

RESEARCH ARTICLE

Present and future downslope windstorms in the Scandinavian Mountains from a kilometre-scale climate model

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Abstract

Climatological studies of downslope windstorms (DWs) in the Scandinavian mountains (Scandes) are rare. Here we use 20-year long simulations with the kilometre-scale regional climate model HCLIM38-AROME to study DWs in the Scandes in present and future climate, their connection to large-scale atmospheric circulation, and their cooling and warming effect. A DW is identified in a model grid point using a two-step conceptual model based on terrain features, and dynamic and thermodynamic variables. We find that DWs occur most frequently in winter, which is therefore the focal season for the analyses. Normally, DWs are predominantly either warming (foehn type) or cooling (bora type), but our results indicate that this distinction does not hold for the DWs in the Scandes. Even though there is a slight overall tendency towards warming, almost all locations with DWs can experience both cooling and warming. The future simulations using two parent global climate models (GCMs) show an overall decrease in the total number of DW occurrences between 7% and 17% toward the end of the century, with a decrease in the northern and central parts of the Scandes and an increase in the southeastern part. These changes can be attributed to two factors: the frequency change of certain circulation types and internal changes within the circulation types. Despite the agreement in the sign of future changes for both GCM forcings, the difference in the contribution of the two factors indicates that different mechanisms are responsible for the total future change in the DW occurrence. The complexity of DWs in the Scandes and their future change indicates that kilometre-scale or finer climate models are required for a proper depiction of DWs and that a larger ensemble of simulations with different GCM forcing is required for evaluation of different mechanisms of the future change.

KEYWORDS

climate change, climate model, downslope windstorm, HCLIM, Scandinavia, wind

1 | INTRODUCTION

A downslope windstorm (DW) is a mesoscale meteorological phenomenon that occurs in mountainous regions. It is a strong wind with mean hourly speeds often exceeding $20 \text{ m}\cdot\text{s}^{-1}$ and gusts of 50 to $70 \text{ m}\cdot\text{s}^{-1}$ (e.g., Doyle & Shapiro, 2000; Grisogono & Belušić, 2009) that originates from airflow over a mountain and a steep lee slope (e.g., Durran, 1990). Some of the globally recognised DWs include the foehn in the Alps, the Chinook in the Rocky Mountains, the Santa Ana winds in southern California, the Zonda in Argentina and the Bora on the eastern Adriatic coast (e.g., Abatzoglou *et al.*, 2021; Kozmar & Grisogono, 2020; Romanic, 2019).

The key mechanism for DWs is the interaction between the flow and the obstacle (Doyle & Shapiro, 2000), which is associated with the generation of waves and their breaking (Klemp & Durran, 1987). Two flow features substantially affect the generation of DWs: a wave critical level and a temperature inversion. A critical level is the height at which the phase speed of vertically propagating waves is equal to the velocity of the surrounding air (e.g., Grisogono & Belušić, 2009), which prevents the vertical propagation of mountain waves and favours their breaking when approaching this level (non-linear effects). The presence of an inversion or a stable air layer also influences the DW dynamics (e.g., Vosper, 2004). The highest wind speeds in the lee are associated with three fundamental situations (e.g., Durran, 1990; Grisogono & Belušić, 2009):

- i. amplification of a stationary mountain wave, its tilting and possible breaking—critical level induced by the wave,
- ii. presence of a critical layer at low altitude (approximately from 0.5–1 vertical wavelength above the terrain; e.g., Enger & Grisogono, 1998; Scinocca & McFarlane, 2000), and
- iii. temperature inversion at the mountain-top level (e.g., Vosper, 2004).

The result of each of these processes separates the low-level, fast flow from the overlying layer. These processes can also occur in combination and lead to even faster winds.

Local effects of DWs can be roughly divided into those of a foehn-like character, which cause the lee area to warm up, and those of a bora-like character, which cool it down. Apart from the local warming or cooling character, there is no significant difference in the dynamics of these winds (Durran, 1990; Gohm & Mayr, 2005). Both wind types have an elevated layer with increased static stability on the windward side, which begins to sink towards the mountain and in the lee. This leads to

an acceleration of the wind on the windward side of the mountain according to Bernoulli's principle and due to the pressure difference.

There are several mechanisms that can be responsible for the warm character of the foehn. Elvidge and Renfrew (2016) show that the most important processes are: (a) isentropic drawdown, (b) latent heating and precipitation, and (c) mechanical mixing due to turbulence. The authors also mention that the areas with the greatest temperature change are not in the jet of the DW, but near it, where the wind is weaker and more turbulent.

Downslope windstorm impacts can be considerable: they can stop traffic (e.g., Kozmar *et al.*, 2012, 2015), melt snow (e.g., Laffin *et al.*, 2023) or rapidly spread fires (e.g., Abatzoglou *et al.*, 2023; Sharples *et al.*, 2010). The effects of climate change on DWs are therefore of high societal and scientific interest. Given that future projections of wind have large uncertainty (e.g., Jaison *et al.*, 2024), it is important to provide uncertainty estimates in such studies. However, climatological studies of DWs in the Scandes have been scarce due to the lack of high-resolution wind data required to analyse such local events. The main focus of research on DWs in the Scandes has so far tended to be on local scales and on case studies, while analyses on larger spatial and temporal scales are not common.

Harstveit *et al.* (1995) analyse five cases of local severe wind in the Scandes and link their occurrence to a layer of increased stability in front of a warm or occluded front moving over the southern Norwegian mountains. They also point out the complexity of analysing the interaction of a front with the orography.

Grønås (1997) points out that processes of all scales influence the flow in the Scandinavian region, which in turn can have an impact on DWs. On a larger scale, the Earth's rotation influences the flow considerably (i.e., the geostrophic wind). The Scandinavian peninsula is surrounded by sea, which favours wind acceleration due to lower friction and static stability, and also serves as a source of moisture. The steep terrain of the Scandinavian peninsula forces an uplift of the air and thus the formation of orographic precipitation, which can be considerable, especially on the west coast. The associated release of latent heat can further influence the flow over the mountains. On smaller scales, flow separation processes occur (e.g., Baines, 1995), which can lead to strong winds in valleys with steep flanks. Grønås (1997) reports that for a case study storm in the Scandes the greatest wind damage occurs in the lee of small mountains where the slope is steep (horizontal extent ~500–2000 m, height ~100–1000 m).

Grønås and Sandvik (1999) performed simulations with a grid spacing of 270 m using MEMO (Mesoscale Model; e.g., Moussiopoulos, 1994, 1995),

a non-hydrostatic numerical meteorological model, to explain the occurrence of strong winds over the steep terrain of Norway. The results show flow separation in a valley with wind reversal at the valley floor, indicating the complexity of local winds in the Scandinavian region and the need for high resolution in DW studies. Cao and Fovell (2016) generally state that model grids with a spacing of less than 4 km are required for DW studies. Sandvik and Harstveit (2005) also use MEMO to study a case of DW in southern Norway.

High-resolution climate modelling studies for the Scandinavian region are limited, especially for wind. Furthermore, we know of only a few studies addressing DWs in future climate in other regions using down-scaled climate projections, with resolutions of 10 km or coarser (e.g., Belušić Vozila *et al.*, 2019; Guzman-Morales & Gershunov, 2019). The NorCP project (Nordic Convection Permitting Climate Projections; Lind *et al.*, 2020, 2023) produced the first 3-km climate simulations over Fenno-Scandinavia using a convection-permitting regional climate model (CPRCM). Several related studies focus on the added value of using a CPRCM compared to a coarser regional climate model (RCM) and global climate models (GCMs), with a focus on precipitation (e.g., Lind *et al.*, 2020; Médus *et al.*, 2022; Olsson *et al.*, 2021). Lind *et al.* (2020) briefly address issues of large-scale circulation and show that the prevailing flow over the Scandinavian peninsula and northern Europe is westerly during the cold season. In the warm season, the mean sea level pressure (MSLP) in northern Europe is generally larger than in the cold season and the pressure gradient and winds are weaker. Lind *et al.* (2023) show decreased MSLP in northern Scandinavia and increased MSLP in continental Europe in both December–January–February (DJF) and June–July–August (JJA) for a high greenhouse gas emission scenario. Previous studies using GCMs show a decrease of DJF wind speed in the northern Scandes and increase in the southern Scandes in the future climate (e.g. Martinez & Iglesias, 2024; McInnes *et al.*, 2011; Reyers *et al.*, 2016). Christensen and Kjellström (2018) show that there is a large spread in average wind speed changes over north Europe in EURO-CORDEX RCMs (e.g., Jacob *et al.*, 2014), especially in winter. In summer, there is a more consistent, but small, decrease.

The availability of NorCP data allows us to study the dynamics and climatology of DWs over a 20-year period in the past and provide insights into the changes of DWs towards the middle and end of the 21st century. Making use of these high-resolution data, this study investigates the DW climate in the Scandes in the present and future climate, the connection of DWs to the large-scale atmospheric circulation, and the cooling and warming effect of DWs at the time of their onset.

2 | DATA AND METHODS

2.1 | HCLIM38

The regional dynamical downscaling in this study was performed using HARMONIE-Climate (HCLIM; Lindstedt *et al.*, 2015; Belušić *et al.*, 2020). HCLIM is a regional numerical climate model based on the numerical weather prediction (NWP) system HARMONIE-AROME (Bengtsson *et al.*, 2017), which is part of the ACCORD NWP system (A Consortium for CONvection-scale modelling Research and Development, i.e., a European consortium for limited-area NWP models). HCLIM38 is the 38th cycle (i.e., the 38th version) of HCLIM.

The focus of HCLIM is on climate modelling at smaller spatial scales resolving deep convection (HCLIM38-AROME), although the RCM configuration with parameterised convection (HCLIM38-ALADIN) is also widely used (Belušić *et al.*, 2020). HCLIM38-AROME is used with non-hydrostatic dynamics (Bengtsson *et al.*, 2017). Most convective processes occur on scales of the order of 1–10 km, which is comparable to the grid spacing of HCLIM38-AROME. Convective processes on smaller scales are taken into account by the parameterisation of shallow convection. Simulations with coarser resolution (>10 km) are usually performed with HCLIM38-ALADIN, where a parameterisation of deep convection and the hydrostatic approximation are used (Termonia *et al.*, 2018). The processes at the land surface are simulated with SURFEX (Masson *et al.*, 2013). Further details on the modelling system can be found in Belušić *et al.* (2020).

2.2 | NorCP

The historical and future regional climate simulations analysed in this study were performed within the NorCP project. A detailed description of the NorCP project can be found in Lind *et al.* (2020, 2023) and we only provide a brief overview here. As part of the project, simulations were carried out with HCLIM38-AROME, with a horizontal resolution of 3 km, 65 vertical levels and a time step of 75 s (hereafter HCLIM3). As Matte *et al.* (2017) report, an intermediate step is required to bridge the large grid spacing of the GCM with the small grid spacing of the CPRCM, as the direct transition is too abrupt. Therefore, the intermediate step was set to 12 km with the HCLIM38-ALADIN model (hereafter HCLIM12). HCLIM12 has 65 vertical levels and a time step of 300 s, and provides the nested HCLIM3 grid with boundary data every three hours. For the evaluation simulation, ERA-Interim reanalysis data (e.g. Dee *et al.*, 2011) with a horizontal resolution of about 80 km

TABLE 1 HCLIM simulations analysed in this study. Used global climate models (GCMs) are EC-Earth2 and GFDL-CM3.

HCLIM12 & HCLIM3 runs	Period	Forcing
Evaluation	1998–2017	ERA-Interim
Historical	1986–2005	GCM
Mid-century	2041–2060	GCM-RCP8.5
End-century	2081–2100	GCM-RCP8.5

and a time step of six hours were used as the boundary data for HCLIM12. The simulation covers the period from 1998 to 2018, but in this paper we use 20 years, 1998–2017.

For the scenario simulations downscaling GCMs, there are three runs (Table 1): historical from 1986 to 2005, mid-century from 2041 to 2060 and end-century from 2081 to 2100, the latter two using the RCP8.5 scenario (e.g., Moss *et al.*, 2010), i.e., a high greenhouse gas emission scenario.

For each simulation, the boundary data from the CMIP5 (Taylor *et al.*, 2012) GCMs EC-Earth2 (e.g., Hazeleger *et al.*, 2010, 2012) and GFDL-CM3 (e.g., Donner *et al.*, 2011; Griffies *et al.*, 2011) are used (hence, HCLIM3-ECE and HCLIM3-GFDL onward). While generating a larger CPRCM ensemble was computationally too prohibitive, the uncertainty range of projections, which can be large for wind (e.g. Jaison *et al.*, 2024), has been estimated here using two parent GCMs that span a larger part of the ensemble spread: EC-Earth2 is close to the CMIP5 ensemble mean precipitation and temperature change by the end of the century, while GFDL-CM3 is at the higher end of these changes (see Lind *et al.*, 2023 for more details). Furthermore, the EC-Earth model family has been placed among top-performing climate models in terms of reproducing large-scale circulation types (Brands, 2022; Brands *et al.*, 2023), which is ultimately related to local winds. For all simulations, the model is run for one year before each period to spin up the soil. HCLIM12 covers almost all of Europe and the northeast Atlantic, while HCLIM3 covers Fenno-Scandinavia and the northern continental European countries (Figure 1).

Given that the Scandes extend over about 10 degrees in latitude with a total length of around 1700 km, they cover a large variety of climatic and terrain conditions. Therefore, we separate them into four subdomains for the analysis (Figure 1): Northern (N), Central (C), Southwestern (SW) and Southeastern (SE).

2.3 | Observational data and evaluation

To evaluate the modelled wind, the hourly 10-m wind observation data are retrieved from four observational

stations, which are shown in Figure 1: Fokstugu (62.11° N, 9.29° E), Juvvasshoe (61.68° N, 8.37° E) Makkaur Fyr (70.71° N, 30.07° E) in Norway and Stora Sjöfallet (67.49° N, 18.29° E) in Sweden. These are the mountain stations where DW events occur in the model, and the horizontal dimensions of surrounding terrain features are sufficiently large to be resolved by the 3-km model. For each observational station, the four closest model grid points are considered as potential candidates for evaluation, and finally the grid point with the largest number of DW occurrences is selected. Since the climate model simulations cannot be assumed to match the observed reality from case to case, but they rather develop their own circulation, the observations are used solely for statistical validation of the simulations. The observed and modelled wind speed distributions are compared using quantile–quantile (Q–Q) plots. We want to focus the evaluation of wind speed only on DWs. Since DWs are not detected from the observations, we determine preferred DW wind directions for each observational station from the model data. The preferred DW wind directions are determined as the union of 22.5° ranges from the HCLIM12 and HCLIM3 DW wind roses where the DW frequency exceeds 30%. The simulated wind speed data from the preferred direction ranges are compared with observed data from the same wind direction ranges for the period from 2008 to 2017. After evaluating the overall statistics here, only the model output is used for subsequent analyses.

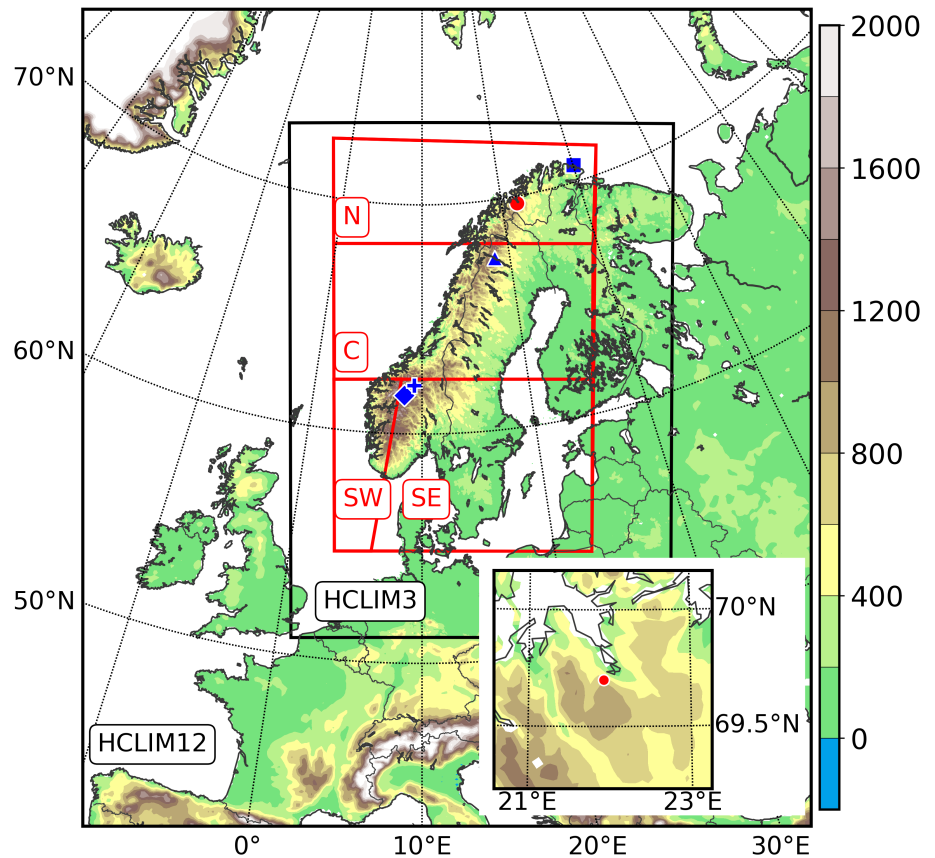
2.4 | Downslope windstorm identification

Abatzoglou *et al.* (2021) developed a two-step conceptual model to identify DWs based on terrain characteristics, horizontal and vertical wind speed and static stability. In the first step, the model identifies DW-prone areas as sites meeting both a 2% minimum local slope threshold and a vertical distance of at least 200 m beneath the local (in the 50-km radius) mountain top.

For grid points that fulfil the terrain criteria, the second step of the model is activated. It is based on three factors: (1) mean cross-barrier wind at the mountain top (v_{cb}), (2) maximum static stability at or above the mountain top ($\delta\theta/\delta z$), and (3) maximum vertical velocity at or below the mountain-top level (ω).

Cross-barrier wind is calculated as the mean wind perpendicular to the slope in the 100-hPa deep layer whose bottom is at the pressure level closest to the local mountain top. The static stability is taken as the maximum value of $\delta\theta/\delta z$ from the same layer. All isobaric surfaces from the local mountain top to the ground level in the model

FIGURE 1 Domains of HCLIM12 and HCLIM3 (black rectangles). The HCLIM3 domain is divided into four subdomains for analyses (red polygons): Northern (N), Central (C), Southwestern (SW) and Southeastern (SE). The colour bar represents orography height (m). The red dot denotes the analysed grid point and blue symbols denote the locations of meteorological stations (plus, Fokstugu; diamond, Juvvasshoe; triangle, Stora Sjöfallet; and square, Makkaur Fyr). A zoomed-in area surrounding the analysed grid point is indicated in the lower right corner of the HCLIM12 domain.



were considered for the maximum vertical velocity. A DW is detected when threshold values for all three factors are reached. In Abatzoglou *et al.* (2021), who used ERA5 data (e.g. Hersbach *et al.*, 2020) for the analysis, the thresholds were set based on published articles analysing DW velocities and the chronology of regional DW in southwestern California, as well as the horizontal resolution of ERA5 (~ 30 km). The thresholds were: $v_{cb} \geq 13 \text{ m}\cdot\text{s}^{-1}$, $\omega \geq 0.6 \text{ Pa}\cdot\text{s}^{-1}$, $\delta\theta/\delta z \geq 6 \text{ K}\cdot\text{km}^{-1}$.

The threshold value for the vertical velocity is revised for this study due to the higher resolution of the model used (3 km): the vertical velocity must be less than $-1 \text{ m}\cdot\text{s}^{-1}$ to detect DW (i.e., the subsidence is greater than $1 \text{ m}\cdot\text{s}^{-1}$; also note that the original threshold was given in pressure coordinates, which would now approximately be $12.6 \text{ Pa}\cdot\text{s}^{-1}$). The revised value is obtained by first increasing the original threshold by a factor of 10 (corresponding to the change in model resolution) and then additionally increasing the threshold to reduce the exaggerated number of DWs detected. In the absence of suitable data on DWs in the Scandes, the threshold is estimated based on the characteristics of the modelled DWs. We additionally tested a range of vertical velocity thresholds around $-1 \text{ m}\cdot\text{s}^{-1}$ and despite the change in the number of DWs detected, the main conclusions of the study did not change.

2.5 | Circulation type classification

Downslope windstorms are usually embedded in synoptic systems, so the classification of synoptic circulation types (CTs) could help to group DWs according to their local characteristics. In the classification of CTs, a time series of spatial fields (e.g., geopotential or MSLP) is grouped according to certain similarities and differences. The Euclidean distance (e.g., Philipp *et al.*, 2007) and the pattern correlation coefficient (e.g., Lund, 1963) are usually used to quantify similarity. However, these two similarity measures tend to provide very different classification results, which can be mitigated by using spatial anomalies instead of raw data as input for classification methods (Hansen & Belušić, 2021). Here, we use spatial anomalies of the MSLP as input data for the automatic classification method SANDRA (Simulated ANnealing and Diversified RAndomisation; Philipp *et al.*, 2007) to obtain 10 CTs. The classification is based on the MSLP from HCLIM12 for the period 1998–2017. Every day in all considered simulations is assigned to one of these 10 CTs. Hansen and Belušić (2021) indicate that very similar results are obtained if instead of MSLP the 850-hPa geopotential height is used. Similarly to Hansen *et al.* (2024), 10 CTs are used based on visual inspection of the results with different numbers of CTs, providing low variability

within a CT and sufficient differences between the CTs. The visual inspection is used because objective measures (e.g., the Explained Variance, the Silhouette Index) failed to point to an optimal number of CTs.

Assuming that the frequency of occurrence of DWs (i) depends on CTs and (ii) changes in the future climate, the question is whether the DW frequency change is due to CT frequency change or dynamical and thermodynamical changes within CTs. Therefore, the total change in the occurrence of future DWs, ΔC , is decomposed into two parts: one part resulting from the change in the frequency of a particular CT, and the other from the change within a particular CT (e.g., Barry & Perry, 1973; Kjellström *et al.*, 2022), following the expression:

$$n \cdot \Delta C = \sum_{i=1}^G [\Delta F_i(C_i + \Delta C_i) + F_i \Delta C_i] \quad (1)$$

where

- G is the number of CTs (here 10),
- F_i is the frequency of CT i during the past period,
- $F_i + \Delta F_i$ is the frequency of CT i during the future period,
- n is the number of time steps,
- C_i is the DW frequency in the CT i (total number of DW in CT i / times steps when CT i is present) for the past period, and
- $C_i + \Delta C_i$ is the DW frequency for the future period in CT i .

The first term on the right-hand side (the CT frequency change) implies a dependence between the variable and the CTs. The second term (within-CT change) may include dynamic and climatic variability not related to CT frequency changes but can also suggest a weaker connection between the variable and the CTs (note that in the extreme case of complete lack of connection, the second term explains the entire change of the variable).

3 | RESULTS

3.1 | Model evaluation

Evaluation of the model wind speed from the local prevailing DW directions for four chosen stations in complex terrain is shown in Figure 2. The DW directions in HCLIM3 and HCLIM12 are generally similar, with differences in Figure 2a where HCLIM12 does not have a prevailing wind direction (>30%), and Figure 2d with HCLIM12

having a WNW and HCLIM3 a NW direction. The union of both directions is used for this evaluation. It shows that HCLIM12 underestimates wind speed, especially for stronger winds, while HCLIM3 reproduces well the entire wind speed distribution of the measured data (Figure 2). These results indicate the added value and necessity of the high-resolution HCLIM3 model for local wind research in complex terrain. Therefore, we use HCLIM3 in the remainder of this study. It should be noted that the added value of CPRCMs has already been established in previous studies on near-surface wind speed, especially for extreme winds (e.g., Belušić & Lind, 2025; Belušić Vozila *et al.*, 2024; Molina *et al.*, 2024), although in some of these studies the added value was not as evident as seen here.

3.2 | Downslope windstorm occurrence

To investigate the seasonality of DWs, the percentage of DW occurrences over the 20-year period is calculated for each season. This is done by summing all DW occurrences every six hours, across all grid points within the domain where DW occurs at least once, and dividing by the product of the total number of such grid points and the total number of six-hour time steps in the season.

The frequency of DW is 0.7% in spring (March–April–May, MAM), 0.2% in summer (JJA), 1% in autumn (September–October–November, SON), and 1.8% in winter (DJF). The annual cycle of DW frequency on a monthly resolution can be seen in Figure S1. This study therefore focuses on the DJF season and all subsequent results relate to this season. The highest DW occurrence in the DJF season is in the central subdomain (2%), followed by northern and southwestern (1.9%), and southeastern (1.2%).

To relate DW occurrence to the larger-scale atmospheric circulation, 10 CTs are computed over the domain as described in Section 2.5 but only for DJF. Figure 3 shows the 10 CTs over Scandinavia for the DJF season. The CTs are ordered by frequency of occurrence, so that CT 1 is the most frequent type and CT 10 the least frequent. The classification is based on the MSLP from HCLIM12 for the period 1998–2017.

Figure 4 depicts the occurrence of DWs for DJF for each CT. DWs are associated with certain CTs more than others. For example, the most common CT, CT 1, does not trigger a large number of DWs, only 140 per CT occurrence. The rarest CT, CT 10, triggers the relatively largest number of DWs with 719 DWs per CT occurrence. CT 10 is followed by CT 7 with 649 and CT 3 with 395 DWs per CT occurrence. All three CTs with the highest average number of DWs have approximately the same ratio of occurrences between the subdomains.

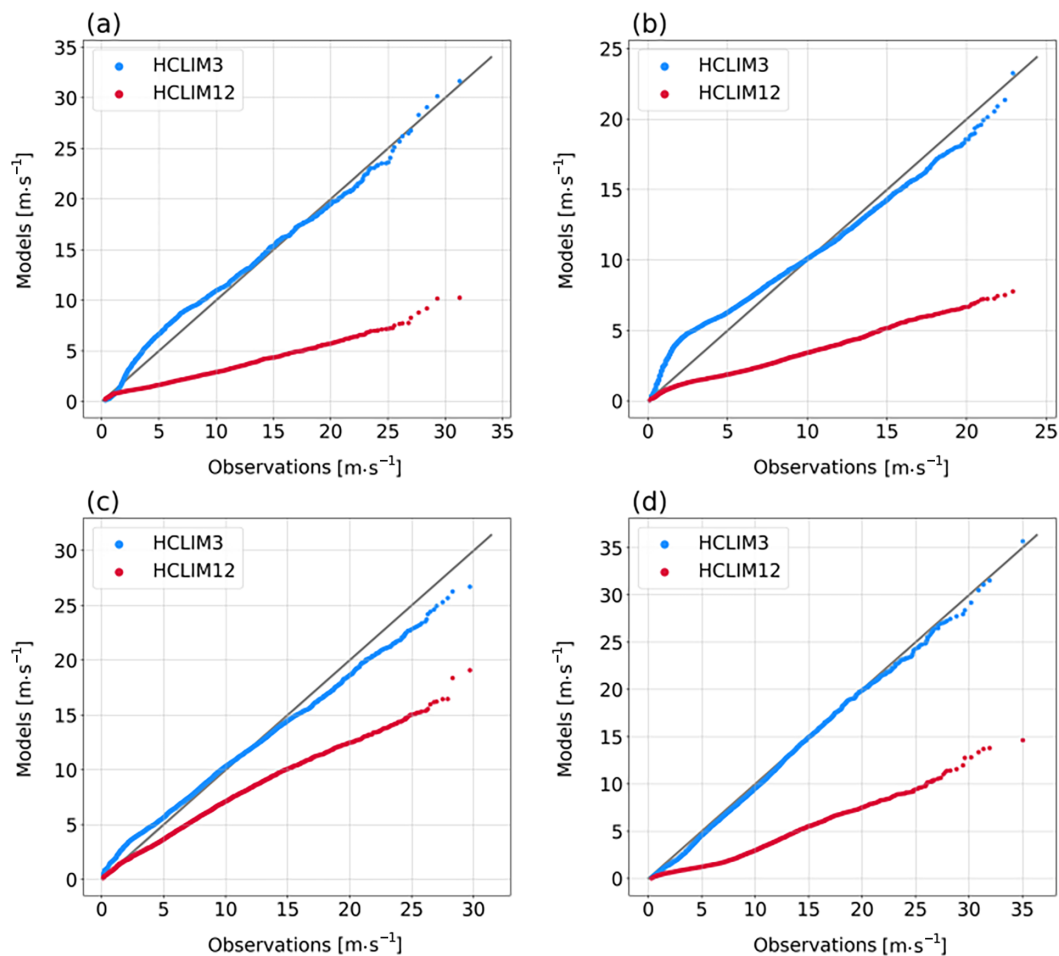


FIGURE 2 Quantile–quantile (Q–Q) plots of wind speed from the local prevailing downslope windstorm (DW) direction (see text for details) for 2008–2017 between the model results (HCLIM12 and HCLIM3) and observations at four observational stations: (a) Juvvashoe (prevailing wind direction S), (b) Fokstugu (prevailing wind directions SSE and S), (c) Makkaur Fyr (prevailing wind directions SSW and SW) and (d) Stora Sjöfallet (prevailing wind directions WNW and NW).

The CTs that are most favourable for the occurrence of DWs have different synoptic settings, but they are all associated with the strong cross-mountain pressure gradient (Figure 3). On the other hand, CT 4 is the circulation type with the lowest average number of DWs, due to much weaker pressure gradient compared to the other CTs, hindering the crossing of air over the mountain.

3.3 | Wind direction

The relationship between the local DW wind directions and the CTs is shown in Figure 5. For the three CTs that most favour the occurrence of DWs (3, 7 and 10), the predominant wind direction of DWs in all subdomains is southerly and southeasterly, as expected from the synoptic setting. Namely, in CT 7, the synoptic wind is predominantly from the south or southwest. In this CT, the airflow

is expected to follow the isobars until it approaches the mountains and then be deflected towards the Atlantic due to the pressure gradient. For CT 3, the surface synoptic flow on average splits into two currents: one west of the coast along the Norwegian coastline and the other over the land. In this case, the expected DW directions have a southerly component, ranging from southeasterly to southwesterly. In situations with CT 10, a pronounced southerly to southeasterly flow would be expected on a synoptic scale. As the isobars run locally parallel to the Scandes, there is also a stronger pressure gradient across the mountains, which makes it easier for the air to cross them. For other CTs, DWs in the SE subdomain have more frequently westerly directions than in the other subdomains. The reason for this lies in the orientation of the local orography and its influence on the airflow as well as in the definition of the subdomains. The mountain peaks are predominantly located in the western part of the SE subdomain, with the mountain slopes orientated to

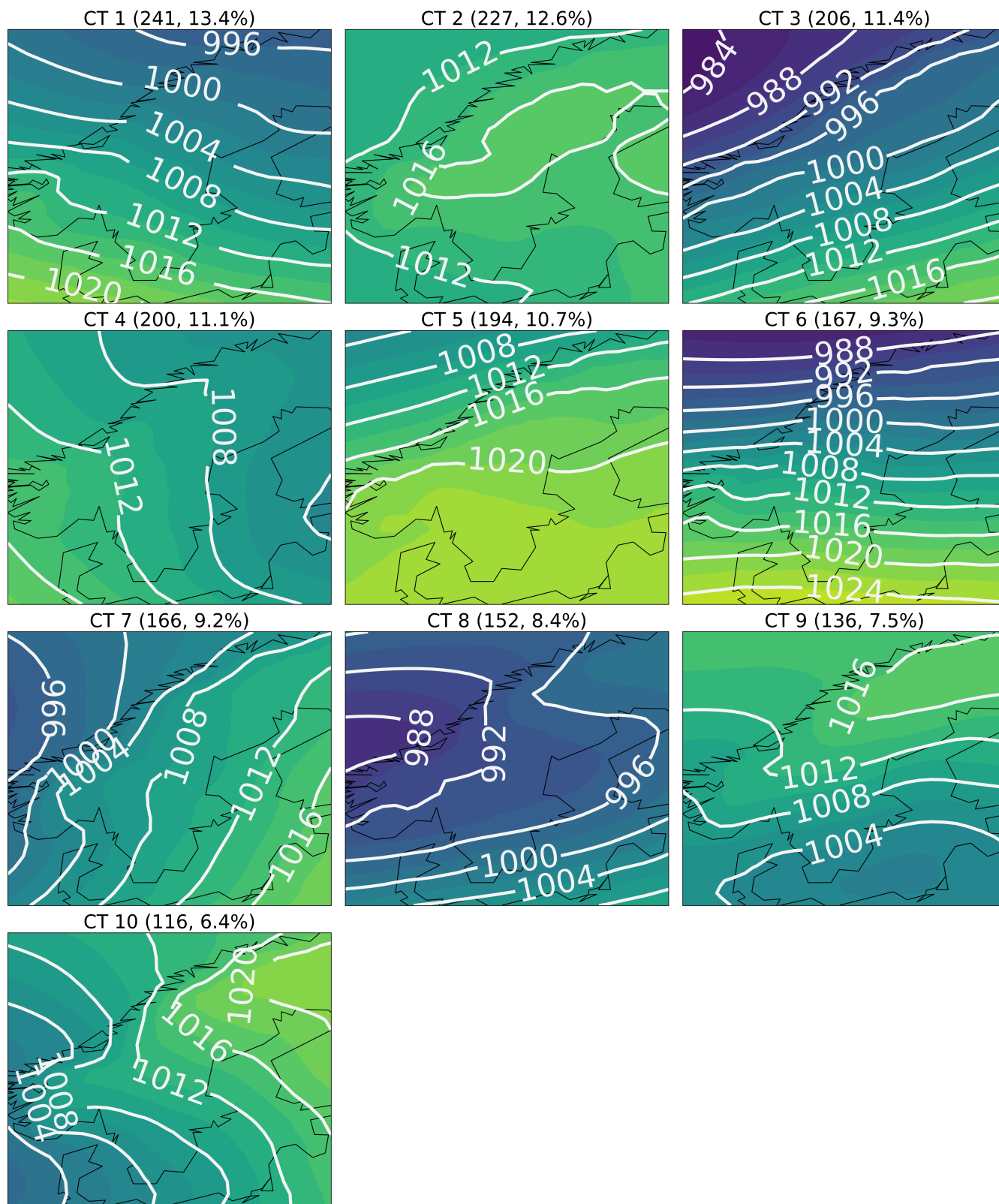


FIGURE 3 Ten circulation types (CTs) over the Scandinavian peninsula for December–January–February (DJF). The isobars and colour show mean sea level pressure (MSLP) in hPa. CT 1 is the most common, and CT 10 the least. The number, and a percentage of days (1998–2017) classified as certain CT are written in brackets above every figure.

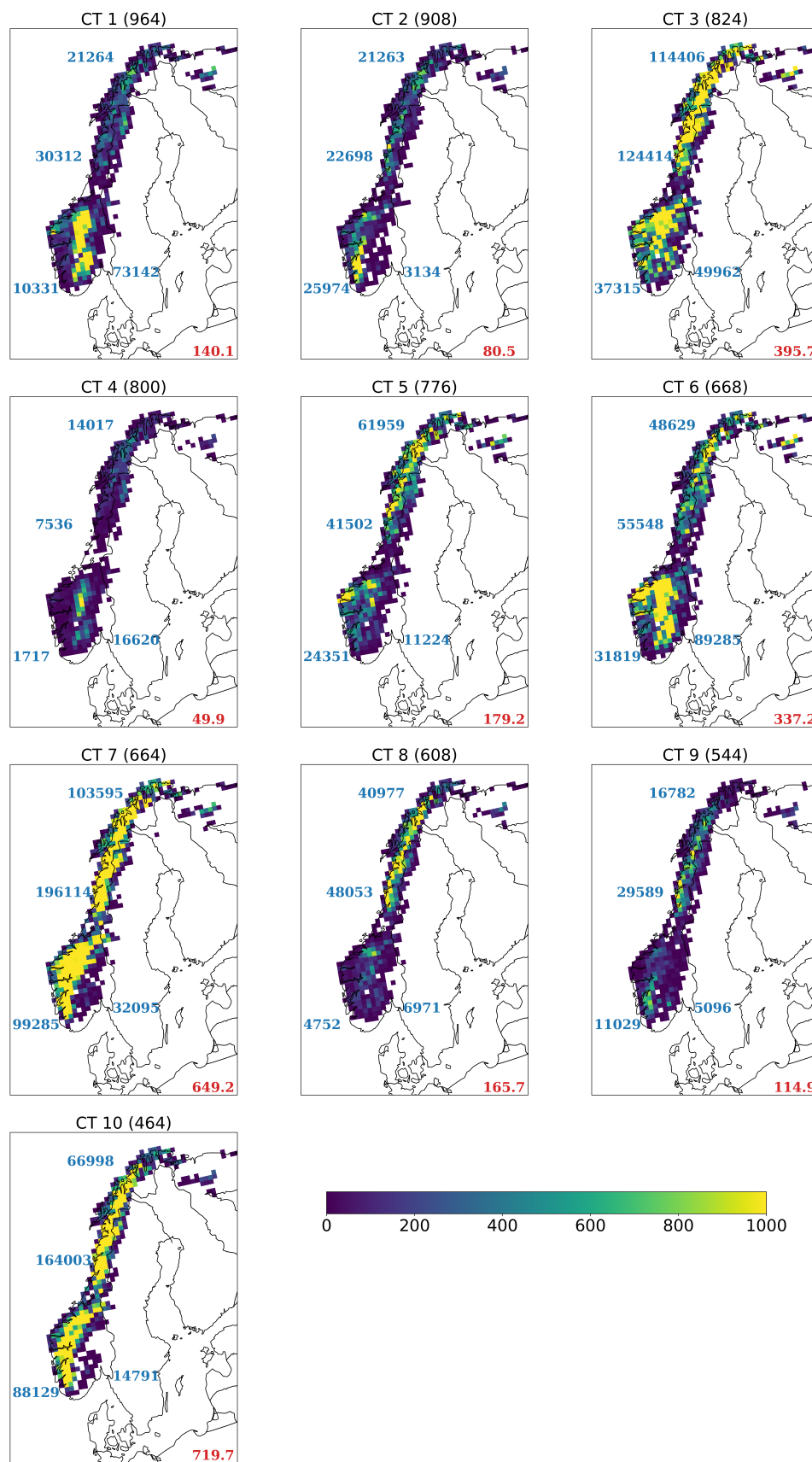


FIGURE 4 Occurrence of downslope windstorms (DWs) in each circulation type (CT) for December–January–February (DJF) in the evaluation simulation (1998–2017). The occurrence is summed over 10×10 grid points (i.e., on a $30 \text{ km} \times 30 \text{ km}$ resolution) for better visualisation. The total number of DWs in 20 years in each subdomain is indicated in blue font, and the average number of DWs per CT (the quotient of the total number of DWs over the entire domain for the CT and the frequency of occurrence of the CT) in red font in the bottom right corner. The frequency of occurrence of a CT (number of six-hour instances classified as certain CT) is indicated in brackets above every figure.

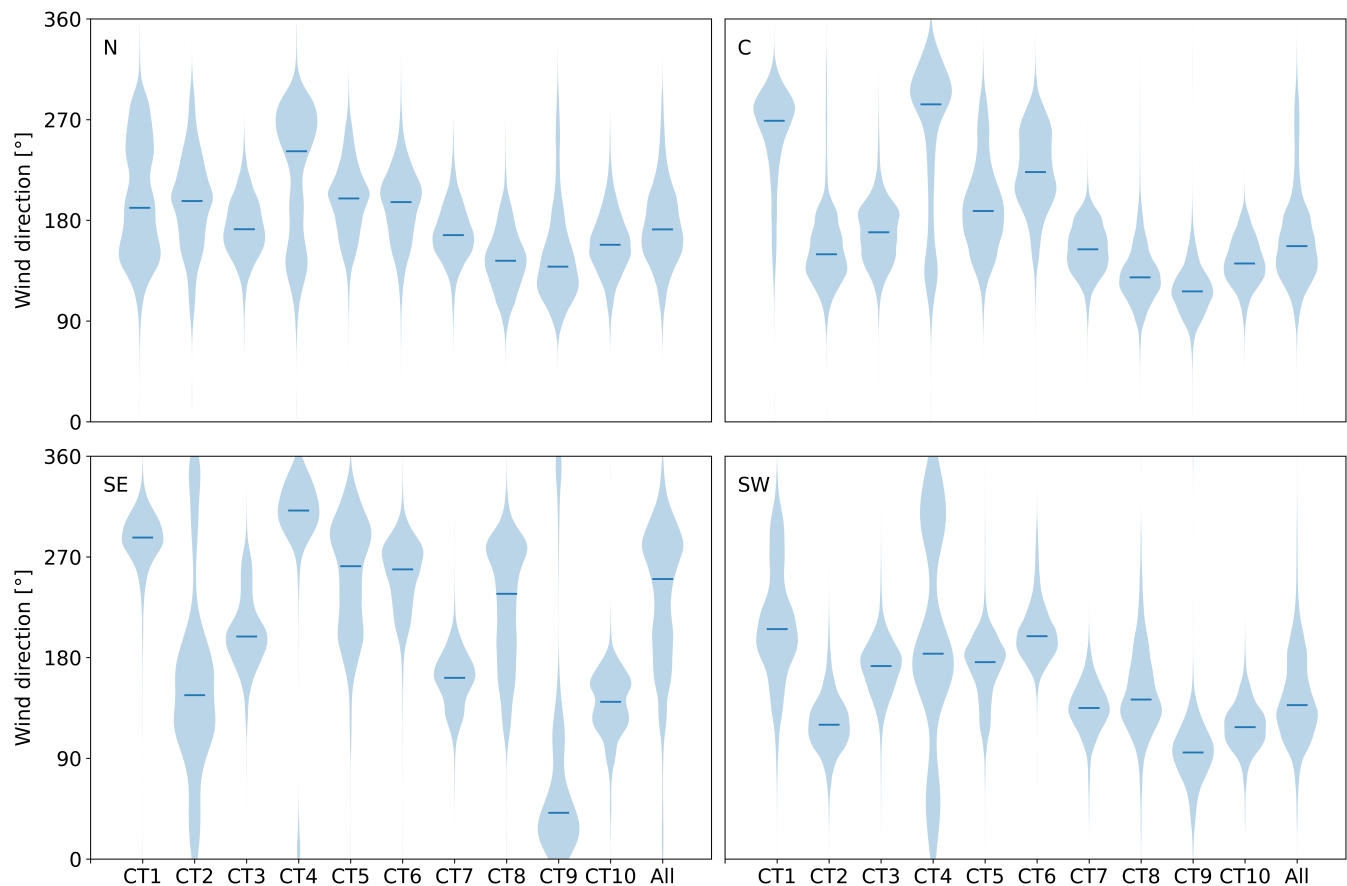


FIGURE 5 Violin plot of downslope windstorm (DW) wind direction in December–January–February (DJF) for every subdomain (N, C, SE, SW) and for every circulation type (CT) in the evaluation simulation (1998–2017).

the east. Therefore, western DWs predominate in the SE subdomain. Likewise, eastern DWs are more common in the SW subdomain.

It can be seen that CT 1 and CT 4 are quite similar in the wind direction violin plots for all subdomains. Looking at Figure 3, they are indeed similar in the synoptic setting above the mountains, except that CT 1 has a stronger pressure gradient.

What also stands out is CT 9, which is the only CT that has northeasterly wind in the SE subdomain. The reason for this is the forcing from prevailing wind in the area because of the pressure field pattern – high pressure above northern Scandinavia and low pressure above the southern Baltic sea.

The reason for generally predominantly southerly DW directions can primarily be found in the prevailing air-flow and the pressure gradient across the mountains, but also in topography (e.g., the asymmetry of the mountain). Previous studies indicate that the slope and asymmetry of a mountain can influence DWs (e.g., Miller & Durran, 1991), and this could be the case here because generally the northwest side of the Scandes is steeper than the southeast side.

Overall, DWs mostly blow from directions with a southerly component, with the exception of the SE subdomain.

3.4 | Temperature change

To analyse the effects of DWs on the local temperature, we calculate temperature change at the DW onset and compare it to temperature change in the gridpoints where DW occurred at least once, but at times when there is no DW onset (non-DW onset). The temperature change at the onset of DW is calculated as the difference in 2-m temperatures between the time of DW onset and the previous time step, which is six hours earlier, with climatological diurnal cycle removed. For non-DW onsets, consecutive six-hour time steps are subtracted. The mean value of temperature change in six hours for non-DW onsets equals $\Delta T = (-0.01 \pm 2.22)^{\circ}\text{C}$ and a median value of -0.07°C . For DW onsets, $\Delta T = (0.83 \pm 1.96)^{\circ}\text{C}$ and a median value of 0.58°C . Due to the large number of data points, this difference is statistically significant (e.g., p-value is about 0.0001 using a *t*-test with 10,000 permutations [e.g., Efron & Hastie, 2016;

chapter 4.4]). Figure S2 shows violin plots for DW and non-DW onsets for every subdomain and every CT. Generally, DW onsets have a tendency to cause local warming. In the Q–Q plot of ΔT for DW onsets versus non-DW onsets (Figure S3), there is a general positive displacement for DW onsets quantiles, but for the lowest quantiles that displacement is more prominent. For example, for ΔT below the 10th percentile, the wind direction histogram (Figure S4) for non-DW onsets shows higher frequency of the northerly component compared to DW onsets. The reason for the DW predominantly southerly component is explained in the previous section and does not pertain to non-DW onsets. Towards the higher percentiles of ΔT , the wind direction histograms are more similar to each other, which explains the better agreement in Q–Q plot for higher percentiles of ΔT , contrary to the lowest 10 percentiles.

The only non-positive medians are for CT 9 (-0.6°C in the SE subdomain and -0.2°C in the SW subdomain) and CT 2 (0°C in the SE subdomain). Additionally, CT 9 in the SE subdomain is the only one where the mean value of DW onsets temperature change is lower than for non-DW onsets. However, all CTs can lead to both cold and warm DWs, and there is no clear distinction between the effects of the different CTs. Additionally, inspection of individual model grid points with DWs showed that most of them have both cold and warm DWs (“mixed” grid points), regardless of the subdomain. This finding differs from other DWs in the world, where a fixed location tends to experience predominantly either a bora (cooling) or foehn (warming) type of DW. Therefore, we selected a grid point that is representative of most of these mixed grid points to analyse the similarities and differences between individual warming and cooling cases.

3.5 | Case studies

An exemplary grid point with both cold and warm DWs is selected for specific case studies. The events shown below are more pronounced (in terms of wind speed and warming or cooling) so that the mechanisms are easier to recognise. Not every grid point has such distinct DWs, but the mechanisms are generally the same. The grid point is at 21.927°E , 69.699°N , located in the northern part of the Scandes, with a model elevation of 731 m and a land fraction of 100%. To the north and northeast of our grid point is a fjord, as can be seen in Figure 1.

During the entire period of 20 winters, there were 176 warm and 226 cold DW events at the grid point with an average duration of 18 hours (three six-hour instances).

The average rate of warming at DW onset is 1.8 K and that of cooling is -1.5 K .

Two cooling and two warming DW cases were selected for further analysis, one of which is mainly synoptically induced and the other more locally generated (Figures 6 and 7).

3.5.1 | Synoptic warming

The first DW case occurred on 9 December 2004 at 0600 UTC and is embedded in a broad synoptic-scale situation (Figure 6a) classified as CT 5. The centre of the cyclone is located northwest of the Scandinavian peninsula and north of Iceland. The occluded front extends along the northwest coast and a dry line runs across the Scandinavian peninsula (not shown). On this day, the sun is below the horizon at the selected grid point for the entire day. The temperature change at the DW onset is 9.2°C and the wind speed is $18\text{ m}\cdot\text{s}^{-1}$. In the southern part of the peninsula, the wind blows predominantly from the southwesterly direction, while in the northern parts it becomes more southerly (Figure 7a). Due to the synoptic conditions (Figure 6a), the wind flows from the Atlantic into the interior of the peninsula in the C subdomain and southern part of the N subdomain. After entering the peninsula, the winds begin to take a more southerly direction due to the pressure field and the influence of the terrain. Since the ocean is warmer than the continent, the air arriving at the grid point produces local warming. A widespread warming of $5\text{--}10^{\circ}\text{C}$ can be observed around the grid point (Figure 7b).

Six hours before the detection of the DW, the near-surface wind around the grid point had been calm, while winds up to 4 km above mean sea level (MSL) did not exceed $8\text{ m}\cdot\text{s}^{-1}$ (not shown). At the onset of the DW (Figure 8a), the winds up to a height of 1300 m above MSL are stronger than $12\text{ m}\cdot\text{s}^{-1}$, being the strongest in the lowermost 500 m with a speed maximum of $21\text{ m}\cdot\text{s}^{-1}$ 50 m above the grid point. The winds are around $4\text{ m}\cdot\text{s}^{-1}$ above 2-km MSL. The drawdown of the isentropes is visible in the lee of the summit, with a hydraulic jump taking place further downstream. The region where the waves break can be recognised by low wind speeds about 500 m above the grid point and higher. In this case the large-scale warm advection and DW dynamics are responsible for the warming.

3.5.2 | Local warming

The second case occurred on 12 February 2001, starting at 0000 UTC. The prevailing wind is cyclonic and the centre of the cyclone is west of the peninsula (Figure 6b).

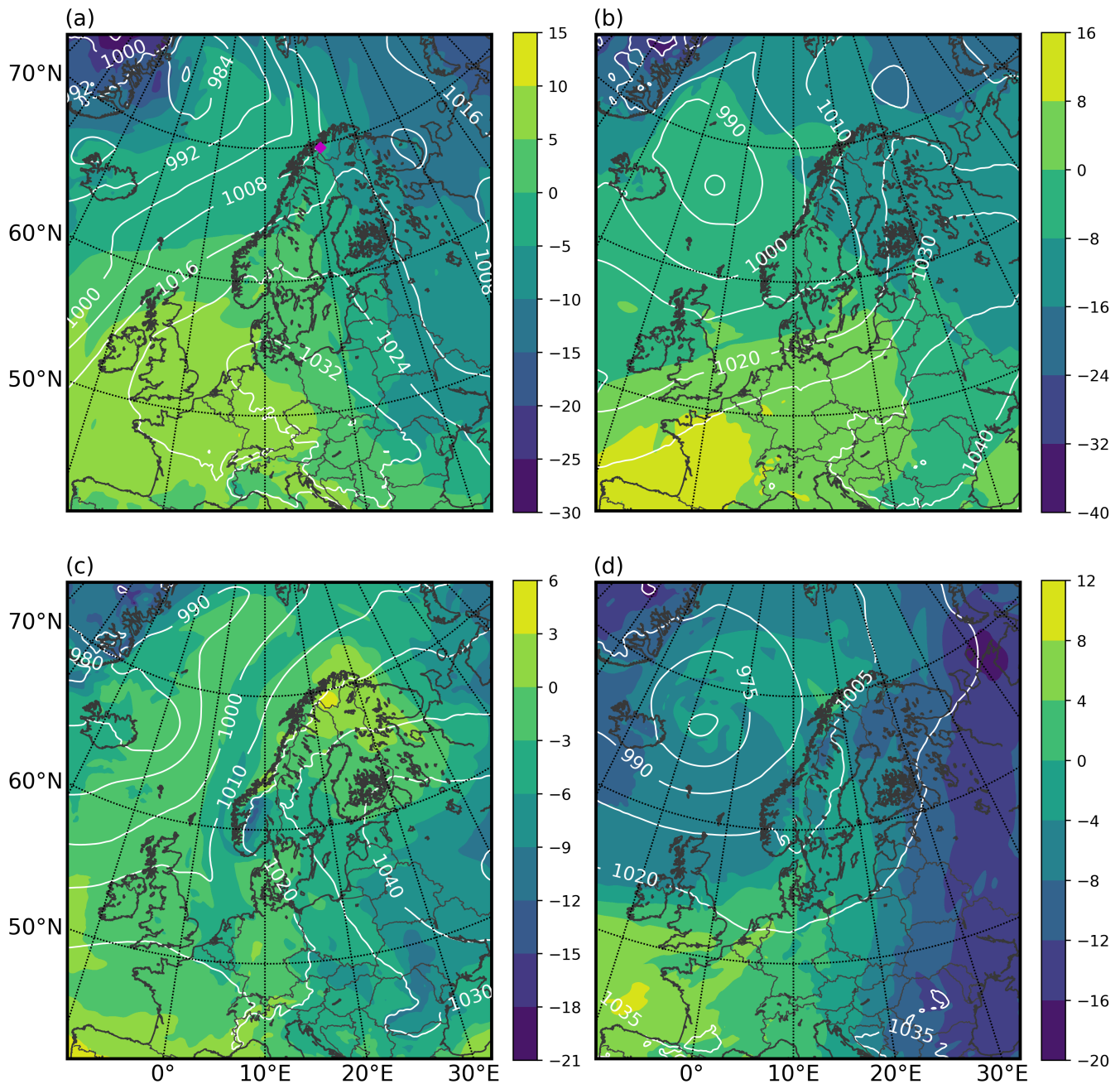


FIGURE 6 Synoptic situations for the four selected downslope windstorm (DW) cases on the HCLIM12 domain: (a) synoptic warming, (b) local warming, (c) synoptic cooling, (d) local cooling. Shown are the 850-hPa temperature (colour, with colour bar in °C) and mean sea level pressure (MSLP) (isolines, 10 hPa interval). The case study grid point is indicated by the magenta circle in panel (a).

These synoptic conditions forced the wind to move across the Scandinavian peninsula from the Gulf of Bothnia into the Norwegian Sea. The wind direction at the grid point is southerly with a small easterly component and the wind speed is 20.6 m s^{-1} (Figure 7c). The synoptic situation is classified as CT 3. The local character of the warming is indicated by the presence of both warming and cooling patches near the selected grid point. Upstream of the grid point there is an area where cooling occurs or the temperature change is very small. The temperature

change at the grid point is 6.7° C in the last six hours (Figure 7d).

The period six hours before the onset of the DW is characterised by weak and variable horizontal winds on the entire vertical section, not exceeding 7.5 m s^{-1} , and only 0.7 m s^{-1} at the grid point – the DW is not yet developed (not shown). During the DW (Figure 8b), an isentropic drawdown is visible in the lee, but it does not reach the ground. The vertical structure is typical for a DW – strong wind over the mountain with the maximum in the lee,

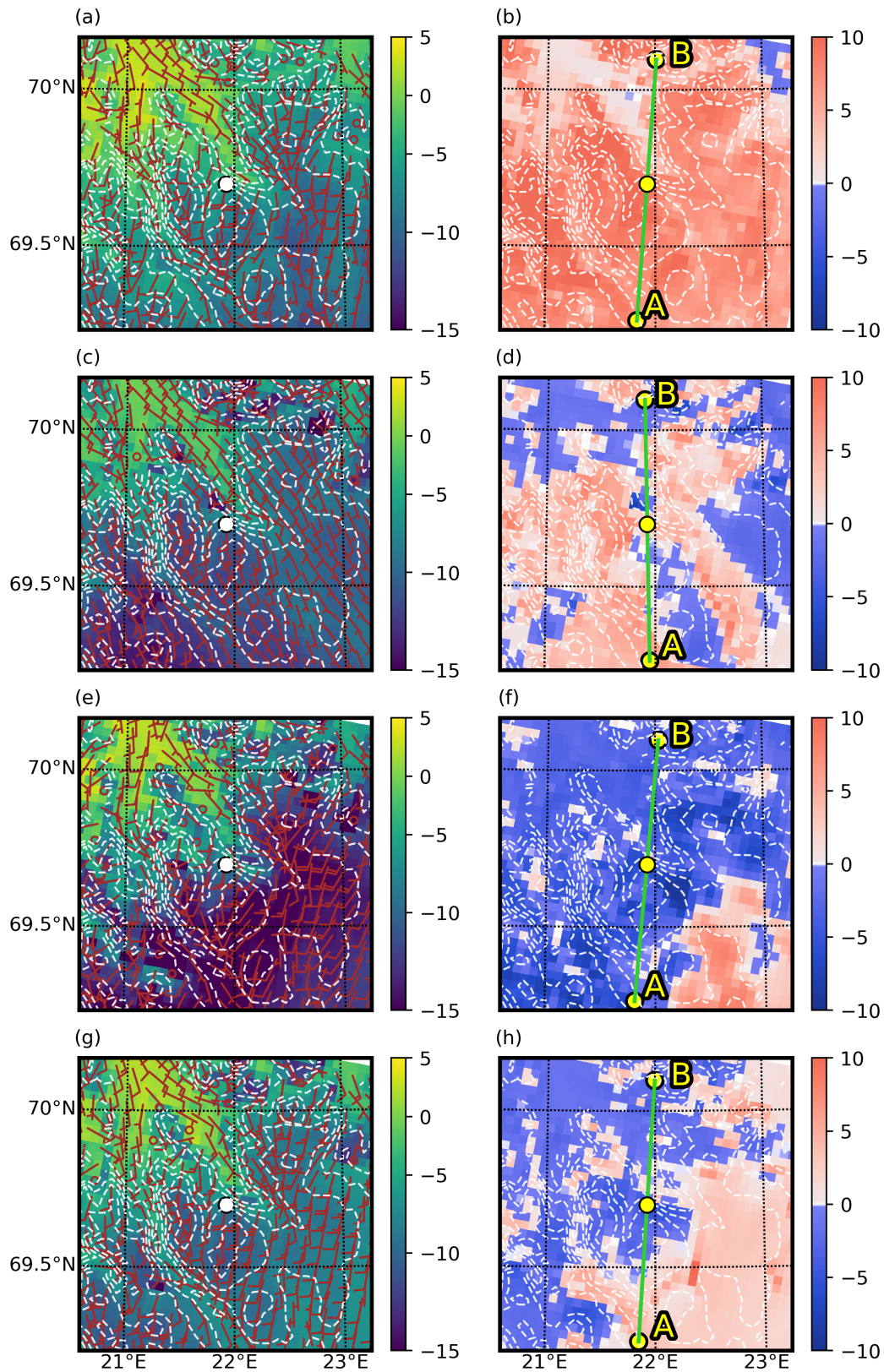


FIGURE 7 Zoomed-in view around the selected grid point (located in the center of each panel) for the four downslope windstorm (DW) cases. Left panels: temperature (colour in $^{\circ}\text{C}$), orography (white dashed lines) and wind barbs (standard increments of $5\text{ m}\cdot\text{s}^{-1}$ for half line, $10\text{ m}\cdot\text{s}^{-1}$ for full line and $50\text{ m}\cdot\text{s}^{-1}$ for a flag). Right panels: the temperature change in the last six hours (colour, with colour bar in $^{\circ}\text{C}$) and orography (white dashed lines). The green lines from A to B indicate the vertical cross-sections in Figure 8. The four DW cases are: (a,b) synoptic warming, (c,d) local warming, (e,f) synoptic cooling, (g,h) local cooling. See Figure 1 for the location and orography.

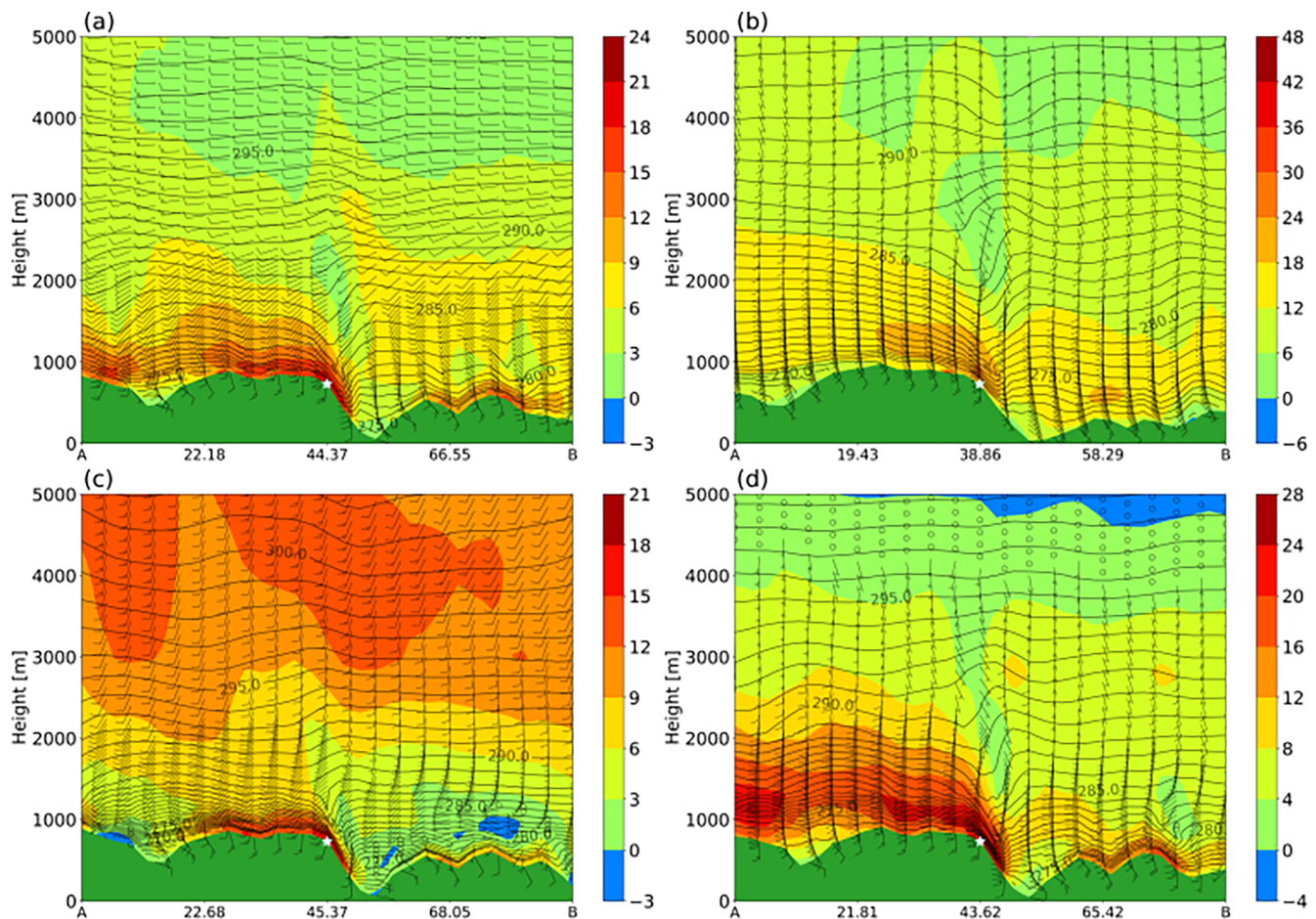


FIGURE 8 Vertical cross-sections from point A to point B for the four selected downslope windstorm (DW) cases (respectively indicated in Figure 7), showing isentropes (isolines, 1-K interval), the wind component along the cross-section (colour bar in $\text{m}\cdot\text{s}^{-1}$) and horizontal wind by wind barbs (standard increments of $5\text{ m}\cdot\text{s}^{-1}$ for half line, $10\text{ m}\cdot\text{s}^{-1}$ for full line and $50\text{ m}\cdot\text{s}^{-1}$ for a flag). The horizontal axis distance is in km from point A. The grid point is marked with a white star. (a) Synoptic warming, (b) local warming, (c) synoptic cooling, (d) local cooling.

isentropic drawdown and a visible hydraulic jump. The vertically propagating gravity waves can be seen in the potential temperature field, and the area where they break about 2 km above the grid point is clearly visible. Since the isentropic drawdown does not reach the surface, and there is no precipitation around the grid point (not shown), the most likely warming mechanism is turbulent mixing (see the listed warming mechanism in the Introduction; Elvidge & Renfrew, 2016).

3.5.3 | Synoptic cooling

In this broader cooling case (Figure 7e,f), which occurred on 17 January 2010 at 1200 UTC, the centre of the cyclone is located near Iceland and a strong anticyclone forms in the east (Figure 6c). The synoptic situation is classified as CT 7. The prevailing winds are southwesterly and turn southeasterly above the Scandes before crossing them. The

DW lowered the local temperature at the grid point by 5.1°C , while the wind speed of the DW is $15.9\text{ m}\cdot\text{s}^{-1}$. The cold air from the interior of the Scandinavian peninsula is forced to the northern, relatively warmer area, as can be seen in Figure 6c in the 850-hPa temperature field. As a result, the entire warm area experiences cooling. At the DW onset, the winds are accelerated in the entire area around the grid point (compared to six hours earlier).

The isentropes in the vertical cross-section (Figure 8c) increase in height and move closer together than six hours earlier, before the onset of the DW (not shown), indicating that cold air is being advected near the surface, as previously explained using the synoptic setting. The wave-breaking region in the lee is very close to the grid point – both horizontally and vertically. The turbulent wake region with weak winds is quite low, at 800 m above MSL (while our grid point is at 731 m). The hydraulic jump is particularly pronounced.

3.5.4 | Local cooling

The synoptic situation for this particular case, which occurred on 18 January 2012 at 1800 UTC (Figure 6d), is similar to the case of local warming (Figure 6b). Both cases are classified as CT 3. The centre of the cyclone is located east of Iceland. An occluded front runs through the of the Scandinavian peninsula, that is, it is located south of the grid point (not shown).

As the centre of the cyclone is west of the peninsula, the wind is driven northwards near the grid point. There are both positive and negative temperature changes near the observed grid point (Figure 7h), but the grid point and the narrow area around it stand out as the cooling area. The temperature change is -5.5°C and the wind speed is $21.4\text{ m}\cdot\text{s}^{-1}$. The wind in the narrow area around the grid point is mainly southerly, with a weak westerly component at the grid point, as can be seen in Figure 7g. A pool of cold air can also be recognised upstream, forced to flow down the slope towards the grid point by synoptic and/or mesoscale mechanisms, similar to the bora wind.

The temporal development seen from the vertical cross-sections indicates that the isentropes rise and become denser upstream of and near the grid point compared to six hours earlier (not shown), implying air cooling. During the DW, there is a visible descent of the isentropes above the grid point and in the lee, as well as a distinct hydraulic jump at about 1 km above MSL near the grid point (Figure 8d). At about 1.5 km above MSL directly above the point there is a wake region with weak winds. The waves propagate up to a height of about 3.5 km, above which they weaken as they approach a critical layer. The isentropes' lifting and descent upstream of and near the grid point creates a horizontal pressure gradient, which causes a fast wind according to Bernoulli's principle. The colder air upstream of the summit is pushed over the mountain by the pressure gradient, resulting in the cold, bora-like, DW.

These four cases illustrate that despite having the same local wind direction, DWs can cause both warming and cooling at the same location. Unlike some other DWs that have more persistent local temperature effects and tend to occur in specific synoptic situations (e.g., Grisogono & Belušić, 2009), here DWs appear under a number of different synoptic and regional conditions. The interaction between synoptic advection, local thermal heterogeneity, specifics of DW dynamics and very complex terrain causes a variety of temperature responses at the DW onset.

3.6 | DWs in future climate

In order to explore potential persistent signals in the future change of DWs, we use the scenario simulations, that is,

those downscaling GCMs, for historical and future periods. The DW occurrence for HCLIM3-ECE in the historical period is similar to the evaluation simulation (cf. Figures 9 and S5), which is consistent with the previously established good performance of EC-Earth for CTs (Brands, 2022; Brands *et al.*, 2023). HCLIM3-GFDL has larger DW occurrence in the historical period, particularly in the northern part of the domain (Figure 9), which has been linked to stronger pressure gradients within certain CTs (not shown).

Over all subdomains, there is a 2.5% increase of DW occurrence in the HCLIM3-ECE run and a 12% decrease in the HCLIM3-GFDL run for the mid-century period, and a 5% and 17% decrease, respectively, for the end-century period. Figure 9 indicates that the occurrence of DWs in the northern regions (N and C subdomains) decreases towards both the middle and end of the century. The decrease is evident for both parent GCMs, but it is larger in the HCLIM3-GFDL simulations, which is the warmer of the two simulations. Conversely, there are more DWs in the SE subdomain in the future climates. The SW subdomain shows a negative change in the HCLIM3-GFDL simulations for both future periods, while the HCLIM3-ECE simulations show an increase in the middle of the century and decrease at the end of the century.

The total future DW occurrence change is decomposed into the change due to the change in a CT frequency and the change within a particular CT (e.g., Barry & Perry, 1973; Kjellström *et al.*, 2022). For the HCLIM3-ECE simulations, the contribution of the total change within the CTs to the final result is larger than the change in the CT frequency, and it defines the sign of the change (Figure 10). For the HCLIM3-GFDL simulations, both parts contribute similarly to the final result and it depends on a subarea whether the first or the second part of change defines the sign of total change (Figure 11). The changes in the frequency of CTs are greater in HCLIM3-GFDL than in HCLIM3-ECE, hence the difference in results.

4 | DISCUSSION AND CONCLUSION

This study analyses DWs over the Scandes using an RCM at two resolutions: HCLIM12 with 12-km grid spacing and HCLIM3 with 3-km grid spacing (CPRCM). DWs are assessed in each grid point using a two-step algorithm based on the local terrain properties and the dynamic and thermodynamic atmospheric structure.

The evaluation of the model in relation to the weather station observations in the present climate shows that HCLIM3 reproduces the observed wind speed data well, while HCLIM12 underestimates the wind speed,

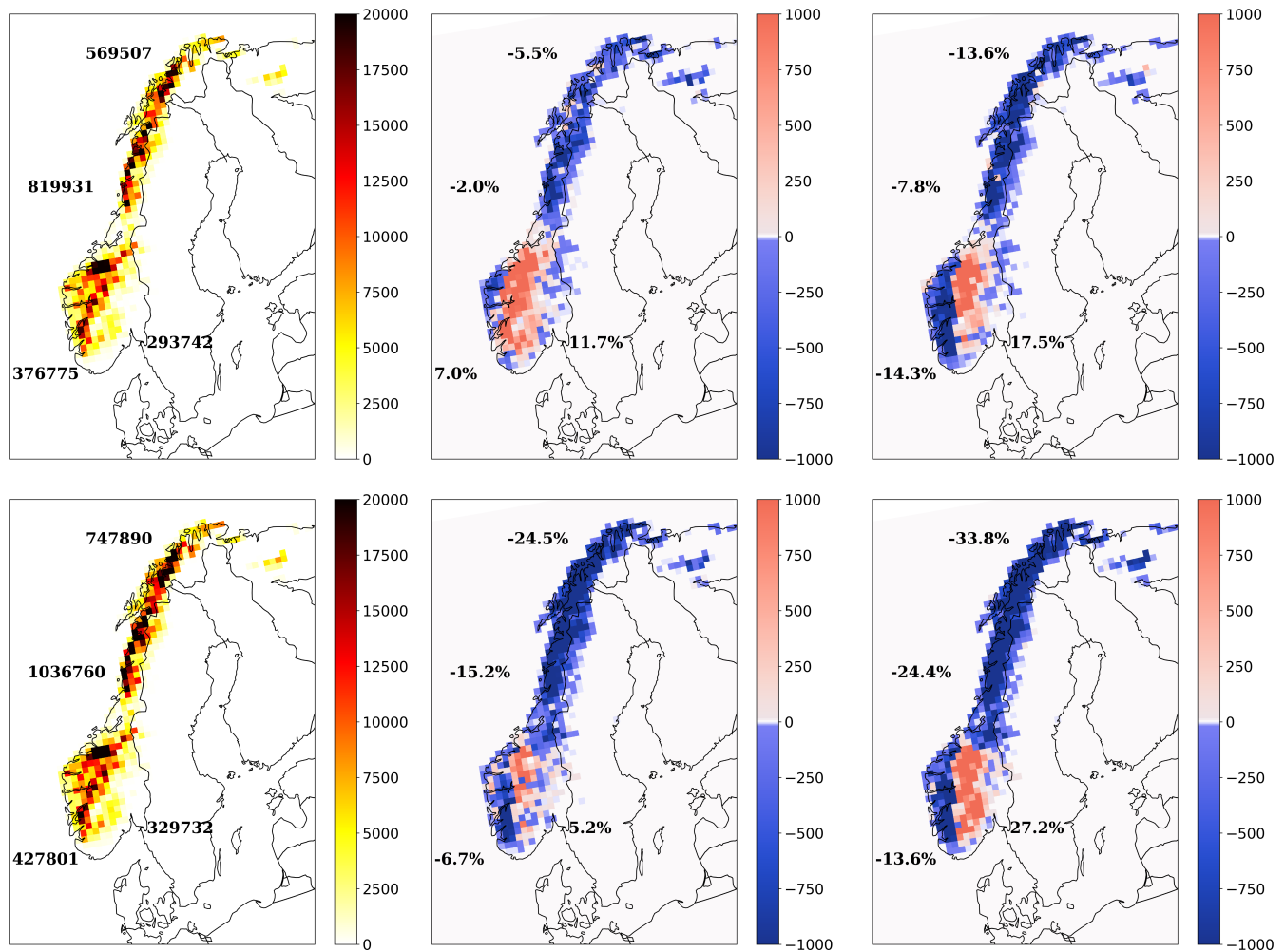


FIGURE 9 Occurrence of downslope windstorms (DWs) for December–January–February (DJF) in the historical period (1986–2005; left), and the change in occurrence from the historical period to the mid-century (2041–2060; middle) and end-century periods (2081–2100; right). The colour bars unit is absolute number of DWs. The occurrence is summed over 10×10 grid points (i.e., on the $30 \text{ km} \times 30 \text{ km}$ resolution) for better visualisation. The total number of DWs in each subdomain is indicated for the historical period, while the future change of DW occurrence is indicated by percentage for every subdomain. Upper panels are for the EC-Earth2 forcing, and lower for the GFDL-CM3 forcing, both for the RCP8.5 scenario.

especially for strong winds. This is consistent with the wind evaluation results in Belušić and Lind (2025), and clearly shows the added value of higher resolution for local wind studies. Some recent studies based on ensembles of CPRCMs report a rather limited benefit of CPRCMs for wind (Molina *et al.*, 2024). The reasons for this discrepancy may be found in one or all of the following: (i) the region of focus is different (Scandes vs an extensive alpine area), (ii) meteorological stations are located in highly complex, mountainous areas, which may influence the assessment results, (iii) Molina *et al.* (2024) used daily wind data, while we use hourly data, and CPRCMs tend to provide larger added value for subdaily data (e.g., Lind *et al.*, 2020; Médus *et al.*, 2022). The latter is consistent with the results obtained over the Adriatic region (Belušić Vozila *et al.*, 2024), which showed that CPRCMs provide

the largest added value for strong winds in complex terrain on a subdaily scale.

Due to their large spatial extent and the resulting different weather dynamics in the different areas, the Scandes are divided into four subdomains for the analysis. We focus on DJF since most DW events occur in this season. To understand the influence of large-scale flow on local DWs, each day is assigned to one of the 10 CTs. The DW wind directions predominantly have a southerly component in almost all subdomains, but they differ depending on CTs. The CT classification groups DWs well in separate predominant directions. The SE subdomain stands out with more westerly and northwesterly events, which is due to the definition of the subdomain. Namely, SE and SW subdomains are separated along the mountain ridge, leaving predominantly eastward-oriented slopes in the SE

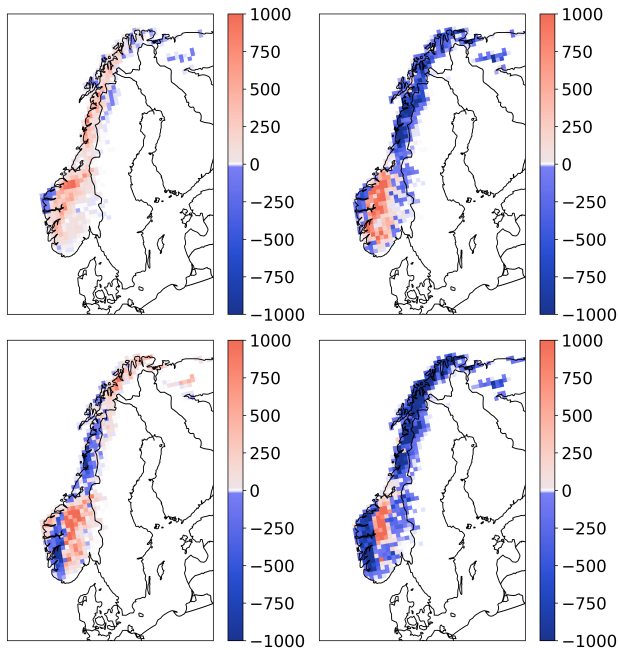


FIGURE 10 Change of occurrence of downslope windstorms (DWs) for December–January–February (DJF) for the HCLIM3-ECE simulations. The change of occurrence is summed over 10×10 grid points (i.e., on the $30 \text{ km} \times 30 \text{ km}$ resolution) for better visualisation. The colour bars' unit is the absolute number of DWs. The left panels show change due to change of frequency of circulation types (CTs), and the right panels show change due to change within CTs. The upper panels are for the mid-century (2041–2060) period, and lower for the end-century (2081–2100), both for the RCP8.5 scenario.

subdomain (and likewise westward-oriented slopes in the SW subdomain).

Since the studied DWs in other parts of the world are locally predominantly either warming or cooling, we analyse the character of the DW by calculating the temperature difference at the beginning of the event. The DWs in most CTs and subdomains have a tendency towards warming, but with a large dispersion and an almost equal frequency of warming and cooling events. The DW onsets are on average significantly warmer than non-DW onsets because DWs have an expressed southerly component, while non-DWs are distributed over all directions, with a pronounced northerly component, especially for lowest percentiles (the coldest cases). The only slightly negative median values for DW onsets are found for CT 9 in the SE and SW subdomains, and only for CT 9 are DW onsets cooler than non-DW onset in terms of the mean value. CT 9 is the only CT that has prevailing wind from the northeast in any of the subdomains. In general, the direction of DWs does not strongly depend on the CT. To understand the different reasons for DW cooling and warming effects, and given the similarity of this result for all subdomains and circulation types, we analysed four specific DW events in a single grid point: synoptic warming, synoptic cooling,

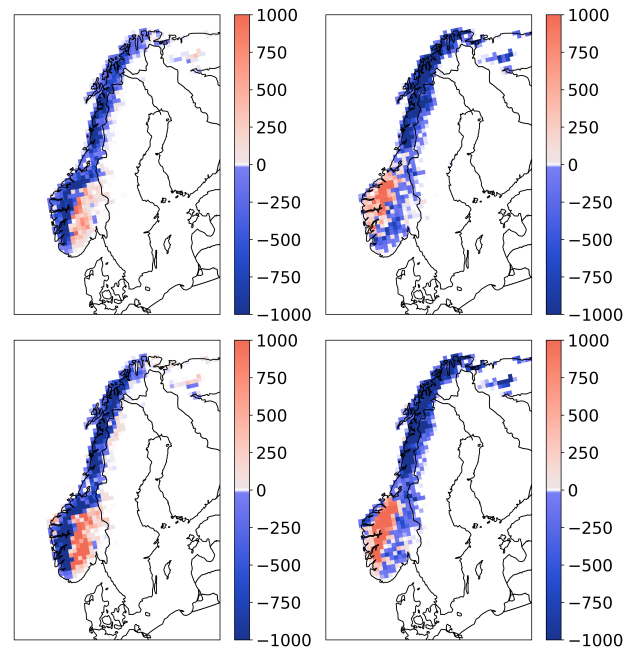


FIGURE 11 As Figure 10, but for the HCLIM3-GFDL simulations.

local warming and local cooling. Each of the four analysed events has typical DW characteristics, but the situations are different. While the synoptic warming is caused by warmer air from the Atlantic, the synoptic cooling is caused by colder air slowly moving from inland towards the ocean. More locally, when a cold air pool is present upstream of a mountain and the synoptic conditions are favourable, the bora-like DW occurs. Warming effects such as isentropic drawdown and turbulent mixing near the mountain top are present in some situations and can be responsible for a warmer DW. In general, the influence of topography on the atmosphere is difficult to predict in such complex terrain.

Future simulations with two GCMs that span a wide uncertainty range of the CMIP5 RCP8.5-scenario future changes, EC-Earth2 and GFDL-CM3, provide insights about the uncertainty. Both GCMs show a decrease in the overall number of DW events in the Scandes towards the end of the century. The decrease is present in all but the SE subdomain. The most likely culprit is the previously noticed dipole change in the large-scale DJF wind speed in the Scandes, with a decrease in the northern parts and increase in the southern parts (e.g., Martinez & Iglesias, 2024; McInnes *et al.*, 2011; Reyers *et al.*, 2016). This change pattern is particularly visible in HCLIM3-GFDL.

Despite the same sign of change in both GCM forcings, the decomposition into internal change within CTs and the change of CT frequency indicates that different mechanisms are responsible for the change. For HCLIM3-ECE, the prevailing influence is from the

internal change of CTs, while for HCLIM3-GFDL the CT frequency change is more pronounced. Another anomaly is the SW subdomain's increase in the occurrence of DWs in the mid-century HCLIM3-ECE run, which is mostly due to the strong within-CT change.

To conclude, we analyse the complexity of DW characteristics in the Scandes and show the need for using CPRCM to reproduce DWs in complex terrain. Unlike some known DWs in other mountainous areas that have a persistent warming (e.g., foehn) or cooling (e.g., bora) effect, or prevailing DW direction, DWs in the Scandes show no such regularity, which is mostly due to very complex terrain and the highly variable synoptic conditions, with frequent passage of new low-pressure systems. As this study focuses on the winter season, the projected changes in DWs do not account for interseasonal variations or changes in DW intensity, which will be investigated in future work. To further study DWs or local winds in the Scandes, even higher resolution could be used to account for the local terrain and effects such as local heating and turbulence. The methodology used here could be applied to other mountainous regions to analyse present and future climate of DWs, and compare the complexity of DWs in different regions.

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DATA AVAILABILITY STATEMENT

The evaluation simulations from the NorCP project are available from the NIRD Research Data Archive at <https://doi.org/10.11582/2025.00046> (The HARMONIE Climate community, 2025a) and <https://doi.org/10.11582/2025.00064> (The HARMONIE Climate community, 2025b). The GCM-forced simulations are planned to be published on NIRD in the near term. In the meantime, these data are available upon request.

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