Projected Changes in European and North Atlantic Seasonal Wind Climate Derived from CMIP5 Simulations

KIMMO RUOSTEENOJA, TIMO VIHMA, AND ARI VENÄLÄINEN

Finnish Meteorological Institute, Helsinki, Finland

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ABSTRACT

Future changes in geostrophic winds over Europe and the North Atlantic region were studied utilizing output data from 21 CMIP5 global climate models (GCMs). Changes in temporal means, extremes, and the joint distribution of speed and direction were considered. In concordance with previous research, the time mean and extreme scalar wind speeds do not change pronouncedly in response to the projected climate change; some degree of weakening occurs in the majority of the domain. Nevertheless, substantial changes in high wind speeds are identified when studying the geostrophic winds from different directions separately. In particular, in northern Europe in autumn and in parts of northwestern Europe in winter, the frequency of strong westerly winds is projected to increase by up to 50%. Concurrently, easterly winds become less common. In addition, we evaluated the potential of the GCMs to simulate changes in the near-surface true wind speeds. In ocean areas, changes in the true and geostrophic winds are mainly consistent and the emerging differences can be explained (e.g., by the retreat of Arctic sea ice). Conversely, in several GCMs the continental wind speed response proved to be predominantly determined by fairly arbitrary changes in the surface properties rather than by changes in the atmospheric circulation. Accordingly, true wind projections derived directly from the model output should be treated with caution since they do not necessarily reflect the actual atmospheric response to global warming.

1. Introduction

Global warming resulting from the increasing concentrations of greenhouse gases affects the geographical and seasonal distributions of multiple additional climate parameters, including atmospheric circulation (Collins et al. 2013). The objective of the present work is to scrutinize future changes in the geostrophic winds over Europe and the North Atlantic Ocean, considering the temporal means, extremes and, in particular, the joint distribution of the wind speed and direction. Projections are derived from the output of 21 global climate models (GCMs) participating in phase 5 of the Coupled Model Intercomparison project (CMIP5).

In addition to the modeled geostrophic winds, we discuss complexities related to the near-surface true

wind projections. The true winds are strongly affected by nonmeteorological geographical factors, particularly the surface roughness. Therefore, model-derived wind speed projections do not necessarily represent the pure climate change signal. Geostrophic winds are far less sensitive to such complications (e.g., Feser et al. 2015).

Surface measurements reveal a general decreasing trend in the wind speeds in Europe since 1979, but this trend has been ascribed to an increasing roughness of the surface due to vegetational changes (Vautard et al. 2010). According to the review paper of Feser et al. (2015), storm activity in northern Europe and the North Atlantic area has fluctuated substantially on an interdecadal time scale without any clear long-term trend. Brönnimann et al. (2012) reported an increasing trend in the occurrence of wintertime high-speed winds in central Europe and the ocean areas to the west between 1950 and 2008; elsewhere in Europe and the North Atlantic area, the trends were in general not statistically significant. The conflicting trends disclosed in the various studies may result from the different time spans examined, owing to the widespread increase in wind speeds before the 1990s and a decrease thereafter

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Corresponding author: Kimmo Ruosteenoja, kimmo.ruosteenoja@fmi.fi

(Brönnimann et al. 2012). Consistently, the maximum windstorm damages in northern and western European forests occurred particularly in the 1990s (Gregow et al. 2017).

To give some examples of recent observational studies, wind speeds have shown a negative trend in Finland (years 1959–2015) throughout the year (Laapas and Venäläinen 2017) and in southwestern Germany (years 1974–2013) in spring and summer (Kohler et al. 2018). In the Mediterranean area during 1979–2014, by contrast, the observational trends have been spatially divergent (Soukissian et al. 2018).

Future projections of the wind climate given in a multitude of previous studies have been summarized in the review papers of Feser et al. (2015) and Mölter et al. (2016). In many model simulations, the North Atlantic storm track is projected to extend farther eastward, leading to increasing storminess in western and central Europe. Conversely, the southern European wind climate is stalling.

Zappa et al. (2013b) reported a general future decrease in the number and intensity of extratropical cyclones over the North Atlantic Ocean and its adjacent continental areas, with the exception of northern central Europe in winter and the areas surrounding Greenland in summer. Decreasing numbers of strong and explosively deepening cyclones were also uncovered by Chang (2018) and Seiler and Zwiers (2016), even though according to the latter paper such cyclones would intensify on the northern edge of the storm track.

Accordingly, future changes in the true and geostrophic wind speeds and related changes in cyclone climatology have been studied extensively, but potential changes in the directional distribution have been addressed less comprehensively (Pryor et al. 2012). Even so, there are numerous applications for which information about the wind direction is essential, and in general the importance of the directional distribution is emphasized under strong wind speeds. For example, the fetch and directional spectrum of sea surface waves are determined by the wind speed and direction (e.g., Kahma and Pettersson 2005; Pettersson et al. 2010). Winds likewise affect the spatial distribution of the sea level rise (e.g., Johansson et al. 2014), the occurrence of coastal storm surges (e.g., Gaslikova et al. 2013; Mäll et al. 2017; Särkkä et al. 2017), coastal upwelling (Goubanova et al. 2011; Alvarez et al. 2017), accumulation of pack ice on downwind coasts (Ronkainen et al. 2018), and opening of coastal polynyas during offshore winds (Stössel et al. 2011).

In land areas, forests downwind of recently clear-felled areas are highly vulnerable to wind damages (e.g., Peltola et al. 1999). Moreover, winds influence the

spread of wildfires (Coen 2018) and the transport of fire-generated smoke and other air pollutants (Anttila et al. 2008). The speed and direction of the wind are likewise of key importance for the spatial distribution of snow accumulation in mountain areas (Minder 2010) and the drift of convective snowfall from watersheds into land areas (Notaro et al. 2015).

Further, both the speed and direction of the wind determine the conditions of wind power production. The potential power production depends very nonlinearly on the wind speed (Sempreviva et al. 2008) and, as wind speeds in the atmospheric boundary layer are typically stronger over surfaces with a small aerodynamic roughness length, the distribution of wind directions has to be taken into account when planning the placement of wind power plants (Tammelin et al. 2013). For example, in a region where westerly winds prevail, coasts facing west are optimal.

The frequencies of the various wind directions can also be utilized in studying climate conditions during the preinstrumental era. For example, Mellado-Cano et al. (2018) scrutinized wind reports in ship logbooks to figure out weather conditions that prevailed in Europe during the coldest phase of the Little Ice Age around 1700.

Future changes in the directional distribution of winds have been studied primarily for the North Sea area where the frequency of strong westerly winds is projected to increase (de Winter et al. 2013; Ganske et al. 2016). Pryor et al. (2012) explored changes in the directional distribution over Scandinavia, but their analysis was based solely on a single model simulation; no drastic changes were reported. McInnes et al. (2011) examined changes in the time-mean wind vector, which in southern Europe appeared to rotate clockwise in winter and anticlockwise in summer.

In this paper, section 2 introduces the model runs utilized in the analyses; also, the modeled geostrophic winds are compared with corresponding estimates derived from reanalysis data. For a background information for the following more detailed analyses, section 3 briefly examines the time-mean responses of the geostrophic wind speed and its components, paying attention to the robustness of the responses in terms of intermodel agreement. Changes in the extreme geostrophic wind speeds are considered in section 4. Section 5 deals with the joint changes of the speed and direction of the geostrophic wind, quantified by analyzing the bivariate frequency distributions of the zonal and meridional component. Challenges related to the near-surface true wind speed projections are examined in section 6. The findings are discussed in section 7 and the key conclusions summarized in section 8.

TABLE 1. Global climate models explored, showing the model acronym and the country of origin. EC-EARTH has been developed by a consortium of European countries. The remaining columns state the number of parallel runs available for the historical period and the RCP4.5 and RCP8.5 scenarios, separately for the geostrophic (N^g) and true winds (N^t). (Expansions of acronyms are available online at <http://www.ametsoc.org/PubsAcronymList>.)

Model	Country	N^g_{hist}	$N^g_{4.5}$	$N^g_{8.5}$	N^t_{hist}	$N^t_{4.5}$	$N^t_{8.5}$
MIROC5	Japan	5	3	3	5	3	3
MIROC-ESM	Japan	3	1	1	3	1	1
MRI-CGCM3	Japan	1	1	1	1	1	1
BCC-CSM1.1	China	1	1	1	1	1	1
INM-CM4	Russia	1	1	1	1	1	1
NorESM1-M	Norway	1	1	1	-	-	-
HadGEM2-ES	United Kingdom	4	4	4	1	1	1
HadGEM2-CC	United Kingdom	3	1	3	3	1	3
MPI-ESM-LR	Germany	3	3	3	3	3	3
MPI-ESM-MR	Germany	3	3	1	3	3	1
CNRM-CM5	France	1	1	1	1	1	1
IPSL-CM5A-LR	France	5	4	4	5	4	4
IPSL-CM5A-MR	France	1	1	1	3	1	1
CMCC-CM	Italy	1	1	1	1	1	1
CMCC-CMS	Italy	1	1	1	1	1	1
GFDL CM3	United States	2	2	1	5	3	1
GFDL-ESM2M	United States	1	1	1	1	1	1
NCAR-CCSM4	United States	3	3	3	—	—	—
CanESM2	Canada	5	5	5	5	5	5
ACCESS1.0	Australia	1	1	1	—	—	—
EC-EARTH	Europe	5	4	3	—	—	—

Most of the previous studies have explored anticipated changes in wind climate on the annual-mean level only (e.g., de Winter et al. 2013; Kumar et al. 2015; Ganske et al. 2016; Seiler and Zwiers 2016) or by focusing on one or two individual seasons (e.g., McInnes et al. 2011; Zappa et al. 2013b; Chang 2018). In the present paper, by contrast, projections are given separately for all four seasons. Such a seasonally specific analysis entails significant benefit since the projected changes turn out to depend strongly on season. Moreover, in several of the applications listed above, the importance of the wind direction is highest for strong winds. It is therefore of particular interest to investigate the entire bivariate distribution of the wind speed and direction rather than merely to look at changes in the time-mean wind vector or its components.

As another focal point of the work, we adduce caveats concerning the interpretation of the near-surface true wind projections. It will be shown that in some GCMs the true wind speed response is actually determined by changes in the surface roughness in the course of the model run rather than changes in the atmospheric circulation.

2. Climate model data

a. Climate models and processing the data

The wind climate projections analyzed in the present study are founded on 21 CMIP5 GCMs (Table 1). The

CMIP3 and CMIP5 GCMs have been used widely for studying future wind conditions (e.g., McInnes et al. 2011; Gregow et al. 2012; Kumar et al. 2015; Chang 2018). The models utilized in this work have previously been found to be able to simulate European recent past climate reasonably well (Luomaranta et al. 2014). We only included those models for which the sea level pressure and near-surface temperature data are available at a daily level. Furthermore, to avoid giving excessive weight for any individual modeling center, no more than two model versions have been incorporated from any center.

Geostrophic winds ($\mathbf{V}_g$) were calculated from the modeled daily sea level pressure (p_s) and 2-m air temperature (T_{2m}) fields by using the standard definition:

$$\mathbf{V}_g = (u_g, v_g) = \frac{RT_{2m}}{2\Omega \sin\phi p_s} \hat{\mathbf{k}} \times \nabla_z p_s, \quad (1)$$

where u_g and v_g stand for the zonal and meridional components of $\mathbf{V}_g$, R is the gas constant of air, Ω the rotation angular speed of Earth, and ϕ the latitude. Note that $\mathbf{V}_g$ was calculated on the native grid of each individual GCM.

The availability of the geostrophic and near-surface (10-m height) true wind velocity data is shown in Table 1. Geostrophic winds were analyzed for all 21 GCMs while the true wind data were available for 17 GCMs. Several models provide data for multiple parallel runs that use

FIG. 1. Seasonal mean geostrophic wind speeds (in m s^{-1}) for the period 1971–2000; averages over the 21 GCMs.

the same greenhouse gas forcing but different initial conditions; parallel runs have been utilized in this work to improve the robustness of the analyses.

The daily scalar geostrophic wind speeds were derived from the component representation (1) by

$$V_g = \sqrt{u_g^2 + v_g^2} \quad (2)$$

and the daily true wind speeds were calculated analogously. Scalar winds speeds approximated by the length of the time-mean wind vector have been widely used when studying future changes in the wind speed and related quantities, applying either a 24-h (e.g., Gastineau and Soden 2009; McInnes et al. 2011; Gregow et al. 2012; de Winter et al. 2013) or 6-h (e.g., Zappa et al. 2013b; Chang 2018) averaging interval. Since the vector components calculated by (1) represent the daily means rather than instantaneous values, (2) tends to underestimate the daily mean when compared to the exact value obtained by averaging the momentary scalar wind speeds. Nevertheless, as the resulting reference-period and future wind speeds are both biased in a similar manner, it is possible to obtain realistic estimates for the change. In section 6b we shall show that the 24-h averaging only induces a negligibly small bias in the scalar wind speed response, given that the response is expressed in percent.

For the components u_g and v_g , no systematic error is engendered, and changes in these quantities can thus be expressed in absolute terms (meters per second).

Future projections are formulated for two representative concentration pathways (RCPs), RCP4.5 representing moderate and RCP8.5 large emissions of greenhouse gases (van Vuuren et al. 2011). To obtain the largest achievable signal-to-noise ratio, the focus is on the response to RCP8.5 over the period of 2070–99. Projections for RCP4.5 are presented in the online supplemental material. Projected changes are calculated with respect to the period 1971–2000, with the exception of wind speed extremes (section 4).

In calculating the time-mean changes, we first averaged the responses derived from the individual parallel runs of each model, only including the equivalent parallel runs for the historical and scenario periods. Thereafter, the responses were interpolated onto a common $2.5^\circ \times 2.5^\circ$ grid and multimodel means were calculated by giving an equal weight for all GCMs. The seasonal multimodel-mean geostrophic wind speeds for the baseline period, calculated by using the approximations (1) and (2), are shown in Fig. 1.

For performing an initial evaluation of the model data, we surveyed the simulated temporal means for the period 1971–2000 and changes for 2070–99 under RCP8.5 for every GCM run. The geostrophic winds and

FIG. 2. Modeling bias in the temporally averaged sea level pressure (in hPa; shading) and geostrophic wind (in m s^{-1} ; vectors): seasonal differences between the 21 GCM (Table 1) mean and the ERA-40 reanalysis for the period 1971–2000: (top left) December–February, (bottom left) March–May, (top right) June–August, and (bottom right) September–November.

their projected changes proved to be reasonable with the exception of the areas where large orographic variations occur (section 2b). In the true wind speed projections, significant deficiencies were found in some models. This issue will be discussed more profoundly in section 6a.

b. Comparison of modeled baseline period $\mathbf{V}_g$ with reanalysis data

Here, the spatial distributions of the climatological multimodel means of p_s and $\mathbf{V}_g$ are compared with their observational counterparts derived from the ERA-40 reanalyses (Uppala et al. 2005). This reanalysis dataset covers, unlike several more recent reanalyses, the entire baseline period 1971–2000.

The ERA-40 analyses are represented at a 6-h interval. To ensure an explicit comparability with the present GCM-derived geostrophic winds [(1) and (2)], we first determined daily means from the reanalyzed p_s and T_{2m} and used them to calculate $\mathbf{V}_g$. Seasonally averaged differences between the multimodel mean and reanalysis are shown in Fig. 2. In winter, the model-derived p_s is lower than the corresponding observational estimate in a zone extending from the British Isles to central and eastern Europe. This broadly coincides with

the area where Zappa et al. (2013a) report an overestimation in the number of extratropical cyclones in the CMIP5 simulations. In the high-latitude ocean areas where Zappa et al. (2013a) found the simulated cyclonic activity to be underestimated, the GCMs simulate an overly high p_s . In accordance with biases in p_s , geostrophic winds are too westerly south of $\sim 50^\circ\text{N}$ and too easterly farther in the north. Qualitatively similar albeit weaker biases occur in spring and autumn (Fig. 2).

Close to elevated areas, in particular Greenland and the Middle East mountains, the modeled geostrophic winds tend to be excessively strong (see also Fig. 1). A similar feature is apparent near the coasts of Greenland in Fig. 11 of Mioduszewski et al. (2018). In high-lying areas p_s has been calculated through reduction of pressure to the sea level, which involves an extrapolation of the data far below the ground (both in GCMs and reanalyses). Geostrophic winds have been derived from the resulting artificial subterranean pressure fields.

Figure S1 in the online supplemental material displays the seasonal cycle of the components and speed of the geostrophic wind at two grid points, the northern point residing in the zone with an easterly bias and the southern one in the zone with a westerly bias in u_g . At the northern

FIG. 3. Projected multimodel mean change in the time-mean scalar geostrophic wind speed from 1971–2000 to 2070–99 under RCP8.5 in (top left) December–February, (bottom left) March–May, (top right) June–August, and (bottom right) September–November. The contour interval is 2.5%. Areas where more than 80% of the 21 GCMs agree on the sign of change are hatched. The Greenland area where the model-derived geostrophic winds are unreasonable (see Fig. 2) is masked out.

example point, the multimodel mean reproduces the observational seasonal cycle of all three variables rather well. At the southern point the biases are far larger, and for the model-derived u_g the seasonal cycle is incorrect even in qualitative terms. Nevertheless, at that point the observational estimate fits in the uncertainty interval of the GCM simulations.

3. Time-mean responses in the geostrophic wind speed and its components

Projected multimodel-mean seasonal changes in the geostrophic wind speed for the period 2070–99 are shown in Fig. 3. The response is generally fairly modest, in the majority of the domain far smaller than $\pm 10\%$. In relative terms, the changes are much smaller than those anticipated for precipitation, for instance (Collins et al. 2013, Fig. 12.22).

The time-mean V_g is projected to weaken throughout the year over the North Atlantic Ocean between 30° and 40°N and also in the Mediterranean area, particularly in winter. Conversely, the pronounced strengthening of the geostrophic wind speeds apparent in southern Europe in summer is evidently spurious, resulting from spatial variations in orography and the associated below-ground

extrapolation of the surface air pressure. This topic will be discussed in more detail in section 4.

In northern Europe in autumn and central Europe in winter, V_g appears to strengthen by a few percent, albeit in the latter case intermodel agreement is fairly low. Close to 70°N , geostrophic winds weaken in all seasons apart from summer whereas farther in the north they strengthen slightly. In winter in particular, the geographical pattern of change in V_g (Fig. 3) is in a broad agreement with the previous projections for the track density and dynamical intensity of extratropical cyclones of Zappa et al. (2013b).

The westerly component of the geostrophic wind (u_g) strengthens in the North Atlantic area in all seasons (Fig. 4). In autumn the area of a positive tendency likewise covers northern Europe and in winter central Europe. In the southernmost parts of Europe and in the ocean areas to the west, a more easterly flow is projected. These features are in line with previously reported changes in lower-tropospheric jet streams (Woollings and Blackburn 2012; Simpson et al. 2014).

For the meridional component v_g (Fig. 5), the model projections are mainly less consistent than those for u_g . Over the bulk of the European continent, the time-mean

FIG. 4. Projected multimodel mean change (in m s^{-1}) in the zonal component of the geostrophic wind (u_g) from 1971–2000 to 2070–99 under RCP8.5. The contour interval is 0.5 m s^{-1} . For further information, see the caption of Fig. 3.

flow is anticipated to become more northerly than it is at present.

For the moderate-emission RCP4.5 scenario, the geographical pattern of the responses is generally qualitatively very similar to that under RCP8.5 but the amplitude is smaller (Figs. S2–S4). This kind of behavior, termed the pattern scaling approximation, is typical for multiple climatological quantities (Collins et al. 2013, their section 12.4.2).

4. Extreme wind projections without directional separation

In this section we examine future changes in high geostrophic wind speeds, without yet paying attention to the directional distribution. In climate research, the 98th (Schwierz et al. 2010; Brönnimann et al. 2012, and references therein) or 99th (McInnes et al. 2011) percentile of the frequency distribution is commonly used as a threshold for extremity. To assess the robustness of the resulting inferences, we studied here four different percentiles (the 95th, 98th, 99th, and 99.5th) for the geostrophic wind speed. The percentiles were first calculated separately for every GCM and calendar month, considering an extended baseline period 1961–2005 and the projection period 2070–99 under RCP8.5. To further

consolidate statistical significance, the time series of V_g from all the parallel runs available for the period (e.g., for MIROC5, five runs for the baseline and three runs for the scenario period; see Table 1) were aggregated and percentiles were determined from this pooled manifold. As for the long-term means in section 3, the monthly percentiles were finally averaged to obtain seasonal means and then interpolated onto the $2.5^\circ \times 2.5^\circ$ grid to calculate multimodel means.

Projected seasonal changes in the 99th percentile of V_g are shown in Fig. 6. For the other percentiles explored, the changes are qualitatively and in wide areas even quantitatively very similar (Figs. S5 and S6). Moreover, changes in the extreme and time-mean (Fig. 3) geostrophic wind speeds exhibit quite a similar geographical distribution. Accordingly, increases (decreases) in the mean wind speed tend to relocate the entire frequency distribution toward higher (lower) values. If the frequency distribution followed the Weibull distribution and the shape parameter of the density function remained unchanged along with an increase/decrease of the scale parameter [which is proportional to the expectation; see Eq. (1) of Sempreviva et al. (2008)], this deduction would be exactly valid.

With the exception of the areas surrounding the Mediterranean Sea in summer, projected changes in high

FIG. 5. Projected multimodel mean change (in m s^{-1}) in the meridional component of the geostrophic wind (v_g) from 1971–2000 to 2070–99 under RCP8.5. The contour interval is 0.5 m s^{-1} . For further information, see the caption of Fig. 3.

wind speeds are generally modest, ranging from a 10% decrease to a 3% increase. It is noteworthy that, in all seasons, a major part of the study area is characterized by a decrease rather than an increase of extreme geostrophic wind speeds (Fig. 6); the contrast is particularly striking when focusing on areas of high intermodel agreement. In southern Europe, the Middle East, and northern Africa in summer, geostrophic wind projections are contaminated by the reduction of surface pressure onto the sea level. This phenomenon can be illustrated by the Iberian thermal low. In summer, the Iberian Peninsula is projected to warm much more than its adjacent sea areas (Kirtman et al. 2013, their Fig. 11.10). This acts to diminish the computational surface pressure difference between the Iberian highland and the standard sea level, leading to enhanced horizontal gradients in the sea level pressure between the peninsula and the nearby sea areas. The resulting increase in cyclonicity is clearly discernible in the changes of the time-mean u_g and v_g (Figs. 4 and 5, upper-right panels). Accordingly, the drastic increase in the (mostly subterranean) mean and extreme sea level geostrophic wind speeds (Figs. 3 and 6) is an artifact that does not manifest itself in the near-surface true wind speed changes (section 6). In winter, no such disparity between the ocean and land-area warming occurs in southern

Europe, and consequently the geostrophic wind projections are sound.

In the review article of Mölter et al. (2016), storminess was anticipated to increase in central and western Europe and over the North Atlantic Ocean to the west of these areas and decrease in southern Europe and in the Atlantic area north of 60°N . Seiler and Zwiers (2016) and Chang (2018) project a decreasing occurrence of strong and explosive cyclones for the northwestern North Atlantic area. In Feser et al. (2015), an eastward extension of the North Atlantic storm track was reported. The present projections with a modest increase in high wind speeds around 50°N in winter and near 60°N in autumn and the simultaneous decreases elsewhere (e.g., in areas to the east and south of Greenland) are in reasonable agreement with these findings. By studying the 25-, 50-, and 100-yr return levels, Kumar et al. (2015) likewise concluded that changes in high wind speeds will be relatively small.

5. Changes in the joint distribution of u_g and v_g

a. Compiling the frequency distributions

In this section, bivariate frequency distributions are elaborated for u_g and v_g . Prior to that, u_g and v_g are

FIG. 6. Projected multimodel mean change (in %) in the 99th percentile of the geostrophic wind speed from 1971–2000 to 2070–99 under RCP8.5 in (top left) December–February, (bottom left) March–May, (top right) June–August, and (bottom right) September–November. The contour interval is 2.5%. Areas where more than 17 GCMs out of the 21 agree on the sign of change are hatched.

normalized by the baseline-period (1971–2000) monthly mean of the geostrophic wind speed ($\bar{V}_g$):

$$u_g^{\text{norm}}(\phi, \lambda, t, r) = u_g(\phi, \lambda, t, r) / \bar{V}_g(\phi, \lambda, m, r), \quad (3)$$

$$v_g^{\text{norm}}(\phi, \lambda, t, r) = v_g(\phi, \lambda, t, r) / \bar{V}_g(\phi, \lambda, m, r), \quad (4)$$

where u_g^{norm} and v_g^{norm} are the components of $\mathbf{V}_g$ after normalization. The indices ϕ , λ , t , m , and r refer to the latitude, longitude, time, calendar month, and model run, respectively. An idea of the magnitude and spatial distribution of $\bar{V}_g$ can be obtained from the multimodel mean shown in Fig. 1, although in (3) and (4) $\bar{V}_g$ represents the individual model runs.

The objective of normalization is to make the simulated geostrophic winds in the different models, regions, and months commensurate. Without normalization, only those models that simulate the strongest geostrophic wind speeds compared to the remaining GCMs would be represented in the histogram classes of the largest wind speeds. The resulting changes in the occurrence of high wind speeds would then not be representative for the entire model ensemble. Another rationale for normalization is that local ecosystems and human infrastructure have acclimatized to the

prevailing wind conditions (e.g., Brönnimann et al. 2012). Consequently, the threshold wind speed that causes severe impacts depends on the region; after normalization, winds in the various areas are more readily comparable.

The terms u_g^{norm} and v_g^{norm} were both grouped into 16 histogram classes: < -3.5 , $[-3.5, -3.0]$, ..., ≥ 3.5 . For each class, the frequencies were calculated for the tridecadal baseline and scenario periods and transformed into percentages by dividing by the total number of days belonging to the respective month during that 30-yr period; after that, these percent frequencies were averaged over the parallel runs. Finally, the frequency distributions were regridded onto the common $2.5^\circ \times 2.5^\circ$ grid using the nearest-neighbor method and averaged over the 21 GCMs.

b. Projected changes in the frequency distributions

We first examine, as an illustration, the winter joint distribution of u_g^{norm} , v_g^{norm} at a single grid point situated in the southern North Sea between England and the Netherlands (Fig. 7). At that point, light to fresh westerly geostrophic winds ($0 < V_g^{\text{norm}} < 1.5$) are most widespread, the frequency of each histogram class being 5%–9% of all incidences (Fig. 7a). Conversely, high wind speeds with $V_g^{\text{norm}} > 2.5$ are rare and mainly occur in the westerly and

FIG. 7. The bivariate frequency distribution of the u and v components of the normalized geostrophic wind at 52.5°N , 2.5°E in December–February in (a) 1971–2000 and (b) 2070–99 under RCP8.5, and (c) their difference; a mean of 21 GCMs. In (a) and (b), the frequencies are given in percent. In (c), for each pixel of the distribution, the number denotes the change in absolute terms (in percentage points) and the color in percentage terms (see the color bar). For example, the relative frequency of normalized velocities belonging to the class $u_g^{\text{norm}} \in [-2.0, -1.5]$, $v_g^{\text{norm}} \in [-1.0, -0.5]$ (west-southwesterly winds with the normalized speed about 2) has increased by 0.41 percentage points from 1.38% to 1.79%; in relative terms this corresponds to an increase of 30%. Circles around the origin represent the normalized scalar wind speeds of 1, 2, and 3.

southwesterly sectors. As the multimodel-mean geostrophic wind speed at that point in winter is about 13 m s^{-1} (Fig. 1), normalized speeds above 2.5 correspond to dimensional speeds of $V_g > 32 \text{ m s}^{-1}$.

The model-derived frequency distribution of u_g^{norm} , v_g^{norm} is qualitatively similar for the baseline (1971–2000) and future (2070–99) periods (Figs. 7a,b), but in quantitative terms, substantial changes are nevertheless evident (Fig. 7c). The frequencies of westerly geostrophic winds between the directions 190° and 315° increase while the northerly, easterly, and southerly wind directions become less common. In absolute terms (denoted by the numbers in the pixels), the largest increases are projected for the west-southwesterly winds with $0.5 < V_g^{\text{norm}} < 2$. The

proportion of very weak winds with $V_g^{\text{norm}} < 0.5$ decreases in all directional classes. These frequency changes accord with Figs. 3 and 4 (top-left panels), which reveal increases in both the time-mean scalar wind speed and its westerly component for this area.

To more readily discern changes in the histogram classes having a small number of incidences, the responses were likewise considered in relative terms:

$$\Delta f = 100 \frac{f_{\text{sce}} - f_{\text{bas}}}{f_{\text{bas}}}, \quad (5)$$

where f_{bas} and f_{sce} denote the absolute frequencies derived for the baseline and scenario period. The largest

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FIG. 8. Projected changes (in percent) in the bivariate frequency distribution of $(u_g^{\text{norm}}, v_g^{\text{norm}})$ from 1971–2000 to 2070–99 under RCP8.5 at selected grid points in September–November; a mean of 21 GCM. Pixels depicting histogram classes in which more than 80% of all GCMs (≥ 17 GCMS) agree on the sign of change with the multimodel mean are marked by black dots; pixels with $\geq 66\%$ (≥ 14 GCMS) are marked by gray dots. For detailed technical information, see the caption of Fig. 7.

percentage changes are seen in high wind speeds (colors in Fig. 7c). Westerly and southwesterly geostrophic winds with $V_g^{\text{norm}} > 2$ become more common by more than 20%, in some pixels even $>50\%$. Concurrently, moderately strong winds ($V_g^{\text{norm}} \in [1, 2]$) from easterly directions become less frequent by $\sim 20\%$. The most extreme wind speeds, with $V_g^{\text{norm}} > 2$ for the easterly and $V_g^{\text{norm}} \approx 3$ for the westerly sector, are so infrequent that changes in these categories are largely random, showing very large positive and negative relative changes without coherence. Histogram classes with a relative frequency of $<0.006\%$ during the baseline period are flagged by omitting the number denoting the absolute change.

The projected changes shown in Fig. 7 are in fair agreement with the previous model simulations for the North Sea area; these have revealed a general increase in the proportion of westerly winds (Ganske et al. 2016), especially for strong wind speeds (de Winter et al. 2013).

Figures 8–11 show projected seasonal changes in the joint distribution of u_g and v_g at 20 regularly placed grid points covering the entire European area. For clarity, changes are shown in relative terms alone, omitting very sparsely populated classes with baseline-period frequencies $<0.006\%$.

In autumn, at the northernmost latitude (67.5°N) considered, there is a coherent signal (changes are of the

DEC–FEB

FIG. 9. As in Fig. 8, but for December–February.

same sign at neighboring pixels) of an increasing proportion of westerly and a decreasing proportion of easterly geostrophic winds (Fig. 8). The changes are most pronounced among strong winds, up to >50%. A similar but somewhat weaker tendency is apparent at the second northernmost latitude, apart from its easternmost point. Note that these are the very areas where the models agree on a positive trend in the time-mean u_g (Fig. 4, bottom right panel). At the points 47.5°N, 17.5°W and 37.5°N, 12.5°E, the share of weak winds increases (the red/yellow area in the middle of the diagram) and moderate to strong winds become less common regardless of the direction (blue areas in the periphery); in these regions, the mean geostrophic wind speed decreases as well (Fig. 3, bottom right panel). In

western France and over the Aegean Sea, an increasing share of winds blow from the northeast or east, while all the other directions are decreasing (compare with Figs. 4 and 5). At some points in central and eastern Europe, the patterns are fairly diffuse, without a distinct tendency of any particular direction to become more/less frequent.

In winter (Fig. 9), projected changes in the frequency distributions are somewhat more heterogeneous than in autumn. Substantial increases in the occurrence of strong westerly geostrophic winds are seen (in Iceland, southern Scandinavia, and north of the Black Sea, for instance). In northeastern Europe, the proportion of intense northerly/northwesterly winds increases while in the western Mediterranean area southerly or southeasterly

MAR–MAY

FIG. 10. As in Fig. 8, but for March–May.

winds become more frequent. Again, these changes are in general accordance with above-discussed changes in the temporal means of u_g , v_g , and V_g .

In spring (Fig. 10), over the majority of the European continent northwesterly winds are projected to become more prevailing; in the Mediterranean area, southerly to easterly winds will be more likely. In summer (Fig. 11), the large changes apparent in central and southern Europe are evidently artifacts caused by the reduction of pressure in elevated areas (see discussion in section 4). Over the Norwegian Sea, the frequency of westerly winds increases substantially in all speed classes. However, in summer the mean wind speeds are far weaker than in the other seasons (Fig. 1), and consequently this does not imply a frequent occurrence of devastating westerly storms.

In most cases, proportional changes are largest in the categories of strong wind speeds. Furthermore, at very many sites the changes are coherent (i.e., the sign of change is the same in multiple adjacent bins of the bivariate distribution). Accordingly, although changes in the mean and extreme scalar geostrophic wind speed are fairly modest (Figs. 3 and 6), remarkable increases and decreases (in many cases about 50% or even more) are to be anticipated in the frequency of strong winds from particular directions.

At many grid points, the intermodel agreement on the sign of change in the individual pixel frequencies is fairly low. In addition to the histogram classes that show small relative changes in the frequency (light-colored pixels in Figs. 8–11), the agreement is typically weak at the

JUN–AUG

FIG. 11. As in Fig. 8, but for June–August.

fringes of the distribution (i.e., for the strongest wind speeds). In these histogram classes, in individual GCMs the number of occasions is small, and thus the impact of stochastic variations is substantial; in particular, in those 13 GCMs that are represented by only a single parallel run (Table 1). Nonetheless, in the present analysis we have been able to improve the signal-to-noise ratio by normalizing the wind data [(3) and (4)] and then pooling the frequencies derived from the 21 models before calculating the ultimate frequency distributions. At many grid points this has resulted in responses in which changes in the adjacent histogram classes are coherent. In the future, the intermodel agreement can potentially be improved if the subsequent model generations provide larger numbers of parallel runs per a GCM.

Bivariate projections of u_g and v_g under RCP4.5 (Figs. S7 and S8) are in most cases qualitatively similar to those under RCP8.5, but the magnitude is lower and the coherence and intermodel agreement weaker.

We made a trial to find out whether changes in the bivariate distributions of (u_g, v_g) might be explained simply by changes in the time-mean wind vector. For that purpose, we first subtracted the vectorial change (2070–99 minus 1971–2000) in the time-mean geostrophic wind from the daily u_g and v_g data of the period 2070–99, separately for every month and model run. The resulting modified frequency distributions for the scenario period were then used to calculate changes from the baseline frequencies. A comparison between the unmodified and modified responses at three example

grid points is shown in Fig. S9. While the responses in the frequency distributions are notably weaker when using the modified wind data for the future, substantial changes continue to exist. One reason for this is that the pure vectorial change in the mean wind velocity does not properly reveal changes in the scalar wind speed. For example, in Fig. S9d the modified response is indeed only weakly directionally dependent, but the negative trend in the mean wind speed still manifests itself as increasing (decreasing) frequencies in histogram classes representing low (high) wind speeds.

It should be emphasized that in most cases discerning the very large relative changes in the frequency of high geostrophic wind speeds is only viable when inspecting the various winds directions separately. Without directional segregation, proportional changes in the frequencies of strong wind speeds are generally smaller (disregarding the areas surrounding the Mediterranean Sea in summer where the large increases are evidently spurious), although nonnegligible increases of $>20\%$ occur in northern central Europe in winter, for instance (Fig. S10).

For clarity, in Figs. 8–11 changes in the joint distribution are given only for selected grid points. The reader can independently produce bivariate distributions analogous to those in Fig. 7 for an arbitrary point of the $2.5^\circ \times 2.5^\circ$ grid by using the data file and Grads script that are available from <https://etsin.fairdata.fi/dataset/087f9d18-5196-4e6e-83a0-dfe2d94c5aa6> or from the authors by request.

6. True wind speed changes

a. Issues related to near-surface true wind projections

The geostrophic wind vector is determined, by definition, purely by the prevailing meteorological conditions, primarily by the synoptic- and planetary-scale distribution of the sea level pressure. Near-surface true winds, by contrast, are strongly affected by the surface roughness and boundary layer stratification. For instance, Vautard et al. (2010) noticed a widespread decreasing trend in the observed wind speeds in Europe, which could not be explained by changes in the atmospheric conditions. They attributed the trend to afforestation and other recent changes in the vegetation cover.

To exemplify a dissimilar model response in the true and geostrophic wind speed, Fig. 12 displays changes in the two quantities inferred from a simulation performed with the GFDL-ESM2M model. In ocean areas, the projections for both quantities are roughly similar, but over the European continent large differences are evident. The geostrophic wind speeds are simulated to increase or decrease by a few percent at maximum. Conversely, the temporally averaged true wind speeds

appear to change intensely, the response ranging from a 15% decrease in western Russia to a more than 30% increase in southern Finland.

In the CMIP5 archive, no data were available for surface roughness. In GFDL-ESM2M, however, the height of vegetation is parameterized to vary purely as a function of plant biomass (Dunne et al. 2013). We therefore chose to study the mass of carbon in vegetation as a proxy for the roughness length. Figure 12 indicates that the vegetational carbon diminishes drastically (at 62°N , 26°E , by more 80% of its original amount) exactly in the area where the simulated increase in the near-surface true wind speed is largest. In fact, the vegetation carbon mass and wind speed changes are nearly perfect mirror images over the entire European continent, down to the small-scale spatial details. This strongly indicates, that in this model, simulated changes in the wind speed are not primarily induced by changing meteorological conditions but rather result from modeled trends in the parameterized vegetation and surface roughness.

In Earth system models (ESMs), vegetation is permitted to adjust to the prevailing climatic conditions, and thus it evolves divergently in the different models (Flato et al. 2013; Dunne et al. 2013). Furthermore, changes in vegetation are caused by anthropogenic land-use changes. Although the land-use scenarios employed in the various models follow a common harmonized dataset, their implementation and the resulting changes in the vegetation cover vary substantially across the models (Hurtt et al. 2011). Accordingly, the fidelity of any single model in predicting regional vegetational changes should be deemed rather low. The same inference applies to the accompanying changes in true wind speed.

In some other models, the discrepancy between the geostrophic and true wind speed responses is even more severe than in GFDL-ESM2M. The MIROC-ESM model simulates true wind speed decreases of 50%–70% (Fig. S11) and MRI-CGCM3 increases of 100%–240% (Fig. S12) for large areas of Europe. In both models, the geostrophic wind speed response is gentle.

Data for the vegetation carbon mass were not available for all the models examined (e.g., the data were absent for MRI-CGCM3), and consequently we restricted ourselves to study the evolution of this quantity in eight models originating from different centers. In all these models, the carbon mass varied substantially during the model run. Nevertheless, in models other than GFDL-ESM2M, we could not find any straightforward connection between the evolution of the vegetational carbon and near-surface wind speed. In MIROC-ESM, for instance, the vegetational carbon

FIG. 12. Changes in the temporally averaged annual mean (top left) geostrophic (in %) and (top right) near-surface true wind speeds (in %) and (bottom) the mass of carbon in vegetation (kg m^{-2}) from 1971–2000 to 2070–99 in the simulation of the GFDL-ESM2M model under RCP8.5.

decreases substantially in northern and increases in central Europe, but these changes do not coincide with projected large changes in the true wind speed (Fig. S11). Presumably, in most models the carbon content alone does not account for the principal characteristics of the parameterized vegetation. Besides, the surface roughness is affected by factors other than vegetation, such as urbanization (Li et al. 2018). A more profound analysis of the topic is beyond the scope of the present work. Note, however, that in many other models the near-surface true wind responses were of reasonable amplitude despite substantial trends in vegetation; IPSL-CM5A-MR is given as an example (Fig. S13).

Multimodel-mean annually averaged changes in the true and geostrophic wind speeds are shown in Figs. 13a and 13c. Over ocean areas the distributions are rather similar, with the exception of the Arctic Ocean, where the near-surface true winds intensify more than the

geostrophic ones. Presumably this is due to the retreat of sea ice, which tends to reduce the surface roughness and, above all, destabilize the atmospheric boundary layer, which results in an enhancement of the transfer of momentum from the free atmosphere into the surface layer (Mioduszewski et al. 2018). In ocean areas south of Iceland, the projected relatively weak warming of the surface (Collins et al. 2013, their Fig. 12.11) acts to increase the static stability; the true winds consequently abate somewhat more than the geostrophic winds. In continental areas, the spatial pattern of the geostrophic wind response is smooth, apart from the areas affected by orography near Greenland and the southern edge of the domain. Conversely, for the near-surface true winds the pattern is noisy. In light of the above examples (Fig. 12; see also Figs. S11 and S12), the pattern is very evidently affected by model-specific changes in the surface roughness, despite the smoothing effect of

FIG. 13. Annually averaged 17 model mean change (in %) in the true wind speed from 1971–2000 to 2070–99 under RCP8.5, calculated by two methods: wind speeds derived (a) from the daily means of the u and v components (see section 2) and (b) from the monthly mean scalar wind speeds downloaded from the CMIP5 archive (variable `sfcWind`). (c) Corresponding 21 GCM mean change in geostrophic wind speed derived from the daily u_g and v_g . The contour interval is 2.5%. To facilitate discerning of finescale details, no information about intermodel agreement is included.

multimodel averaging. Note that the above-discussed changes in the lower-atmospheric stability, induced by the sea ice retreat and ocean current changes, can be regarded as genuine components of the changing climate system. In contrast, changes in the land surface roughness in the various models are quite contradictory (e.g., Fig. 12; see also Figs. S11 and S13) and thus largely unpredictable at present.

Figure 14 depicts seasonal projections for the 99th percentile of the near-surface true wind speed. Compared to the corresponding geostrophic pattern (Fig. 6), the distributions are once again fairly similar in ocean areas, apart from high-latitude oceans where extreme true wind speeds increase substantially. Over the European continent, the response is noisy, and the finescale noise has a similar spatial structure in all seasons, even if the large-scale pattern of change is seasonally dependent. The intermodel agreement on these finescale features is low. Moreover, the finescale pattern is similar in the high and time-mean wind response (Figs. 13a and 14). Many geographical details can be traced to the individual model simulations considered in Fig. 12 and Figs. S11 and S12. This supports the above-stated notion

that, in the true-wind response, the contributions of nonclimatological changes in the surface roughness are indeed important.

Kumar et al. (2015) ascertained changes in the extreme true wind speeds (annual maxima and the 25- to 100-yr return levels) using an ensemble of 15 CMIP5 GCMs, partially containing the same models as employed in the present study. In many continental areas, including Europe, the resulting response patterns were rather noisy. It is plausible that the noisiness is contributed by model-specific changes in the surface roughness.

b. Applicability of daily mean model output to calculating wind speed responses

As stated in section 2, in this study daily scalar wind speeds (both for the geostrophic and true winds) are approximated by the magnitude of the daily mean wind vector [(2)]. Compared to a temporal average calculated from instantaneous wind speeds, this results in an underestimation. By using the component data at a higher temporal resolution—for example, with 6-h intervals as in Zappa et al. (2013b) and Chang (2018)—the underestimation could be reduced to some extent but not eliminated.

FIG. 14. Projected seasonal multimodel mean changes (in %) in the 99th percentile of the true wind speed from 1971–2000 to 2070–99 under RCP8.5. The contour interval is 2.5%. Areas where more than 14 GCMs out of the 17 agree on the sign of change are hatched.

In any case, it is essential to find out whether this approximation engenders bias in the future projections.

In addition to the daily means for the u and v components, the CMIP5 archive provides data for the monthly mean scalar wind speeds as prefabricated by the modeling groups (the variable with an acronym “sfcWind”). According to the CMIP5 documentation table, sfcWind “is the mean of the speed, not the speed computed from the mean u and v components of wind.”

Percent changes in the mean true wind speed calculated by the two methods, the present approximation (2) and using the downloaded sfcWind data, are compared in Figs. 13a and 13b. The patterns of change are nearly identical. Accordingly, using the daily mean wind vectors [(2)] tends to produce a similar systematic error in the present and future time mean wind speeds, and no remarkable error is induced in the projected change. However, on account of the common bias, the changes definitely need to be expressed in relative rather than absolute terms.

c. Comparison of the joint distributions of (u , v) for the true and geostrophic wind

In practical applications, near-surface true winds are of a larger interest than the geostrophic ones. In this section, we compare projected changes in the bivariate

frequency distributions of the u and v components of the true and geostrophic wind at two example grid points, one representing land and the other ocean. To approximate the actual climatologically induced response, the distributions of both variables were now derived from 13 GCMs, by omitting MIROC-ESM, MRI-CGCM3, GFDL CM3, and GFDL-ESM2M—that is, the four models most severely affected by changing surface properties (section 6a).

At the land grid point examined (62.5°N, 27.5°E; Fig. 15a), pronounced increases in the absolute frequency of the geostrophic wind (by more than 0.2 percentage points) are clustered on the directional angles between about 200° and 330°. For the true winds (Fig. 15b), significant increases correspondingly occur in the sector from ≈160° to 290°. Accordingly, it can be concluded that the response pattern for the true winds has twisted approximately 40° anticlockwise relative to the corresponding pattern of the geostrophic winds. A similar inference holds for decreasing frequencies that are accumulated on the easterly sector for the geostrophic and northeasterly sector for the true winds. Of course, this kind of subjective evaluation only yields a tentative approximation for the angle between the frequency distributions. Changes in weak winds are consistent as well; considering the sum for the four

FIG. 15. Projected changes in the bivariate distributions of $(u_g^{\text{norm}}, v_g^{\text{norm}})$ from 1971–2000 to 2070–99 under RCP8.5 at (a) 62.5°, 27.5°N and (c) 40°N, 22.5°W in September–November; a mean of 13 GCMs (see the text). Corresponding responses in the frequencies of the near-surface true wind are given in (b) and (d). For further technical information, see the caption of Fig. 7.

histogram classes representing the smallest wind speeds, the summed decrease in the frequency is 0.4 percentage points for the geostrophic and 0.3 percentage points for the true winds. For strong wind speeds, some tendency toward increasing frequencies can be seen in the westerly sector for the geostrophic winds and in the southwesterly sector for the true winds, but the changes are less coherent than those for the weak and moderate winds.

The ocean grid point 40°N, 22.5°W is located in an area of weakening mean wind speeds (Fig. 3). At this point, the response patterns for the true and geostrophic winds agree well (Figs. 15c and 15d). The directional angle with the maximum increase in the frequency of the true winds is 10° to 20° anticlockwise compared to that for the geostrophic winds. Consistently with the weakening of the time-mean wind speed, the proportions

of weak true and geostrophic winds both increase substantially, while strong winds tend to become less frequent.

Note that the true-wind frequency responses given in Figs. 15b and 15d should not be regarded as trustworthy future projections. First, the number of models inspected ($N = 13$) is rather small, which downgrades the coherence in the responses, particularly for high wind speed. Even more seriously, there is no guarantee that even the models retained in this analysis are adequately free from the influence of changing surface properties. On the contrary, in several models the simulated changes in the true and geostrophic scalar wind speed at the land example grid point were, though modest, still of opposite sign. We made a trial to repeat the analysis by omitting such models, but the resulting response was excessively noisy.

The anticlockwise turning of the true-wind frequency distribution is in agreement with the generally known tendency of surface friction to rotate winds toward a lower pressure, more so in land than ocean areas. Besides, the projected decreases/increases in the frequencies of weak true and geostrophic winds are consistent and also in accordance with corresponding changes in the time-mean wind speed.

7. Discussion

To obtain a high signal-to-noise ratio, this work primarily deals with projections for the late-twenty-first-century climate under the large-emission RCP8.5 scenario. Projections for the moderate-emission RCP4.5 scenario are consistently weaker but the geographical pattern is similar to that under RCP8.5. If international climate policy were successful and global warming restrained below 2°C or even below 1.5°C relative to the preindustrial era, the changes would presumably be qualitatively similar but their intensity and signal-to-noise ratio would be even weaker. Anyhow, even quite modest changes in the strength and frequency of extreme winds blowing from a risky direction may substantially affect the devastating power. Also, the influence of fairly minor changes may be notable for the potential of wind power production, which is proportional to the third power of the wind speed; besides, power plants have to be shut down under winds larger than about 25 m s^{-1} , depending on the wind turbine type (Sempreviva et al. 2008; Tammelin et al. 2013).

Projected changes in temporally averaged geostrophic winds (Figs. 3–5) and the joint distribution of the speed and direction (Figs. 8–11) are affected by climate change-induced shifts in the jet streams. According to Butler et al. (2010) and McGraw and Barnes (2016), increasing low-latitude diabatic heating in the upper troposphere acts to shift the jets northward and lower-tropospheric heating in high latitudes southward. The resulting net impact is determined by the balance between these competing forcing factors. Over the North Atlantic area the situation is further complicated by the weakening oceanic thermohaline circulation that reduces warming at the surface. Woollings et al. (2012) suggest that this tends to increase baroclinicity on the southern edge of the zone of small warming and expand the North Atlantic winter storm track toward northwestern Europe. This would engender a westerly mean-flow response that is primarily maintained by momentum flux convergence in high-pass filtered transient eddies (Simpson et al. 2014). However, the stronger time-mean westerlies in northern and central Europe may also be related to an anomalous wave train extending from

the western Pacific Ocean over North America into Europe. Such a wave train is generated as a response to global warming even in model simulations forced by a spatially uniform increase in the sea surface temperature (Karpechko and Manzini 2017, their Fig. 5).

In this work, the geostrophic and true wind speeds are approximated by the magnitude of the daily mean wind vector [(1) and (2)]. The validity of this approximation in calculating future percent changes in the mean wind speed was confirmed in section 6b. Consequently, we found it relevant to apply those approximate wind speeds in studying the joint distributions of the direction and speed as well; note that in this analysis wind speeds are normalized by their long-term means, which further acts to eliminate the impact of any systematic bias. The comparison presented in section 6b likewise confirms the validity of the findings of several previous studies that have used this approximation (e.g., McInnes et al. 2011; de Winter et al. 2013).

The present geostrophic wind projections are internally consistent in several respects. First, changes in the temporal means and extremes (measured by four different percentiles) exhibit a similar geographical distribution. Thereby, the frequency distribution of wind speeds seems to shift toward higher or lower values in its entirety, in accordance with changes in the mean speed. Second, in ocean areas where any potential spurious trends in the surface roughness do not interfere, changes in the true and geostrophic wind speeds are largely similar. The emerging differences are in line with the deduced changes in the static stability of the atmospheric boundary layer (e.g., those induced by the diminishing sea ice cover). This is valid for the temporally averaged (Fig. 13) as well as for the extreme winds (Figs. 6 and 14). Third, changes in the joint distribution of the components of $\mathbf{V}_g$ and time-mean changes in u_g and v_g generally agree well. In addition, the responses to the RCP4.5 and RCP8.5 forcing are consistent.

Krueger and von Storch (2011) showed that the annual frequency distributions of the true and geostrophic wind speeds correlate fairly well. Nevertheless, geostrophic winds alone do not yield a comprehensive picture of wind climate and its future changes. Small and mesoscale phenomena, such as convective storms, polar lows, and typhoons, are not captured. The actual intensity of synoptic-scale storms is affected, besides the geostrophic wind speed, by the ageostrophic portion of the wind (e.g., the isallobaric wind) (Knox et al. 2011). The strongest horizontal pressure gradients and geostrophic wind speeds typically occur close to cyclone centers where the cyclonic curvature makes the true wind strongly subgeostrophic. This may explain why

very strong geostrophic winds with $V_g^{\text{norm}} \sim 3$ are more common than normalized true winds of a similar magnitude; this feature is apparent in Fig. 15 at both the example grid points. Accordingly, changes in the frequencies of the strongest geostrophic winds do not necessarily represent well corresponding changes in strong true winds.

Furthermore, Willison et al. (2015) suggested that the spatial resolution of the present GCMs is not yet necessarily adequate to simulate the storm tracks perfectly. Another topic potentially requiring attention is the interaction between the stratosphere and troposphere (Manzini et al. 2014). Even so, all these restrictions influence both the baseline-period and future climate, allowing us to satisfactorily project changes in large-scale wind climate. Mountainous areas (e.g., southern Europe particularly in summer) constitute an exception to this inference; geostrophic wind projections are disturbed by the extrapolation of pressure from the actual ground level to the standard sea level.

In interpreting the present findings, it is essential to recall that the directional distributions of the true and geostrophic winds tend to differ systematically as surface friction acts to rotate the near-surface true-wind vector anticlockwise in relation to the geostrophic wind. At the land grid point examined in Fig. 15, the turning is several tens of degrees. For example, if the direction with the most pronounced increase in the geostrophic winds frequencies is west, we can roughly judge that for the true winds the largest increases occur in the southwesterly sector. At the ocean example point, by contrast, the turning of the pattern proved to be minor.

8. Conclusions

As far as the various wind directions are considered collectively, multimodel-mean changes in the true and geostrophic wind speeds are generally fairly modest. This holds both for the temporally averaged and high wind speeds, the latter being represented by the 95th–99.5th percentiles of the frequency distribution. In this respect, the present findings are in line with previous research (e.g., Gregow et al. 2012; de Winter et al. 2013; Kumar et al. 2015). Over the majority of the domain, extreme geostrophic winds tend to weaken somewhat rather than strengthen (Fig. 6).

However, the signal is generally far more pronounced when geostrophic winds are segregated according to direction: in many areas, strong winds from particular directions are projected to become substantially more frequent than in the recent past. For example, in northern Europe and the northern North Atlantic area in autumn, the frequency of strong westerly geostrophic

winds increases by $\sim 50\%$. Nearly equally large changes occur in the areas surrounding the North Sea in winter. In these cases, the signal is fairly robust; that is, the projected changes are of the same sign in the adjacent bins of the joint distribution of u_g and v_g . In these areas and seasons the models likewise agree well on an increase of the temporally averaged u_g .

There are several applications for which it is crucial to consider not only the bare wind speed but also the direction from which the strong wind is blowing. For example, in forests the risk of wind damage is highest when there has been an abrupt jump in wind loading to which the trees have not yet acclimatized (e.g., Peltola et al. 1999). Such conditions occur on the down-wind edge of recently clear-felled areas. The forest areas most severely exposed to wind damages can be mapped by utilizing topographic and land-use data, and such accounts have been developed, for example by Venäläinen et al. (2017) and Suvanto et al. (2018). Significant changes in the directional distribution of extreme winds would necessitate recalculation of the risk mapping tools.

To mention other examples, the damages caused by storm surges exacerbate on the down-wind shore of a sea where strong winds from the sea become more common. According to the present projections, westward-facing coastal areas on the eastern side of the North Sea and Baltic Sea are likely to become more exposed to strong westerly winds, while on the western side the risk is mitigated. On the other hand, on the western and northwestern edges of the sea upwelling of cool nutrient-rich deep water increases, which enhances the risks of harmful algae blooms (Kudela et al. 2010) and thick fog events hazardous for navigation. Furthermore, sea ice is driven by the winds; in the Baltic Sea, the projected increase of strong northwesterly winds during the ice winter (the December–February and March–May seasons) would aggregate pack ices on the coasts of the Baltic countries, the eastern Gulf of Finland, and the Finnish side of the Gulf of Bothnia. Simultaneously, however, rising temperatures will mitigate the ice conditions (e.g., Luomaranta et al. 2014). In buildings, in northern Europe wind-driven rains (Blocken and Carmeliet 2004) would increasingly seriously moisten the outer walls exposed to westerly and southwesterly winds.

Trends in the near-surface true wind speeds are influenced, besides atmospheric circulation conditions, by changes in the surface roughness. In Fig. 12 we provided an example of a model simulation in which changes in the land surface properties completely hid the meteorologically induced wind speed change signal. The surface roughness is primarily determined by vegetation cover while vegetational changes are induced by anthropogenic land use and shifts in the climate zones.

Both phenomena are modeled quite divergently in the different climate models. Moreover, in reality human-induced land-use changes will be determined by societal development and human decisions, such as deforestation, afforestation, and urbanization, that are hardly foreseeable in the long run. Changes in the near-surface true wind speeds that are derived directly from the model output files therefore vary quite arbitrarily across the models. Inspection of the geostrophic winds reveals the actual meteorological climate change signal much more confidently (Feser et al. 2015). Of course, surface properties can affect the sea level pressure and geostrophic wind fields indirectly, such as by modifying the horizontal divergence of the boundary layer winds. Nonetheless, this influence is far weaker than the direct effect of surface roughness on the true near-surface wind.

One viable approach to retrieve the purely meteorological component of the near-surface true wind response is to use a regional climate model (RCM) for which boundary conditions are adopted from a global model but vegetation and other factors affecting the roughness length are kept identical in the baseline-period and future simulations. To ensure the robustness of the projection, such RCM downscalings would have to be performed for a large ensemble of independent GCMs. Alternatively, near-surface winds might be simulated by a one-dimensional atmospheric-column model that is forced by free-atmosphere conditions simulated by the GCMs.

To summarize, the key findings of the present study are that 1) while the mean and extreme scalar geostrophic wind speeds do not change much in response to global warming, substantial projected changes in the occurrence of high wind speeds are nevertheless identified when studying winds from the different directions separately, and 2) it is questionable to derive wind projections for continental areas directly from the modeled near-surface true-wind data.

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REFERENCES

- Alvarez, I., M. N. Lorenzo, M. deCastro, and M. Gomez-Gesteira, 2017: Coastal upwelling trends under future warming scenarios from the CORDEX project along the Galician coast (NW Iberian Peninsula). *Int. J. Climatol.*, **37**, 3427–3438, <https://doi.org/10.1002/joc.4927>.
- Anttila, P., U. Makkonen, H. Hellén, K. Kyllönen, S. Leppänen, H. Saari, and H. Hakola, 2008: Impact of the open biomass fires in spring and summer of 2006 on the chemical composition of background air in south-eastern Finland. *Atmos. Environ.*, **42**, 6472–6486, <https://doi.org/10.1016/j.atmosenv.2008.04.020>.
- Blocken, B., and J. Carmeliet, 2004: A review of wind-driven rain research in building science. *J. Wind Eng. Ind. Aerodyn.*, **92**, 1079–1130, <https://doi.org/10.1016/j.jweia.2004.06.003>.
- Brönnimann, S., O. Martius, H. von Waldow, C. Welker, J. Luterbacher, G. Compo, P. Sardeshmukh, and T. Usbeck, 2012: Extreme winds at northern mid-latitudes since 1871. *Meteor. Z.*, **21**, 13–27, <https://doi.org/10.1127/0941-2948/2012/0337>.
- Butler, A. H., D. W. J. Thompson, and R. Heikes, 2010: The steady-state atmospheric circulation response to climate change-like thermal forcings in a simple general circulation model. *J. Climate*, **23**, 3474–3496, <https://doi.org/10.1175/2010JCLI3228.1>.
- Chang, E. K.-M., 2018: CMIP5 projected change in Northern Hemisphere winter cyclones with associated extreme winds. *J. Climate*, **31**, 6527–6542, <https://doi.org/10.1175/JCLI-D-17-0899.1>.
- Coen, J., 2018: Some requirements for simulating wildland fire behavior using insight from coupled weather–wildland fire models. *Fire*, **1**, 6, <https://doi.org/10.3390/fire1010006>.
- Collins, M., and Coauthors, 2013: Long-term climate change: Projections, commitments and irreversibility. *Climate Change 2013: The Physical Science Basis*. T. F. Stocker et al., Eds., Cambridge University Press, 1029–1136.
- de Winter, R. C., A. Sterl, and B. G. Ruessink, 2013: Wind extremes in the North Sea Basin under climate change: An ensemble study of 12 CMIP5 GCMs. *J. Geophys. Res. Atmos.*, **118**, 1601–1612, <https://doi.org/10.1002/JGRD.50147>.
- Dunne, J. P., and Coauthors, 2013: GFDL’s ESM2 global coupled climate–carbon Earth System Models. Part II: Carbon system formulation and baseline simulation characteristics. *J. Climate*, **26**, 2247–2267, <https://doi.org/10.1175/JCLI-D-12-00150.1>.
- Feser, F., M. Barcikowska, O. Krueger, F. Schenk, R. Weisse, and L. Xia, 2015: Storminess over the North Atlantic and north-western Europe—A review. *Quart. J. Roy. Meteor. Soc.*, **141**, 350–382, <https://doi.org/10.1002/qj.2364>.
- Flato, G., and Coauthors, 2013: Evaluation of climate models. *Climate Change 2013: The Physical Science Basis*. T. F. Stocker et al., Eds., Cambridge University Press, 741–866.
- Ganske, A., B. Tinz, G. Rosenhagen, and H. Heinrich, 2016: Interannual and multidecadal changes of wind speed and directions over the North Sea from climate model results. *Meteor. Z.*, **25**, 463–478, <https://doi.org/10.1127/metz/2016/0673>.
- Gaslikova, L., I. Grabemann, and N. Groll, 2013: Changes in North Sea storm surge conditions for four transient future climate

- realizations. *Nat. Hazards*, **66**, 1501–1518, <https://doi.org/10.1007/s11069-012-0279-1>.
- Gastineau, G., and B. J. Soden, 2009: Model projected changes of extreme wind events in response to global warming. *Geophys. Res. Lett.*, **36**, L10810, <https://doi.org/10.1029/2009GL037500>.
- Goubanova, K., V. Echevin, B. Dewitte, F. Codron, K. Takahashi, P. Terray, and M. Vrac, 2011: Statistical downscaling of sea-surface wind over the Peru–Chile upwelling region: Diagnosing the impact of climate change from the IPSL-CM4 model. *Climate Dyn.*, **36**, 1365–1378, <https://doi.org/10.1007/s00382-010-0824-0>.
- Gregow, H., K. Ruosteenoja, N. Pimenoff, and K. Jylhä, 2012: Changes in the mean and extreme geostrophic wind speeds in Northern Europe until 2100 based on nine global climate models. *Int. J. Climatol.*, **32**, 1834–1846, <https://doi.org/10.1002/joc.2398>.
- , A. Laaksonen, and M. E. Alper, 2017: Increasing large scale windstorm damage in Western, Central and Northern European forests, 1951–2010. *Sci. Rep.*, **7**, 46397, <https://doi.org/10.1038/srep46397>.
- Hurt, G. C., and Coauthors, 2011: Harmonization of land-use scenarios for the period 1500–2100: 600 years of global gridded annual land-use transitions, wood harvest, and resulting secondary lands. *Climatic Change*, **109**, 117–161, <https://doi.org/10.1007/s10584-011-0153-2>.
- Johansson, M. M., H. Pellikka, K. K. Kahma, and K. Ruosteenoja, 2014: Global sea level rise scenarios adapted to the Finnish coast. *J. Mar. Syst.*, **129**, 35–46, <https://doi.org/10.1016/j.jmarsys.2012.08.007>.
- Kahma, K. K., and H. Pettersson, 2005: Wind-wave systems. *Measuring and Analysing the Directional Spectrum of Ocean Waves*, D. Hauser et al., Eds., Office for Official Publications of the European Communities, COST Action 714, Working Group 3, chap. 1.3, Part I, 11–14.
- Karpechko, A. Y., and E. Manzini, 2017: Arctic stratosphere dynamical response to global warming. *J. Climate*, **30**, 7071–7086, <https://doi.org/10.1175/JCLI-D-16-0781.1>.
- Kirtman, B., and Coauthors, 2013: Near-term climate change: Projections and predictability. *Climate Change 2013: The Physical Science Basis*, T. F. Stocker et al., Eds., Cambridge University Press, 953–1028.
- Knox, J. A., J. D. Frye, J. D. Durkee, and C. M. Fuhrmann, 2011: Non-convective high winds associated with extratropical cyclones. *Geogr. Compass*, **5**, 63–89, <https://doi.org/10.1111/j.1749-8198.2010.00395.x>.
- Kohler, M., J. Metzger, and N. Kalthoff, 2018: Trends in temperature and wind speed from 40 years of observations at a 200-m high meteorological tower in Southwest Germany. *Int. J. Climatol.*, **38**, 23–34, <https://doi.org/10.1002/joc.5157>.
- Krueger, O., and H. von Storch, 2011: Evaluation of an air pressure-based proxy for storm activity. *J. Climate*, **24**, 2612–2619, <https://doi.org/10.1175/2011JCLI3913.1>.
- Kudela, R., S. Seeyave, and W. Cochlan, 2010: The role of nutrients in regulation and promotion of harmful algal blooms in upwelling systems. *Prog. Oceanogr.*, **85**, 122–135, <https://doi.org/10.1016/j.pocean.2010.02.008>.
- Kumar, D., V. Mishra, and A. R. Ganguly, 2015: Evaluating wind extremes in CMIP5 climate models. *Climate Dyn.*, **45**, 441–453, <https://doi.org/10.1007/s00382-014-2306-2>.
- Laapas, M., and A. Venäläinen, 2017: Homogenization and trend analysis of monthly mean and maximum wind speed time series in Finland, 1959–2015. *Int. J. Climatol.*, **37**, 4803–4813, <https://doi.org/10.1002/joc.5124>.
- Li, Z., L. Song, H. Ma, J. Xiao, K. Wang, and L. Chen, 2018: Observed surface wind speed declining induced by urbanization in East China. *Climate Dyn.*, **50**, 735–749, <https://doi.org/10.1007/s00382-017-3637-6>.
- Luomaranta, A., K. Ruosteenoja, K. Jylhä, H. Gregow, J. Haapala, and A. Laaksonen, 2014: Multimodel estimates of the changes in the Baltic Sea ice cover during the present century. *Tellus*, **66A**, 22617, <https://doi.org/10.3402/tellusa.v66.22617>.
- Mäll, M., Ü. Suursaar, R. Nakamura, and T. Shibayama, 2017: Modelling a storm surge under future climate scenarios: Case study of extratropical cyclone Gudrun (2005). *Nat. Hazards*, **89**, 1119–1144, <https://doi.org/10.1007/s11069-017-3011-3>.
- Manzini, E., and Coauthors, 2014: Northern winter climate change: Assessment of uncertainty in CMIP5 projections related to stratosphere–troposphere coupling. *J. Geophys. Res. Atmos.*, **119**, 7979–7998, <https://doi.org/10.1002/2013JD021403>.
- McGraw, M. C., and E. A. Barnes, 2016: Seasonal sensitivity of the eddy-driven jet to tropospheric heating in an idealized AGCM. *J. Climate*, **29**, 5223–5240, <https://doi.org/10.1175/JCLI-D-15-0723.1>.
- McInnes, K. L., T. A. Erwin, and J. M. Bathols, 2011: Global climate model projected changes in 10 m wind speed and direction due to anthropogenic climate change. *Atmos. Sci. Lett.*, **12**, 325–333, <https://doi.org/10.1002/asl.341>.
- Mellado-Cano, J., D. Barriopedro, R. García-Herrera, R. M. Trigo, and M. C. Álvarez-Castro, 2018: Euro-Atlantic atmospheric circulation during the Late Maunder Minimum. *J. Climate*, **31**, 3849–3863, <https://doi.org/10.1175/JCLI-D-17-0261.1>.
- Minder, J. R., 2010: The sensitivity of mountain snowpack accumulation to climate warming. *J. Climate*, **23**, 2634–2650, <https://doi.org/10.1175/2009JCLI3263.1>.
- Mioduszewski, J., S. Vavrus, and M. Wang, 2018: Diminishing Arctic Sea ice promotes stronger surface winds. *J. Climate*, **31**, 8101–8119, <https://doi.org/10.1175/JCLI-D-18-0109.1>.
- Möller, T., D. Schindler, A. T. Albrecht, and U. Kohnle, 2016: Review on the projections of future storminess over the North Atlantic European region. *Atmosphere*, **7**, 60, <https://doi.org/10.3390/atmos7040060>.
- Notaro, M., V. Bennington, and S. Vavrus, 2015: Dynamically downscaled projections of lake-effect snow in the Great Lakes basin. *J. Climate*, **28**, 1661–1684, <https://doi.org/10.1175/JCLI-D-14-00467.1>.
- Peltola, H., S. Kellomäki, H. Väisänen, and V. P. Ikonen, 1999: A mechanistic model for assessing the risk of wind and snow damage to single trees and stands of Scots pine, Norway spruce and birch. *Can. J. For. Res.*, **29**, 647–661, <https://doi.org/10.1139/x99-029>.
- Pettersson, H., K. K. Kahma, and L. Tuomi, 2010: Wave directions in a narrow bay. *J. Phys. Oceanogr.*, **40**, 155–169, <https://doi.org/10.1175/2009JPO4220.1>.
- Pryor, S. C., R. J. Barthelmie, N. E. Clausen, M. Drews, N. MacKellar, and E. Kjellström, 2012: Analyses of possible changes in intense and extreme wind speeds over northern Europe under climate change scenarios. *Climate Dyn.*, **38**, 189–208, <https://doi.org/10.1007/s00382-010-0955-3>.
- Ronkainen, I., J. Lehtiranta, M. Lensu, E. Rinne, J. Haapala, and C. Haas, 2018: Interannual sea ice thickness variability in the Bay of Bothnia. *Cryosphere*, **12**, 3459–3476, <https://doi.org/10.5194/tc-12-3459-2018>.
- Särkkä, J., K. Kahma, M. Kämäräinen, M. Johansson, and S. Saku, 2017: Simulated extreme sea levels at Helsinki. *Boreal Environ. Res.*, **22**, 299–315.

- Schwierz, C., P. Köllner-Heck, E. Zenklusen Mutter, D. N. Bresch, P.-L. Vidale, M. Wild, and C. Schär, 2010: Modelling European winter wind storm losses in current and future climate. *Climatic Change*, **101**, 485–514, <https://doi.org/10.1007/s10584-009-9712-1>.
- Seiler, C., and F. W. Zwiers, 2016: How will climate change affect explosive cyclones in the extratropics of the Northern Hemisphere? *Climate Dyn.*, **46**, 3633–3644, <https://doi.org/10.1007/s00382-015-2791-y>.
- Sempreviva, A. M., R. J. Barthelmie, and S. C. Pryor, 2008: Review of methodologies for offshore wind resource assessment in European seas. *Surv. Geophys.*, **29**, 471–497, <https://doi.org/10.1007/s10712-008-9050-2>.
- Simpson, I. R., T. A. Shaw, and R. Seager, 2014: A diagnosis of the seasonally and longitudinally varying midlatitude circulation response to global warming. *J. Atmos. Sci.*, **71**, 2489–2515, <https://doi.org/10.1175/JAS-D-13-0325.1>.
- Soukissian, T., F. Karathanasi, P. Axaopoulos, E. Voukouvalas, and V. Kotroni, 2018: Offshore wind climate analysis and variability in the Mediterranean Sea. *Int. J. Climatol.*, **38**, 384–402, <https://doi.org/10.1002/joc.5182>.
- Stössel, A., Z. Zhang, and T. Vihma, 2011: The effect of alternative real-time wind forcing on Southern Ocean sea ice simulations. *J. Geophys. Res. Oceans*, **116**, C11021, <https://doi.org/10.1029/2011JC007328>.
- Suvanto, S., H. M. Henttonen, P. Nöjd, and H. Mäkinen, 2018: High-resolution topographical information improves tree-level storm damage models. *Can. J. For. Res.*, **48**, 721–728, <https://doi.org/10.1139/cjfr-2017-0315>.
- Tammelin, B., and Coauthors, 2013: Production of the Finnish Wind Atlas. *Wind Energy*, **16**, 19–35, <https://doi.org/10.1002/we.517>.
- Uppala, S. M., and Coauthors, 2005: The ERA-40 Re-Analysis. *Quart. J. Roy. Meteor. Soc.*, **131**, 2961–3012, <https://doi.org/10.1256/qj.04.176>.
- van Vuuren, D. P., and Coauthors, 2011: The representative concentration pathways: an overview. *Climatic Change*, **109**, 5–31, <https://doi.org/10.1007/s10584-011-0148-z>.
- Vautard, R., J. Cattiaux, P. Yiou, J.-N. Thépaut, and P. Ciais, 2010: Northern Hemisphere atmospheric stilling partly attributed to an increase in surface roughness. *Nat. Geosci.*, **3**, 756–761, <https://doi.org/10.1038/ngeo979>.
- Venäläinen, A., and Coauthors, 2017: Estimation of the high-spatial-resolution variability in extreme wind speeds for forestry applications. *Earth Syst. Dyn.*, **8**, 529–545, <https://doi.org/10.5194/esd-8-529-2017>.
- Willison, J., W. A. Robinson, and G. M. Lackmann, 2015: North Atlantic storm-track sensitivity to warming increases with model resolution. *J. Climate*, **28**, 4513–4524, <https://doi.org/10.1175/JCLI-D-14-00715.1>.
- Woollings, T., and M. Blackburn, 2012: The North Atlantic jet stream under climate change and its relation to the NAO and EA patterns. *J. Climate*, **25**, 886–902, <https://doi.org/10.1175/JCLI-D-11-00087.1>.
- , J. M. Gregory, J. G. Pinto, M. Reyers, and D. J. Brayshaw, 2012: Response of the North Atlantic storm track to climate change shaped by ocean atmosphere coupling. *Nat. Geosci.*, **5**, 313–317, <https://doi.org/10.1038/ngeo1438>.
- Zappa, G., L. C. Shaffrey, and K. I. Hodges, 2013a: The ability of CMIP5 models to simulate North Atlantic extratropical cyclones. *J. Climate*, **26**, 5379–5396, <https://doi.org/10.1175/JCLI-D-12-00501.1>.
- , —, —, P. G. Sansom, and D. B. Stephenson, 2013b: A multimodel assessment of future projections of North Atlantic and European extratropical cyclones in the CMIP5 climate models. *J. Climate*, **26**, 5846–5862, <https://doi.org/10.1175/JCLI-D-12-00573.1>.