INTERNATIONAL JOURNAL OF CLIMATOLOGY
Int. J. Climatol. **26**: 159–178 (2006)
Published online 24 November 2005 in Wiley InterScience (www.interscience.wiley.com). DOI: 10.1002/joc.1254

THE SURFACE WINDS OF SWEDEN DURING 1999–2000

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Received 12 November 2004
Revised 12 July 2005
Accepted 13 July 2005

ABSTRACT

This study aims at increasing our understanding of the regional wind climate in Sweden. Spatial and temporal patterns of the surface winds are presented for the years 1999–2000. Annual mean wind speeds range between 2 and 5 m/s with high values at exposed mountainous sites and on islands off the coast. Combining wind speed and direction into mean wind velocities shows that flow conditions are stronger and more coherent in space in southern Sweden than in central and northern Sweden. The spatial scale, defined as the distance between stations when the correlation for wind speed drops to ∼0.37, was determined by pairwise correlations between all possible station pairs. Scales range from 38 to 530 km for wind speed and from 40 to 830 km for wind direction depending on the region. They tend to be smaller in central and northern Sweden, where the more pronounced relief has a larger influence on the local wind conditions. The strength and the timing of the annual and diurnal wind speed cycle have been estimated for each station. Amplitudes of the annual cycle are greater at exposed sites and correlate generally well with annual mean wind speeds. Monthly mean wind speeds peak in winter in southern Sweden, but peak in other seasons in the remaining regions. In winter, weaker pressure gradients over northern Sweden and surface-near temperature inversions contribute to weaker surface winds. Diurnal cycles vary in strength between summer and winter. Compared to the last normal climate period (1961–1990), 1999–2000 is characterized by the increased occurrence of westerly and southerly geostrophic flow. Copyright © 2005 Royal Meteorological Society.

KEY WORDS: surface winds; spatial patterns; temporal patterns; spatial scale; Sweden

1. INTRODUCTION

Studies of the regional climate in Sweden have up to now mainly focused on the temperature and precipitation climate at various scales in time and space (e.g. Chen and Hellström, 1999; Bergström and Moberg, 2000; Busuioc *et al*., 2001a,b; Alexandersson, 2002; Moberg *et al*., 2002; Chen and Johansson, 2003). In the light of the ongoing global climate change, temperature and precipitation are considered as key variables because of their crucial impact on biological processes, ecosystems and human activity. Today's access to long and quality-controlled, digitized series of instrumental temperature and precipitation observations obtained from an increasing number of places around the world has facilitated the investigation on the regional as well as the global scale. The situation, however, is different regarding the investigation of surface winds.

Wind is an important climate variable with respect to many practical and ecological applications (wind energy generation, air quality purposes). Better understanding of today's surface wind conditions is also important from a climate change point-of-view. During the last years, parts of the country have experienced some severe storms that have caused damage to infrastructure and forested areas by wind throw. Such events raise the question whether climate change due to increasing atmospheric greenhouse-gas concentrations may

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also alter the surface winds. Climate change experiments made with global climate models (GCMs) predict changes in circulation patterns (Räisänen, 2000), a shift in cyclone path toward the northeast (Knippertz *et al.*, 2000) and enhanced westerlies between 50° and 70 °N (Andersen *et al.*, 2001). These changes are expected to have an impact on surface winds. According to a regional climate change scenario for Scandinavia, mean and extreme wind speeds would increase by *ca* 5–10% during the next 100 years (Räisänen *et al.*, 2003). However, since the model results are generally not well verified because of the lack of readily available reliable wind observations, the reliability of these results is largely unknown.

In Sweden, digitally stored wind observations are available from SMHI (Swedish Meteorological and Hydrological Institute) for the period from around 1960 and onwards. Before SMHI gradually built up its automated station network during the 1990s, wind observations were mainly collected from instruments located on lighthouses or from meteorological stations at airports. Observations were therefore concentrated along the coastline and in southern Sweden, while the observation density was very sparse in northern Sweden. In addition, uncertainties concerning sensor height above ground make it difficult to derive homogenized and quality-controlled data series, which is a prerequisite for systematic climatological investigations, especially for climate change studies. For the purpose of the present study, which is the identification of main spatial and temporal patterns of the observed surface wind conditions in Sweden, it was therefore decided to use only quality-controlled data from the modern, automated station network. This choice allows for best possible geographical coverage, but the short observation period does not allow studies on the long-term climate change.

The existing observation-based studies on surface winds in Sweden either focus on regional and local winds or on practical applications. Jönsson and Fortuniak (1995) studied the wind climate in Lund, southern Sweden and Jönsson and Holmquist (1995) investigated wind direction variability since the eighteenth century in this region. In addition, the risk for wind erosion on agricultural fields in Scania was estimated by Achberger *et al.* (2002) and Ekström (2002). The frequency of gales since 1860 along the Swedish west coast was investigated by Franzén (1991) and was found to be related to forest damage, while Gustavsson *et al.* (1995) and Borne *et al.* (1998) studied the land and sea breeze system along the west coast. Investigations regarding the wind conditions of entire Sweden, however, are restricted to compilations of basic descriptive statistics for wind speed and direction at some selected stations by Raab and Vedin (1995), Taesler (1972) and Ångström (1974).

This work aims at increasing our understanding of regional wind conditions in Sweden by characterizing spatial and temporal patterns of the recent surface winds. To ascertain a constant and high number of wind stations throughout the whole study period, the analysis was restricted to the years 1999–2000. These stations are quite evenly distributed over the entire country, and the station density is considerably higher compared to the earlier network. Specifically, the objectives of the study are (1) to identify and explain spatial patterns of wind speed and wind direction as well as their joint spatial variability when combined into wind velocity, (2) to identify the typical spatial scale for wind, and (3) to establish the strength and timing of the seasonal and diurnal cycles in wind speed across the country. Since only 2 years of data are used, the climatological representativeness of the period is evaluated as well. The paper is organized as follows: Section 2 addresses relevant factors in influencing the surface winds in Sweden, Section 3 describes the methods, Section 4 presents and discusses the results, and Section 5 summarizes the findings with concluding remarks.

2. CONTROLS OF SURFACE WINDS IN SWEDEN

Two semipermanent pressure systems, the Icelandic low in the north and the subtropical Atlantic high located close to the Azores characterize the large-scale atmospheric pressure distribution over Europe. This pattern causes typically westerly and southwesterly flow conditions over Scandinavia (e.g. Liljequist, 1970; Chen, 2000; Hansen-Bauer and Førland 2000). The position of both pressure systems varies with season, as well as with their strength. Pressure gradients are largest during winter, causing well-developed zonal flow conditions whose fluctuation in time is known as the North Atlantic Oscillation (NAO) (Hurrell and van Loon, 1997). The NAO is linked to the large-scale atmospheric dynamics and is considered as the main mode for winter temperature and precipitation variability over extended regions of the Northern Hemisphere (Hurrell *et al.*, 2003). It highly influences Swedish winter temperatures (Chen and Hellström, 1999) and precipitation in northern Europe (Hansen-Bauer and Førland, 2000; Busuioc *et al*., 2001a; Uvo, 2003). In accordance with the strong meridional pressure gradient in the cold season, wind speeds undergo typical variations over the course of a year, i.e. they are stronger in winter than in summer (Troen and Petersen, 1989; Yin, 2000). At the synoptic scale, cyclones and anticyclones cross Europe along various tracks and control speed and direction of the surface winds in Sweden. Depending on their strength as well as their position and size in relation to Sweden's geographical location, these transient pressure systems cause the surface winds to blow from varying directions and strengths. In the following parts of the paper, when discussing large-scale forcings, the pressure gradients at the top of the planetary boundary layer are implied. These can be represented by circulation indices based on sea-level pressure (SLP) as outlined in Section 3.3.

Turning to Sweden's physiographical conditions, the Scandes are the dominating topographical feature in central and northern Sweden. This north-to-south-directed mountain range, whose main part is located in Norway, reaches up to about 2000 m above sea level at the highest elevations. Numerous parallel river valleys oriented in northwest–southeast direction dissect the region east of the Scandes and partly forces surface winds to follow valley orientation (Figure 1). The mountains create a barrier for the maritime air masses from the west, causing more continental climate conditions east of the Scandes. During winter, strong and persistent temperature inversions develop in the valleys supported by the short daylight length (Liljequist, 1970; Ångström, 1974; Lindkvist and Chen, 1999; Chen and Johansson, 2003), but they may also occur in other seasons. Persistent stable conditions in winter strongly reduce ventilation and have implications for air quality (Sjöberg *et al*., 2003). Compared to central and northern Sweden, the relief of southern Sweden is mostly flat despite some elevated regions of minor extent. The region is more directly exposed to winds from the west because of the absence of protecting mountains in the west.

Temperatures in Sweden decrease from south to north and the length of the vegetation period varies in accordance. It is approximately 7 months in the south but gradually decreases to 3 months in the north. Throughout the country, the forest is the dominating surface cover except in southernmost Sweden, along the west coast, in some regions in central Sweden and in regions above the timber lines which lie at 700–1100 m

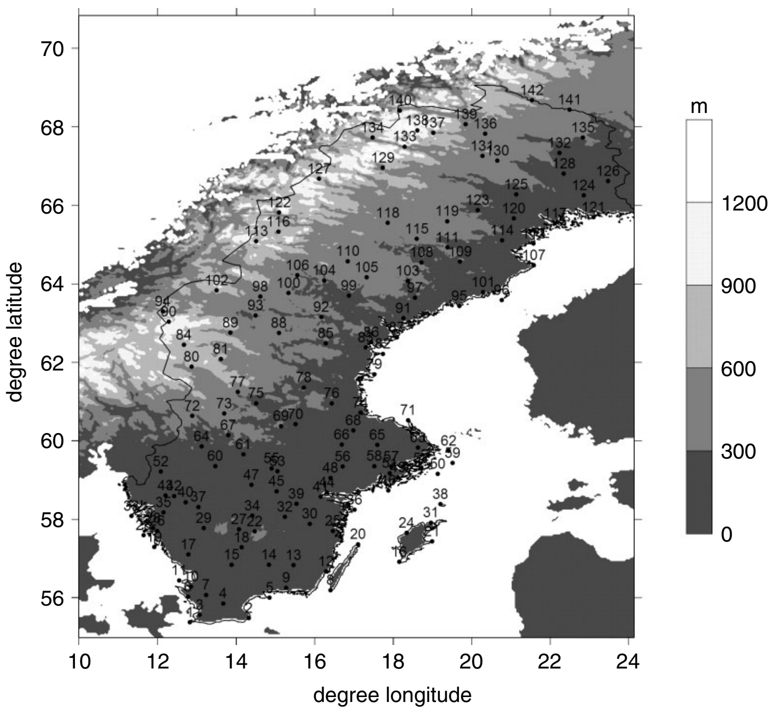

Figure 1. Location of the 142 stations observing wind in 1999–2000. The gray scale indicates the altitude

(Kullman, 2001). The nonforested areas in the southern part of Sweden are characterized by open landscape that is used for agricultural purposes, such as cultivation of grain and animal husbandry, while wetlands dominate the nonforested areas in the northern half. Large parts of Sweden are snow covered during extended periods. The highest mountainous areas close to the Norwegian border have a snow cover for more than 200 days a year, but this number gradually decreases toward the south and the east. Least days with snow (25 days) are experienced in the coastal areas in southernmost Sweden (Raab and Vedin, 1995). The transition from smoother surfaces in winter (snow cover, less dense vegetation) toward increased roughness in summer can to some extent modulate seasonal wind speed variations forced by seasonal pressure gradient variations.

Wind speeds typically vary over the course of a day. Daytime heating causes variations in the turbulent state of the atmospheric boundary layer with stronger turbulence and increases downward mixing of momentum during day (Stull, 1988; Arya, 1988). Consequently, winds speeds are typically higher during day. In addition, local forcings such as differential heating associated with topography, and adjacent land and water surfaces (Hsu, 1988; Dai and Deser, 1999) generate daily wind variations. With regard to Sweden, the country's long coastline (>50% of Sweden's border consists of coastline, *ca* 2600 km; Sporrong and Wennerström, 1990), the large number of lakes and the topography of the Scandes favor the development of local wind systems. Local winds are, however, often obscured by stronger regional winds (Borne, 1998).

3. DATA AND STATISTICAL ANALYSIS

3.1. Wind observations

Wind speed and direction observations for the years 1999–2000 are obtained from 142 synoptic stations maintained by SMHI (Figure 1). Both variables are measured at 10 m above ground and are available with 3-hourly resolution. The stored values are an average over samples measured during the 10-min period preceding the observation hour. Wind speeds are measured in meters per second and rounded toward the nearest integer; wind direction is given in degrees and stored in 10°-steps between 10° and 360°. Calm conditions are coded with 0. Stations contain at most 15% of missing values but generally the amount is much lower: 70% of the stations have <2% missing data, while 10% of the stations have >6% missing data.

3.2. Statistical analysis of wind

Wind observations commonly consist of simultaneous speed and direction readings. These can be analyzed either separately or together as a vector-valued variable. In the first case, common scalar statistics is applied to wind speed, whereas the circular nature of the directional data requires circular statistics (Batschelet, 1981; Essenwanger, 1986). As an alternative, speed and direction are decomposed into the Cartesian components u (west–east) and v (north–south) and vector-based statistics are applied. It allows the analysis of speed and direction together, which may be regarded as a more 'holistic' approach. Scalar, directional and vector statistics provide different, but complementary, information (Klink, 1998). In the following text, different kinds of statistics are applied. 'Speed' and 'velocity' are often used as synonyms, but here 'speed' refers always to the scalar wind speed, while 'velocity' refers to the magnitude of the wind vector.

3.2.1. Vector statistics and wind steadiness. The 3-hourly wind speed and direction observations were decomposed into the Cartesian u- and v-components and averaged over the 2 years. Since the vector representation reflects the combined effect of speed and direction, the interpretation of the results may not be straightforward (Essenwanger, 1986). The reason is that wind components blowing approximately equally strong and equally often from opposing direction sectors are canceled out because of averaging in time. This results in weak wind velocities, whereas winds blowing from similar directions but with varying speed do not cancel out. The temporal means of velocity and wind speed are in this case of similar magnitudes. Small velocities may either imply weak winds from similar directions or winds of varying speed but from opposing directions. To facilitate the interpretation of winds in vector representation, the steadiness should be analyzed in parallel. It is the ratio between the mean wind velocity and the mean wind speed for a given averaging period (here 2 years) and it ranges between 0 and 1 (Klink, 1998). The maximum value of 1 indicates that winds always blow from the same directions, while the minimum value of 0 implies that winds blow from opposing directions. Intermediate values result from variations in both wind direction and speed. The steadiness was calculated for all stations.

3.2.2. Directional statistics. The circular nature of the wind direction data requires the application of circular statistics (Batschelet, 1981; Essenwanger, 1986). Here, circular statistics was used to correlate wind direction observations pairwise from all the stations in order to estimate the spatial coherence of the wind directions. In addition, the wind direction standard deviation (*wdir_std*) at all the stations was calculated. Both statistics were obtained following the approach by Batschelet (1981). The calculation of the circular correlation coefficient is based on δ_i, which can either be the difference between the two circular variables ϕ_i and ψ_i or their sum (i.e. $\partial_i = \phi_i - \psi_i$ or $\partial_i = \phi_i + \psi_i$. If two circular series are correlated, it can be expected that the magnitude of the angles δ_i are similar, i.e. they group around a certain mean angle. The degree of clustering around the mean of δ_i reveals, therefore, the strength of the correlation and can be expressed as r, the mean vector length of δ_i:

$$r = \frac{1}{n}\left[\left(\sum \cos\delta_i\right)^2 + \left(\sum \sin\delta_i\right)^2\right]^{1/2} \tag{1}$$

where n is the number of samples. This measure ranges between 0 (uncorrelated) and 1 (perfectly correlated) and reaches a maximum if δ_i remains constant. Whether $\partial_i = \phi_i - \psi_i$ or $\partial_i = \phi_i + \psi_i$ is used depends on whether ϕ decreases or increases with increasing ψ. The first (second) case indicates a negative (positive) relation between the two series. As this might not be known in advance, r should be calculated for both cases. Two different coefficients are obtained: r_- for $\partial_i = \phi_i + \psi_i$ and r_+ for $\partial_i = \phi_i - \psi_i$. This leads to the odd fact that positive and negative correlations can exist at the same time (Batschelet, 1981). Batschelet has therefore suggested following Mardia's idea, namely, to choose the larger correlation coefficient of the two as the final one: $r = \max(r_+, r_-)$.

Directional standard deviation is based on the mean vector length r_s for a series of directional data. Since r_s decreased from 1 to 0 with increasing scatter around the directional mean, it is a measure of the dispersion around the angular mean. The *wdir_std* was calculated as:

$$wdir\_std = \frac{180^\circ}{\pi}[2(1 - r_s)]^{1/2} \tag{2}$$

where r_s is calculated using Equation (1) but δ_i is replaced by ϕ_i

3.2.3. Spatial coherence of wind. The spatial coherence estimates the degree of similarity between observations obtained from different locations. This measure can be quantified in several ways, but a common approach is to study how quickly a variable changes with distance from a certain location (Gunst 1995; Robeson and Shein, 1997). Here, the distance–decay relationship is determined by means of spatial autocorrelation (correlograms). Pairwise correlations are calculated between variables at one reference station and all neighboring stations. The procedure can be repeated until correlations for all possible station pairs are calculated. Plotting correlation values *versus* separation distance of the station pairs illustrates how quickly the connection between stations decreases with increasing distance. Distance–decay relationships have previously been established for temperature (e.g. Briffa and Jones, 1993), precipitation (Osborn and Hulme, 1997) and wind speed (Robeson and Shein, 1997).

Here, spatial decay is estimated separately for wind speed and direction (with 142 stations, there are 10.011 station pairs for each variable). Pairwise correlations for speed and direction were calculated using the 3-hourly data and were plotted against the separation distance between the stations. As the decrease in correlation with increasing distance often follows an exponential curve, an exponential function was fitted to the data following the approach by Briffa and Jones (1993) and Osborn and Hulme (1997):

$$r = e^{-x/x_0} \tag{3}$$

where x_0 is the correlation decay length at the distance when the correlation falls to 1/e (∼0.37). This measure was used here to define the spatial scale of wind speed and wind direction.

3.2.4. Estimation of annual and diurnal wind speed cycles. To estimate the strength of the annual cycle, a time series of monthly mean wind speeds were calculated from January 1999 to December 2000 for each station. The amplitude and the phase of the annual cycle were estimated by means of Fourier analysis. Compared to a Fourier analysis of the original time resolution, averaging into monthly means has the advantage that gaps in the series due to missing values are removed and the signal of the annual cycle is 'cleaned' from variability at higher frequencies. A disadvantage is, however, that amplitude and phase of the annual cycle is estimated from rather few data points, making the spectral estimate more uncertain.

The daily cycle has been calculated separately for each month by averaging over all observations belonging to a certain observation hour, i.e. 0, 3, ..., 21 h GMT. Then, their amplitude and phase are estimated by Fourier analysis to determine the strength and phasing of the diurnal variation. To avoid errors related to missing observations, diurnal cycles were only calculated from those days with eight valid observations. Some stations had rather many 'incomplete' days with <8 observations and were therefore excluded from the analysis.

3.3. Quantifying the large-scale forcing

The climatological representativeness of the winds in 1999–2000 was evaluated by means of comparing the large-scale forcing during these 2 years with the conditions for 1961–1990. It is assumed that differences in the surface winds between the two periods are mainly caused by variations in the air pressure distribution, i.e. if the surface wind conditions in 1999–2000 are different from the long-term mean, there should also be differences in the large-scale forcings. The reason for applying such an indirect comparison is that the surface wind measurements obtained from the manual station network are associated with a number of uncertainties, which would make a direct comparison rather questionable.

Here, the large-scale forcings in 1999–2000 and 1961–1990 were quantified by means of circulation indices that can be calculated from a gridded air pressure data set (e.g. Briffa, 1995). These indices quantify the strength and the direction of the geostrophic flow and the vorticity over a large area. Chen (2000) and Linderson (2001) describe the calculation of indices for southern and central Sweden, but they can also be calculated for other areas. These indices are frequently used to relate surface climate variability or other environmental parameters to the large-scale atmospheric circulation (e.g. Hellström *et al.*, 2001; Bleckner and Chen, 2003). Here, the index calculation is based on a 6-hourly mean SLP taken from the NCEP-NCAR reanalysis project (Kistler *et al.*, 2001). The annual cycles and the frequency distribution of the geostrophic flow (w_g) and its westerly (u_g) and southerly (v_g) components in 1999–2000 and 1961–1990 are calculated for two domains, one centered at 60 °N, 15 °E representing southern Sweden (domain 1) and another one centered further north at 65 °N, 17.5 °E (domain 2).

4. RESULTS

4.1. Spatial distribution of mean wind speeds and variability

Annual mean wind speeds (Figure 2(a)), standard deviation (Figure 2(b)), the 99th percentile speeds and the percentage of calm events (the latter two are not shown) were calculated for each station. Annual speeds range from 1.2 m/s for the most sheltered location (station 72 in Figure 1) to 7.7 m/s for the two most exposed sites (station 59 and 62 in Figure 1) located on small islands in the Baltic Sea. At inland stations, annual means lie between 2 and 5 m/s. Southern Sweden (south of a line located at ∼58°–60 °N) experiences slightly stronger winds compared to the northern part. Annual speeds decrease rapidly when moving inland from the coast. The spatial pattern of the wind speed variability (Figure 2(b)), extreme winds (not shown)

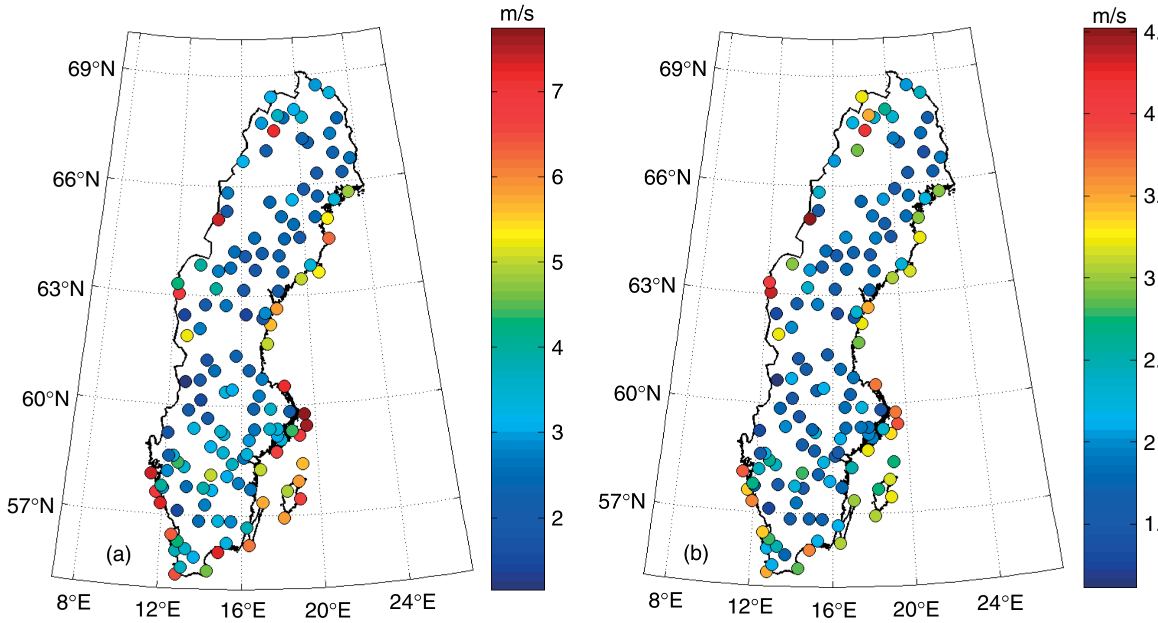

Figure 2. Wind speed distribution across Sweden: (a) annual mean wind speeds (m/s) at all stations; (b) standard deviation of annual mean wind speeds (m/s). This figure is available in colour online at www.interscience.wiley.com/ijoc

and the frequency of calm events (not shown) are basically the same. The means and the variability of wind are thus closely linked.

The decreasing trend in wind speeds from the south to the north cannot only be explained by weaker large-scale forcing over northern Scandinavia. The reason could, however, be the Scandinavian mountains providing large-scale shelter to areas east of the mountains north of ∼58–60° and increasing large-scale roughness because of their more rugged topography. Since winds are predominantly from westerly directions, the effect of the Scandes is noticeable. Furthermore, temperature inversions frequently developing in valleys of central and northern Sweden increase the number of calm conditions thereby lowering annual wind speed means. On a global scale, Yin (2000) found that annual means of surface wind speeds are generally <6 m/s. He also presents higher annual means in southern Sweden compared to those for regions further north. Pryor and Barthelmie (2003) studied winds over the Baltic region using 850-hPa winds from an NCEP-NCAR reanalysis data set. They found gradually declining wind speeds over Scandinavia from the southwest to northeast.

An important spatial feature is the rapid change in wind speed statistics from the coast toward inland, caused by the steplike change in surface roughness. As an example, along the west coast, the difference in annual mean wind speeds between station 26 (Gothenburg) and station 23 (located on a small island) is almost 3 m/s over a distance of about 30 km, while along the east coast the difference amounts to 5.4 m/s over a distance of around 40 km (for station 63 and 62). Similar differences for annual means over sea and land areas (4–4.5 m/s over a distance of 30–40 km) were found by modeling boundary-layer winds over a coastal area in southeastern Sweden (Bergström, 1996).

4.2. *Surface flow pattern*

Resultant wind vectors were plotted on a map (Figure 3) illustrating the joint pattern of flow strength and direction averaged for 1999–2000. The strength of the flow (as indicated by the length of the arrows) shows a similar geographical trend as the distribution of annual speeds (Figure 2(a)) with stronger winds in southern Sweden south of ∼58°–60 °N. This difference is, however, more pronounced in the vector plot. South of this line, winds are predominantly from the southwest and there is little variation in space. Here, topography is of minor importance and there is no mountain range to the west providing shelter to the region. It allows the surface flow field to reflect the trajectories of large-scale airflows. Also, coastal stations and some mountainous locations appear with increased flow strength. North of the line at ∼58°–60 °N, wind directions are from the west and southwest, but vector strength and direction are less coherent. At some scattered stations in the northernmost regions (north of 66 °N), winds are from the northwest.

Are weaker velocities north of ∼58°–60 °N simply a consequence of lower wind speeds or are they caused by stronger variability in wind direction? Since the differences in wind speeds at inland stations are small between southern and northern Sweden (Figure 2(a)), varying wind direction is assumed to play a decisive role. The geographical pattern of the steadiness over the 2-year period sheds light into this finding (Figure 4). Steadiness ranges between 0.05 and 0.5 across the country and the above-mentioned boundary is again apparent. Northern Sweden is characterized by lower steadiness, while the southern region and some selected locations in the mountain range experience comparatively high steadiness. The pattern is in good agreement with the flow field of Figure 3, revealing the relationship between steadiness and the consistency of wind direction.

To further confirm this relation, the *wdr_std* was calculated over the 2 years for each station. The results (though not shown here) clearly show that *wdr_std* in the inland of middle and northern Sweden is relatively high. Steadiness and *wdr_std* are also highly negatively correlated (the correlation coefficient is −0.82), i.e. higher wind direction variability decreases wind velocities. One might now ask how wind direction distributions differ between stations characterized by both high steadiness and low *wdr_std* or vice versa and why such differences exist. In the following text, wind directions at some selected stations are, therefore, presented.

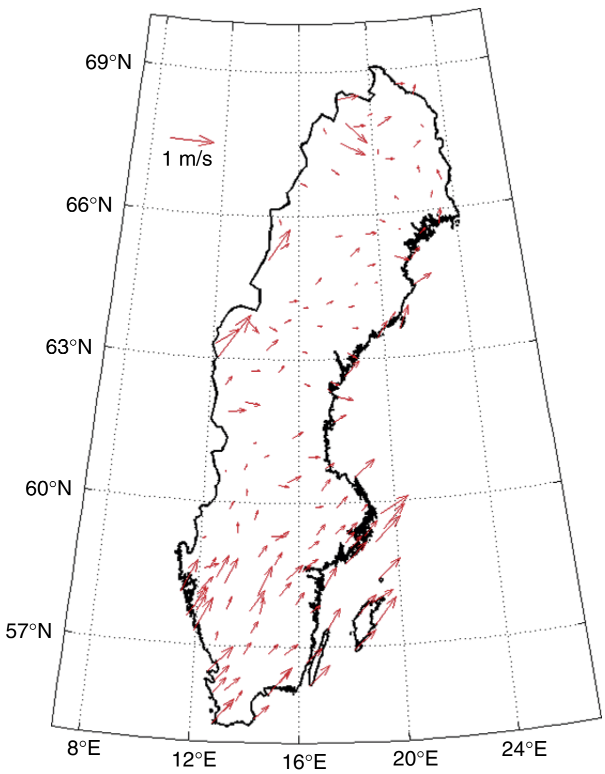

Figure 3. 1999–2000 wind velocity field. The length of the arrows indicates the flow strength. The arrows indicate the direction of the wind. This figure is available in colour online at www.interscience.wiley.com/ijoc

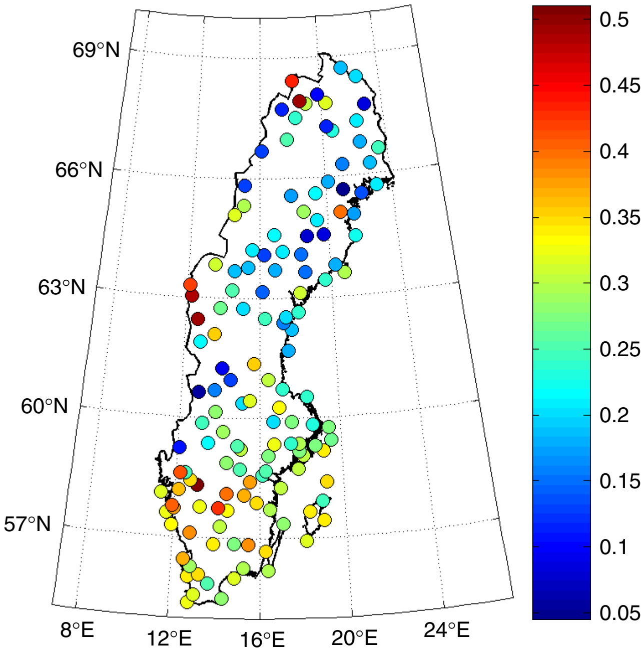

Figure 4. Wind steadiness (dimensionless) for the averaging period 1999–2000. This figure is available in colour online at www.interscience.wiley.com/ijoc

4.3. Spatial distribution of wind direction

Common wind direction statistics are presented as histograms or wind roses showing how frequently winds blow from various direction sectors. Here, wind direction observations for selected stations are binned into 12 sectors of 30° each, ranging between 0° and 330° and are presented as wind roses in Figure 5. The stations were chosen in order to represent different geographical regions as well as different ranges of steadiness, i.e. <0.15, 0.2–0.3 and >0.35. At stations characterized by low wind steadiness (stations 52, 104, 117, 135), winds blow rather equally often from opposite direction sectors. In contrast, for stations at which the steadiness is higher (station 34, 38, 84, 140) winds tend to have one or more dominating wind directions sectors, while stations in the intermediate range are characterized by winds blowing from a wider range of different sectors.

In southern Sweden, winds are often from westerly and southwesterly sectors, while winds from northern directions are rather rare. North of ~60 °N, northwesterly and/or northerly winds become more dominant, and at the coastal stations winds from all wind direction sectors may occur. The frequency of northerly geostrophic winds is slightly smaller over domain 2 compared to those over domain 1 (not shown here), suggesting increased influence of the terrain on surface wind direction north of ~60 °N. Depending on station location, winds are to a varying extent influenced by local physiographical conditions. Station 104, for example, is located close to a NW-SE-orientated lake explaining the dominance of winds from NW and SE directions. The strong dominance of winds from W and N at station 140 are due to the station's location in a WE-directed valley and a lake north of the station. At the coastal stations, winds tend to follow the coastline as was also described by Ångström (1974).

4.4. Spatial coherence of wind speed and direction

Correlograms for wind speed and wind direction derived from 3-hourly measurements are given in Figure 6, showing 'global' distance-decay relationships, i.e. all possible station pairs are included in the analysis. For both variables, correlations decrease with increasing distance as expected. Also included in the figures are

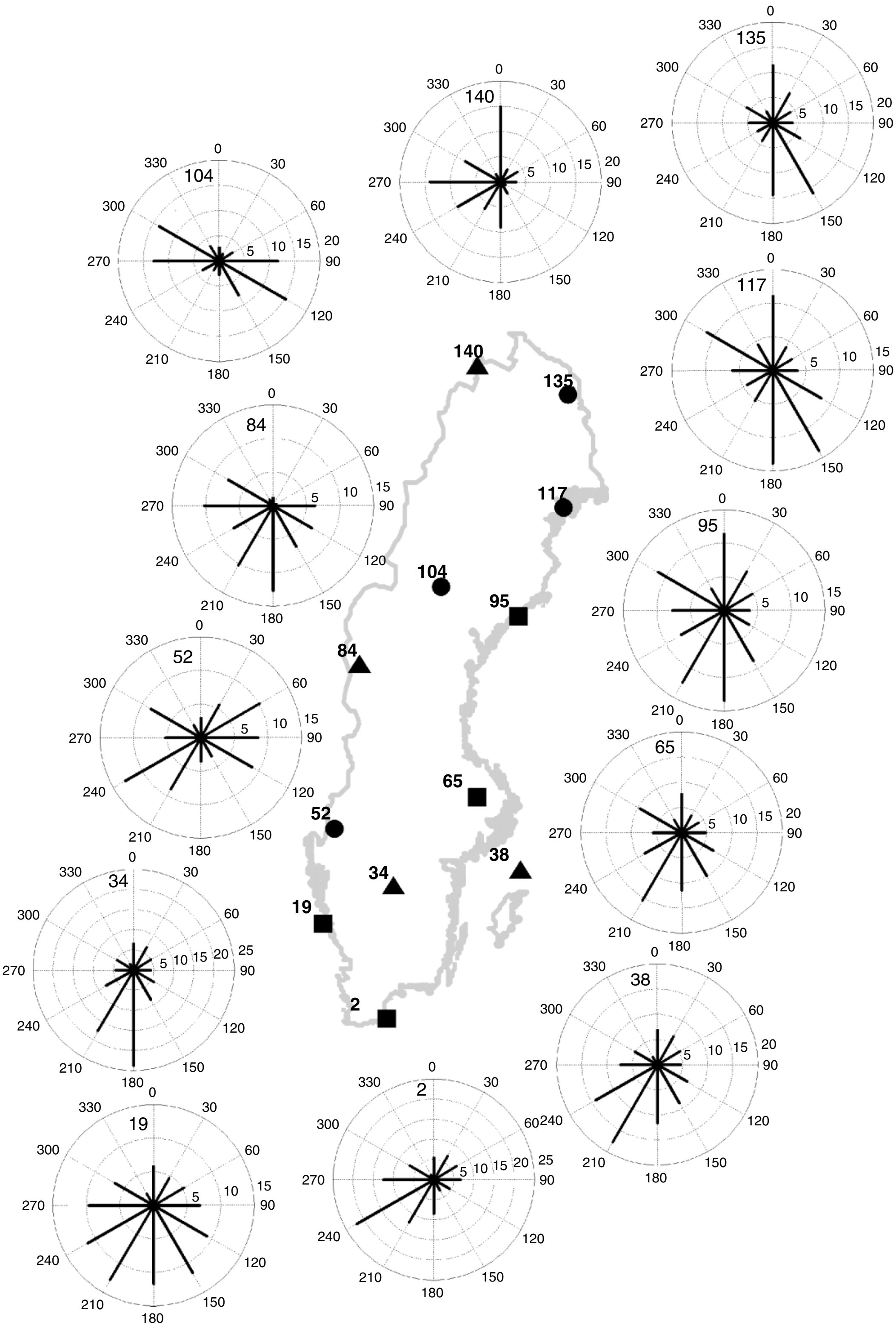

Figure 5. Wind directions at selected stations in 1999–2000 presented as wind roses. The symbols indicate the range of wind steadiness, circle: <0.15, square: 0.2–0.3, triangle: >0.35

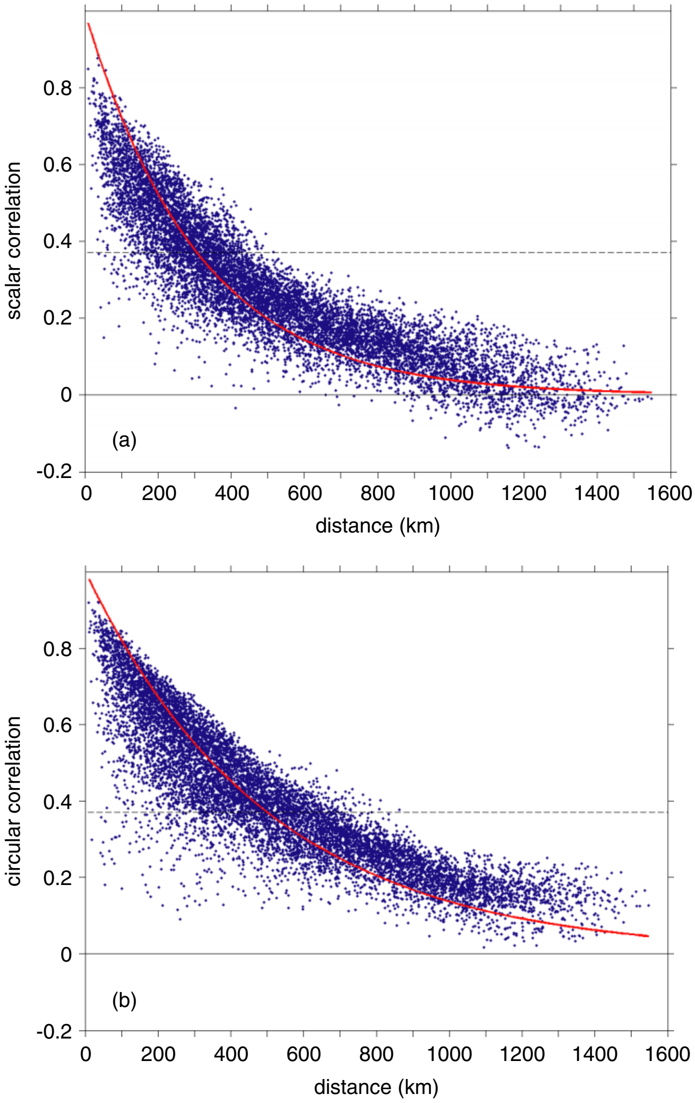

Figure 6. 'Global' distance–decay relationship for: (a) wind speed and (b) wind direction. An exponential curve was fitted to the scatter points. This figure is available in colour online at www.interscience.wiley.com/ijoc

exponential curves fitted to the scatter points. Using x_0 as a measure for the spatial scale reveals a larger spatial scale for wind direction than for wind speed. This implies that wind directions to a higher degree covary in time with neighboring stations over larger distances compared to wind speed.

Both wind speed and direction correlograms show considerable scatter in the distance–decay relationship suggesting geographical differences in the spatial scale of wind. Fitting the exponential function separately for each station, local x_0 were obtained (not shown here). They vary between 40 and 830 km for wind direction and 38 to 530 km for wind speed. In general, x_0 is longer in southern Sweden south of $\sim 58°–60°$N, but it varies considerably even over short distances. Local conditions around the station site determine to which degree winds are representative for a larger area. Moving from the south to the north of Sweden, the relief gets more pronounced north of $\sim 58°–60°$N (Figure 1). This change in landscape characteristics coincides well with the general decrease in x_0 in both speed and direction.

An issue that was raised by Gunst (1995) and Robeson and Shein (1997) is concerned with the bias that can be introduced in correlograms because of nonstationarities in the time series. The presence of seasonal and diurnal cycles as well as trends causes nonstationarities that can lead to systematic overestimations of the distance–decay relationship. In particular, the risk is high if a large part of the total variability is related to these periodicities and when cycles vary in phase. To avoid this problem, seasonal and diurnal wind speed cycles were removed prior to the calculation of the wind speed correlogram shown in Figure 6. The calculation was also repeated without removing these cycles (results not given here). For Swedish conditions, including nonstationarities introduces only a very small bias. With regard to wind direction, no cycles were removed. The reason is that circular correlation is based on the differences (or sums) between the two series. According to Robeson and Shein (1997), these are less affected by the presence of nonstationarities than by deviations from a temporal mean, which are the base for scalar correlation. Comparing the wind speed correlograms with those presented by Robeson and Shein (1997) for the United States, Swedish wind speeds have a slightly smaller scale. Possible explanations for this difference could be that their correlograms are biased by the presence of seasonal and diurnal cycles (it is not mentioned whether nonstationarities are removed) or that stations were selected from places with relatively little influence from the surroundings.

4.5. Diurnal and annual cycles of wind speed

This section looks at the strength of the seasonal and diurnal cycles in wind speed and the time when the strongest winds occur. Spatial differences in both strength and the timing across Sweden are analyzed.

4.5.1. Annual cycle. Amplitudes and phases of the annual cycles are presented in Figure 7. Amplitudes (Figure 7(a)) are strongest (>1.5 m/s) at coastal stations and at two mountainous locations and are weakest at stations in the inner regions of middle and northern Sweden. The annual cycle peaks in winter for regions south of 61°–62°N at all coastal stations and at some locations in the mountains (Figure 7(b)). In the inner regions of central and northern Sweden, annual cycles mainly peak in late spring or early summer. At these

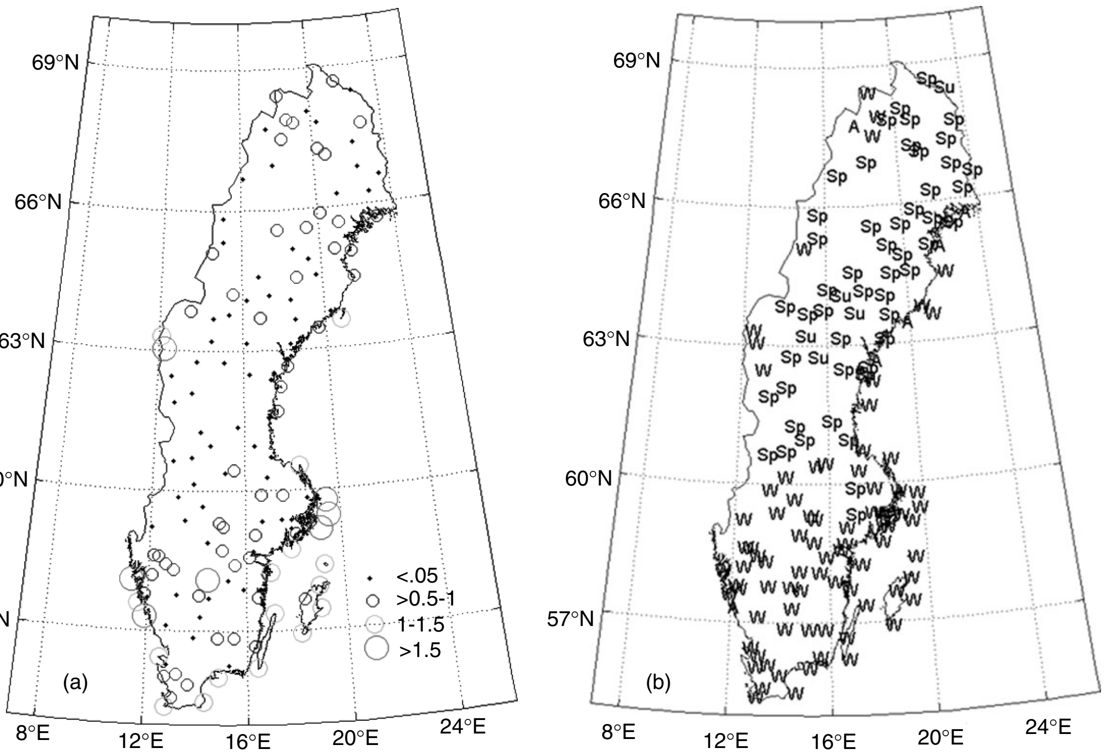

Figure 7. Strength and timing of annual cycle: (a) amplitude (m/s); (b) season when cycle peaks (W: winter, Sp: spring, Su: summer, Au: autumn)

locations, annual cycles are in general weak and winds are weakest in winter despite stronger winter pressure gradients. Figure 8 shows annual cycles for selected stations located in different regions and with different timing of the maximum: winter (33), spring (92) and summer (99). The seasonal wind speed variation at station 33, located on a small island off the west coast, is mainly controlled by the annual cycle in the large-scale forcing (see for comparison Figure 11(c)). At stations 92 and 99, however, the annual cycles are also influenced by other processes like seasonal variations in the atmospheric stability. The increased wind speeds in June are probably caused by stronger large-scale forcing during 1999–2000 as will be shown in Section 4.6.

The geographical distribution of the strength of the cycles (Figure 7(a)) suggests greater amplitudes at exposed sites where the annual means are high compared to more sheltered locations. This finding is in good agreement with Børresen (1987), studying the Norwegian wind climate, but this applies even to winds on a global scale. According to Yin (2000), annual amplitudes of monthly mean wind speeds are generally <1 m/s and are related to annual mean wind speeds with a correlation coefficient of 0.5. For Swedish winds, the correlation coefficient amounts to 0.85. Regional differences in the strength of the annual cycle as well as in the timing suggest differences in atmospheric stability over land and water and seasonal stability variations that highly influence wind speeds over the course of a year. Owing to the larger heat capacity of water compared to land, the seasonal variation in stability over the sea is smaller than over land. In autumn and winter, the ice-free sea is warmer than the land surfaces and the atmosphere is less stably stratified over water. This situation favors downward mixing of momentum over the sea and at coastal locations, which together with the stronger pressure gradient in winter enhances wind speeds. In contrast, the more stable conditions over land limit downward mixing of momentum in winter, which reduces the seasonal wind speed variations. Inland stations in middle and northern Sweden, especially when located in valleys and basins, experience strong temperature inversions in winter when outgoing long-wave radiation exceeds incoming shortwave radiation over extended periods (Liljequist, 1970). This leads to spatial differences in the season of peak monthly mean wind speeds which are absent south of ∼60 °N.

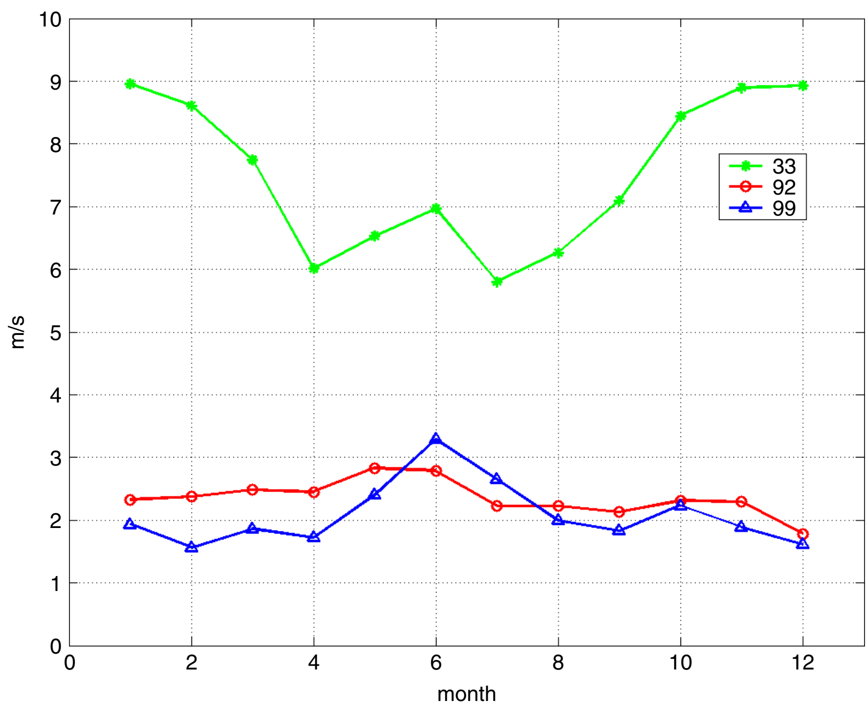

Figure 8. Annual cycle in 1999–2000 for three selected stations. This figure is available in colour online at www.interscience.wiley.com/ijoc

4.5.2. Diurnal cycle. The amplitudes and phases of the diurnal cycles were calculated for each month. As an example, the results are shown for December (Figure 9) and June (Figure 10). Only stations with at least 50% complete days are included in the figures. Diurnal wind speed amplitudes in December (Figure 9(a)) are at most locations very weak (<0.25 m/s); larger amplitudes (0.25–0.75 m/s) are only found at coastal stations and some scattered stations in the mountainous area. The time of the day when the maximum occurs is given in Figure 9(b), though only for cycles with amplitudes >0.25 m/s (the time when even weaker cycles peak is very hard to define). Winds along the east coast south of 61 °N have their daily maximum in the morning hours between 7 am and 12 am; north of 61 °N, daily cycles peak in the afternoon between 1 pm and 5 pm. In contrast, along the west coast and partly in the mountainous stations, the daily cycle peaks during night. With the arrival of spring and summer, the number of stations with amplitudes >0.25 m/s increases and they become progressively stronger (not shown here). In June, amplitudes generally range between 0.25 and 1.25 m/s (Figure 10(a)). Amplitudes >1.25 m/s are found at some scattered stations close to the coastline as well as at inland stations, while very weak diurnal cycles are almost entirely restricted to stations located on small islands. Here, little diurnal variation in wind speed is expected since the surrounding water surfaces hold a relatively constant temperature over the course of a day, restricting diurnal air temperature variations in the boundary layer above. In general, wind speeds reach their maximum in the afternoon hours despite few exceptions when the timing of the maximum occurs during the morning or even during the night (Figure 10(b)). With the arrival of autumn and winter, the reversed process as in spring takes place, i.e. progressively decreasing diurnal amplitudes (not shown here).

Lager diurnal amplitudes at coastal places in December (Figure 9(a)) are probably caused by the difference in air temperature between the cooler land surface and the warmer ice-free sea surface. This gives rise to local and mesoscale circulation systems and differences in atmospheric stability in the coastal zone (Hsu, 1988). Holmer and Haeger-Eugensson (1999) studied the winter land breeze in the Gothenburg area along the west coast during three winter seasons. They found that during occasions of land breeze, typically air temperatures over sea were 4–5 K higher than those over land with the largest temperature difference during

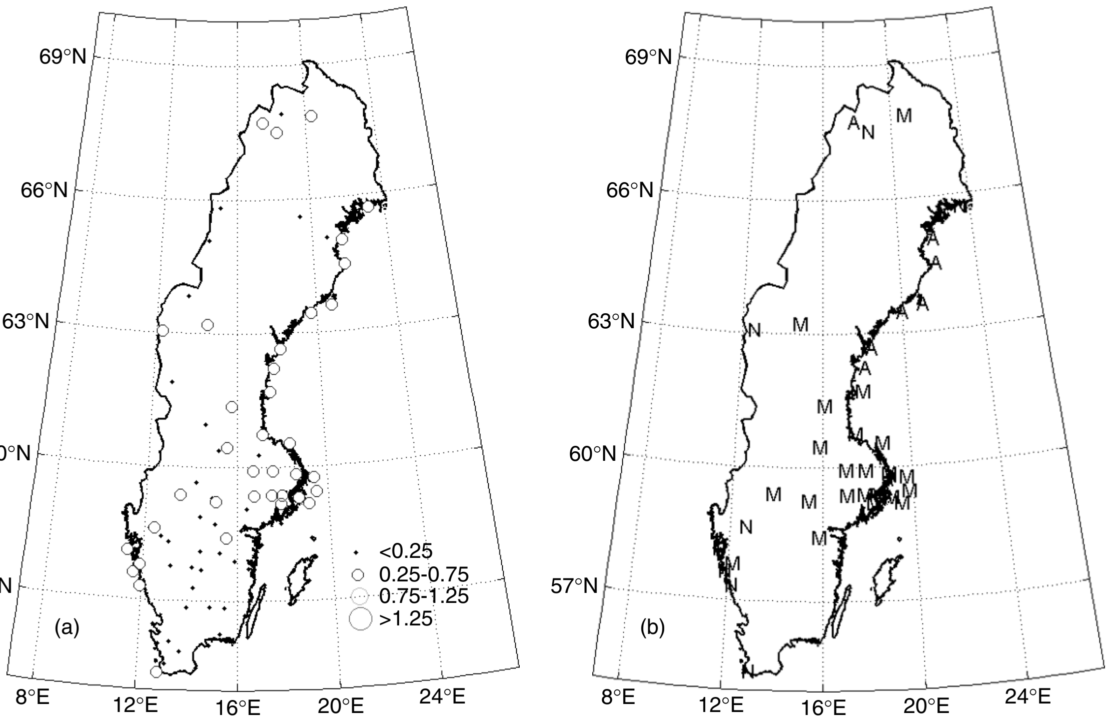

Figure 9. Daily wind speed cycle in December: (a) amplitude (m/s); (b): time of day when cycle peaks (only given for amplitudes >0.25 m/s) (N: peaks at night (6 pm–6 am), M: peaks in the morning (7 am–12 noon), A: peaks in the afternoon (1 pm–5 pm)

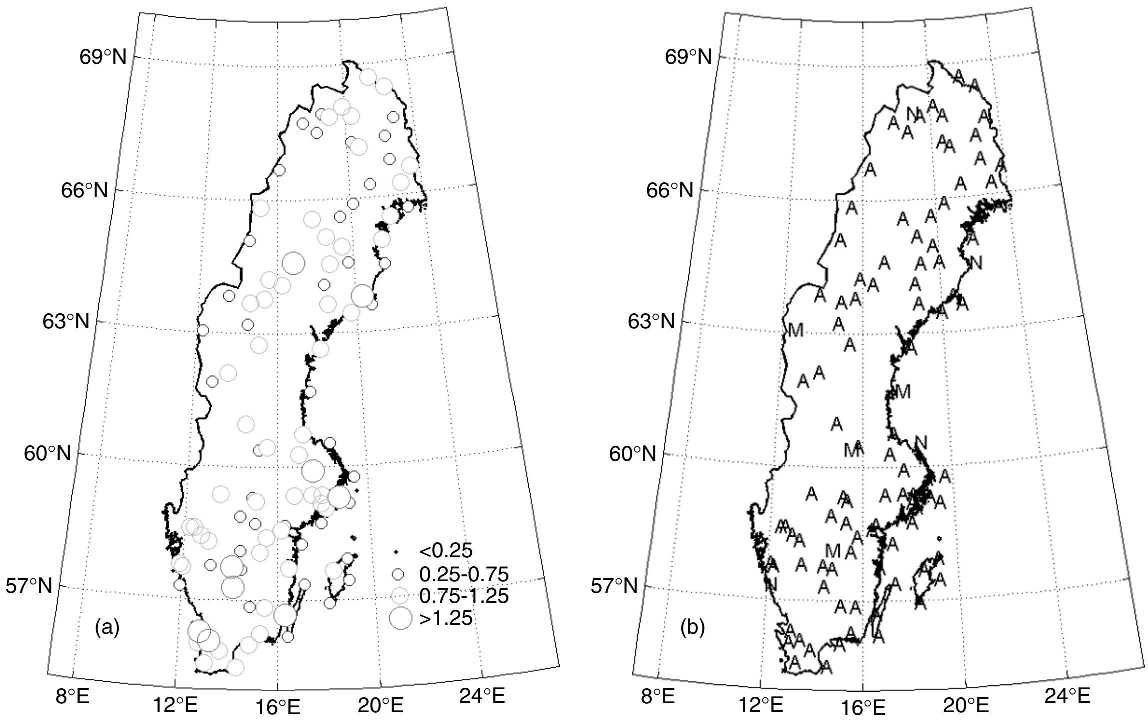

Figure 10. Same as Figure 9 but for June

midnight. As a consequence, differences in atmospheric stability over land and over water are expected to be largest in the night or early morning, which could explain the occurrence of the reversed diurnal cycle along the west coast. With regard to the sea-ice conditions along the Swedish coast, on average, sea ice starts building up in the northernmost part of the Baltic Sea at the end of November. South of 63 °N, the east coast is normally ice-free until the end of December, and freezing is even more delayed in the south and along the west coast (SMHI, 2004). The winters 1999/2000 and 2000/2001 were mild, which delayed the buildup of sea ice (FIMR, 2004), implying that similar conditions prevailed along all the whole coastline of Sweden.

Inversed diurnal cycles were observed at the Danish Vindeby wind farm in all seasons except in winter (Barthelmie *et al.*, 1996) and along the Dutch coast (Coeling *et al.*, 1998). Both studies stress the importance of wind direction (i.e. fetch) and station location relative to the coastline for the strength and phase of the diurnal cycle. This also became obvious when calculating diurnal cycles for different wind direction sectors at some selected stations in Sweden (not shown here). A more systematic study of the diurnal wind speed variation with respect to station location, daily mean wind speed and wind direction would provide deeper insight into the processes, but is beyond the scope of this work.

Owing to Sweden's elongated shape in north–south direction, the number of hours when the sun is above the horizon varies greatly with latitude. In December, the sun is above the horizon for about 7 h in southernmost Sweden (but at a very low angle), whereas the northernmost region only gets a few hours of twilight. Conditions are very different in June when the sun never sets in northern Sweden but is below the horizon for about 4 hours in southern Sweden. During winter, the short daylight length largely restricts the boundary-layer growth, reducing diurnal wind speed variations. Investigating diurnal variations in surface winds on a global scale, Dai and Deser (1999) found that seasonal variations in the diurnal cycle amplitude are only significant over land areas and outside of the tropics. According to their study, typical amplitudes for Scandinavia range between 0.6 to 0.8 m/s in winter and between 1.4 and 1.8 m/s in summer.

4.6. Representativeness of winds in 1999–2000

With only 2 years of data, it may be questioned to what degree the results can be regarded as climatologically representative, and additional analyses were therefore carried out. First, the annual cycles from 39 stations belonging to the earlier as well as the automated station network were calculated for the period 1963–1990 and compared to the 1999–2000 cycles. Second, the circulation in 1999–2000 over the two domains located over southern and northern Sweden were compared with the conditions in 1961–1990 (Figure 11).

With regard to the annual cycles for 1963–1990, the uncertainties related to the earlier data are considered as systematic (e.g. measurement heights may vary from site to site but remain stable in time). It is thus argued that spatial patterns of amplitude and timing found in the short observation period can be confirmed as long as the pattern's relative, and not its absolute, magnitude is of main interest. Despite the much sparser station coverage, especially in northern Sweden, the results (not given here) show the same general trend with stronger developed annual cycles along the coast and weaker cycles at inland stations. Monthly wind speeds generally peak in winter except at one inland station in northern Sweden when the peak occurs in May. The spatial variability in amplitude and timing derived from the 2 years is therefore considered as representative.

For both domains, shape and strength of annual cycles of the total geostrophic flow w_g are similar in both periods (Figure 11 (c) and (f)). With regard to the southern domain, autumn, winter and June 1999–2000 differed from the long-term mean with partly stronger w_g (Figure 11(c)), while w_g during the 2 years was weaker in December over the northern domain (Figure 11 (f)). Stronger w_g in June 1999–2000 was probably responsible for the secondary peaks in the annual cycles of surface winds shown in Figure 8 (station 33 and 99), while the weaker w_g in December 1999–2000 probably caused the drop in surface wind speed at station 141 (Figure 8). The annual cycles of u_g and v_g for the southern domain revealed increased westerly winds in late winter (Figure 11(a)) and increased southerly winds in autumn (Figure 11(b)). In the northern domain, v_g in November 1999–2000 deviated from the long-term mean (Figure 11(e)). The relative frequency distribution of u_g and v_g for both domains (not shown here) was characterized by more frequent westerly and southerly winds in 1999–2000. Very strong easterly and northerly winds were absent, i.e. the 2 years did not entirely cover the whole range of u_g and v_g in 1961–1990.

According to Pryor and Barthelmie (2003), 850-hPa wind speeds over the Baltic increased significantly during the latter half of the twentieth century. They relate these changes to an increase in westerly and northwesterly circulation types, which in turn is caused by the NAO. Since the 1970s, the NAO is mainly in its positive phase and has grown more positive during the past 20 years (e.g. Hurrell and van Loon, 1997). In 1999 as well as in 2000, the NAO was positive, which explains the increased frequency of westerly and southerly geostrophic winds.

5. CONCLUDING REMARKS

Recent surface wind conditions in Sweden are studied for a 2-year period, focusing on the spatial variability of surface winds in Sweden. In the following text, the main findings are summarized. Typical annual mean wind speeds in Sweden range between 2 and 5 m/s with higher values at exposed sites in the mountains or on islands. Southern Sweden south of $\sim$58°–60 °N experiences, in general, higher annual means since this region is more directly exposed to winds from the west and the southwest. Annual means, wind speed variability and extremes decrease quickly with increasing distance from the coast. This feature is attributed to the abrupt change in surface roughness at the coastline.

The mean annual wind velocities combining wind speed and direction information provide additional information on the spatial distribution of winds. Flow velocities are stronger and more coherent in direction in southern Sweden than further north. There exists a rather clear boundary dividing the flow regimes along a southwest-to-northeast-directed line located at $\sim$58°–60 °N. South of this line, flow directions are predominantly from SW, and the flow field reflects the trajectories of common air masses. Central and northern Sweden's flow directions vary more in space and wind steadiness is generally lower. This feature is explained by increased variability in wind directions, which is due to a more pronounced influence of

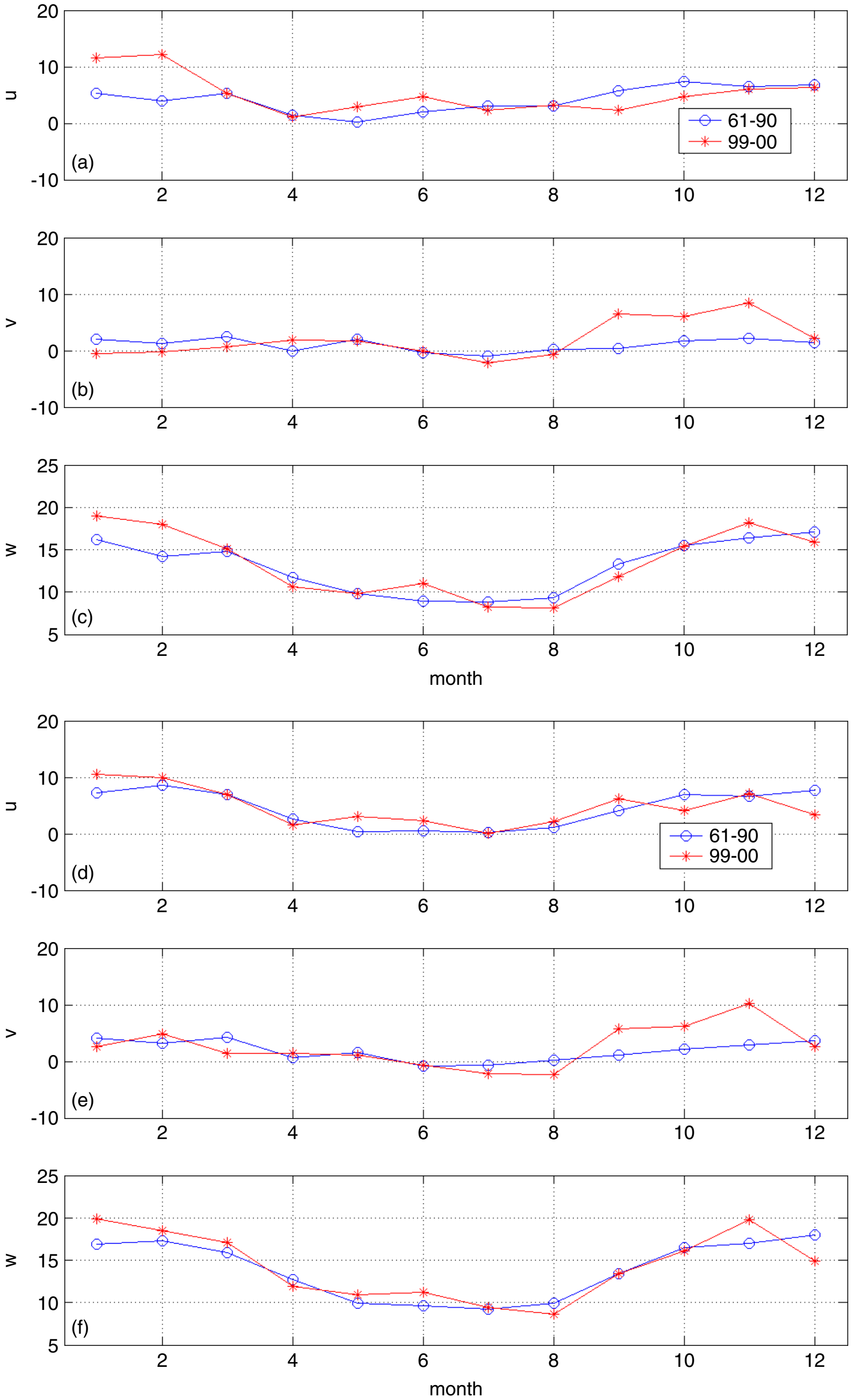

Figure 11. Seasonal variation in the large-scale circulation in 1961–1990 and 1999–2000 for two regions represented by u_g, v_g and w_g: (a–c) southern domain; (d–f) northern domain. The units are in hectopascal per 10° latitude at 60 °N. This figure is available in colour online at www.interscience.wiley.com/ijoc

the topography restricting surface winds to certain wind direction sectors. It also explains the weaker flow strength in central and northern Sweden.

The spatial coherence of wind speed and direction is determined by means of spatial autocorrelations. Exponential functions are fitted to the 'global' and 'local' distance–decay relationship. The spatial scale is defined as the distance at which correlation has dropped to ∼0.37. Using this measure, the 'global' scale of wind direction amounts to ∼500 km and is somewhat larger than that for wind speed (∼350 km). 'Local' distance–decay relationships for both variables change considerably depending on location. These range from 38 to 530 km for wind speed and from 40 to 830 km for wind direction. In general, 'local' scales tend to be larger south of ∼58°–60 °N because of weaker topographical influence.

Surface wind speeds vary over the course of a year in accordance with seasonal variations in atmospheric pressure patterns and surface boundary-layer dynamics. The strength of the annual cycles generally varies between 0.25 and 1.5 m/s. Amplitudes are greater at exposed sites and correlate generally well with annual mean wind speeds. Owing to variation of the large-scale forcing with season, monthly mean wind speed normally peaks during winter at all stations in the southern part of the country, at stations near the coast and at some locations in the mountains. In addition, many stations experienced a secondary peak in summer because of stronger than normal (1961–1990 means) geostrophic flow conditions in June 1999 and 2000. Winds in the inner regions of central and northern Sweden peak in early summer despite the stronger pressure gradients in winter. Temperature inversions persisting over extended regions in northern and central Sweden during the cold season decouple surface winds from the large-scale forcing. It clearly demonstrates the special role of atmospheric stability on the winter surface winds.

Wind speeds vary over the course of a day as a consequence of the diurnal atmospheric boundary-layer evolution. Owing to the large differences in the length of the daylight in Sweden between summer and winter, the strength of the diurnal wind variations differs considerably. Diurnal wind speed variations in December are very small (<0.25 m/s) except for some stations at the coast and in the mountains, where there exists a stronger coupling between large-scale forcings and surface winds. The timing of the peak may occur either during the daytime or at night (west coast). In June, the diurnal amplitudes generally amount to 0.25–1.25 m/s with stronger amplitudes at some coastal and inland stations. Very weak diurnal cycles are restricted to stations located on small islands.

The comparison of the circulation indices for 1999–2000 and 1961–1990 shows both similarities and deviations. With regard to the total geostrophic flow, the annual cycle as well as the frequency distribution are rather similar in the two periods, but stronger deviations were found when comparing the components separately. Stronger westerly (southerly) flows were found in winter and June (November) in 1999–2000, leading to increased geostrophic forcing. As a consequence, secondary peaks during these months have been observed in the annual cycles of the surface winds, which may not be a typical feature in the climatological sense. The 1999–2000 large-scale forcings are characterized by an increased occurrence of westerly and southerly winds which probably is related to the positive NAO phase.

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