 m, and a shoreline retreat of 18 m.

According to the NCEP–NCAR reanalysis data, the storm originated along the Arctic front over Siberia around 145.6°E on 2 October 1963 (Figs. 1a,b). Over the next 24 h it traversed to the coast of the East Siberian Sea and continued northward, as is typical for such systems. However, shortly after 0600 UTC 3 October, the storm turned eastward and commenced a rapid deep-
enning, reaching an analyzed central pressure of 988.7 hPa at 1800 UTC 3 October while located in the Beaufort Sea north of Barrow (Figs. 1b and 2). Based on the NCEP–NCAR reanalysis, the cyclone continued to intensify as it traversed eastward, reaching its analyzed peak intensity of 981.0 hPa when it reached the Canadian Archipelago at 1200 UTC 4 October. It should be noted that it is likely that the minimum central pressure of the cyclone was deeper than indicated by the NCEP–NCAR reanalysis. Schafer (1966) shows an approximate synoptic analysis of the storm (reproduced in Fig. 2c) based on surface data from the National Weather Service (NWS). This “hand analysis” indicated that the storm reached an intensity of 976 hPa, at 2000 UTC 3 October. This intensity is more consistent with the 3-m storm surge observed. The location of the Schafer (1966) analyzed storm center at this time is shown in Fig. 1b—its position to the northeast of Barrow suggests that the NCEP–NCAR reanalysis shows the storm moving somewhat too slowly. Further, Schafer (1966) suggests that it is likely that the storm was decaying by 0000 UTC 4 October, not continuing to deepen as shown in the NCEP–NCAR reanalysis. Because of the low spatial resolution, sparse observations, and model biases typical of the polar regions (see e.g., Cullather et al. 2000; Francis 2002; Walsh et al. 2002) it is not surprising that details of the storm in the NCEP–NCAR reanalysis are not consistent with independent observations. In any case, the cyclone was of sufficient strength to cause the extreme winds measured at Barrow as it traversed the Chukchi and Beaufort Seas, particularly because of the presence of strong high pressure over the Aleutian Islands (central pressure 1030 hPa).

The storm of 10–11 August 2000 (Figs. 1c and 3) produced official record wind gusts of 29 m s\(^{-1}\) and sustained winds of 25 m s\(^{-1}\), according to the NWS anemometer. The peak wind intensity in Barrow occurred at 0000 UTC on 11 August 2000. It has been suggested that the NWS anemometer may have “pegged out” —a maximum wind gust of 33 m s\(^{-1}\) was reported at the NOAA Climate Monitoring and Diagnostics Laboratory (CMDL) weather station (D. Endres 2001, personal communication) although the official CMDL peak 1-s wind was recorded at 31 m s\(^{-1}\). Emergency management teams had insufficient notice to mobilize heavy equipment to build temporary protective berms along
the coast. The storm eroded the beach to within 100 m of a main junction location of the underground utility corridor, sunk the dredge barge, washed out a boat ramp, and removed roofs from 40 buildings. A tidal gauge at Prudhoe Bay reported a storm surge of 1.46 m, within 6 cm of the 100-yr event (A. Morkill 2001, personal communication). The final cost of this storm, primarily due to the damaged dredge barge Qayuutaq, was $7.7 million.

The August 2000 storm originated around 145.1°E but in this case the Arctic front was situated considerably closer to the East Siberian Sea coast than the 1963 storm (Figs. 1a,c). Twelve hours later, around 0600 UTC 9 August 2000, it crossed the coastline at the same location as the 1963 storm (within the roughly 200-km resolution of the analyses), then continued directly eastward. As the cyclone traversed from the Chukchi to the Beaufort Sea it continued to deepen, reaching an intensity of 989.3 hPa, 290 km to the north of Barrow, around 0000 UTC 11 August, and then dissipating rapidly before reaching the Canadian Archipelago. Figure 3c shows the storm at 1200 UTC 11 August from Advanced Very High Resolution Radiometer (AVHRR) data, from the AVHRR Polar Pathfinder dataset, developed as part of the NOAA–NASA Pathfinder effort (Fowler et al. 2000). The image is a 25 km × 25 km product averaged from composited channel-4 (thermal) and channel-2 (reflectance) 5 km × 5 km images. From this imagery, the storm is clearly mesoscale in nature [the organized cloud bands span a diameter of O (400 km) at the time shown], and shows the ECMWF analysis tracks it rather too rapidly (Fig. 1c). It is clear from the storm track climatology (Serreze 1995), also shown in Fig. 1a, that the trajectories of the October 1963 and August 2000 storms were relatively unusual.

The intensity of Beaufort–Chukchi sector cyclones has significantly increased over the past 40 yr in summer, but not in other seasons, and the number of cyclones in this sector has not increased in any season (LCBM). Not all high wind events in Barrow are associated with Beaufort–Chukchi cyclones. Barrow high wind events can be linked to a range of different synoptic conditions, including large, deep Aleutian cyclones and strong Beaufort–Chukchi ridges. LCBM analyzed the occurrence of high wind events in Barrow over the past 55 yr, using available NWS observations. The number of high wind events each year decreased throughout the ’50s, ’60s and ’70s, and have increased through the ’80s and ’90s. Average winds have increased in winter (significant at the 99% confidence level) and spring (significant at the 95% confidence level), but not in autumn or summer. There is little apparent linear trend in the highest winds in any season, and the cyclone of August 2000 is a clear outlier in the August record (Fig. 4a), and in fact in the summer record. The daily average winds associated with the October 1963 cyclone do not represent such a strong departure from climatological intensity (Fig. 4b), and although the highest observation is significantly stronger than any other October, several other years have suffered fall storms in other months with comparable average daily wind maxima (LCBM).

4. Simulation and analysis

a. October 1963 case

Simulation of the October 1963 case study commenced at 0000 UTC 1 October 1963. The cyclone formed to the southwest of the model domain, and first appears 24 h after initialization at the domain’s western boundary, over Siberia. It then travelled very rapidly northeastward to the coast, crossing the coast around 1800 UTC 2 October, just slightly ahead of the analyzed position (Fig. 1b). Rather than migrating northward, as is typical for such systems, the simulated cyclone tracked eastward over the next day, reaching maximum intensity of 977.3 hPa at 1500 UTC 3 October slightly to the west of directly north of Point Barrow, whereupon the system started to dissipate (Figs. 2 and 5). Throughout this period the cyclone was over sea ice, and continued to develop in a confluent region of the large-scale flow, which is consistent with the zonally elongated structure that resulted (Schultz et al. 1998; Sinclair and Revell 2000). The system was strongly steered by the 500-hPa wind, and, unusually for cyclones that form along the Arctic front (Keegan 1958), had formed a warm seclusion by the time it neared Barrow (Fig. 5). The simulated central pressure at maximum intensity was comparable to analyses performed by Schafer (1966; Fig. 2). In contrast, the NCEP–NCAR analyzed cyclone did not reach as low a central pressure and continued to intensify as it continued toward the Canadian Archipelago. As noted, however, the NCEP–NCAR reanalyses should not be considered reliable for the details of this storm, although the large-scale forcing of the boundaries of the mesoscale model appears to be adequate. The simulated wind at Barrow reached a peak of 24.2 m s⁻¹, in good agreement with Hume and Schalk (1967) and NWS observations.

Given that the Polar MM5 simulates adequately the development and path of the storm, at least up until the closest approach to Barrow, it is possible to draw conclusions regarding the mechanisms involved in storm development, based on the increased information available from the model output (e.g., Giordani and Caniaux 2001). Figure 6a shows the 850-hPa vertical velocity and mean sea level pressure simulated by the Polar MM5 at the location of maximum uplift in the storm throughout its life cycle, starting at 1300 UTC 2 October. At this time, the simulated cyclone is over land, progressing to the coast within the next 6 h. During this early period, simulated vertical velocities are quite strong, up to 2 cm s⁻¹. As the cyclone progresses across the Chukchi Sea, the rapid and consistent deepening of the simulated cyclone is clear, with uplift between 2 and 3 cm s⁻¹.
Fig. 4. Highest wind events for (a) Aug and (b) Oct, from 1945 to 2001, showing the highest daily avg wind speed reported for the season (blue); the highest sustained wind speed reported for the season (purple; 1965–79, 1984–2001); the highest observation reported for the season, which is a good estimate for sustained wind (pink; 1945–64, 1980–83), and the highest wind gust reported for the season (orange; 1976–2001).
associated with the most intense phase of the storm. Also shown is the vertical velocity as calculated by the quasigeostrophic method described in section 2b using the approximation $\omega = -\nabla \times \mathbf{v}$, and the contributions to this velocity from the convergence of $\mathbf{Q}$, the latent heat release from grid-scale and convective processes at 850 hPa, and the contribution from the boundary layer turbulent fluxes of sensible and latent heat. Apart from the first 6 h while the cyclone is close to the lateral boundaries, the calculated vertical velocity tracks the model vertical velocity quite accurately, although it slightly underestimates the magnitude. Inaccuracies in this method arise from the assumption of small departures from geostrophic flow, the assumption of zero fall velocity for condensates, truncation and convergence limitations in the numerical method, and the use of hourly model output for the calculation to compare with instantaneous vertical velocities.

As the cyclone moves out over the ice (Fig. 6b), the most important sources of uplift in the system are the boundary layer sensible heat fluxes, large-scale horizontal convergence of moisture, and subgrid-scale convection, consistent with the findings of other authors (e.g., Mak 1998; Kuo et al. 1991; Leslie et al. 1987). During this period, vertical motion forced by the term $-2\mathbf{v} \cdot \nabla \mathbf{Q}$ (that is, the adiabatic processes of differential vorticity advection and thermal advection) actually drives subsidence, and initially the effects of this synoptic environment overwhelm other sources of uplift. As the system continues eastward over sea ice that approaches 100% concentration, the influence of turbulent fluxes and latent heat release progressively diminishes, and the convergence of $\mathbf{Q}$ in the middle troposphere becomes of increasing importance, intensifying the uplift. Sensible heat fluxes from the surface are of consistent, though small, importance until around 1200 UTC 3 October, contributing around 15% of the total vertical velocity throughout this period. Once cyclogenesis commences, the forcing due to differential vorticity and thermal advection diminishes rapidly. It is clear that diabatic effects are not sufficient to sustain the cyclone except in the earliest phase of cyclogenesis, and hence this system can be classed as a frontal cyclone rather than a polar low.

b. August 2000 case

Simulation of the August 2000 case study commenced at 0000 UTC 7 August 2000—the simulated cyclone track is compared to the analyzed track in Fig. 1c. The simulated cyclone follows the track of the ECMWF analysis very closely. At 1200 UTC 10 August (Fig. 3) the simulated cyclone is slightly to the
northwest of the analyzed position, and more intense. The peak intensity near Barrow was 988.0 hPa at 1900 UTC 10 August, but central pressures within 0.5 hPa of 988.0 hPa were maintained for 9 h, encompassing the peak of 989.3 hPa in the 12 hourly ECMWF analysis data at 0000 UTC 11 August (Fig. 3). The location of the cyclone at this time was identical to the analyzed location within the grid spacing of the analysis data. In general, the analyzed cyclone deepens later and more rapidly than the simulated cyclone. The August 2000 cyclone represents a “classic” type of open wave cyclone, which is more typical of this region (Keegan 1958; Fig. 7). The cyclone formed on the northern side of the jet exit region associated with the Arctic front, and intensified as it was steered eastward by the upper-level winds (not shown).

The development of this system was more complex than the October 1963 cyclone. Figure 8a shows the 850-hPa vertical velocity and mean sea level pressure simulated by the Polar MM5 at the location of maximum uplift in the storm throughout its life cycle, starting at 0100 UTC 9 August. At this time, the simulated cyclone had just entered the domain—12 h later it passed over the coast and was situated over the coastal lead (indicated by diagonal hashes in the timeline, Fig. 8b). Also shown is the vertical velocity as calculated by the quasigeostrophic method described in section 2b, and the contributions to this velocity from $-2\nabla \cdot Q$, the latent heat release from grid-scale and convective moist processes at 850 hPa, and the contribution from the boundary layer turbulent fluxes of sensible and latent heat. As noted, the simulated cyclone undergoes consistent strengthening from the time of cyclogenesis until it reaches its peak intensity at 1900 UTC 10 August (Fig. 8a). Following this, the cyclone gradually and consistently weakens. Until a few hours after this cyclosis commences, vertical velocities at 850 hPa tend to be above 1 cm s$^{-1}$, with the most vigorous vertical motion occurring while the cyclone is at the leading edge of the open water along the Siberian coast, from 0600 to 1700 UTC 9 August 2000. Maximum gradients in the surface heat fluxes are present at this location (not shown). The quasigeostrophic vertical velocity is quite close to the modeled vertical velocity and the pattern of development is similar, and again, given the approximations inherent in the method, the errors are acceptable. Early in the development of the cyclone, uplift as strong as 11 cm s$^{-1}$ is simulated, associated primarily with convection and also the surface fluxes of latent heat. Later, as the cyclone tracks eastward over the ice, which is generally between 50% and 65% concentration, the convergence of $Q$ is of most consistent importance, with contribu-
tions of convection, surface sensible heat fluxes, and latent heat release due to the conversion of horizontally converged moisture of minimal importance. Only surface latent heat fluxes continue to play a role, albeit sporadically, comparable with the synoptic environment. As the cyclone then weakens, all diabatic terms drop to near-zero, and the $-2\nabla \cdot Q$ term contributes actively to subsidence for almost 7 h during this phase. Thus, the pattern noted in the October 1963 case is also seen here, but more dramatically, with surface fluxes very important during the early intensification, and adiabatic forcing becoming dominant for the most intense phase of the cyclone. Cyclolysis commences when this additional forcing diminishes, as the system undergoes a classic frontal cyclone progression (e.g., Schultz et al. 1998). However, given the importance of surface fluxes and convection during the initial development of this cyclone, the system could be considered to be a “multitype” polar low (Rasmussen 1985).

Two additional experiments were performed on the August 2000 cyclone. The first used the sea ice cover observed during August 1998, which saw a strong retreat of sea ice on the Beaufort Sea north of Barrow (Maslanik et al. 1999). The second experiment used the sea ice cover of August 1990, during which time there was a record retreat of sea ice in the East Siberian Sea (Lynch et al. 2001). In the first case, in which sea ice cover was reduced later in the development cycle of the system, there was minimal impact on the simulation (not shown). The resulting cyclone followed the same track and reached the same maximum intensity. The position at closest approach to Barrow was the same. In the second case, the impact on the simulated cyclone was small but discernible, particularly in the early stages. The cyclone moved across the coast slightly earlier than the control case. While over the coastal lead, the surface fluxes were increased by as much as 100 W m$^{-2}$ for the sensible heat flux and 75 W m$^{-2}$ for the latent heat flux (Fig. 9) and the horizontal gradient of surface fluxes also increased markedly. During this period, vertical velocity forced by subgrid-scale convective processes was stronger and persisted about twice as long, and the influence of the horizontal convergence of moisture on the forcing of vertical velocity by latent heating was enhanced. Further, the propagation of the cyclone slowed while over the coastal lead. With the additional forcing on vertical
velocity, the cyclone was somewhat deeper at this stage (Fig. 9c). As the cyclone tracked across the Chukchi Sea, the importance of the synoptic environment on the development of the cyclone was preeminent, just as in the control case, and hence the simulation from this time forward was very similar to the control. This pattern of behavior in the presence of low ice concentrations is consistent with the results of, for example, Angel and Isard (1997). While the closest approach to Barrow occurred at around the same time (0000 UTC 11 August), the cyclone was around 1 hPa deeper and around 50 km closer to Barrow, with concomitant stronger winds at Barrow. Cyclolysis commenced at the same time, and proceeded at the same rate in both the control and the experiment simulations.

5. Conclusions
The Beaufort–Chukchi cyclones of October 1963 and August 2000 produced the highest winds ever officially recorded in Barrow, Alaska. The October 1963 storm produced a significant storm surge and caused extensive flooding, whereas the August 2000 storm caused more limited flooding, but had a large financial impact due to the damage to a dredge and the consequent ending of the beach nourishment program (LCBM). The cyclones also differed in that the October 1963 system was a long-lived, warm core, zonally elongated cyclone that traversed around the Arctic basin through the Canadian Archipelago, whereas the August 2000 system was an open wave cyclone, which dissipated rapidly into a weak, cold core eddy in the Alaskan sector of the Beaufort Sea.

Despite these differences, there were several important commonalities between the two cyclones. Both formed along the Arctic frontal zone over Siberia and traversed northeastward to the East Siberian Sea. At this point, rather than tracking into the central Arctic Ocean as is typical for such systems, the cyclones turned in a more eastward direction, steered by the upper-level flow, into the Chukchi Sea and thence to the Beaufort Sea. As they moved from land to ocean, the cyclones intensified initially due to surface fluxes and convection over lower-concentration ice, and subsequently as a result of thermal advection and differential vorticity advection above the boundary layer. The August 2000 storm underwent more significant intensification associated with the more open coastal lead in the East Siberian Sea at this time, and for this reason could be considered to be a “multitype” polar low.

The primary impetus for this analysis is the concern of Barrow residents that retreating sea ice in the Chukchi and Beaufort Seas (Gloersen et al. 1999) may contribute to an increase in the impacts of severe storms on Barrow. This concern arises not only because sea ice serves as a significant protection against storm surge, waves, and coastal erosion, but also because there is a perception that open water provides a significant source of energy for the intensification of cyclones. However, concern is mitigated by the additional perception that cyclones in this region track the ice edge, and hence a retreating sea ice may cause the few cyclones that track eastward from the East Siberian Sea to shift further north. The study of these two systems does not bear out these perceptions completely. Neither cyclone tracked the ice edge—they were steered by a strong upper-level flow on their unusual eastward tracks. While the cyclone of August 2000 did intensify as it passed over the open coastal lead in the East Siberian Sea, the location of the ice edge north of Barrow had little impact on the course of cyclone development. This is borne out by the additional experiments that showed that the simulation of this cyclone was insensitive to the specification of the ice edge in this area. However, retreating sea ice in the East Siberian Sea had some impact on the cyclone, increasing the intensity early in its lifecycle and causing it to track more closely to Barrow in its subsequent development. This result is consistent with the analysis of Rogers (1978) who found that while the amount of open water in late summer and autumn in the East Siberian Sea is associated with greater cyclonic activity, open water in the Beaufort Sea has little or no influence upon subsequent local surface winds or the sea level pressure distribution. Thus, the retreat of sea ice in the Beaufort–Chukchi region will continue to have implications for the impacts of storms on the Alaskan North Slope coastal zone, but it is the sea ice retreat in the East Siberian Sea that should be considered a possible influence on the trends in the number and intensity of Beaufort–Chukchi cyclones.

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APPENDIX

List of Symbols

\[ c_p \] specific heat of dry air \( (\text{J} \, \text{kg}^{-1} \, \text{K}^{-1}) \)
\[ \Delta C \] total change of condensate in the 800–900-hPa-layer grid cell, including cloud water and cloud ice, assuming a fall velocity of zero \( (\text{kg} \, \text{kg}^{-1}) \)
\[ f_0 \] Coriolis parameter, constant for the domain \( (\text{s}^{-1}) \)
\[ F_s \] sensible heat flux from the surface \( (\text{W} \, \text{m}^{-2}) \)
\[ H \] diabatic heating rate \( (\text{W} \, \text{kg}^{-1}) \)
\[ L_v \] latent heat of vaporization \( (\text{J} \, \text{kg}^{-1}) \)
pressure (Pa)

\( \Delta d_{\text{conv}} \) change in water vapor in the 800–900-hPa layer grid cell due to subgrid-scale convection (kg kg\(^{-1}\))

\( Q \)

\[-(\partial \mathbf{v} / \partial x) \cdot \nabla(-\partial \Phi / \partial p), \quad -(\partial \mathbf{v} / \partial y) \cdot \nabla(-\partial \Phi / \partial p)\]

\( R \)

gas constant for dry air (J kg\(^{-1}\) K\(^{-1}\))

\( \Delta t \)

model time step (s)

\( v \)

geostrophic wind velocity vector (m s\(^{-1}\))

\( \Delta Z \)

thickness of the 800–900-hPa layer (m)

\( \rho \)

density (kg m\(^{-3}\))

\( \sigma \)

static stability (m\(^2\) Pa\(^{-2}\) s\(^{-2}\))

\( \Phi \)

geopotential height (m)

\( \omega \)

vertical velocity, in the form \( dp/dt \) (Pa s\(^{-1}\))

REFERENCES


