



Future changes in Antarctic near-surface winds: regional variability and key drivers under a high-emission scenario

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Abstract. Antarctic near-surface winds play a key role in shaping the local climate of Antarctica. For instance, they trigger drifting snow and reduce the amount of precipitation reaching the ground. Despite their importance, substantial uncertainties remain regarding their future changes over the continent, especially in winter, under a warming scenario. Here, we analyse projections of winter near-surface winds in Antarctica produced by four CMIP6 Global Climate Models downscaled by a regional atmospheric model adapted for the study of polar regions. Our analysis first demonstrates that the downscaling helps to improve the representation of near-surface winds at present day. On the continent, projected changes in July wind speeds between the late 21st and 20th centuries reveal considerable regional variability, with opposing trends depending on the area and model used. Nevertheless, the 4 models used agree on a significant strengthening of near-surface winds in Adélie Land, Ross-ice shelf and Enderby Land and a significant weakening in some coastal areas, such as Shackleton ice shelf, Pine Island Glacier and Ronne ice shelf. Using the momentum budget decomposition, we separate and quantify the contributions of different drivers to future changes in wind speed. These drivers include katabatic and thermal wind accelerations (which are related to the net radiative cooling by the iced surface) as well as large-scale forcing. We project a significant decrease of both katabatic and thermal wind accelerations. Because in a warming climate they act to increase the wind speed in opposite directions, we find an overall compensation effect of the changes in katabatic and thermal wind at the margins of the continent, while large-scale forcing exhibits both significant increases and decreases depending on the location. Ultimately, we find that most significant strengthening of near-surface winds originates from strengthening in the large-scale forcing while most significant weakening of near-surface winds can be attributed to changes in the surface forcing.

1 Introduction

The extraordinarily strong and persistent winds are a defining characteristic of Antarctica's climate. They include powerful westerlies on the ocean and easterlies at the ice sheet margins. In the interior, near-surface winds are predominantly directed downslope and play a major role in shaping the Antarctic climate as they trigger drifting snow (Amory, 2020), they indirectly



influence sea ice formation (Holland and Kwok, 2012), the amount of precipitation reaching the ground (Grazioli et al., 2017), the stability of the boundary layer (Vignon et al., 2017) and they can play a determining role in triggering rapid ice shelf collapse (Cape et al., 2015).

25 Near-surface Antarctic winds result from both large-scale and surface pressure gradients (Van den Broeke and van Lipzig, 2002; Bintanja et al., 2014a; Davrinche et al., 2024), whose relative magnitudes in future projections are yet uncertain. Large-scale forcing is intrinsically linked to one of the leading modes of variability in the southern hemisphere, the Southern Annular Mode (SAM), whose strength is characterized by the SAM index, defined as the zonally averaged mean sea-level pressure gradient between 40 and 65 °S (Marshall, 2003). On the other hand, the surface forcing includes a gravitational katabatic
30 pressure gradient that is proportional to the strength of the temperature inversion and a thermal wind that acts to replenish the pressure low created by the downslope displacement of air.

In future projections, the evolution of each family of forcing and their relative magnitude remains uncertain. On the one hand, the greenhouse warming causes an increase of the incoming longwave radiation. As a consequence, the temperature inversion and thus the katabatic forcing, which is proportionate to the strength of the temperature inversion, should decrease
35 (Van den Broeke and van Lipzig, 2002; Bintanja et al., 2014b). On the other hand, the increase in GHG concentration drives the SAM towards a more positive phase by the end of the 21st century (Miller et al., 2006; Fogt and Marshall, 2020; Goyal et al., 2021). Thus, models predict a strengthening and poleward shift of the westerlies, and a weakening of coastal off-shore easterlies during summer (Bracegirdle et al., 2008; Langlais et al., 2015; Hazel and Stewart, 2019; Neme et al., 2022). As a result, the large-scale forcing is expected to increase over the ocean and decrease at the coastal margins in summer, but its
40 trend over the continent itself is unknown. In winter however, changes in the zonally averaged SAM are weaker. Bracegirdle et al. (2008) hypothesize that the impact of the SAM does not have the ability to penetrate sufficiently southward to influence the large-scale forcing of coastal on-shore and mid-slope easterlies. However, under a doubling of CO₂, Van Den Broeke et al. (1997) and Turner et al. (2013) showed that the circumpolar trough is locally enhanced in specific locations where sea ice is completely removed (e.g. north of Ross and Amery ice shelves and north of the Peninsula. Although there is a consensus on the
45 reduction of surface forcing in climate projections, large uncertainties remain regarding the evolution of the large-scale forcing around the coastlines of Antarctica in winter, and even more in the interior. Because of the zonal asymmetries in the changes of sea-level pressure around Antarctica, we expect to find zonal asymmetries in the evolution of the on-shore large-scale forcing as well.

Global Climate Models (GCMs) have limited capacity to resolve each family of pressure gradients, particularly local pressure
50 gradients that influence surface wind speed on the continent, due to complex topography, land–sea contrasts, boundary-layer, and convective processes (Di Virgilio et al., 2022; Smith and Polvani, 2017), leading to large uncertainties regarding the evolution of near-surface winds in future projections. An approach that circumvents these limitations is dynamical downscaling, which involves using high-resolution regional climate models (RCMs) driven at their upper and lateral boundaries with a GCM or reanalysis data. This ensures a better resolution of the ice sheet topography as well as a physically realistic simulation of
55 boundary-layer dynamics achieved through adapted parametrizations of the interactions between the snow/ice surface and the atmosphere.



Here we investigate the projected changes of Antarctic winter near-surface winds under a high-emission scenario, focusing on the respective response of katabatic and large-scale forcings. We focus on the Antarctic continent, which is the source region of the katabatic forcing, and on the winter season, as it is the season for which both the katabatic forcing and the mean wind speed are the highest (Davrinche et al., 2024). We mitigate GCM limitations used in previous studies by using the regional atmospheric climate model MAR to dynamically downscale four recent CMIP6 GCMs carefully selected on their ability to represent the large-scale circulation in polar regions. Using the momentum budget decomposition, we analyse how each family of drivers evolves in the different downscaled GCMs.

2 Materials and Methods

2.1 Selection of AWS using ERA5

In this section, we describe our protocol for the evaluation of near-surface winds in both GCMs and their downscaling with MAR against a subset of automated weather stations (AWS). Because of their resolution, GCMs are not expected to perform well in locations with complex topography. Therefore, we select a subset of AWS based on i) the ERA5 reanalysis' ability to represent the mean wind speed and variability outside of areas of complex topography, and ii) the length of available winter time series to evaluate GCMs on a representative climatic time scale.

2.1.1 The AntAWS dataset

We use the monthly AntAWS dataset provided by Wang et al. (2023) that compiles all the available Automatic Weather Station data in Antarctica from 1980 to 2021. For all stations (except Zhongshan which is on a mast at ~ 10 m from the ground), data are collected at a height of ~ 3 m above ground level (agl), although the height of the wind sensor is poorly controlled and varies greatly between 1 and 6 m (Wang et al., 2023), depending on the initial sensor height and snow accumulation rate. According to the logarithmic theoretical profile of wind speed in the boundary layer, with a constant roughness length $z_0 = 1$ mm (Vignon et al., 2017), we estimate the correction to be between -11 % and 7 % of the theoretical value:

$$\text{correction}_{6-3} = \frac{\log(\frac{6}{z_0})}{\log(\frac{3}{z_0})} = 1.07 \quad (1)$$

$$\text{correction}_{1-3} = \frac{\log(\frac{1}{z_0})}{\log(\frac{3}{z_0})} = 0.90 \quad (2)$$

Data are collected every 3 hours and monthly averages are computed when at least 75 % of the 3-hourly observations are available in a month, based on Kittel et al. (2021). An additional quality control is performed in which wind speed exceeding 60 m s^{-1} or equal to 0 m s^{-1} are discarded. If wind speed and direction remain constant for 2 consecutive timesteps, values are discarded, as it might be due to sensors being frozen. Other values were flagged and validated or discarded based on a visual comparison with reanalysis datasets (ERA5). This includes rapidly changing values of wind speed (i.e. two consecutive values with a difference greater than 74 km h^{-1}) and values outside of the likelihood interval of 3 standard deviations from the mean value, based on the criteria described in Lazzara et al. (2012).



2.1.2 ERA5 reanalysis

We use the ERA5 reanalysis dataset to select the AWS in relevant locations for the evaluation of the GCMs. For the representation of near-surface winds in Antarctica, ERA5 (produced by the European Centre for Medium-Range Weather Forecasts (ECMWF), Hersbach et al. (2020)) performs best for the monthly averaged wind speeds in comparison to other reanalysis products (MERRA-2, JRA-55, ERAI, NCEP2 and CFSR) as demonstrated by Dong et al. (2020). ERA5 does not assimilate wind speed measurements from AWS, and can therefore be used to screen observational stations (see Sect. 2.1.4).

The horizontal spatial resolution of ERA5 is ~ 31 km and outputs are given at a hourly frequency. The assimilation system (IFS Cycle 41r2 4D-Var) uses 10 members to produce a 4D-Var ensemble of data assimilation (Hennermann and Guillory, 2019).

2.1.3 Selection of AWS based on dataset length, and computation of the reference climatology

We want to create a climatology of the winter wind speed in Antarctica in order to have a reference to study potential evolution of wind speed by the end of the 21st century. Therefore, we need datasets long enough to accurately represent the historical climatology. AWS data are only available during austral summers for almost 50 % (128 out of 267) of stations. As our study focuses on the winter month of July, we screen for the availability of observations during this month. In order to test whether datasets are long enough to be representative of a climatological period, we compute using ERA5 the minimum value of N_{July} for which the relative uncertainty on the mean value of the July wind speed between 1980 and 2020 is lower than 5 % (See Supplementary Section S1.1). We conclude that selecting stations for which the number of July observations at each station N_{July} is greater than 10 is a reasonable criterion that enables a fair representation of the climatology of July wind speeds (Figure S1). As a result, out of 267 stations listed in the AntAWS dataset, we consider that only 28 of them are suitable to evaluate GCMs. These stations are presented in Fig. 1 and their elevation ranges from 30 to 3350 m above sea level (Table 1). For the 28 pre-selected AWS stations, the datasets exhibit no significant trend between 1980 and 2020, with values of the linear trend computed with ERA5 monthly July wind speed ranging between -0.08 and $0.1 \text{ m s}^{-1} \text{ decade}^{-1}$.

Furthermore, we compare the averaging of ERA5 wind speed over the 1980-2020 period or over the period available for each AWS, and we find differences lower than 0.4 m s^{-1} in absolute value or 5 % of the mean value over 40 years (Figure S2). Therefore, we are confident that we can use the climatology at the 28 selected stations of the AntAWS dataset to evaluate the climatological historical mean of the GCMs over the period 1980-2000.

2.1.4 Exclusion of sites near complex topography based on performance of ERA5

GCMs have limited capacity to resolve local processes that influence regional climate, such as complex topography, land-sea contrasts and boundary layer convective processes (Di Virgilio et al., 2022). For a fair evaluation of GCMs, we do not want to analyze locations for which the topography is too specific and the resulting atmospheric dynamics will not be resolved by the models, e.g. close to the Transantarctic mountains or at the boundary between the ocean and the continent. We decided to exclude stations for which ERA5 wind speed in the nearest grill cell shows poor agreement with observed wind speed, as

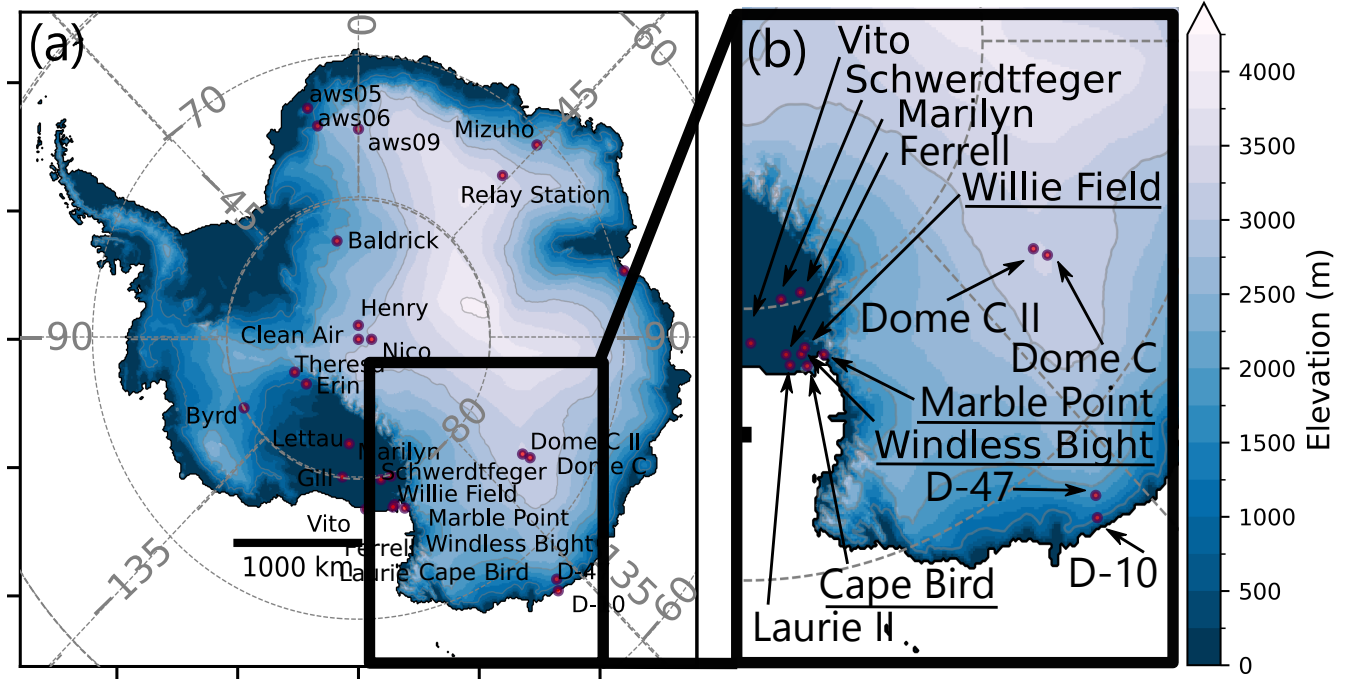


Figure 1. Elevation, from Bedmachine (a) over all Antarctica, (b) zoomed on the black rectangle area. Superimposed are the 28 pre-selected AWS. Stations that have been discarded because of the inability of ERA5 to properly represent winds at these locations (see Sect. 2.1.4) are underlined.

we do not expect GCMs to perform better than the reanalysis over the period of available AWS observations. We consider the following metrics, computed for monthly or annual means:

- the Pearson correlation coefficient (R) of ERA5 and AWS mean wind speed
- the normalized bias $B = (|\overrightarrow{V_{ERA5}}| - |\overrightarrow{V_{AntAWS}}|) / |\overrightarrow{V_{AntAWS}}|$
- and the normalized standard deviation $\sigma_N = \sigma_{ERA5} / \sigma_{AntAWS}$

We compute these three metrics for July, December and annual means for each station. Then we attribute a score for each metric, each station and each period, equals to 1 if $|R| > 0.5$ or $B \leq 30\%$ or $0.5 < \sigma_N < 1.5$, and -1 otherwise. Finally, we combine the scores into one total performance score (TPS) per station, computed as the sum of each individual performance score. This TPS is comprised between -9 and 9 (three metrics times three seasons which can get scores between -1 and 1). Results are presented in Table 1 and Figure S3. We discard stations with a TPS inferior to 3, as it corresponds to half of the metrics exhibiting a poor performance score (Cape Bird, Windless Bight, Willie Field and Marble Point, Table 1 and Fig. 3, Figure S4 and Figure S5). These four stations exhibit the largest biases in terms of temporal variability ($R < 0.3$ and $\sigma_N > 2$, which indicates that the variability in ERA5 is underestimated) and mean amplitude ($B > 30\%$, which indicates that ERA5



overestimate the mean value of the wind speed) and are all located at the foot of the Transantarctic mountains (Fig. 1). The 24 remaining stations, which cover locations from the coast to the plateau) are then listed in Table 1, above the double horizontal line.

135 2.2 Climate models

2.2.1 The regional atmospheric model MAR

The Regional Atmospheric Model MAR is a polar-oriented model which includes snowpack physics and its interactions with the atmosphere. It is a hydrostatic model whose primitive and prognostic equations have been extensively described in Gallée and Schayes (1994) and Gallée (1995). The turbulent scheme is well adapted to stable boundary layers, which is well suited
140 for the study of polar regions. Additionally, the roughness length is parameterized as a function of surface air temperature to take into account the effect of sastrugis and is fitted to match observations of the temporal variability of wind speed in Adélie Land (Amory et al., 2017; Vignon et al., 2017; Agosta et al., 2019). The topography of the model is fixed, and derived from Bedmap 2 (Fretwell et al., 2013). We use 3-hourly model outputs on the standard Antarctic polar stereographic grid at a horizontal resolution of 35 km. The vertical spacing is in σ coordinates with 12 levels between ~ 2 m and ~ 1000 m above
145 ground level. MAR is forced every 6 hours at the top of the atmosphere (wind and temperature, above 10 km) and at its lateral boundaries by large-scale atmospheric fields (wind, temperature, specific humidity, pressure, sea surface temperature, and sea ice concentration).

2.2.2 Selection of four Global Climate Models among CMIP6

We forced MAR with four GCMs from CMIP6: IPSL-CM6A-LR (Boucher et al., 2020), UKESM1-0-LL (Sellar et al., 2019),
150 MPI-ESM1-2-HR (Mauritsen et al., 2019) and CNRM-CM6-1 (Voldoire et al., 2019), referred to in this paper as IPSL, UKESM, MPI and CNRM. CMIP6 models are the latest GCM simulations from the Coupled Model Intercomparison Project (Eyring et al., 2016).

CMIP6 models are selected based on their ability to represent the current climate at both poles ($> 50^\circ$ N in the Arctic and $< 40^\circ$ S in Antarctic). For this selection, nine metrics are considered: annual 500 hPa geopotential height, annual sea level
155 pressure, summer sea surface temperature, winter sea ice concentration, annual and summer temperatures at 850 and 700 hPa. Implausibility is defined for each metric as the portion of the surface where the difference between historical averages in the model and ERA5 is greater than 3 times the ERA5 interannual standard deviation (Agosta et al., 2022).

We chose to study CMIP6 GCMs that are representative of a large range of climate sensitivity typical of CMIP6, and have a low fraction of implausibility for both poles and for all metrics (Agosta et al., 2022), which leads us to select IPSL-CM6A-
160 LR, UKESM1-0-LL, MPI-ESM1-2-HR and CNRM-CM6-1. The choice of these four models for our study is supported by another study by Williams et al. (2024) where these models were classified among the best performing ones in winter when comparing their sea ice extent, surface air temperature, zonal wind at 850 and 50 hPa to ERA5. Note that all of these models are



Table 1. List of AWS used to evaluate July wind speed and associated characteristics: longitude (Lon), latitude (Lat), elevation in MAR, real elevation, local slope in MAR and Total Performance Score (TPS, as described above). In the station name column, bracketed (C) corresponds to location where the corresponding grid-point of the model is at the interface between the continent and the ocean and bracketed (TM) correspond to locations close to the Transantarctic Mountains. The stations below the double horizontal line were excluded from the analysis, based on their low Total Performance Score (TPS, see Sec. 2.1.4)

Station name	N_{July}	$\sigma/ \overline{V} $ (%)	Lon (°)	Lat (°)	Elevation (m, MAR)	Real elevation (m)	Slope (m km ⁻¹)	TPS
Schwerdtfeger (TM)	32	19.7	170.36	-79.82	60	50	0	9
D-47	14	7.6	138.73	-67.39	1630	1560	7	9
Relay Station	20	10.9	43.06	-74.02	3350	3350	2	9
Laurie II (C, TM)	13	17.8	170.74	-77.43	0	30	0	9
Henry	18	12.8	-0.41	-89.0	2830	2880	1	9
Ferrell (C, TM)	14	18.1	170.82	-77.78	40	40	4	9
Erin	13	8.3	-128.87	-84.9	920	990	6	9
Mizuho	14	9.7	44.29	-70.7	2280	2260	4	9
D-10 (C)	14	6.2	139.84	-66.71	320	240	15	9
Clean Air	17	15.6	0.0	-90.0	2800	2840	2	9
Byrd	15	11.8	-119.44	-80.01	1520	1540	2	9
Vito (C)	11	15.0	177.83	-78.41	50	50	0	7
aws06	11	10.4	-11.52	-74.47	1050	1160	9	7
Baldrick	12	6.7	-13.05	-82.77	1970	1970	3	7
Lettau	21	19.4	-174.59	-82.48	60	40	0	7
aws09	20	16.7	0.0	-75.0	2870	2900	1	7
Elaine	12	22.1	174.24	-83.07	70	60	1	5
Marilyn (TM)	21	17.8	165.77	-79.9	60	60	0	5
Gill	12	15.1	-178.54	-79.82	50	50	0	3
Theresa	20	13.5	-115.85	-84.6	1740	1450	10	3
Nico	20	15.0	90.02	-89.0	3020	2980	2	3
aws05	14	11.8	-13.17	-73.1	450	360	8	3
Dome C	13	16.5	123.0	-74.5	3230	3280	1	3
Dome C II	23	18.4	123.35	-75.11	3260	3250	0	3
Willie Field	16	13.9	166.92	-77.87	20	10	3	1
Marble Point	34	15.1	163.75	-77.44	70	110	10	-1
Windless Bight	15	15.0	167.67	-77.73	30	40	6	-5
Cape Bird	16	18.5	166.44	-77.22	0	40	5	-7



Table 2. List of selected GCMs with climate characteristics: Earth's equilibrium Climate Sensitivity (ECS) (Flynn and Mauritsen, 2020), horizontal resolutions (Williams et al., 2024), and storyline of projected Sea Ice Extent (SIE) and Stratospheric Polar Vortex (SPV) strength (Williams et al., 2024). SIE + (SIE-) corresponds to a storyline with a low (strong) projected SIE (when compared to the multi-model mean of CMIP6) while SPV+ (SPV-) corresponds to a storyline with a strong (weak) projected SPV strength.

Model	Institution	Resolution	ECS	Winter storyline	
				SIE	SPV
IPSL-CM6A-LR	IPSL	250 km	4.50	+	+
UKESM1-0-LL	MOHC	250 km	5.31	+	-
MPI-ESM1-2-HR	MPI-M	100 km	2.84	-	+
CNRM-CM6-1	CNRM- CERFACS	250 km	4.81	-	+

Earth System Models, except for CNRM-CM6-1 which does not include interactive ocean biogeochemistry nor atmospheric chemistry (Voldoire et al., 2019).

Furthermore, these models are representative of different storylines for Antarctica (Williams et al., 2024). They have different Earth's Equilibrium Climate Sensitivity (ECS, corresponding to the change in temperature at equilibrium that would result from a doubling of CO₂), which is a proxy for the intensity with which the model warms the Earth's surface temperature. While UKESM has one of the strongest ECS of all CMIP6 models, MPI exhibits one of the lowest. They are also associated with either large or small projected Sea Ice Extent (SIE) in the future and strong or weak projected Stratospheric Polar Vortex (SPV, linked to the strength and position of the surface westerlies, Table 2) during winter. Note that, unlike in Williams et al. (2024), we classify models based on their projected Sea Ice Extent (SIE) instead of the SIE change divided by the calculated global warming in each model.

2.2.3 Experiments

We use a high emission scenario (SSP585) to test the sensitivity of wind speed to climate change with a strong warming of the continent. The expected global radiative forcing by 2100 with this scenario is +8.5 W m⁻² (IPCC AR6, 2023). We then force MAR by one member of each of the four GCMs (r1i1p1f1 for all models except CNRM-CM6A-1, which is forced by r1p1i1f2). Here, we define the historical reference period as 1980-2000 and compare this period with the end of the 21st century (2080-2100), as in Bracegirdle et al. (2020). We study the change in the monthly-mean July near-surface wind speed at 10 m (sfcWind in CMIP6) averaged over 20 years, between these two periods.

2.2.4 Statistical significance

In order to test the statistical significance of changes in 10 m wind speed or any related variable between the end of the 21st and the 20th century, we apply the non-parametric Kruskal-Wallis test (Kruskal and Wallis, 1952). This test (also called one-way ANOVA on rank) is performed at a level of significance of 80 %. It has been used in multiple previous studies to assess past or



future changes (Machado and Calliari, 2016; Marshall et al., 2017; Casado et al., 2023, e.g.). This test assesses that one sample
185 (e.g. July mean monthly wind speed between 2080 and 2100) has significantly higher or lower values than another one (e.g. July mean monthly wind speed between 1980 and 2000).

2.3 Momentum budget decomposition

2.3.1 Equations

The momentum budget decomposition is a useful tool for identifying the drivers of wind speed variability in Antarctica
190 (Van den Broeke and van Lipzig, 2002; Bintanja et al., 2014b). The method is described extensively in Davrinche et al. (2024). We compute the momentum budget in the cross- and downslope directions and we decompose it into 6 different accelerations, defined as follows:

	Horizontal advection	Coriolis	Vertical advection & Turbulence	Large-scale	Thermal wind	Katabatic
195 Cross-slope:	ADVH	COR	TURB	LSC	THW_{TD}	KAT
	$\frac{\partial U}{\partial t} = -U \frac{\partial U}{\partial x} - V \frac{\partial U}{\partial y}$	$+fV$	$-W \frac{\partial U}{\partial z} - \frac{\partial \bar{u}w}{\partial z}$	$-fV_{LSC}$	$+ \frac{g}{\theta_0} \frac{\partial \hat{\theta}}{\partial y}$	
Downslope:	$\frac{\partial V}{\partial t} = -U \frac{\partial V}{\partial x} - V \frac{\partial V}{\partial y}$	$-fU$	$-W \frac{\partial V}{\partial z} - \frac{\partial \bar{v}w}{\partial z}$	$+fU_{LSC}$	$+ \frac{g}{\theta_0} \frac{\partial \hat{\theta}}{\partial y}$	$+ \frac{g}{\theta_0} \Delta\theta \sin(\alpha)$

(3)

with (U, V) the horizontal components of the wind in the cross- and downslope direction. α is the local slope, θ is the
200 potential temperature, and θ_0 is the background potential temperature described in Davrinche et al. (2024), which represents the extrapolation down to the surface of the potential temperature in the upper part of the atmosphere, where surface processes do not come at play. $\Delta\theta$ represents the temperature deficit, i.e. the difference between the background and the actual potential temperature. $\hat{\theta}$ is the vertically integrated potential temperature deficit from the top of the boundary layer. Above the boundary layer, as $\theta = \theta_0$, both $\Delta\theta$ and $\hat{\theta}$ become zero. While the latter are linked to the influence of the surface on the vertical potential
205 temperature profile, θ_0 is related to the synoptic forcing and is used in the computation of the large-scale components of the winds V_{LSC} and U_{LSC} :

$$\frac{\partial U_{LSC}}{\partial \ln(p)} = + \frac{R_d}{f} \left(\frac{p}{p_0} \right)^{\frac{R_d}{C_p}} \left(\frac{\partial \theta_0}{\partial y} \right)_p$$

$$\frac{\partial V_{LSC}}{\partial \ln(p)} = - \frac{R_d}{f} \left(\frac{p}{p_0} \right)^{\frac{R_d}{C_p}} \left(\frac{\partial \theta_0}{\partial x} \right)_p$$

where p is the pressure (in hPa), p_0 the standard reference pressure (equals to 1013.2 hPa), R_d and C_p are respectively the
210 gas constant and specific heat capacity of dry air ($R_d = 287 \text{ J kg}^{-1} \text{ K}^{-1}$ and $C_p = 1005.7 \text{ J kg}^{-1} \text{ K}^{-1}$).



Further descriptions of the equations and validation of the method is performed in Davrinche et al. (2024).

2.3.2 Description of the six accelerations

The pressure gradient force in the momentum budget equation is divided into three accelerations reflecting the origin of the driver: the large-scale acceleration, katabatic acceleration, and the thermal wind acceleration. The large-scale acceleration (LSC) represents the portion of the pressure gradient force that originates from the synoptic forcing above the boundary layer. The katabatic acceleration (KAT) represents the gravity-driven motion induced by the temperature inversion over a sloping surface. It is especially strong during the austral winter in a narrow band close to the coastal margins and displays a strong diurnal cycle in summer and seasonal cycle throughout the year. The thermal wind acceleration (THW_{TD}), related to the temperature deficit, is sometimes referred to as shallow baroclinicity (Caton Harrison et al., 2024) or integrated temperature deficit (Parish and Cassano, 2003). It corresponds to the near-surface baroclinicity induced by changes in the depth of the temperature deficit layer.

The horizontal advection (ADVH) corresponds to the horizontal transport of momentum budget by the wind itself. It is weak in comparison to the other terms of the momentum budget equations but can sometimes become significant in coastal areas or in topographically complex zones such as valleys, or at the foot of the mountains. The Coriolis acceleration (COR) is a deviation induced by the Earth's rotation and it results in a rotation of the wind by 90° to the west in comparison to its acceleration. The residual term (TURB) encompasses both the vertical advection (which is weak) and the turbulent drag that opposes the other accelerations and is strong when the wind speed is high.

Katabatic, thermal wind and large-scale accelerations are considered active because they are produced by a forcing, either large-scale or surface pressure gradients, while turbulence, Coriolis, and advection are passive terms, that form as a reaction once motion has been triggered by an active term.

2.3.3 Attribution of changes in wind speed using the Momentum Budget Decomposition

In winter, the first order temporal derivatives of the wind vector ($\frac{\partial U}{\partial t}$ and $\frac{\partial V}{\partial t}$) are 5 orders of magnitude smaller than the other accelerations (Fig. 2). Therefore, we can assume stationary conditions and rewrite Eq. (3) in a "quasi-geostrophic" form:

$$\begin{cases} U = \underbrace{\frac{1}{f}(-U\frac{\partial V}{\partial x} - V\frac{\partial V}{\partial y})}_{U_{ADVH}} + \underbrace{\frac{1}{f}(-W\frac{\partial V}{\partial z} - \frac{\partial v\bar{w}}{\partial z})}_{U_{TURB}} + \underbrace{\frac{1}{f}(fU_{LSC})}_{U_{LSC}} + \underbrace{\frac{1}{f}(\frac{g}{\theta_0}\frac{\partial \hat{\theta}}{\partial x})}_{U_{THW}} + \underbrace{\frac{g}{f\theta_0}\Delta\theta\sin(\alpha)}_{U_{KAT}} \\ V = \underbrace{-\frac{1}{f}(-U\frac{\partial U}{\partial x} - V\frac{\partial U}{\partial y})}_{V_{ADVH}} - \underbrace{\frac{1}{f}(-W\frac{\partial U}{\partial z} - \frac{\partial u\bar{w}}{\partial z})}_{V_{TURB}} + \underbrace{\frac{1}{f}(fV_{LSC})}_{V_{LSC}} - \underbrace{\frac{1}{f}(\frac{g}{\theta_0}\frac{\partial \hat{\theta}}{\partial y})}_{V_{THW}} \end{cases} \quad (4)$$

The vectorial form of this equation is :

$$\vec{V} = \vec{V}_{ADVH} + \vec{V}_{TURB} + \vec{V}_{LSC} + \vec{V}_{THW} + \vec{V}_{KAT}, \quad (5)$$

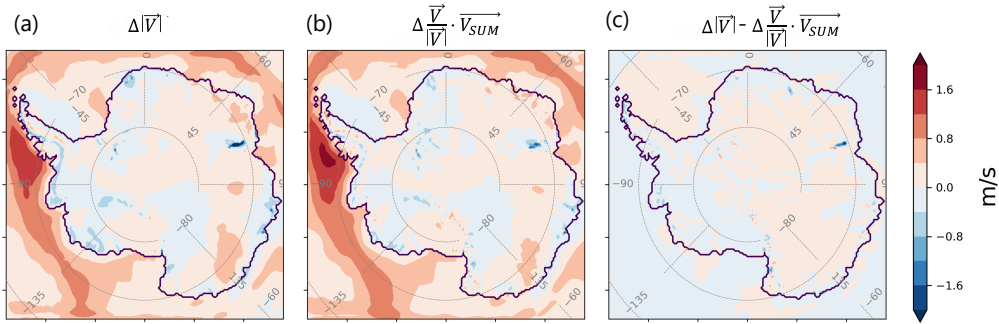


Figure 2. (a) Changes in multi-model mean (MAR-IPSL, MAR-UKESM, MAR-MPI and MAR-CNRM) 10-m wind speed between 2080-2100 and 1980-2000, (b) Changes in multi-model mean (MAR-IPSL, MAR-UKESM, MAR-MPI and MAR-CNRM) 10-m scalar product of the sum of the accelerations with the wind direction $\frac{\vec{V}}{|\vec{V}|}$, i.e. $\frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{SUM} = \frac{\vec{V}}{|\vec{V}|} \cdot (\vec{V}_{ADVH} + \vec{V}_{TURB} + \vec{V}_{LSC} + \vec{V}_{THW} + \vec{V}_{KAT})$, between 2080-2100 and 1980-2000 and (c) Difference of (a) and (b)

with $\vec{V}$ the total wind vector, of components (U, V) in the cross- and downslope coordinate system, and $\vec{V}_{ACC}$ the wind that would be in geostrophic balance with the corresponding acceleration **ACC** (i.e. Coriolis acceleration balances **ACC**), of components (U_{ACC}, V_{ACC}) shown in Eq. (4). Note that the wind vector associated to each acceleration corresponds to a rotation to the left of the acceleration, with the norm divided by $1/f$. E.g. the **KAT** acceleration is downslope, but its contribution to the wind vector $\vec{V}_{KAT}$ is in the cross-slope direction due to its deviation by Coriolis.

We define $|\vec{V}|$ as the norm of the wind vector (i.e. the wind speed). This norm can be written as the scalar product of the wind direction $\frac{\vec{V}}{|\vec{V}|}$ with the wind vector, which enables us to decompose the wind speed into a sum of contributions:

$$|\vec{V}| = \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V} \quad (6)$$

$$\Rightarrow |\vec{V}| = \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{ADVH} + \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{TURB} + \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{LSC} + \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{THW} + \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{KAT}. \quad (7)$$

Projected changes in near-surface wind speed between the end of the 21st and the end of the 20th century $\Delta|\vec{V}|$ can be decomposed as the sum of changes in the mean value of the scalar product computed on 3-hourly values of each accelerations with the wind direction vector:

$$\Delta|\vec{V}| = \Delta\left(\frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{ADVH}\right) + \Delta\left(\frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{TURB}\right) + \Delta\left(\frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{LSC}\right) + \Delta\left(\frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{THW}\right) + \Delta\left(\frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{KAT}\right) \quad (8)$$

Therefore, changes in near-surface wind speed between the end of the 21st and the end of the 20th century can be decomposed as a sum of scalar product (Fig. 2). In the rest of the paper, we will note ΔACC the "changes in wind speed due to a specific acceleration between 2080-2100 and 1980-2000", with **ACC** being the specific term considered (LSC, THW, KAT, ADVH,

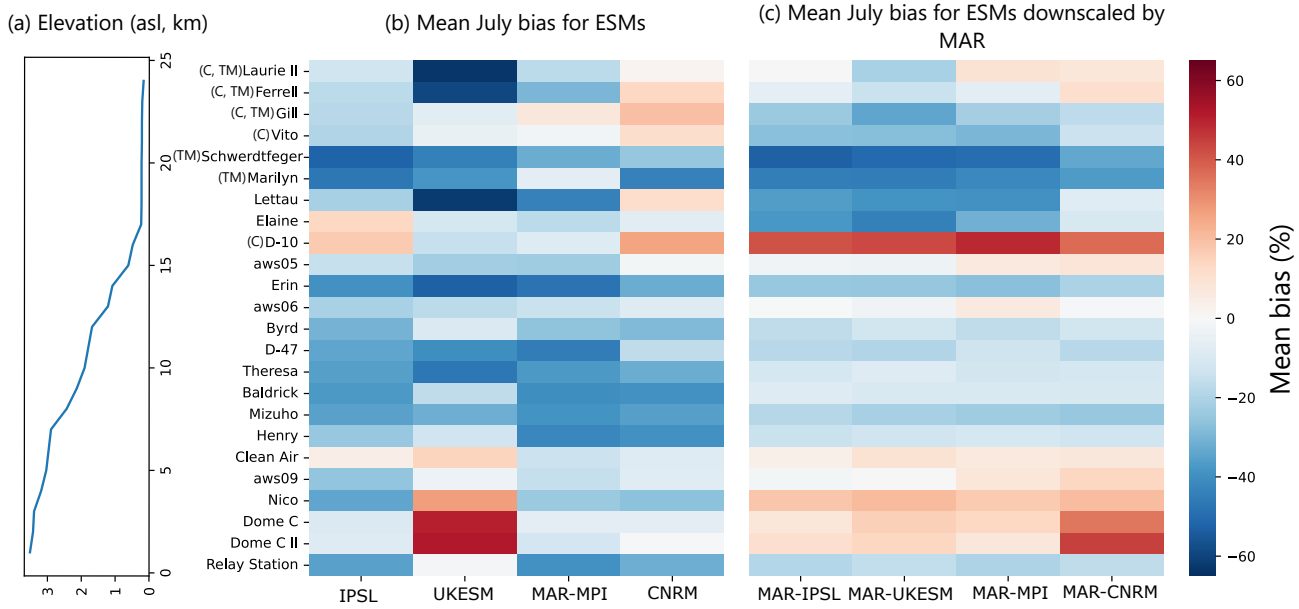


Figure 3. (a) Altitude of the selected stations. Mean normalized bias (B) for wind speed with regard to the AntAWS observations ($B = (|\overrightarrow{V_{GCM}}| - |\overrightarrow{V_{AntAWS}}|) / |\overrightarrow{V_{AntAWS}}|$ for (a) and $B = (|\overrightarrow{V_{MAR-GCM}}| - |\overrightarrow{V_{AntAWS}}|) / |\overrightarrow{V_{AntAWS}}|$ for (b)) for the 24 selected AntAWS stations, computed for July (b) using the GCMs, (c) using the GCMs downscaled by MAR.

TURB), that we define as follows:

$$\Delta ACC = \frac{\overrightarrow{V}}{|\overrightarrow{V}|} \cdot \overrightarrow{V_{ACC}}(2080 - 2100) - \frac{\overrightarrow{V}}{|\overrightarrow{V}|} \cdot \overrightarrow{V_{ACC}}(1980 - 2000). \quad (9)$$

255

3 Results

3.1 Evaluation of the models ability to represent near-surface winds in Antarctica

Most studies about the future evolution of near-surface winds in Antarctica across different models focus on direct monthly wind speed output of GCMs (Neme et al., 2022; Bracegirdle et al., 2008). Davrinche et al. (2024) showed the importance of boundary layer processes in representing surface wind accurately. However, GCMs often do not include an appropriate representation of the physics of the Antarctic boundary layer. Smith and Polvani (2017) show evidence of misrepresentation of the west-east Antarctica differences in the near-surface temperature field while Cuxart et al. (2000) mentions that GCMs commonly fail to represent the stability of the boundary layer. Here, we alleviate this shortcoming of GCMs by dynamically

260



downscaling GCM with the polar-oriented regional atmospheric model MAR (Section 2.2). We evaluate the value of the
265 downscaling by comparing biases in monthly mean 10-m wind speed computed between weather station observations (see
Sect. 2.1) and GCMs alone or downscaled by MAR (Fig. 3).

Overall, all GCMs tend to underestimate the mean wind speed, with the mean normalised bias across the 24 stations ranging
from -24 % for MPI, which demonstrates a consistent negative bias at all stations, to -13 % for CNRM. The latter exhibits
indeed a slight positive bias in coastal locations (Gill, Vito and D-10) that is compensated for by a negative bias everywhere
270 else (Fig. 3). In contrast, UKESM shows an inverse pattern, displaying substantial negative biases in coastal stations that are
partially offset by a pronounced positive bias at Dome C on the plateau.

We observe that biases are more similar between models downscaled by MAR than for raw GCMs (except for Dome C
and Dome C II in MAR-CNRM). Furthermore, the downscaling by MAR significantly reduces the mean bias compared to the
different GCMs in the sloped regions of Antarctica i.e. from AWS05 at 360 m above sea level to Henry at 2880 m above sea
275 level), where topography plays an important role in shaping the wind field. However, there is a consistent overestimation of
the weak winds of the Plateau across all downscaled models and an underestimation of the stronger winds in coastal areas.
Downscaling by MAR reduces the regional variability in wind speed bias on the continent.

Overall, downscaling by MAR significantly reduces the mean biases of the different GCMs, with the exception of stations
situated at the interface between the continent and the ocean (i.e. D-10) or in the Transantarctic mountains (Schwerdtfeger,
280 Marilyn and Lettau) (Fig. 3). With these coastal and Transantarctic AWS, there is a significant improvement of the mean
normalised bias for all models in December and annually, but in July, improvements are not statistically significant (Table 3).
However, if we discard the coastal and Transantarctic AWS, there is a significant improvement of the mean normalised bias for
all models and in all seasons.

To conclude, downscaling with a regional climate model significantly improves the representation of near-surface winds. A
285 finer resolution helps with topographic forcing, but the improved physics likely provides benefits in sloped terrains and on the
plateau.

3.2 Projected changes in near-surface winds by the end of the 21st century

In winter, all downscaled GCMs project a strengthening and poleward shift of the westerlies over the ocean (Fig. 4a and b),
more pronounced in IPSL and UKESM, which are also the models with the strongest changes in sea ice concentration (IPSL
290 and UKESM).

On the continent, changes are weaker, with larger differences among the downscaled models. Each of them features approx-
imately 50 % of the continental grid-cells exhibiting an increase and 50 % exhibiting a decrease in wind speed by the end of
the 21st century (Table 4). The ratio of significant decrease and significant increase remains approximately equal, both under
20 % except for MAR-IPSL which exhibits more significant increases (40 %) than significant decreases (6 %). Regions of sig-
295 nificant changes greatly vary among the downscaled models, with more significant decrease in coastal areas for MAR-UKESM
and MAR-CNRM, large patches of significant increases on the East Antarctic Plateau for MAR-IPSL and smaller-size sparse
patches for MAR-MPI (Fig. 4).

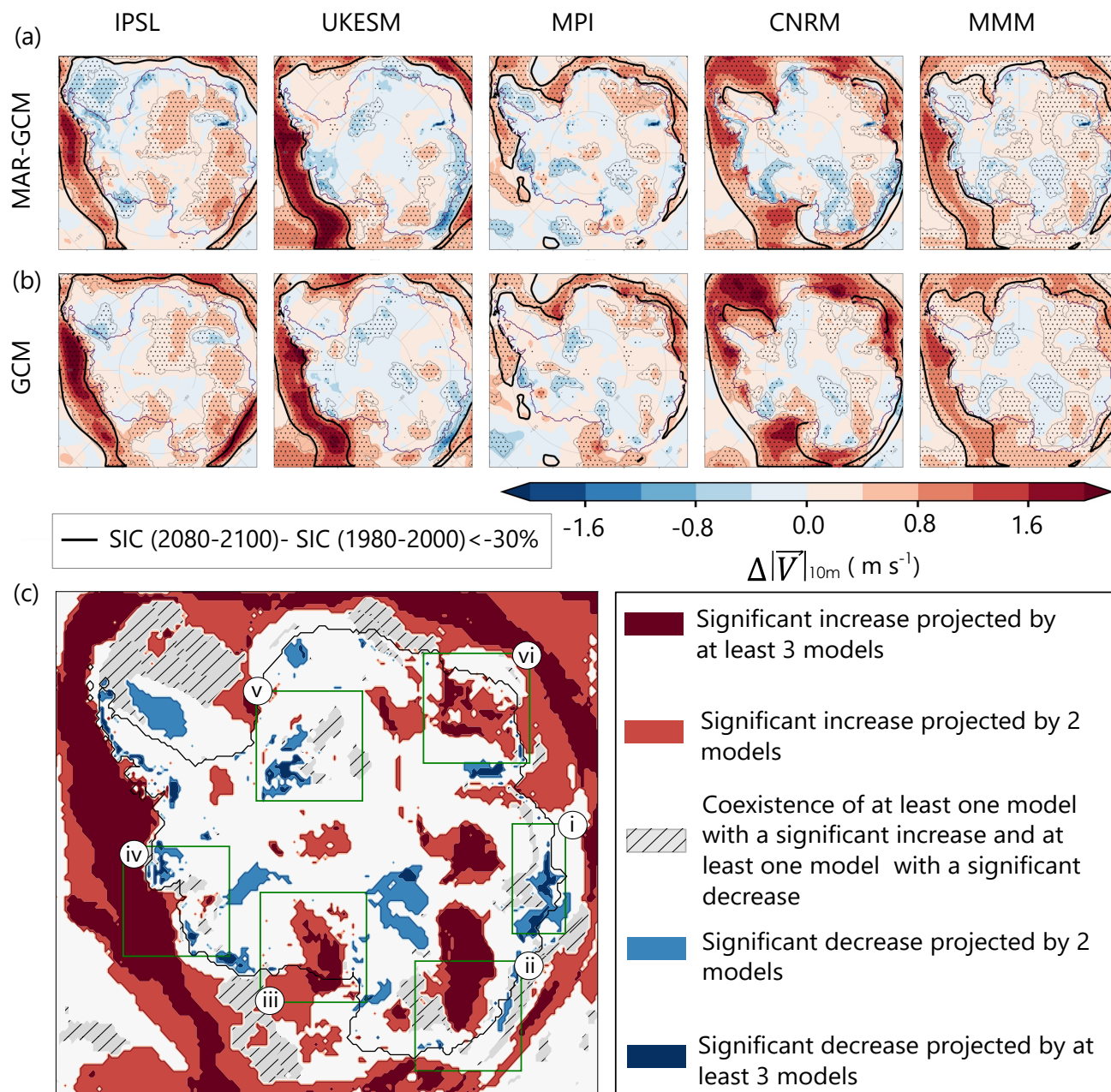


Figure 4. Projection of 10-m July wind speed changes between 2080-2100 and 1980-2000 ($\Delta|\vec{V}|_{10m}$) for GCMs downscaled by MAR (a) and for GCMs (b). MMM refers to the multi-model mean. Superimposed is the contour line at -30 % of the difference in Sea Ice Concentration (SIC) between July 2080-2100 and July 1980-2000 (black dashed line). (c) Map of the zones of significant near-surface wind speed changes between 2080-2100 and 1980-2000. Dark red (blue) areas represent zones for which at least 3 GCMs downscaled by MAR project a significant increase (decrease) of near-surface wind speed. Light red (blue) areas represent zones for which 2 models project a significant increase (decrease) of near-surface wind speed. Hashed grey areas indicate locations for which there is a significant disagreement between at least two models regarding the sign of evolution of near-surface wind speed. Green squares define 6 zones of interest which are used in the rest of the article: (i) Shackleton ice shelf, (ii) Adélie Land, (iii) Ross ice shelf, (iv) Pine Island Glacier (PIG), (v) Ronne ice shelf and (vi) Enderby Land.



Table 3. Improvement of the mean bias due to downscaling of the 4 GCMs computed as the difference between the absolute values of mean normalised bias of the July monthly wind speed output of GCMs (compared to AWS measurements) and the absolute values of mean normalised bias of the July monthly wind speed output of GCMs downscaled by MAR ($|B_{GCM}| - |B_{MAR-GCM}|$ in %). Positive values indicate an improvement due to downscaling while negative values indicate a decline. Significant improvements due to downscaling (computed using a t-test with a significance level of 0.1) are denoted by an asterisk (*). Values are given for the 28 AWS for which there is enough July months to create a climatology, for the 24 AWS presented in Table 1 that exhibit a coherent representation of the wind in ERA5 and for the 18 stations listed in Table 1 that are not in the Transantarctic mountains, nor on the shore (without TM/C)

	July			December			Annual		
Improvement due to downscaling (%)	28 AWS	24 AWS	without (TM/C)	28 AWS	24 AWS	without (TM/C)	28 AWS	24 AWS	without (TM/C)
IPSL	+4.4	+6.8*	+9.3*	+12.8*	+10.8*	+14.6*	+9.0*	+8.8*	+11.7*
UKESM	+1.3	+8.1	+9.8*	+23.4*	+21.8*	+20.1*	+7.0*	+11.1*	+12.1*
MPI	+0.2	+4.8	+10.7*	+11.1*	+12.1*	+16.0*	+8.1*	+10.0*	+16.0*
CNRM	-0.2	+1.8	+3.1*	+6.6*	+5.6*	+7.5*	+1.8*	+1.6*	+4.1*

Table 4. Percentage of continental grid-cells (including ice shelves) exhibiting an increase in July wind speed between 2080-2100 and 1980-2000 (significant or not, $\Delta|\vec{V}| > 0$), a significant increase in wind speed ($\Delta|\vec{V}| > 0^*$), no significant change in wind speed ($\Delta|\vec{V}| \sim 0$), a significant decrease in wind speed ($\Delta|\vec{V}| < 0^*$) and a decrease in wind speed (significant or not, $\Delta|\vec{V}| < 0$), for MAR-IPSL, MAR-UKESM, MAR-MPI, MAR-CNRM, for at least 3 downscaled models ($> 3M$) and for the multi-model mean (MAR-MMM)

Model	$\Delta \vec{V} > 0$	$\Delta \vec{V} > 0^*$	$\Delta \vec{V} \sim 0$	$\Delta \vec{V} < 0^*$	$\Delta \vec{V} < 0$
MAR-IPSL	71 %	40 %	55 %	6 %	29 %
MAR-UKESM	42 %	11 %	76 %	13 %	58 %
MAR-MPI	49 %	16 %	66 %	18 %	51 %
MAR-CNRM	52 %	18 %	72 %	11 %	48 %
$> 3M$	41 %	8 %	90 %	2 %	35 %
MAR-MMM	57 %	23 %	63 %	14 %	43 %

However, some areas display similar changes in all downscaled GCMs and in the multi-model mean (MAR-MMM, see right column in Fig. 4a). There is a significant increase on Ross ice shelf (Fig. 4c(iii)) for all models except MAR-MPI, a significant increase on Enderby Land (Fig. 4c(vi)) for all models except MAR-UKESM, a significant increase in Adélie Land (Fig. 4c(ii)) for all models and a significant decrease for all models except MAR-IPSL on Shackleton ice shelf (Fig. 4c(i)), Ronne ice shelf (Fig. 4c(v)) and on Pine Island Glacier (PIG) (Fig. 4c(iv)).

Even though downscaling by MAR significantly improves representation of near-surface winds (Sec. 3.1), projected 10-m wind speed changes between 2080-2100 and 1980-2000 using GCMs not downscaled by MAR show similar patterns of



Table 5. Percentage of the continental (including ice shelves) grid-cells exhibiting an increase or a decrease in the scalar product of wind direction and large-scale wind (first three columns), katabatic wind (columns 4 to 6), thermal wind (column 7 to 9) and the sum of katabatic and thermal wind (column 10 to 12). Metrics are computed as differences of the average values over the months of July between 2080-2100 and 1980-2000 for different models. MMM indicates changes in the multi-model mean while >3M indicates significant changes observed in at least 3 downscaled GCMs.

Model	Δ LSC			Δ KAT			Δ THW			Δ SURF		
	> 0*	< 0*	~ 0	> 0*	< 0*	~ 0	> 0*	< 0*	~ 0	> 0*	< 0*	~ 0
MAR-IPSL	38 %	2 %	60 %	3 %	36 %	61 %	22 %	11 %	67 %	8 %	34 %	58 %
MAR-UKESM	20 %	4 %	76 %	4 %	37 %	59 %	19 %	12 %	69 %	5 %	33 %	62 %
MAR-MPI	29 %	2 %	69 %	5 %	50 %	45 %	22 %	13 %	65 %	6 %	48 %	46 %
MAR-CNRM	25 %	8 %	67 %	4 %	52 %	44 %	29 %	7 %	64 %	12 %	43 %	45 %
>3M	9 %	0 %	91 %	1 %	33 %	66 %	11 %	2 %	87 %	3 %	27 %	70 %
MAR-MMM	48 %	5 %	47 %	5 %	66 %	29 %	34 %	13 %	53 %	10 %	59 %	31 %

305 evolution (e.g. an increase on Ross ice shelf and in Adélie Land) but however miss out on most of the significant decreases in near-surface winds (compare Table 4 with Table S1).

3.3 Projected changes in the components of near-surface winds

We investigate the drivers of changes in near-surface winds by performing a momentum budget decomposition (Section 2.3), allowing us to isolate the respective contributions of large-scale forcing, surface forcing (katabatic and thermal wind), and
 310 passive terms (advection, coriolis, and turbulence). The changes in each term are presented in Figure 5, and summary statistics of the prevalence of increases or decreases in the components of the wind in Table 5.

3.3.1 Changes in large-scale circulation

For the four GCMs downscaled by MAR, the increase of wind speed on the ocean is associated with an increase of the large-scale contribution (Fig. 5b), which is partially offset by an associated increase in turbulence (Fig. 5f). The Pearson correlation
 315 coefficient (R) between changes in wind speed on the ocean and changes in wind speed due to large-scale is greater than 0.7 for all models (Table S2). Note that MPI displays the weakest poleward shift and strengthening of the surface westerlies. It is also the model with the lowest ECS (Table 2), and the largest sea ice extent at present day.

This result is in agreement with previous studies that showed that the already observed increasing positive trend of the Southern Annular Mode (SAM index, computed as the zonally averaged pressure gradient between 40 °S and 65 °S (Marshall,
 320 2003; Miller et al., 2006)) will likely continue in response to increasing greenhouse gases and after the recovery of the ozone hole (which offsets the strengthening of the SAM (Bracegirdle et al., 2008)). As a consequence of the increased pressure gradient between the mid-latitudes and 65 °S, westerlies are strengthening and shifting poleward (Goyal et al., 2021; Fyfe, 2006).

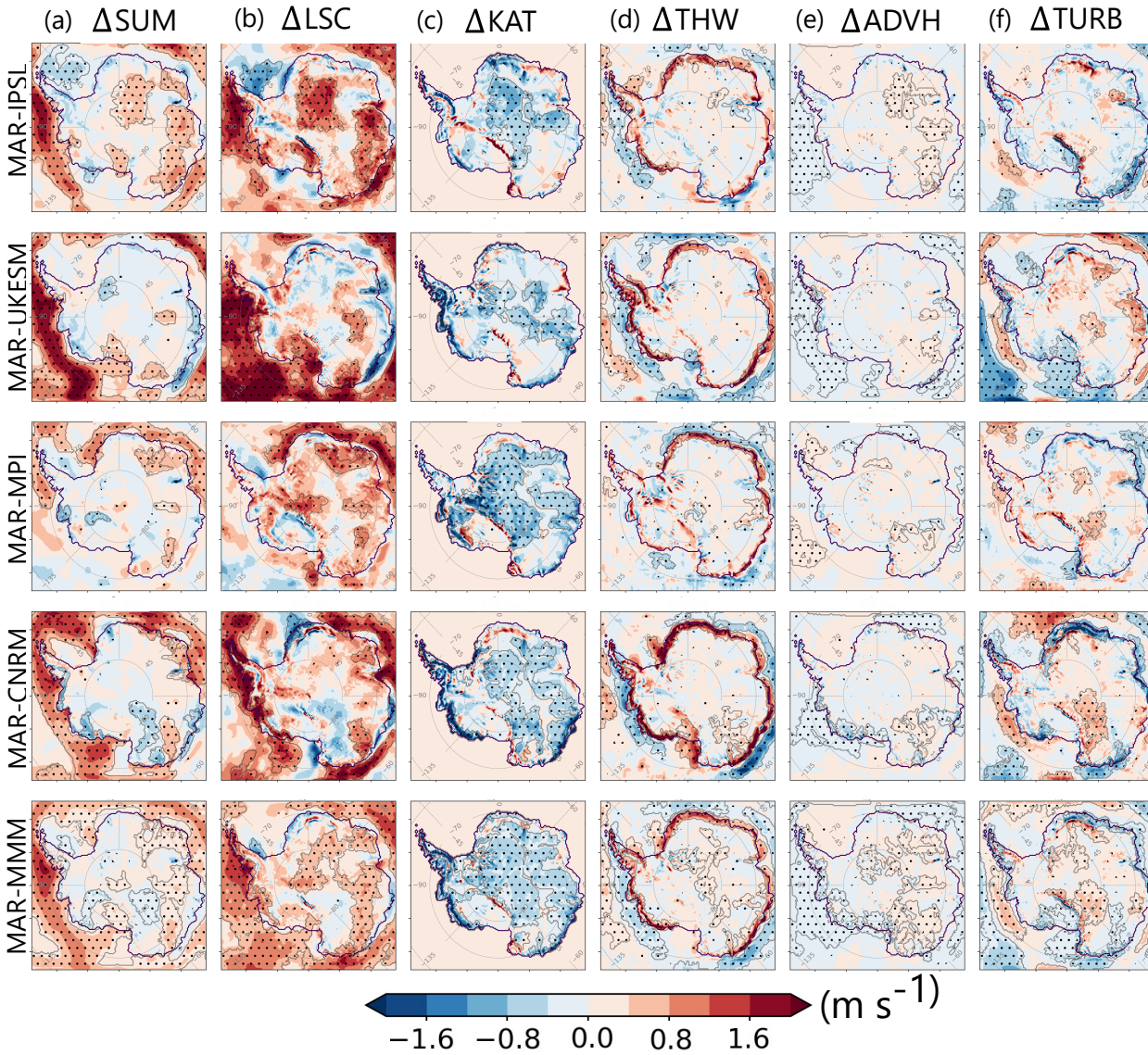


Figure 5. Projection of changes in 10-m wind speed between 2080-2100 and 1980-2000 associated with large-scale forcing ($\Delta LSC = \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{LSC}(2080-2100) - \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{LSC}(1980-2000)$, column b), katabatic forcing ($\Delta KAT = \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{KAT}(2080-2100) - \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{KAT}(1980-2000)$, column c), thermal wind forcing ($\Delta THW = \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{THW}(2080-2100) - \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{THW}(1980-2000)$, column d), advection ($\Delta ADVH = \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{ADVH}(2080-2100) - \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{ADVH}(1980-2000)$, column e) and turbulence ($\Delta TURB = \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{TURB}(2080-2100) - \frac{\vec{V}}{|\vec{V}|} \cdot \vec{V}_{TURB}(1980-2000)$, column f) and sum of all the above-mentioned forcings (large-scale, katabatic, thermal wind, advection and turbulence), which is equivalent to changes in wind speed ($\Delta SUM = \Delta LSC + \Delta KAT + \Delta THW + \Delta ADVH + \Delta TURB$, column a), see Fig. 2. Dotted areas indicate locations for which changes are significant at a 80 % level, significant area larger than 350 km² are highlighted with a grey solid line.



The increase in westerlies is also closely related to changes in the extent of sea ice, shown in thick black lines in Figure 4. For GCMs with low sea ice loss (IPSL and UKESM), the poleward shift of the westerlies does not extend up to the coastline in the Indian sector (20-90° E) in East Antarctica, while it does for models with strong sea ice extent loss (MPI and CNRM). MPI retains a significant amount of sea ice in the Pacific sector at the end of the 21st century, where other models show a retreat, and thus does not show an increase in the large-scale wind as others do.

On the continent, the results are much less homogeneous. Most significant changes in large-scale acceleration are positive (48% in the MMM, Table 5) and some locations such as Adélie Land (Figure 7) or Enderby Land exhibit a significant increase in large-scale forcing in all models. Aside from these areas, models disagree on the exact location of significant changes: MAR-IPSL and MAR-MPI project for example a significant strengthening of large-scale acceleration on Ross ice shelf while MAR-MPI and MAR-CNRM projects a non significant weakening. Everywhere else in Antarctica, MAR-IPSL and MAR-MPI project an overall increase of large-scale acceleration while MAR-UKESM and MAR-CNRM exhibit some significant weakening of coastal easterlies on Shackleton ice shelf and in Queen Maud Land (between Ronne ice shelf and Enderby Land). From Fig. 5b, we also observe that the largest inter-model differences in the forcing of wind changes originate from differences in the large-scale pattern of change.

These inconsistencies are related to variable trends in large-scale pressure gradients that are different between models. Although the trend in SAM is well understood, and reproduced by most models (MPI does not show a clear trend), the changes in the pressure gradient between the circumpolar trough at 65 °S and the pole is much less clear, and inconsistent between models. In Antarctica, computing the pressure gradient based on the mean sea level pressure results in strong biases because of the extrapolation of the pressure under the surface layer. Instead, we looked directly at the difference between the mean geopotential height and mean geopotential height at 65 °S at 500 hPa (Figure S11). For MAR-UKESM, on the interior, the difference with the geopotential height at 65 °S becomes more negative at the end of the 21st, meaning that the polar cell is strengthening. It is the opposite for MAR-IPSL and MAR-MPI, and there is on average no change for MAR-CNRM. However, we found no evidence of a correlation between a strengthening of the polar cell and an intensification of the large-scale pressure gradients at the surface. The attribution and robustness of changes in the large-scale pressure gradients remain to be evaluated.

3.3.2 Changes in surface forcing

On the continent, for all GCMs downscaled by MAR, we find a consistent weakening of the katabatic forcing (Fig. 5b). This decrease is large on the coast in the Amundsen sea sector and in Adélie Land for MAR-CNRM, MAR-MPI and MAR-UKESM. Across all downscaled models, changes are also large and significant in the interior, even in locations where slopes are gentle.

Katabatic forcing is indeed computed as the product of the slope and the strength of the inversion layer ($\Delta\theta$ in Eq. (3)). Here, as the surface slope is not changing, the significance of changes in $\Delta KAT = \frac{\bar{\nabla}}{|\bar{\nabla}|} \cdot \overrightarrow{\nabla_{KAT}}(2080 - 2100) - \frac{\bar{\nabla}}{|\bar{\nabla}|} \cdot \overrightarrow{\nabla_{KAT}}(1980 - 2000)$ (see Eq. (2.3.3)) reflects the significance of changes in the inversion strength due to Antarctic surface warming. These changes are larger in areas where the inversion strength is large at present day ($\Delta\theta > 20^\circ C$): the high plateau and the ice shelves (Figure S6), which explains the significant changes at the center of Antarctica.



Associated with the changes in $\Delta\theta$, the depth of the temperature deficit layer $\hat{\theta}$ also changes. It reduces considerably near the coastline (Figure S7), causing a reduction in thermal wind (Figure 5d). Because the latter on average opposes the direction of the downslope winds (Davrinche et al., 2024), a weakening of the thermal wind increases the resulting wind speed and compensates for the decrease in katabatic acceleration. The compensating effect of thermal wind is particularly pronounced in coastal East Antarctica where it often surpasses the decrease in katabatic forcing (Figure S8). As thermal wind and katabatic forcing both result from the forcing by the surface ($\text{SURF} = \text{KAT} + \text{THW}$), in the rest of the study we will call "changes in the forcing by the surface" (ΔSURF) the changes in wind speed linked to changes in the sum of katabatic and thermal wind forcings. Overall, SURF increases on the coastline and decreases elsewhere.

3.3.3 Changes in passive terms

Turbulence, Coriolis, and advection accelerations can be viewed as passive terms, as they only come into play once the motion has been triggered by an active term such as katabatic, large-scale or thermal wind (Van den Broeke and van Lipzig, 2002). Here, the contribution of horizontal advection is negligible almost everywhere, except on the Amery ice shelf. Unlike the advection, the turbulent forcing is strong and encompasses surface drag. Therefore, it resembles (but with an opposite sign) changes in the sum of the dominant active accelerations. Changes in the scalar product of turbulent wind vector and the wind direction (ΔTURB , Fig. 5f) are positive when friction decreases and negative when friction increases. ΔTURB increases in all downscaled models on the ocean where westerlies intensify the most, decreases in the coastal margins in locations where easterlies weaken, and increases overall in the interior.

To conclude, Figure 5 shows that, although surface wind changes during the 21st century are small on the continent, and often not consistent between models, they result from the complex interplay between changes in the large-scale forcing that generally induce an increase in wind speed, and changes in the surface forcing that mostly induce a decrease in wind speed under a high emission scenario. The change in surface forcing results from a reduction in the surface temperature inversion, and is consistent between models over the whole continent. The changes in large-scale forcing however, vary greatly between models, with some regions of consistent changes (Adélie Land, Enderby Land, Shackleton, Ross and Ronne ice shelf and Pine Island Glacier). In the following sections, we explore in more details the regions of significant increase and decrease in wind speed across models, to attribute these changes more precisely.

3.4 Drivers of significant regional increases in wind speed across downscaled GCMs

There are multiple zones where GCMs downscaled by MAR agree on an increase in wind speed, including Ross ice shelf, Adélie and Enderby Land, and in these areas, significant increases in wind speed are statistically more linked to changes in large-scale forcing than to other forcings. For all downscaled models, in locations where the increase in wind speed by the end of the 21st century is significant, there is indeed more than 6 times more grid-cells exhibiting a significant increase in large-scale forcing than an increase in forcing by the surface pressure gradients (see Fig. 6a, 6c and Table S3). Furthermore, the proportion of significant increases in the large-scale forcing is greater among grid-cells exhibiting significant increases in

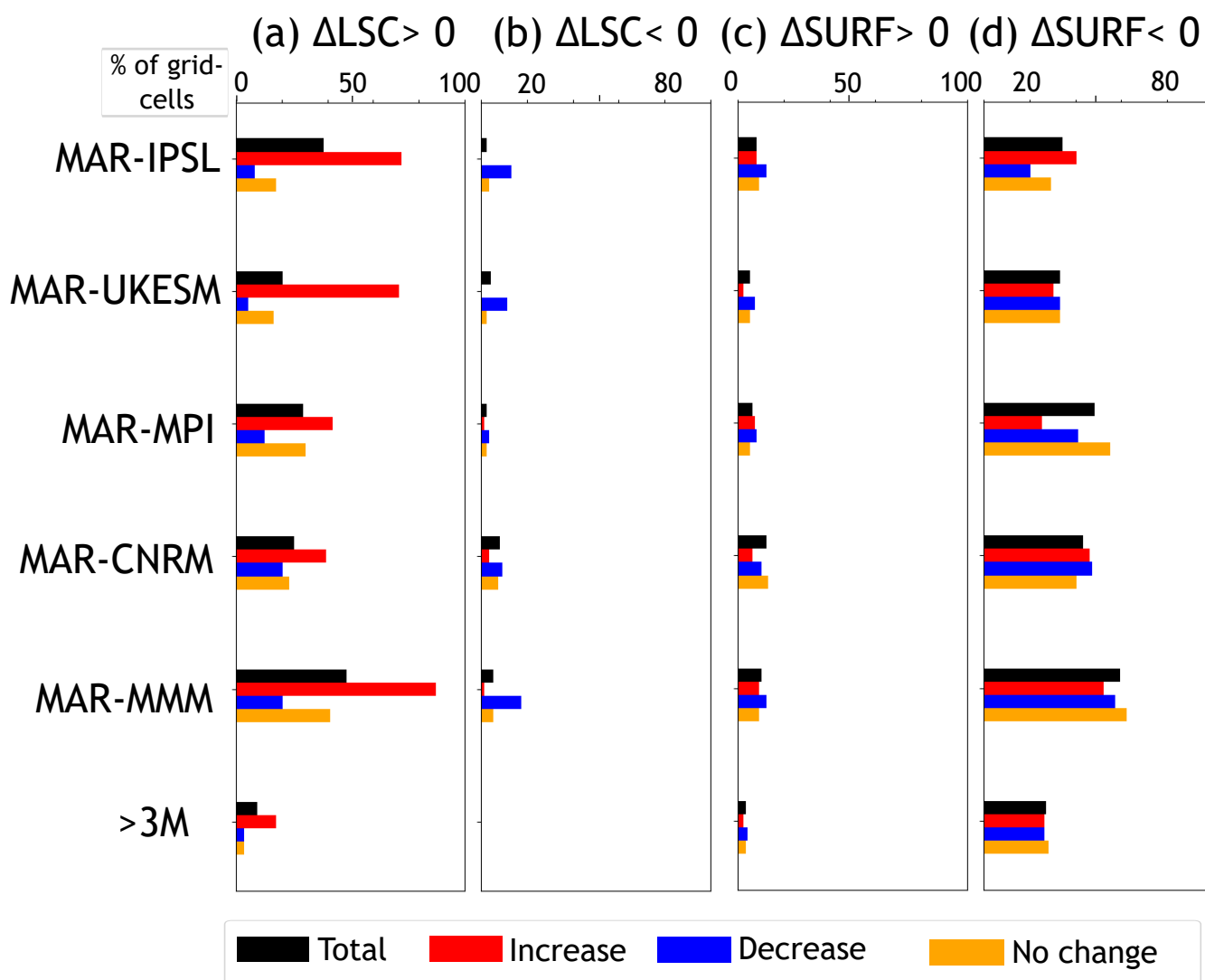


Figure 6. Percentage of the continental grid-cells exhibiting a significant (a) increase or (b) decrease in large-scale forcing or (c) increase or a (d) decrease in surface forcing in all Antarctica (black bars), among grid-cell exhibiting a significant increase (red bars) or decrease (blue bars) or no change (orange bars) in July wind speed between 2080-2100 and 1980-2000. MAR-MMM indicates changes in the multi-model mean (MAR-IPSL, MAR-UKESM, MAR-MPI and MAR-CNRM) while >3M indicates significant changes observed in at least 3 GCMs downscaled by MAR.

wind speed (Fig. 6a) than in all the continental grid-cells. This indicates that significant increases in wind speed are likely
390 linked to significant increases in large-scale pressure gradient forcing.

More specifically, in Adélie Land, there is a large area (denoted by a black and yellow dashed line on Fig. 7) where all GCMs downscaled by MAR agree on a significant increase in both wind speed ($\Delta|\overline{V}| > +0.4 \text{ m s}^{-1}$ for all downscaled models) and



large-scale forcing ($\Delta LSC > +0.6 \text{ m s}^{-1}$ for all downscaled models, see Table S4 and Fig. 7a and b) while changes in the surface forcing (KAT+THW, Figure 7c) are weaker ($-0.2 < \Delta SURF < 0.4 \text{ m s}^{-1}$ for all downscaled models, see Table S4). In this specific area, changes in wind speed are well correlated with changes in large-scale forcing ($R > 0.7$ for all downscaled models except MAR-MPI for which $R \sim 0.3$). The same conclusion can be drawn for Enderby Land (Figure S9 and Table S5).

While it is clear from the analysis of Adélie and Enderby Land that significant increases in the large-scale forcing drives changes in the near surface wind speed, analysis of Ross ice shelf indicates that surface forcing can also contribute to significant wind speed increase. On the Ross ice shelf (Fig. 8), there is a patch for which all GCMs downscaled by MAR project a significant strengthening of wind speed, except MAR-MPI. For MAR-IPSL and MAR-UKESM, significant increases in wind speed are associated with significant increases in large-scale forcing (Fig. 8b): on average, $\Delta LSC > +0.9 \text{ m s}^{-1}$ for these two downscaled models, while $\Delta SURF$ is negative (see Table S6). On the other hand, for MAR-CNRM and MAR-MPI, the increase in wind speed is associated with no change in large-scale forcing ($\Delta LSC \sim 0 \text{ m s}^{-1}$ for all downscaled models, see Table S6) and an increase in surface forcing ($\Delta |\overline{V}| > +0.3 \text{ m s}^{-1}$) for both downscaled models, but only statistically significant for MAR-CNRM (Fig. 8e). Overall, on the Ross ice shelf, trends are not consistent across model for any of the forcings (Fig. 8e).

As a conclusion, significant increases in wind speed are on average more linked to significant increases in large-scale forcing but in some areas, they can also result from the changes in the surface forcing as well. Averaging over the whole continent would mask the influence of the forcing by the surface.

3.5 Drivers of significant regional decrease in wind speed across GCMs downscaled by MAR

For all GCMs downscaled by MAR, significant decreases in wind speed are rarer (14 %) than significant increases (23 %, Table 4). Additionally, in locations where the decrease in wind speed by the end of the 21st century is significant, there is between 1.5 (MAR-IPSL) and 14 (MAR-MPI) times more grid-cells exhibiting a significant decrease in surface forcing than a decrease in large-scale pressure gradients (Fig. 6b and 6d; Table S3 and S7). This indicates that the decreases in total wind speed result from changes in the surface pressure gradients (SURF = KAT + THW) forcing. We have noted before that SURF decreases significantly in more than 30 % of grid cells, but the wind speed is significantly lower in only 6 to 18 % of the grid cells. We hypothesize that the wind speed significantly decreases only when the decrease in SURF is not masked by an increase in large-scale pressure gradients, i.e. where in large-scale pressure gradients are either weak or negative.

On PIG for instance, there is an area (top left on Fig. 9, denoted by a black and yellow dashed line) where all GCMs downscaled by MAR, except MAR-IPSL, agree on a significant decrease in both wind speed and surface forcing (Fig. 9a and e) while changes in the large-scale forcing (Fig. 9b) are weak (for MAR-MPI) to positive (MAR-UKESM and MAR-CNRM). For all continental grid cells in the PIG region exhibiting a decrease in wind speed, changes in the surface forcing is negative in all downscaled models ($\Delta SURF < 0.4 \text{ m s}^{-1}$, see Table S8) while changes in large-scale forcings are mostly positive, except for MAR-MPI ($\Delta LSC = -0.25 \text{ m s}^{-1}$). As a conclusion, on PIG, changes in surface forcing are not masked by changes in large-scale forcing and drive the decrease in near-surface wind.

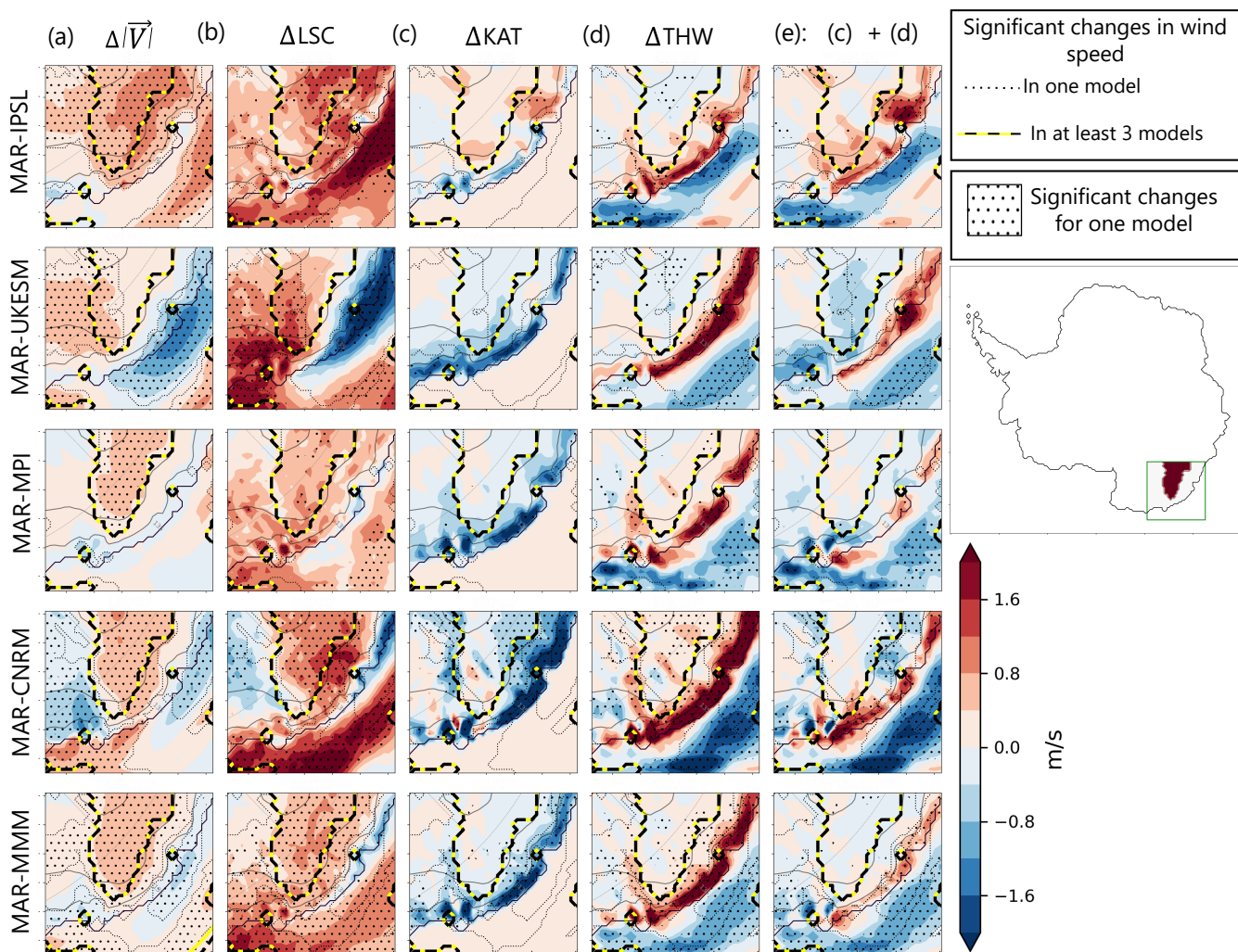


Figure 7. Projections of 10-m changes in July wind speed in Adélie Land between 2080-2100 and 1980-2000 (a), linked to large-scale forcing (column b), katabatic forcing (column c), thermal wind forcing (column d) and total surface forcing (sum of katabatic and thermal wind, column e) for MAR-IPSL (line 1), MAR-UKESM (line 2), MAR-MPI (line 3), MAR-CNRM (line 4) and the multi-model mean of the 4 downscaled GCMs (line 5). Dotted areas indicate locations for which changes are significant at a 80 % level for the metric and the model considered. Dotted lines indicate areas for which changes in wind speed ($\Delta|\vec{V}|$) are significant at a 80 % level for the considered model while dashed black and yellow thick lines indicate locations for which changes in wind speed ($\Delta|\vec{V}|$) are significant at a 80 % across at least 3 downscaled models. Solid grey lines indicate elevation contours (1000, 2000 and 3000 m).

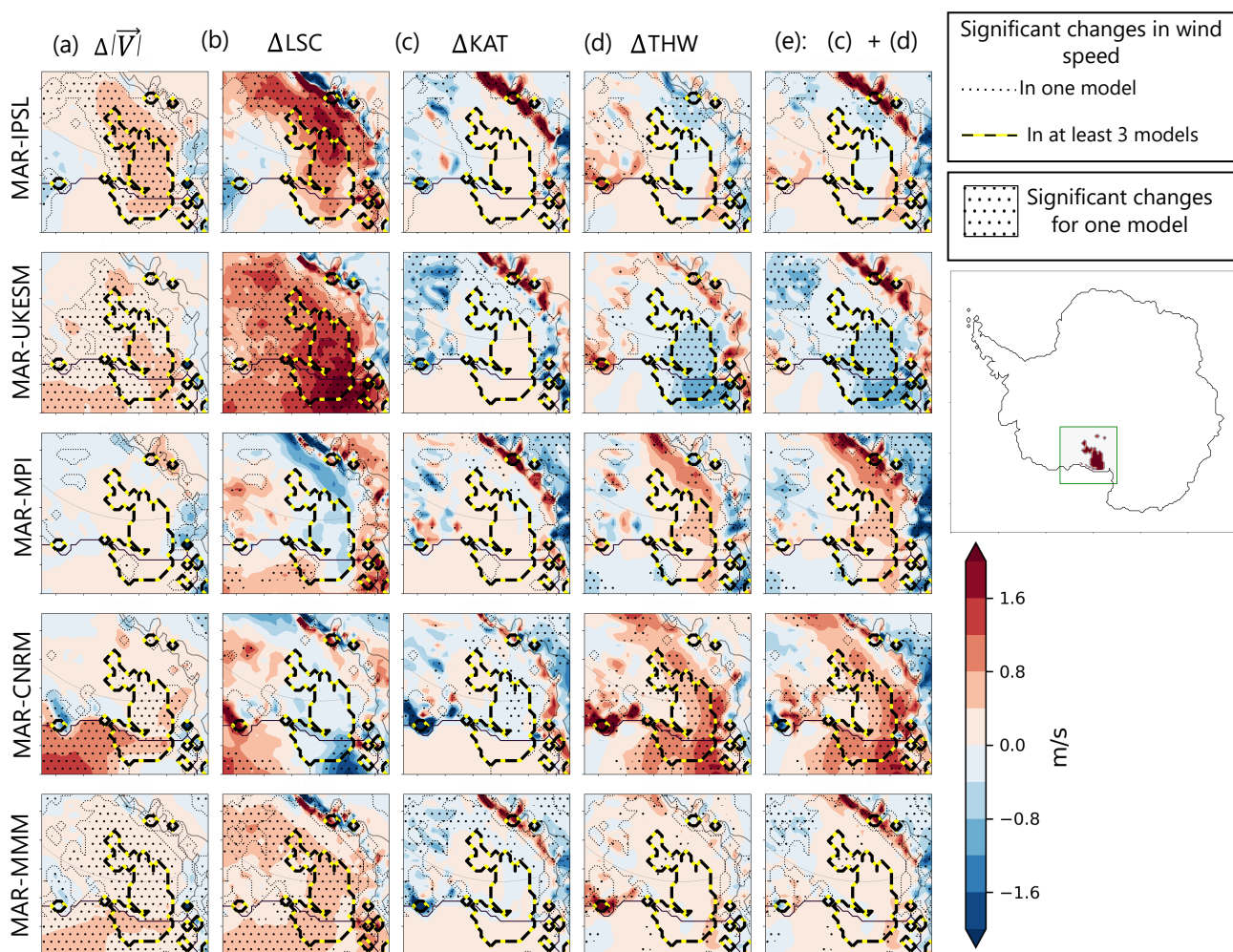


Figure 8. Same as Fig. 7, but for Ross ice shelf.

Similarly, on Shackleton ice shelf (Fig. 10) and on Ronne ice shelf (Figure S10), all downscaled models except MAR-IPSL agree on a significant decrease in both wind speed and surface forcing (Fig. 10a and 10e) while changes in the large-scale forcing 10c) are either positive ($+0.5 \text{ m s}^{-1}$ for MAR-IPSL for continental grid cells exhibiting a significant decrease) or weaker than the changes in near surface forcings (see Table S9). Therefore, changes in surface forcing are not masked by changes in large-scale pressure gradients and drive the decrease in near-surface wind.

As a conclusion, significant decreases in wind speed across multiple GCMs are on average more linked to a significant decrease in surface forcing.

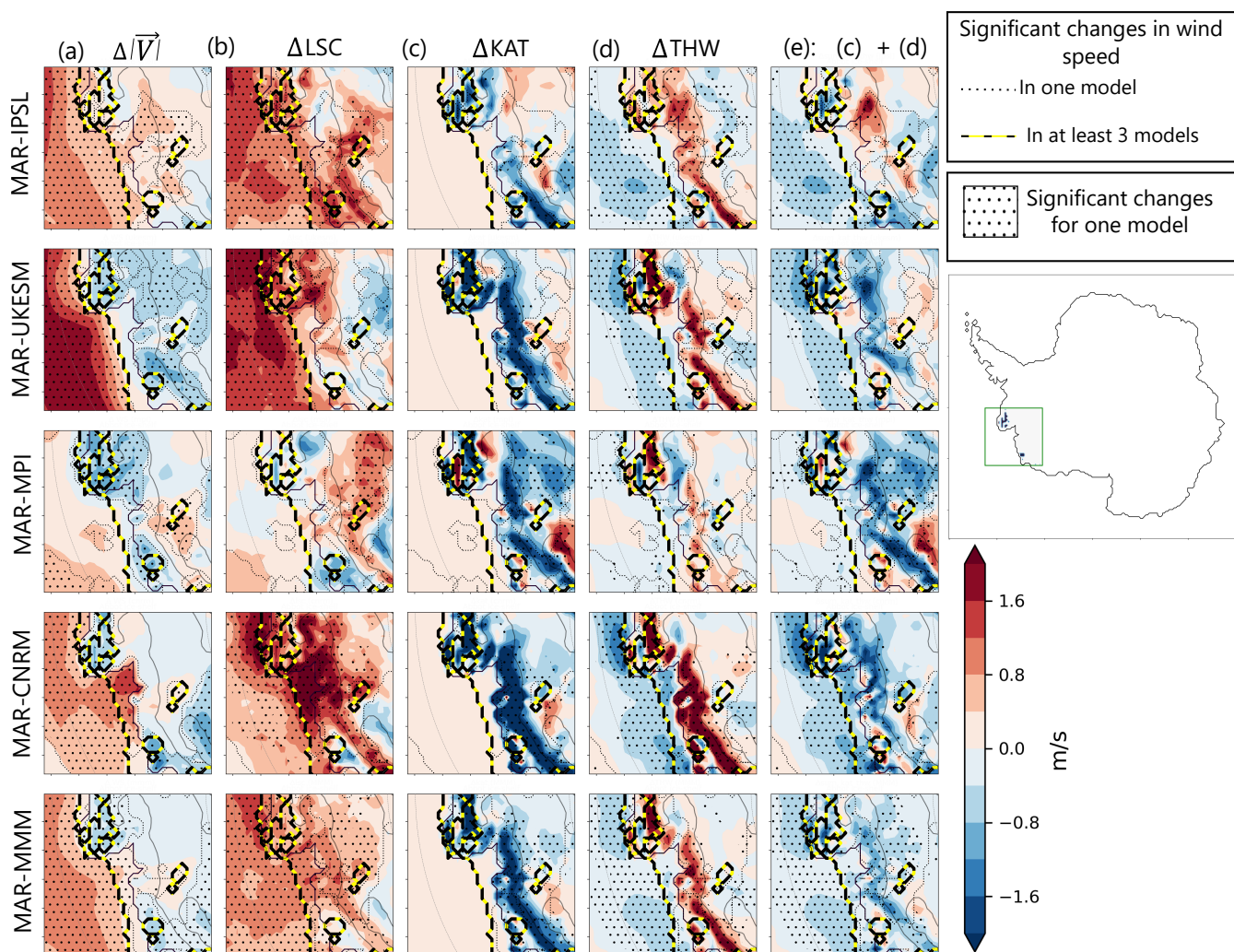


Figure 9. Same as Fig. 7, but for PIG.

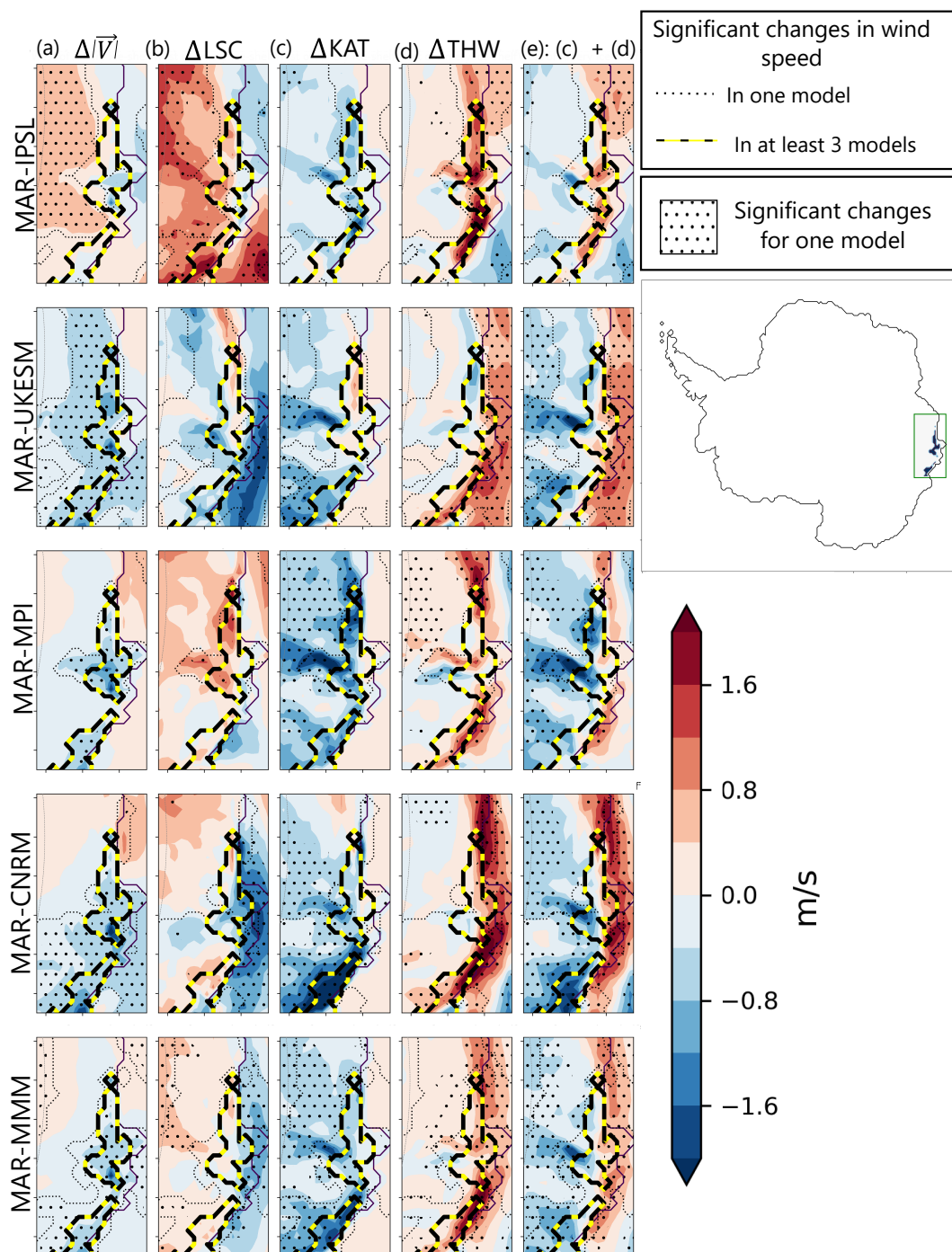


Figure 10. Same as Fig. 7, but for Shackleton ice shelf.



4 Discussion and conclusions

For all four downscaled GCMs, under the SSP585, there is a clear strengthening and poleward shift of the westerlies around Antarctica during the XXIst century, linked to changes in the large-scale forcing. GCMs with strong sea ice loss also exhibit a more pronounced poleward shift, linked to their changes in the SAM.

On the continent, changes in wind speed are much weaker and with regional disparities. Downscaled GCMs agree on a significant strengthening of near-surface wind speed in Adélie Land, on Ross Ice Shelf and Enderby Land. All downscaled models show evidence of decreasing easterlies locally, but their location vary greatly across models (in East Antarctica for MAR-UKESM and MAR-CNRM, west of Dronning Maud Land for MAR-IPSL or west of Ross ice shelf for MAR-MPI), which results in few areas of significant decrease in the multi-model mean.

These patterns of change projected with MAR forced by 4 different GCMs are similar to those projected by the GCMs alone. However, when we look into the details, the GCMs alone do miss a few significant changes both on the continent and on the ocean. Decrease in coastal easterlies in all models are stronger in the MAR downscaling, where changes in the surface forcing are likely better represented.

For all GCMs downscaled by MAR, under the SSP585 scenario, the temperature inversion at the surface of the continent ($\Delta\theta$) weakens (between -6% averaged over the continent for MAR-UKESM and -10% for MAR-MPI). The strongest decrease in $\Delta\theta$ are found in the interior and on the ice shelves (Figure S6). Consequently, there is a significant decrease of the katabatic forcing, consistent across downscaled GCMs, in coastal regions and in the interior as well.

Simultaneously, due to warming of the surface, the ability of coastal margins to accumulate cold air at the foot of the slope is reduced (Figure S7d). Therefore, we observe a significant weakening of thermal wind forcing as well in coastal areas. Our results are consistent with a previous study (Bintanja et al., 2014b), who hypothesized that surface warming induces a decrease in the static stability of the inversion layer, leading to an increase in vertical momentum transfer.

Because the thermal wind opposes the dominant direction of the downslope winds (Davrinche et al., 2024), a weakening of the thermal wind forcing increases the resulting wind speed and compensates for the decrease in katabatic acceleration. The compensating effect of thermal wind is particularly pronounced in coastal East Antarctica where it often surpasses the decrease in katabatic forcing, leading to an overall increase of the wind speed due the surface forcing. For the large-scale forcing, it exhibits larger areas of significant increases than decreases. From our statistical analysis and case studies, we conclude that (i) significant decrease in wind speed are statistically more linked to changes in surface forcing, when not masked by an increase in large-scale forcing (as shown on Shackleton, PIG and Ross ice shelves), and (ii) significant increases in wind speed are statistically more linked to changes in large-scale forcing (as shown in Adélie, Enderby Land and Ronne ice shelf).

We have also investigated the link between the strengthening of the polar cell and the large-scale pressure changes at the surface and could not identify an obvious link between the two of them (Figure S11). The significance of changes in large-scale pressure gradients, as well as their attribution to specific mechanisms remain to be established, with an extension of this study to more models with different dynamical responses to anthropogenic warming.



Changes in the mean value of winter near-surface wind speed are likely to impact the quantity of drifting snow and sublimation, and the stability of the ice shelves through potential enhanced surface melt (Lenaerts et al., 2017). We expect sublimation and drifting snow to be reduced in case of a weakening of the wind speed. However, further studies should be performed to quantify these effects.

470 We have performed this study with a fixed topography on the continent. Therefore, we have not assessed whether changes in both large-scale and surface forcing might be affected by change in topography linked to dynamical losses of the Antarctic ice sheet. Future work should be done to study the effect of a changing topography on the projections of near-surface winds.

Finally, we would like to nuance the findings of Bintanja et al. (2014b) that stated that climate-related (zonally averaged) wind speed changes over the continent are insignificant with respect to the interannual variability and can only be linked to
475 changes in the large-scale forcing. We show evidence that different areas with roughly the same latitude can have opposite but significant projected changes in near-surface winds (namely Adélie Land and Shackleton ice shelf for instance) and that these changes can originate either from changes in the surface forcing or from changes in the large-scale pattern of circulation.

Code and data availability. All Codes and dataset to analyze future changes in near-surface winds in Antarctica under the SSP585 scenario are available at <https://zenodo.org/records/14191007>. Data of the AntAWS are available from Wang et al. (2023) (<https://doi.org/10.48567/key7-ch19>).
480

Author contributions. CD, CeA, and AO designed the study and contributed to the output and observation analyses. CeA, ChA, and CK set up the MAR model for Antarctica with several adaptations. ChA, CK, CeA and CD performed model simulations. CD performed the momentum budget decomposition, post-processed the data, did the bulk of the analysis, and made all the figures. CD wrote the first draft, with input from CeA and AO. All authors contributed to discussions in writing this paper.

485 *Competing interests.* The contact author has declared that none of the authors has any competing interests.

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