2016

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Zhang, Jing; Liu, Fuhong; Tao, Wei; Krieger, Jeremy; Shulski, Martha; and Zhang, Xiangdong, "Mesoscale Climatology and Variation of Surface Winds over the Chukchi–Beaufort Coastal Areas" (2016). *Papers in Natural Resources*. 590.
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Mesoscale Climatology and Variation of Surface Winds over the Chukchi–Beaufort Coastal Areas

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(Manuscript received 25 June 2015, in final form 15 January 2016)

ABSTRACT

The detailed mesoscale climatology of surface winds in the Chukchi–Beaufort Seas and adjacent Arctic Slope region is analyzed using the recently developed Chukchi–Beaufort High-Resolution Atmospheric Reanalysis (CBHAR). Within the study area, surface winds are mainly driven by the prevailing synoptic weather patterns of the Beaufort high and Aleutian low and are further modulated by local geographic features through thermodynamic and dynamic processes. Sea breezes, up- or downslope winds, and the mountain barrier jets are all clearly captured by CBHAR. Sea breezes emerge in June–September and last most of the day, with a maximum spatial extent 100 km inland and 50 km offshore and maximum speed around $1\text{--}3\text{ m s}^{-1}$ in the late afternoon [~ 1500 Alaska standard time (AKST)]. Thermodynamic impacts of mountains on the surface winds vary from time to time. Drainage flows begin to build at the mountaintop in September and reach the strongest during November–February, occupying the entire slope. Upslope winds demonstrate a clear diurnal cycle during summer, starting to build around 0900 local time, reaching the maximum strength around 1500 local time and continuing until 2100 local time. The mountain barrier jets (MBJs) are found to be most active around the Chukotka Mountains during cold seasons. Both sea breezes and MBJs are also subject to variations and changes in response to adjusted large-scale atmosphere circulation. Storm activities can inhibit the development of sea breezes. Different responses from the Beaufort high and Aleutian low to anomalies in large-scale circulations play a vital role in the variations of MBJ activities over the Chukotka Mountains.

1. Introduction

Surface wind is the fundamental dynamic driver for surface energy and water vapor transport, turbulent heat exchange between land/sea/ice and air, pollutant

dispersion, and sea ice and ocean motions. Both prevailing synoptic weather patterns and prominent underlying geographic features play important roles in shaping the wind climate (e.g., [Schwerdtfeger 1974](#); [Kozo 1979, 1980](#); [Liu et al. 2006, 2008](#); [Moore and Pickart 2012](#)). In the Chukchi–Beaufort Seas region, surface winds vary in response to the intensity and location of the Beaufort high and Aleutian low, which are the dominant synoptic-scale weather patterns in the area ([Shulski and Wendler 2007](#); [Overland 2009](#); [Moore 2012](#); [Zhang et al. 2013](#); [Wu et al.](#)

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2014). At the same time, surface winds can be further modulated to develop mesoscale components by local geographic features through both thermodynamic and dynamic processes.

The study area exhibits complex geographic features, including seasonally sea ice–covered seas and mountain ranges with different distance from the coastlines. When a great horizontal temperature gradient exists along the coast, sea breeze develops (e.g., Walsh 1974; Rotunno 1983; Dalu and Pielke 1989; Simpson 1994; Miller et al. 2003). The same mechanism also exists in the mountainous areas, where upslope and downslope winds occur (e.g., Pielke and Segal 1986; Tripoli and Cotton 1989; Wolyn and McKee 1994). During winter, when colder and denser surface air flows off the terrain, gravity drainage develops (e.g., Thorpe et al. 1980; Parish and Bromwich 1987, 1991; Bromwich 1989; Bromwich et al. 2001; Parish and Cassano 2003). When a cold stable air blows toward a mountain, mountain barrier jets can develop (e.g., Parish 1982; Bell and Bosart 1988; Xu et al. 1996; Loescher et al. 2006).

In addition, the study area is located in the marginal ice zone of the Arctic, which is experiencing the fastest rate of sea ice decline (Stroeve et al. 2007; Comiso et al. 2008), as well as enhanced surface winds as sea ice retreats (Stegall and Zhang 2012). However, meteorological observations in the harsh Arctic region are sparse, hindering the detailed analysis of mesoscale winds in the area, which would be important for land–sea heat and moisture exchanges, surface energy budgets, coastal currents, and sea ice edge distribution. The newly developed 31-yr, 10-km horizontal resolution, hourly (high frequency) regional Chukchi–Beaufort High-Resolution Atmospheric Reanalysis (CBHAR) (Zhang et al. 2013; Liu et al. 2014) thus provides a unique opportunity for conducting mesoscale climatology analysis of the surface winds in this area.

The remainder of this paper is organized thusly: Section 2 briefly describes the CBHAR data used in this study, as well the study domain. Section 3 presents the overall climatological features of surface winds in the study area. Mesoscale winds, including sea breezes, up- or downslope winds, and mountain barrier jets are detailed in sections 4–6. The interannual variability of sea breezes and mountain barrier jets are explored in section 7. And a summary of this study is given in section 8.

2. CBHAR data and domain

CBHAR was generated to provide a long-term high-resolution atmospheric reanalysis covering the Chukchi–Beaufort Seas and the adjacent Arctic Slope region for the period from 1979 to 2009. The data have a grid spacing of

FIG. 1. (a) Monthly and (b) hourly averaged RMSEs of wind vector magnitudes (m s^{-1}) in CBHAR (red) and ERA-I (blue) as verified against surface observations.

10-km and 1-hourly temporal resolution, making them able to resolve finescale mesoscale processes. Thus, we will use CBHAR to detail the mesoscale climatology and changes of surface winds in the area.

CBHAR surface winds (i.e., 10-m winds), are first verified against in situ surface observations collected mainly from the National Climatic Data Center, which are taken from over 100 stations across the entire CBHAR domain. To provide a baseline against which CBHAR data could be compared, the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-I; Dee et al. 2011) data that were used as initial and boundary conditions for CBHAR are also verified in the same manner. The statistics of surface wind vector, as shown in Fig. 1, suggests that wind vector root-mean-square error (RMSE) is largest in the winter months, when the variability of weather conditions is high, and declines during the summer, when the variability of weather patterns is at its minimum. Compared to 6-hourly ERA-I data, error reductions in CBHAR wind vector are seen for every season of the year and all four daily hours. These results demonstrate the consistently improved quality of reanalyzed surface winds in CBHAR and give us the confidence to utilize this dataset for a mesoscale climatology analysis of the surface winds in the area.

FIG. 2. CBHAR data domain with vertical cross-section profiles (P1–P4) chosen for various mesoscale wind field analyses; shoreline-parallel sections A, B, C, and D for sea-breeze variation analysis; and the three stations of Barrow, Deadhorse, and Wainwright for sea breeze comparison with observations. Gray shading over the ocean represents annual mean sea ice concentration (SIC) averaged over the entire CBHAR period (lighter colors signify higher SIC) with the 0.8 value highlighted by a dashed line. Gray shading over land represents topography with lighter colors for lower elevations.

To facilitate the analysis and discussion presented in this study, the CBHAR data-covered area is shown in Fig. 2, with superimposed sea ice and geographic features, along with the vertical cross sections and surface sections used in the analyses. The CBHAR domain encompasses the Chukchi and Beaufort Seas, the entire Arctic Slope of Alaska, and the adjacent Brooks Range, as well as portions of the Canadian Yukon and the eastern tip of Siberia. To enhance robustness of analysis results, multiple cross-shoreline/mountain sections (P1–P4 in Fig. 2) are chosen for each of the study areas, and the wind fields averaged across these sections are analyzed. A total of four major cross sections are selected for examination of vertical wind structures, including three coastal areas—one across the Chukotka Mountains (P1), and the other two along the northern Alaska coast (P2 with flat inland and P3 with mountainous inland)—as well as the Brooks Range (P4). Along the vertical sections, the horizontal winds are transformed in order to define components perpendicular and parallel to the vertical section. The wind climatology along the vertical sections is constructed for each hour of the day during different months. To accurately depict the diurnal cycle of wind circulations along P1–P4, the profile local time (PLT), rather than Alaska standard time (AKST), is used to construct the diurnal variation of wind profiles. The shoreline-parallel sections A, B, C, and D as shown in Fig. 2 are selected, where the inland topography is relatively flat so that sea breezes can be isolated from the slope winds. Then the wind fields averaged along these sections at each hour of the day for

the selected months are used to analyze the spatial variations in sea breeze, as impacted by relative distances to the shoreline.

3. Climatology of surface winds in CBHAR: Emergence of mesoscale features

We first analyze the seasonal climatology of surface winds in CBHAR (Fig. 3). Wind speeds are greatest in winter, with a maximum of about 4 m s^{-1} over the Chukchi Sea, northwest Bering Sea, and south slope of the Chukotka Mountains (Fig. 3a). Along the Brooks Range, local topography plays a determining role over surface winds, resulting in intense downslope gravity drainage of cold air with speeds of greater than 4 m s^{-1} . The surface wind maximum for the south slope of the Chukotka Mountains is attributed to both synoptic-scale flow and downslope gravity drainage. Over the north slope of the Chukotka Mountains, the mountain barrier effect turns the northeasterly synoptic flow toward the mountain range into northwesterly winds parallel to the mountain range with enhanced surface wind speeds of about 4 m s^{-1} .

During spring and fall (Figs. 3b,d), downslope winds still play a role in affecting surface winds in the mountain areas, though with reduced wind speed compared to winter. Downslope winds along the Brooks Range are characterized by wind speeds of about 2 m s^{-1} in spring and 3 m s^{-1} in fall. The mountain barrier effect can also be identified in fall along the north slope of the Chukotka Mountains.

In summer (Fig. 3c), surface winds are much weaker compared to other seasons, though sea breezes are active along the Beaufort Sea coastal area, resulting in average northeasterly winds of about 2 m s^{-1} .

Therefore mesoscale features of the area's surface winds include sea breezes, up- or downslope winds, and mountain barrier jets. Among them sea breezes and up- or downslope winds have diurnal variations during summer. Climatological sea breezes and up- or downslope winds can be identified using the following calculations. The surface wind field is first averaged over 1979–2009 at each hour of the day, according to month (such as July), in order to obtain monthly climatological diurnal surface wind. Sea breezes and up- or downslope winds are then identified by subtracting the climatological mean at a given reference time when the local thermodynamic effects are minimal. In this study, 0600 AKST during July is chosen as the reference time, based on the fact that diurnal variation in surface winds is strongest in July for the study area and the areal averaged solar radiative forcing is weakest at 0600 AKST.

FIG. 3. Climatological surface wind (color shading, m s^{-1}) in (a) winter [December–February (DJF)], (b) spring [March–May (MAM)], (c) summer [June–August (JJA)], and (d) fall [September–November (SON)] in CBHAR. Wind speed-colored vectors show the horizontal wind components.

Diurnal variations in surface winds during July occur mainly over mountainous and coastal areas (Fig. 4), confirming that sea breezes and up- or downslope winds are the major contributors. In July, the area's regional-scale winds are dominated by an anticyclonic flow centered in the Beaufort Sea under the governing of the Beaufort high, with several smaller scale cyclonic flows over land and warm water due to thermal lows (Fig. 4a). Because of the polar day and the continuous ice-covered offshore in July, the land surface is warmer throughout the entire diurnal cycle. Thus, even at 0600 AKST, surface winds along some coastal areas are enhanced because of the coupling between regional-scale winds and sea breezes. This constant land–sea temperature contrast suggests that sea breezes occur during the entire daily cycle along the shoreline, though their strength varies temporally (Figs. 4b–h), with weak onshore flow in the early morning and stronger onshore flow in the afternoon. Maximum sea breeze occurs in the late afternoon, at around 1500 AKST, with maximum wind anomalies of $1\text{--}2 \text{ m s}^{-1}$, though this varies by area. Along the Chukchi Sea coastline, the sea breeze can reach 3 m s^{-1} in July, whereas the Beaufort coast experiences a relatively weak sea breeze of about 2 m s^{-1} .

Mountain up- and downslope winds are clearly present over the mountainous areas. Along the Brooks Range, upslope winds develop when the underlying surface is heated. Maximum upslope wind occurs in late afternoon around 1500 AKST, with maximum wind speeds of up to $2\text{--}3 \text{ m s}^{-1}$ in the eastern Brooks Range. Downslope winds emerge around the hours 0000–0600 AKST. Along the Chukotka Mountains, sea breezes couple with upslope winds because of the immediate uplift of terrain close to the coastline. A similar process occurs at the eastern tip of the Brooks Range, where terrain and coastline are in close proximity.

4. Spatial structure of summer sea breezes

Summer sea breezes along the CBHAR coast demonstrate an obvious spatially dependent structure across the shoreline (Fig. 4). To reveal the details of this spatial structure, we examined the diurnal cycle of the surface winds at four selected particular shoreline-parallel sections from A to D, as shown in Fig. 2. Section A represents an offshore location about 50 km from the shoreline; section B is defined directly along the shoreline; sections C and D are located inland about 50 km

FIG. 4. (a) Climatological CBHAR surface wind vectors (m s^{-1}) at 0600 AKST. (b)–(h) Diurnal wind vector anomalies (m s^{-1}) from 0600 AKST (colored vectors, m s^{-1}) at a 3-h interval for July 1979–2009. Wind speed-colored vectors show only the horizontal wind components.

FIG. 5. Climatological diurnal variation in on- and offshore surface wind components (m s^{-1}) in July (a) along sections A, B, C, and D, as shown in Fig. 2, and (b) at three coastal stations Barrow, Deadhorse, and Wainwright and their comparisons with the observations. Positive values represent offshore-directed wind with negative values for onshore wind. Times are given in AKST.

and 100 km from the shoreline, respectively. By comparing the perpendicular components of surface winds relative to the shoreline at these sections we can identify the spatial extension of the climatological coastal mesoscale winds impacted by sea breezes throughout the day.

As indicated by Fig. 4a, offshore winds are present at 0600 AKST along the shoreline of sections A–D, which predominantly result from the combined results of synoptic-scale anticyclonic flows in the north and cyclonic flows in the south. However, these offshore winds obviously weaken or even become onshore winds with increasing radiative heating during the day, reflecting development and contribution of mesoscale sea breezes. This mesoscale feature of the surface winds and its spatial dependence can be clearly depicted by Fig. 5a, showing the average July diurnal cycle of cross-shore surface winds along sections A–D.

At section A, surface winds are always toward the offshore direction but weaker during the day and stronger during nighttime. Along the shoreline section B, offshore wind reaches its maximum of 0.5 m s^{-1} at around 0600 AKST and then begins to decrease in speed. At 1000–1100 AKST, the surface wind turns its direction from offshore to onshore, indicating the beginning of the sea-breeze development. Sea breeze reaches its maximum in the afternoon, at 1600 AKST, resulting in onshore winds at a speed $\sim 1.25 \text{ m s}^{-1}$. On- and offshore winds at the shoreline vary in a range of about 1.75 m s^{-1} from 0.5 to -1.25 m s^{-1} , which is much greater than a range of around 0.5 – 0.75 m s^{-1} at the other sections. The diurnal variation of the mesoscale winds modulated by sea breezes is the most vigorous at the shoreline. In the CBHAR domain, there are a few long-term coastal stations, as labeled in Fig. 2 [~ 31 yr at Barrow and Deadhorse and ~ 10 yr at Wainwright (Baule and Shulski 2014)], from which the observed climatology

of cross-shore winds is compared with the CBHAR winds (Fig. 5b). Since these data have been assimilated into CBHAR, variations in CBHAR and observed winds are in reasonably good agreement. The diurnal ranges of cross-shore winds at Wainwright are very close to that along section B because station Wainwright is close to the southwest end of section B.

Along the inland sections C and D, the diurnal variations of cross-shore winds are much weaker. Similar to the winds at the offshore section A, the cross-shore winds at inland locations C and D are mainly offshore, with weak onshore winds present only around 1800–0000 AKST. The maximum onshore winds at the inland locations pronouncedly lag those along the shoreline by about seven hours. This lag is caused by the gradual extension of the sea-breeze development from the shoreline into inland.

To understand the breeze's vertical spatial distributions, we first examine the two-dimensional vertical and cross-shore wind components associated with sea breeze along the vertical cross section (P2) perpendicular to sections A–D (Fig. 2). Similar to the analysis shown in Fig. 4, the climatological sea-breeze circulation is isolated by subtracting the wind field at 0600 PLT from the wind field at each hour of the day along P2.

The diurnal evolution of climatological sea-breeze circulation along P2 shows that the onshore winds begin to develop at 0900 PLT with a speed increasing to around 0.6 m s^{-1} near the shoreline as land surface warms up (Fig. 6). This is consistent with the decrease in offshore winds at section B (Fig. 5a). At the same time, upward motion also develops gradually from the inland. At noon, a well-organized sea-breeze circulation forms along P2, with onshore winds across an expanse of more than 100 km. The maximum wind speed across the shoreline is greater than 1 m s^{-1} . Over the area where

FIG. 6. Wind anomalies from 0600 PLT in July at (a) 0900, (b) 1200, (c) 1500, (d) 1800, (e) 2100, and (f) 0000 PLT along the vertical cross-sectional profile P2. White wind vectors include both the horizontal wind component (oriented perpendicular to the shoreline) and the vertical component (see legend at bottom lower right). Wind speed along the profile (m s^{-1}) is indicated using colors. Solid contours in white represent a wind speed anomaly of 1.0 m s^{-1} , which is considered as the main extent of the sea-breeze circulation. Dashed contours are for temperature ($^{\circ}\text{C}$). Gray shaded area depicts the profile of topography. The shoreline is located at 0 km.

maximum upward motion occurs, upper-air motion is oriented toward the offshore direction, at about 0.5–1 km above ground, with downward motion developing offshore within 50 km of the shoreline. The sea-breeze

circulation is strengthened further in the following hours because of a continuing warming over land. It should be noted that offshore surface temperatures do not change much throughout the entire diurnal cycle. As a result,

the thermal contrast between land and sea is mainly enhanced or weakened by warming or cooling of the land, though the land surface is always warmer than the sea surface at this season, during the entire diurnal cycle. The strongest sea-breeze circulation occurs in the afternoon, at about 1500 PLT, with maximum onshore sea-breeze wind of $\sim 2 \text{ m s}^{-1}$ located around the shoreline near the surface, elucidating the maximum onshore surface wind at the same time shown in Fig. 5. The maximum return flow of about $0.5 \sim 1 \text{ m s}^{-1}$ appears about 1 km above ground. After 1500 PLT, the sea-breeze circulation weakens gradually over the following 12 h.

Sea breeze along P2 can extend more than 100 km inland over flat onshore topography. But when the coastal mountain is present within this range, such as along the vertical sections P1 and P3, sea breezes will interact with the mountain slope winds. We therefore conduct the same analysis as described above to examine this interplay along P1 and P3, representing the Chukotka Mountain coastal and the Brooke Range coastal areas. Comparison of the development of onshore winds among the vertical sections P1–P3 can particularly help reveal how topography affects intensity and horizontal extension of the sea-breeze-induced onshore winds near the surface and intensity and locations of the associated upward motion and returning flow at the upper level. Since the analyzed physics and results along P1 and P3 are very similar, here we focus on P3.

To compare the horizontal extensions of sea breeze over different topographic features, areas with wind speed anomalies greater than 1 m s^{-1} are highlighted and considered as a core sea-breeze circulation. When examining this highlighted area, we readily find that the maximum extent of the core onshore winds can reach about 70 km inland from the shoreline along P3, where mountain slopes are present (Figs. 7). This shows an obvious difference from that along P2 over the Chukchi Sea coast, where the terrain is flat and the core onshore winds penetrate inland by about 90–100 km during the afternoon hours 1500–1800 PLT. The maximum offshore extension of the core onshore winds along P3 is about 30 km, shorter than the extension of around 50 km along P2 (Figs. 7c,d versus 6c,d). The blocking effects of the Chukotka Mountains and the eastern Brooks Range therefore reduce both the horizontal inland penetration and offshore extension of the near-surface onshore winds by about 20–30 km.

The intensity of upward motion and the location of the maximum onshore wind center also vary when different mountain slopes are present. The mountain slope receives more heat than ambient air during summertime, which produces horizontal surface temperature gradients and, in turn, upslope winds. As a result, the

slope located near the shoreline enhances onshore winds by the coupled sea breeze and upslope winds. Enhanced onshore winds with mountain dynamic uplifting further cause strong upward motion along P3 (Figs. 7). On the other hand, the enhanced upward motion tends to reduce the stability of the lower atmosphere and hence limits the horizontal inland penetration of sea breeze. When the slope is flat along P2, the location of the strongest onshore winds is centered at the shoreline (Fig. 6). When the slope increases, the center of strong onshore winds, as well as all circulation, moves toward the slope (Figs. 7). In addition, if the air mass carried by the onshore wind has sufficiently high humidity, the accompanying upward motion along the slope may result in a low cloud formation. The associated condensation and latent heat release may favor further strengthening of the upward motion. This dynamic–hydrological feedback process would be a very interesting topic, deserving further careful investigation.

Corresponding to the changes in the intensity of upward motion and the location of total breeze circulation forced by the slope, the intensity and location of returning flow at the upper level also vary. When the slope is flat (P2), returning flow is located at $\sim 1 \text{ km}$ above the ground, near shoreline and with maximum wind speed around 0.7 m s^{-1} (Fig. 6). When sea-breeze circulation moves toward inland as slope increases, stronger returning flows occur, with maximum wind speed greater than 1 m s^{-1} along P3 (Fig. 7). Following maximum upward motion, the horizontal location of maximum returning flow moves to about 50 km inland along P3, rather than within 10 km of the shoreline, as along P2. The same mountain-slope-caused shift can also be found along P1 where the Chukotka Mountains exist (not shown).

Although the analysis above focuses on July, sea breeze emerges in June and lasts throughout September, showing a pretty strong variation of spatial extension along all the three sections analyzed in this study. The maximum extension occurs in July as described above, and the offshore extension decreases to 30–40 km in June and August and further to about 10 km in September.

5. Up- and downslope winds over the Arctic Slope

While the dense cold-air flows over the Antarctic and Greenland ice sheets have been studied extensively (e.g., Bromwich et al. 2001; Parish and Cassano 2003), there have been few such studies over the Arctic Slope. Following the method used in section 4, the vertical structures of up- and downslope winds are analyzed by examining the monthly mean profiles of vertical and horizontal wind components along the selected sections

FIG. 7. As in Fig. 6, but along the vertical cross-sectional profile P3.

across the mountains. For convenience of the analysis, we use a threshold of wind speed greater than 1.5 m s^{-1} to represent establishment, development, and weakening of the slope winds.

Along the vertical section across the eastern Brooks Range (P4 in Fig. 2), we can clearly observe seasonal variations of the mountain's thermodynamic effects as

shown by the strength of drainage flows (Fig. 8). The gravity drainage from cold air starts at the mountaintop in September and gradually increases its extent and intensity down the slope in the following cold months as surface temperature decreases. This is because that strong cooling begins at the mountaintop and gradually plunges down the slope, resulting in an accumulation of

FIG. 8. As in Fig. 6, but along profile P4 and for monthly mean winds in each month.

dense cold air along the surface of the mountain slope. The strongest downslope wind occurs during the coldest months of the year—from November to February—with monthly mean downslope winds greater than 2 m s^{-1} spread over the entire slope. During this coldest season, there is no diurnal variation in surface wind because of the lack of solar radiation. After February, downslope winds shrink back to the mountaintops. The identifiable downslope winds last until April, with significantly reduced spatial extent. During the course of downslope wind evolution, the gravity drainage from cold air is trapped within a thin layer less than 300 m from the slope surface, because the extremely cold surface increases the boundary layer stability.

During the summer months from May throughout August, monthly averaged winds are generally very weak near the surface along the slope, because of the cancellation of upslope and downslope winds caused by the diurnal variation of incoming solar radiation. Noticeable weak upslope winds near the mountain slope surface only emerge in June and July and also demonstrate a diurnal variation. To depict the diurnal cycle of these slope winds, we followed the same approach employed in section 4 to analyze the wind profiles along P4 at 3-h intervals relative to that at 0600 PLT in July (Fig. 9). The wind profiles show a well-defined mountain slope circulation, varying with the hour of day. Upslope winds start to build around 0900 PLT. From

FIG. 9. As in Fig. 6, but along the vertical cross-sectional profile P4.

noon to late afternoon, upslope winds occupy the entire slope with wind speeds greater than 1 m s^{-1} . The returning flow is located at about 1000 m above ground over the mountaintop, with a maximum wind speed greater than 1 m s^{-1} at 1500 PLT. The upslope winds continue until 2100 PLT.

Note that, during the cold months, downslope winds are missing over the south slope of the eastern Brooks Range (Fig. 8) and the Chukotka Mountains (not shown) because of interactions between approaching background flows and mountain barriers in the areas (to be discussed in section 6). However, upslope winds appear on the south slope of the eastern Brooks Range during the summer months (Fig. 9).

6. Cold season mountain barrier jets

The mountain barrier effect is another mesoscale process through which topography can affect surface wind fields in the study area (Schwerdtfeger 1974; Kozo 1980). When airflow approaches a mountain, the air is either forced to ascend over the mountain or is blocked, depending on the properties of both the air layer and the topography. When airflow is characterized by a low Froude number, it can be blocked and deflected to the left, and a low-level wind maximum (mountain barrier jet) is formed parallel to the mountain barrier. In the study domain, the occurrence of low-Froude-number flow is common (Serreze and Barry 2005). The mountain

barrier effect is therefore expected to be one of the most important mesoscale processes affecting the surface winds.

Seasonal variation of surface winds over the 31-yr period within CBHAR can help us gain an overall understanding of the occurrence of the mountain barrier effect across the entire study domain. A noticeable strong wind occurs over the Chukchi Sea during the cold months (Fig. 3), which can be attributed to the mountain barrier effect. Under the influence of an intense Beaufort high centered over the northern Chukchi Sea, strong northeasterly synoptic-scale winds blow toward the Chukotka Mountains and are then deflected to flow parallel to the mountain range with enhanced speeds (barrier jets) as a result of the mountain barrier effect. Similarly deflected surface winds also occur at the slope of Brooks Range. To better understand the barrier effect in these two regions, we examine the vertical wind structures along the vertical sections P1 across the Chukotka Mountains and P4 across the eastern Brooks Range for each month of year.

Monthly averaged wind vectors along and wind speeds perpendicular to P1 demonstrate a wind flow toward the mountains during the cold months from September through April, favoring the occurrence of a barrier jet that is characterized by strong winds to the left of the flow (Fig. 10). The strongest mountain barrier jet along the Chukotka Mountains occurs in December and January, when the Beaufort high is strong and located over the northern Chukchi Sea, causing strong northeasterly wind that blows directly to the Chukotka Mountains. In addition, storms from the south are relatively active during these months, which also bring strong northeasterly winds blowing to the mountains. The monthly mean jet's maximum is about 6 m s^{-1} in the area of 40 km inland from the shoreline, where the topography begins to rise dramatically. The maximum winds associated with the barrier jet also extend offshore to a fairly large distance. At the offshore location 50 km away from the shoreline, the wind component parallel to the mountain range is greater than 4 m s^{-1} in January and December. When a wind speed of 2.5 m s^{-1} (red contours in Fig. 10) is used to determine the major upstream extent of the barrier winds, a distance of around 100 km from the foot of the mountain can be found in January and December, which is close to the estimate given by the Rossby radius of deformation $L_R = (1/f)\sqrt{gH(\Delta\theta/\theta)}$, where f is the Coriolis parameter, g is the gravity constant, H is the depth of the cold-air layer ($\sim 500 \text{ m}$), θ is the potential temperature at the bottom of the cold air ($\sim 250 \text{ K}$), and $\Delta\theta$ is the difference of potential temperatures between the top and bottom of the cold-air layer ($\sim 10 \text{ K}$).

From January to April, the jet becomes weaker and the monthly mean wind speed decreases from around 6 m s^{-1} to 3.5 m s^{-1} . Its spatial extension is also reduced to within 50 km of both sides of the shoreline. From September to December, the barrier jet recovers its strength, following a reversal of the almost same process that occurs from January to April. During the warm season from June to August, there are no such strong winds along the mountain ranges.

Over the north slope of the Brooks Range, large-scale flow has only a weak component oriented perpendicularly to the mountain range. Climatological wind distribution thus does not show obvious barrier jets (not shown) similar to that along P1, suggesting the absence of the mountain barrier process and its effect on the surface wind field in the area from a long-term climatological perspective. Nonetheless, the mountain barrier process may still occur occasionally at particular times under favorable synoptic conditions, impacting surface winds. By contrast, relatively strong barrier jets occur over the south slope of the eastern Brooks Range from November to March, with wind speeds around $5\text{--}6 \text{ m s}^{-1}$ (not shown). This is slightly weaker than what has been seen along the Chukotka Mountains in Fig. 10, due to relatively weaker winds blowing toward the south slope of Brooks Range than that blowing toward the Chukotka Mountains.

7. Interannual variability of mesoscale winds in CBHAR

The Chukchi–Beaufort Seas region has experienced dramatic environmental changes. Among these changes, the regional surface wind speeds have increased (Stegall and Zhang 2012), the Beaufort high has been strengthened (Moore 2012; Wu et al. 2014), and the Arctic and North Pacific storm tracks and activities have been altered (Zhang et al. 2004; Serreze and Barrett 2008). As a consequence of these changes, the interannual variability of mesoscale winds results from interactions between the time-varying synoptic systems and the underlying surface dynamic and thermodynamic forcings. The interannual variability of mesoscale winds, specifically sea breezes and mountain barrier jets, and its synoptic driving mechanisms will be analyzed here.

Synoptic weather activities demonstrate considerable seasonal cycle in the study area. There are a maximum number of storm activities in the Arctic during summer, while the Beaufort high reaches its strongest intensity during winter. The variability of the Beaufort high in its intensity and location is strongly modulated by the Arctic Oscillation (AO; Thompson and Wallace 1998) and Arctic Rapid Change Pattern (ARP; Zhang et al.

FIG. 10. As in Fig. 8, but along profile P1 and the colors are for wind speed (m s^{-1}) perpendicular to the profile [negative wind speeds (blue) represent flow toward the reader, and positive (red), away from the reader]. Solid contours in red represent a wind speed of 2.5 m s^{-1} , which is considered as the major extent of the barrier winds.

2008). We therefore examined impacts of Arctic storms on sea-breeze variability and impacts of AO and ARP on the mountain barrier jets.

a. Variability of sea breezes and impacts by summer Arctic storms

According to the climatology of sea breeze presented above, sea breeze reaches its minimum and maximum strength at 0600 and 1500 local time (LT) in summer. We therefore use the difference of the wind component perpendicular to the shoreline between these two times to represent the strength of sea-breeze circulation for the variability analysis:

$$I_{\text{SB}} = \bar{V}_{1500} - \bar{V}_{0600}, \quad (1)$$

where $\bar{V}_{0600}$ and $\bar{V}_{1500}$ are the monthly averaged across-shoreline wind components on each day at 0600 and 1500 LT. In addition, to filter any potential noises at single level, we also made a vertical average of the wind over a shallow near-surface layer (0–40 m) in the calculation of I_{SB} .

The variability analysis is conducted along the vertical section P2 for the month of July, when the most active sea breezes occur (Fig. 11a). Along P2, sole sea breezes can be identified with minimized topographic effects. The normalized I_{SB} demonstrates large interannual

FIG. 11. (a) Annual variations in normalized sea-breeze index I_{SB} along profile P2 and cyclone activity index CAI for the entire study domain during July 1979–2009; the dashed lines are for ± 0.75 standard deviations of I_{SB} to distinguish the strong and weak sea breeze years; and strong I_{SB} years include 1979, 1985–86, 1993, 1999, and 2007, while the weak I_{SB} years are 1984, 1987, 1998, 2003, and 2008. Composite anomalies of lower atmospheric temperature ($^{\circ}\text{C}$) associated with (b) strong and (c) weak sea breeze years along profile P2; and (d) composite anomalies of vertically integrated cloud cover along profile P2 in strong and weak sea-breeze years.

variability from 1979 to 2009, ranging from 2 to -3.5 m s^{-1} . The positive (negative) values suggest stronger (weaker) sea breeze than climatology. The particularly strong sea breezes occur in 1979, while there were generally weak sea breezes in the first five years since the beginning of this century, with the weakest sea breezes in 2003. No obvious long-term trend is found during the 31-yr time period.

Statistical analysis is applied to examine the impact of storms on sea breezes. By using the CBAHR data and adopting the storm identification and tracking algorithm developed by Zhang et al. (2004) and modified by Tao et al. (2016), we calculate the cyclone activity index (CAI) for July in each year, which measures the overall storm activity in the study area by integrating information from cyclone count, intensity, and duration. The CAI and I_{SB} are negatively correlated with a correlation coefficient of -0.38 at the 95% significant level (Fig. 11a). This suggests that strong storm activity in the study area suppresses development of sea breezes.

To further explore the physical mechanism behind the correlative relationship between storm activity and sea breeze, we perform a composite analysis of temperature and cloud cover anomalies associated with interannual fluctuations of sea-breeze strength (Figs. 11b–d). A criterion of ± 0.75 standard deviations of sea-breeze index I_{SB} (dashed lines in Fig. 11a) was used to identify strong and weak sea breeze years. The temperature contrast between land and sea/sea ice plays a deterministic role in generating sea-breeze circulation. During the strong sea breeze years, temperature over the land is warmer than normal by more than 1°C , while sea/sea ice surface temperature is only slightly warmer than normal (Fig. 11b). As a result, the temperature gradient between land and sea is greater, strengthening sea breezes. On the other hand, during the weak sea breeze years, temperature over land is colder (Fig. 11c), and as a result the temperature contrast between land and sea/sea ice is smaller, suppressing the development of onshore winds.

The contrasted surface temperature differences between the years with strong and weak sea breezes can be attributed to cloud radiation effect. The composite analysis of cloud cover demonstrates that, during the years with weak sea breeze, there are anomalously large cloud covers because of high storm activity (Fig. 11d). However, cloud covers are anomalously small during the years with strong sea breezes, particularly over the land area. Cloud cover changes incoming solar radiative heating to the surface and, in turn, surface temperature. In July, most ocean surface in the study area is still covered by sea ice. Thus, lower land surface albedo combined with higher sea ice albedo results in a much greater temperature change over land in response to the change in radiation forcing. This physical process provides an insight into the correlative analysis result above, which indicates that strong (weak) storm activity tends to weaken (strengthen) sea-breeze circulation along the Chukchi–Beaufort coast.

b. Variability of mountain barrier jets and impacts by winter Arctic circulation

As discussed in section 6, the mountain barrier effect along the north slope of the Chukotka Mountains is most significant in the Chukchi–Beaufort Seas region. This is because of strong synoptic northeasterly winds, which are generated by the Beaufort high and reinforced by the Aleutian low and blow directly toward the mountains. To detect the variability of mountain barrier jets, we examine the daily mean wind field along the cross section P1 within 100 km of the shoreline by defining a mountain barrier jet (MBJ) index I_{MBJ} as described below:

- 1) Identifying MBJ events: When the upper-level wind field (1.5–3 km above the surface) is dominated by winds blowing toward the mountain, and surface winds (within 500 m of the mountain surface) turn left to be parallel to the mountain range, we consider an MBJ event to occur.
- 2) Quantifying frequency and intensity of MBJ events: The frequency of MBJ events is defined as the fraction of the days with MBJ events in a month, and the intensity is defined by the monthly averaged wind speed of MBJ (component winds along the mountain range within 500 m above the surface).
- 3) Defining MBJ index: The index measuring monthly integrated feature of the MBJ activity is defined as the product of the frequency and intensity of MBJ events: that is,

$$I_{\text{MBJ}} = \text{frequency of MBJ events} \\ \times \text{intensity of mean MBJ events.} \quad (2)$$

Since the MBJ in January is the strongest along the north slope of the Chukotka Mountains, we use January

I_{MBJ} from each year to detect its variability (Fig. 12a). The I_{MBJ} demonstrates an obvious interannual fluctuation, showing strong MBJ activities in the early 1980s and 2000s. But no visible trend of the MBJ activity is present though the Chukchi–Beaufort seas have been greatly warmed during the study time period.

To understand the interannual variability of the MBJ activities, we examine what roles are played by the leading atmospheric circulation modes of the Arctic Oscillation (Thompson and Wallace 1998) and the Arctic Rapid Change Pattern (Zhang et al. 2008). AO measures atmospheric circulation variability during a long time framework, while the ARP stands for the recent spatial transition of the atmospheric circulation from the conventional AO pattern. These modes substantially modulate the intensity and locations of the Beaufort high and Aleutian low that shape the prevailing northeasterlies toward the Chukotka Mountains. Comparisons between I_{MBJ} and the AO and ARP indices show that opposite phase variations exist between I_{MBJ} and AO and ARP during most years of the study period (Fig. 12a). Correlation coefficients between I_{MBJ} and AO and I_{MBJ} and ARP are -0.34 and -0.48 , respectively, in a 95% significant test, explaining about 11% and 28% of the total variance of the MBJ variability.

To further explore how AO and ARP impact the MBJ activity along the north slope of the Chukotka Mountains, we use the composite analyses of sea level pressure (SLP), 850-hPa winds, and temperature and wind vector profiles crossing the mountain range during the years with strong and weak MBJ activities (Figs. 12b–g). The 850-hPa, rather than surface, winds are analyzed here in order to depict favorable background flow for the development of MBJ events. The years with strong and weak MBJ activities are selected by using I_{MBJ} anomalies greater than ± 1 standard deviation (the dashed lines in Fig. 12a). There are a total of five years with strong MBJ, including 1981, 1985, 1986, 2000, and 2001 and six years with weak MBJ, including 1989, 1993, 1996, 2004, 2005, and 2007.

In the strong MBJ years, the ARP index is generally in the negative phase, suggesting positive SLP anomalies over Eurasia and negative SLP anomalies over the North Pacific and Alaska. Its detailed regional expression in the study domain is that the Beaufort high moves westward over the East Siberian Sea, and the Aleutian low strengthens and expands northward (Fig. 12b). As a result, strong pressure gradients between these two systems generate strong northeast winds blowing toward the Chukotka Mountains directly (Figs. 12b,e), which is one of the required conditions for the formation of MBJ. On the other hand, strong northeast winds also advect cold air from the sea ice surface in the north to

FIG. 12. (a) Annual variations in normalized MBI index I_{MBJ} along profile P1 (black), along with climate indices of AO (blue) and ARP (red) in January 1979–2009; the dashed lines are for ± 1 standard deviation of I_{MBJ} to distinguish the strong and weak MBI years. (b)–(d) Composites of SLP and 850-hPa wind field in years with strong MBI, weak MBI under a near-neutral phase of AO and ARP (type I), and weak MBI under a strong positive phase of AO (type II), respectively. (e)–(g) As in (b)–(d), but for temperature and wind field profiles.

the slope of the Chukotka Mountains (not shown), resulting in a strong inversion near the mountain surface (Fig. 12e). This cold stable air works together with toward-mountain winds to generate strong MBJ events.

During the weak MBJ years, there are two types of circulation patterns. Type I is under a near-neutral phase of AO and ARP (1996, 2004, 2005, 2007, and 2008), and type II is under a strong positive AO phase (1989 and 1993) (Fig. 12a). When there are no significant AO and ARP anomalies (type I), the Beaufort high sits over its climatological location, the north Chukchi Sea, blocking the storm from traveling to the north from the North Pacific. The Aleutian low remains in the south. Accordingly, relatively weak east winds dominate the Chukotka Mountains (Fig. 12c), and the resultant wind component perpendicular to the mountain range is very small (Fig. 12f). The lack of a wind component toward barrier reduces the chance of MBJ formation.

In the years of 1989 and 1993, MBJ activities are weak, but AO is in a very strong positive phase (Fig. 12a). The AO positive phase features strong Arctic cyclone activities (Zhang et al. 2004). Intensified Arctic storm activities diminish the Beaufort high, and a cyclonic circulation dominates the study domain (Fig. 12d). As a consequence, no wind component toward barrier is present over the north slope of the Chukotka Mountains (Fig. 12g). MBJ development is therefore inhibited.

8. Summary and discussion

By using CBHAR which is characterized by the high spatial and temporal resolution, we analyzed the details of mesoscale climatology and changes in the surface wind field over the Chukchi–Beaufort Seas region. Mesoscale winds, including sea breezes, up- or downslope winds, and the mountain barrier jets, and their spatial structures and temporal variability are explored in this study.

The climatology of surface winds displays a strong seasonality over the study area, characterized by stronger winds during cold seasons. The MBJ events and drainage flows, footprints left by the mountain dynamic and thermodynamic effects, are the significant components of strong cold season winds. In summer, winds are calm, but it is also notable that the diurnal variations in surface winds occur mainly over mountainous and coastal areas, revealing that sea breezes and up- or downslope winds are the major contributors to the diurnal variation of surface winds in the area.

Spatial structure analysis of sea breeze reveals that its onshore winds are the strongest right at the shoreline, lasting most of the day and reaching its maximum in the afternoon, around 1500 LT. Terrains near the shoreline

play important roles in affecting sea-breeze circulation by enhancing upward motion and moving the entire circulation toward the slope. Over a relatively flat onshore surface, sea-breeze circulation can extend ~100 km onshore and 50 km offshore. When interacting with the coastal mountain slope, the maximum extent of onshore winds can be reduced by 20–30 km. When the mountain slope is very steep, the offshore extent of onshore winds can also be reduced by about 20–30 km. There also exists strong seasonal dependence of the sea breeze. In the study area, sea breezes emerge in June, with a spatial extent out to 30–40 km offshore, affecting the most extensive areas in July, with an extent out to 30–50 km offshore. Affected offshore distance decreases to about 30 km in August, and further to about 10 km in September.

The thermodynamic impacts of mountains on the surface winds vary from month to month. During cold seasons, drainage flow begins to build at the mountaintop in September, gradually increasing in intensity and extent downward to the slope as surface temperatures continue to decrease. During the area's coldest months of the year, from November to February, the strongest drainage flow can occupy the entire slope; however, flow is trapped within a thin layer less than 300 m from the slope surface because of the extremely strong inversion. During summer, the diurnal variation of surface winds is dominated by the development of mountain–plains solenoidal circulation. Upslope winds near the surface start to build around 0900 LT, reaching their maximum strength around 1500 LT and continuing until 2100 LT. The returning flow is located about 1 km above the mountaintop.

Note that the thermodynamically forced climatological sea breezes or up- or downslope winds analyzed here are just the mesoscale component of total winds. Although its magnitude is generally not greater than 2 m s^{-1} in climatological mean, the mesoscale wind component may intensify or weaken prevailing winds driven by the synoptic-scale weather system or large-scale atmospheric circulation over a particular local area for a particular season or daily time. For example, the peak sea breeze has almost the same direction as the large-scale wind in the Beaufort Sea coastal area, as shown in Fig. 4. The sea breezes therefore can double the actual surface winds in the afternoon and evening during its diurnal cycle. The increased wind speed may enhance the onshore transport of heat (though negative in this case), impacting energy budgets in the coastal land area. Meanwhile, onshore water vapor transport may also be increased, in particular considering the largely retreated sea ice cover during recent decades. When the moist air flows up along the coastal slope,

boundary layer cloud may form. The changes in land surface energy and water budgets because of these processes may have significant climate implications for understanding observed changes in the north-slope land ecosystem (e.g., Stow et al. 2004). In addition, the increased onshore winds may also influence coastal ocean currents and upwelling, as well as float sea ice distribution that is important for sea ice edge positioning (e.g., Pickart et al. 2009). Moreover, when looking at the sea-breeze interannual variability shown in Fig. 11, we can find sea-breeze magnitude fluctuation can reach two or three standard deviations of climatological mean. This greatly fluctuating mesoscale wind component may largely modulate the local surface winds, leading to important climate consequences.

The dynamic impacts of mountains on the area's surface winds are the formation of mountain barrier jets (MBJs), which occur over the north slope of the Chukotka Mountains and the south slope of the eastern Brooks Range from October to March. The MBJ along the Chukotka Mountains is strongest in December and January, with a monthly mean speed around 6 m s^{-1} . The MBJ over the south slope of the eastern Brooks Range is relatively strong, from November to March, with wind speed around $5\text{--}6\text{ m s}^{-1}$.

Sea breezes and MBJs are also subject to variations and changes in response to adjusted large-scale atmosphere circulation. Our analyses show that the sea breeze along the western Beaufort Sea coast is correlated negatively with storm activity in July; that is, strong storm activity can inhibit the development of sea breezes as a result of enhanced clouds associated with storm activity decreasing the thermal contrast between land and sea/sea ice. In winter, when stable air is present in the study area, the MBJ activities along the Chukotka Mountains in general have an opposite phase variation compared to AO and ARP indices. When ARP is in a negative phase, the Beaufort high shifts westward, and the Aleutian low moves northward to the Chukchi Sea, with strong pressure gradients between these two systems generating strong northeast winds toward the mountains and accordingly strong MBJ developing along the slope. But when AO is in a strong positive phase, intensified Arctic storm activities diminish the Beaufort high, and a cyclonic circulation dominates the study area, under which no toward-barrier winds blow toward the Chukotka Mountains, with no MBJ developing, accordingly. When both AO and ARP are in a near-natural phase, the Beaufort high situates over the north of the Chukchi Sea, and the Aleutian low remains in the south. Relatively weak east winds are present over the Chukotka Mountains and reduced toward-barrier winds weaken the strength of the MBJ. Therefore,

different responses from the Beaufort high and Aleutian low to anomalies in large-scale circulations play a vital role in the variations of MBJ activities over the Chukotka Mountains.

Acknowledgments. This work was sponsored by the Bureau of Ocean Energy Management (BOEM) of the U.S. DOI under Contract M06PC00018. Computing resources were provided by the Arctic Region Supercomputing Center (ARSC) at the University of Alaska Fairbanks. CBHAR data are archived at ARSC. We wish to thank two anonymous reviewers for the insightful and constructive comments, which significantly improved the manuscript.

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