Interseasonal Connections between the Timing of the Stratospheric Final Warming and Arctic Sea Ice

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ABSTRACT

Connections across seasons in atmospheric circulation and sea ice have long been sought to advance seasonal prediction. This study presents a link between the springtime stratosphere and Arctic sea ice in summer through autumn. The polar stratospheric vortex dominates the winter stratosphere before breaking down each spring, which is called the stratospheric final warming, as solar radiation returns to the pole. Interannual variability of this breakdown is dynamically driven, leading to different springtime tropospheric and surface circulation patterns. To examine the different impacts of delayed and early final warmings, a multimodel composite was generated from selected CMIP5 models. Additionally, regressions were performed on JRA-55 against an index of springtime polar vortex strength. In both the multimodel composites and reanalysis regressions, significant anomalies in sea ice thickness persist several months following an anomalous timing of the final warming. A later final warming or stronger springtime polar stratospheric vortex leads to negative sea ice thickness anomalies in the East Siberian Sea and positive anomalies in the Beaufort Sea in comparison with an earlier final warming or weaker polar vortex. The spring polar stratospheric vortex is related to spring polar surface circulation patterns. The winds associated with this pattern induce anomalous sea ice motion, moving ice from the East Siberian Sea toward the Beaufort Sea. Reduced sea ice in the East Siberian Sea is linked to anomalous warmth over this region in autumn. Our results suggest that the timing of the stratospheric final warming exerts an influence on the tropospheric circulation and sea ice through autumn, which has implications for seasonal climate prediction.

1. Introduction

The stratospheric polar vortex has a robust annual cycle, spinning up each autumn and breaking down each spring, when the typical wintertime westerlies are replaced by easterlies until the subsequent autumn. This

transition from a winter to a summer circulation regime is known as the stratospheric final warming (SFW) and is driven by solar radiation, though dynamics also have an influence (e.g., [Andrews et al. 1987](#)). SFWs take place every year in both hemispheres, although they show weaker interannual variability in the Southern Hemisphere, where dynamical influences are weaker, than in the Northern Hemisphere ([Vaugh et al. 1999](#)). For instance, the date of occurrence of SFWs oscillates around 2 months in the Northern Hemisphere, but less than 1 month in the Southern Hemisphere ([Vaugh et al. 1999](#)).

Similar to sudden stratospheric warming (SSW) events that induce a negative phase of the Arctic Oscillation (AO) on time scales of 1–2 months ([Baldwin and Dunkerton 2001](#)), SFWs may also exert an influence on

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the troposphere (Black et al. 2006). The tropospheric fingerprint following SFWs resembles the negative AO pattern, but with some differences in the location of its centers of action (Black and McDaniel 2009). Moreover, the large interannual variability in the date or in the vertical structure of the vortex breakup also affects the troposphere, particularly in the month of April (Ayarzagüena and Serrano 2009; Hardiman et al. 2011). In all cases, the pattern of influence has an AO-like structure with an anomalous center at high latitudes.

The AO has also been linked to Arctic sea ice cover changes due to its influence on high-latitude temperature and wind. Rigor et al. (2002) found the winds associated with the positive AO phase cause a wind-driven movement of sea ice, which map onto one of the main Arctic sea ice circulation regimes presented in Proshutinsky and Johnson (1997). The AO induces anomalous sea ice motion and therefore affects sea ice concentration and thickness. More specifically, the AO enhances the Beaufort Gyre and the trans-Arctic drift. The anomalous circulation modes are also seen in the recent work of Park et al. (2018), who additionally found that the winter AO signal is associated with the Arctic sea ice thickness anomalies in the following autumn.

Sea ice, which floats on seawater, is susceptible to both ocean currents from below and atmospheric winds from above. Previously, Thorndike and Colony (1982) found that on monthly and seasonal time scales (those of interest here), the split is approximately 50% between ocean and atmosphere driven, though Rigor et al. (2002) indicates the atmosphere drives most of the variance on time scales shorter than annual. Differences in the relationship between sea ice motion and atmospheric wind can also be due to distance from land, as Thorndike and Colony (1982) found that geostrophic winds are responsible for the wind driving outside of the nearest approximately 400 km to land boundaries. The precise directionality of influence the atmospheric winds have on sea ice motion is determined by how compact the ice cover is (Fissel and Tang 1991). If the ice cover is compact, ice motion is much closer to the wind direction, while free drifting ice will move at an angle to the wind direction (Tansley and James 1999).

The consideration of atmospheric circulation can then help in the prediction of sea ice changes. This prediction is recently attracting large interest from the scientific community, not only because of the impact of sea ice changes on the polar region but also on midlatitudes (e.g., Cohen et al. 2014). However, the low predictability of the polar tropospheric weather beyond 2 weeks hampers longer predictions of the sea ice (Guemas et al. 2014). In addition, most studies devoted to the atmosphere–sea ice coupling have so far only considered the effects of the

troposphere. The inclusion of the stratospheric-induced anomalies over the pole may improve the predictability of the Arctic sea ice. Indeed, recently Smith et al. (2018) analyzed the interaction between the wintertime polar stratosphere and sea ice cover through the lens of strong polar vortex (SPV) events and SSWs. Composites of strong and weak polar vortex events were found to have roughly opposing patterns of sea ice thickness and concentration anomalies, with significant differences in multiple basins. Years with an SSW event featured positive sea ice thickness anomalies in the East Siberian and Laptev Seas, with negative anomalies in the Beaufort Sea, while the opposite was found in years with a wintertime SPV event. Sea ice concentration anomalies, however, were primarily seen in the Sea of Okhotsk and Bearing Sea in winter and spring, though in summer, concentration anomalies are seen in the East Siberian and Laptev Seas, with SSW years having mainly positive anomalies, and SPV years having primarily negative anomalies. Their results from CESM1(WACCM) simulations matched reanalysis for individual basins and selected seasons, though fewer years and nonlinear trends in the reanalysis pose difficulties.

Another recent study by Kelleher and Screen (2018) suggested a stratospheric influence on sea ice area in an ensemble of CMIP5 models, associated with polar vortex perturbations in late winter/early spring. The implications of this result for the forecast of Arctic sea ice might be large because sea ice conditions in spring (the start of sea ice melt) are important for summer/autumn ice (e.g., Kapsch et al. 2013; Schröder et al. 2014; Petrie et al. 2015) due to the long time scale of ice variability. However, Kelleher and Screen (2018) examined the influence of the stratosphere on sea ice area primarily in an Arctic-wide sense, thus more detail is needed on basin specific impacts.

In this study, we aim to analyze in detail the influence of the interannual variability of SFW timing on Arctic sea ice in the following months. To accomplish this, we have used the output of several CMIP5 model preindustrial control simulations and reanalysis data. The methods are described in section 2, the results from CMIP5 composites are described in section 3a, results from reanalysis data are presented in section 3b, with a discussion in section 4, and concluding remarks in section 5.

2. Data and methods

a. SFW identification and stratospheric index

The dates of SFWs were identified in CMIP5 models as the day of the final reversal of zonal mean zonal wind at 60°N and 10 hPa each spring until the subsequent autumn. The computation was performed using monthly data interpolated to daily, similar to the method described in Hardiman et al. (2011), but only in the middle

FIG. 1. Distribution of final warming date (relative to 1 Jan of each year) for each model included in composite analysis and JRA-55. The box represents the interquartile range (IQR) of the model, with the central line indicating the median and the whiskers extending to twice the IQR; outliers are plotted individually beyond this. Mean early and late final warming dates for each model (enumerated in Table 1) are indicated with an orange and a purple mark, respectively. 15 Mar, 15 April, and 15 May for a standard calendar are marked along the top axis.

stratosphere (10 hPa). The use of interpolated monthly data rather than pure daily data helps to identify a clear date for the occurrence of SFW, as it eliminates the small vacillations of the wind around weak westerlies

and easterlies during the SFW (Hardiman et al. 2011). The distribution of SFW date for each model is shown in Fig. 1. The mean date for many models is mid-April to early May, which is similar to the JRA-55 (1979–2015, reanalysis) mean of 16 April (with a standard deviation of 15 days). Based on the SFW date of each year, composites of different fields were constructed. Years were defined as early or late if the final warming date was one standard deviation before or after the mean final warming date for each model (Ayarzagüena and Serrano 2009). The mean dates for these are enumerated in Table 1 along with those of JRA-55. Note that the mean date for the SFW is later in models than in reanalysis, consistent with the findings of Butchart et al. (2011) in Chemistry–Climate Model Validation (CCMVal) models (see their Fig. 13). Nevertheless, the difference in SFW date between models and reanalysis is small (1–2 weeks) compared to the year-to-year variability (2 months), so we do not expect these small biases to be of major importance for our comparison of early versus late SFW. The composites were centered on April of the anomalous final warming year, hereafter referred to as month 0.

In the part of this investigation using reanalysis data, we did not perform the analysis based on SFW dates, given the shortness of the period of study and consequently, the relatively small sample of different types of SFWs. Instead, we calculated linear regressions of different fields on a stratospheric index that describes the intensity of the stratospheric vortex in April, given that this is the month when the vortex has a different state (existent or disappeared) in the early and late years according to Table 1. The stratospheric index is based on the area-weighted mean of temperature at 100 hPa over the polar cap (65°–90°N) in April (ST100 index). The ST100 index is advantageous with respect to the traditional geopotential height in the midstratosphere as it has been shown to

TABLE 1. Mean final warming (FW) dates for early, all, and late warmings for CMIP5 models used and JRA-55. Dates are computed using each model's own calendar. Also included are the number of final warmings falling into the late and early categories, as well as the number of each FW type per decade.

Model	Years	Early	Overall	Late	No. of late/early FW	Late/early FW (decade ⁻¹)
CanESM2	995	26 Mar	22 Apr	14 May	156/137	1.6/1.4
CESM1(WACCM)	199	7 Apr	26 Apr	18 May	30/40	1.5/2.0
CMCC-CESM	276	7 Apr	29 Apr	21 May	40/64	1.4/2.3
GFDL CM3	499	8 Apr	27 Apr	15 May	59/101	1.2/2.0
HadGEM2-CC	239	17 Mar	14 Apr	3 May	50/29	2.1/1.2
IPSL-CM5A-LR	999	8 Apr	30 Apr	17 May	94/170	0.9/1.7
IPSL-CM5A-MR	299	3 Apr	28 Apr	18 May	24/56	0.8/1.9
MIROC-ESM	629	23 Mar	17 Apr	6 May	98/91	1.6/1.4
MIROC-ESM-CHEM	254	24 Mar	17 Apr	6 May	46/33	1.8/1.3
MPI-ESM-MR	999	28 Mar	18 Apr	5 May	183/136	1.8/1.4
JRA-55	36	22 Mar	16 Apr	6 May	10/3	2.8/0.8

TABLE 2. Description of models included in the subset. Those with available sea ice motion data are marked with an asterisk (*).

Model	Atmosphere resolution	Atmosphere levels (Model top)	Sea ice resolution	Sea ice model
CanESM2*	64×128	35 (0.5 hPa)	64×128	CanCM
CESM1(WACCM)	96×144	66 (5.1×10^{-6} hPa)	384×384	CICE4
CMCC-CESM*	48×96	39 (0.01 hPa)	149×149	LIM2
GFDL CM3	90×144	48 (0.01 hPa)	200×200	SIS
HadGEM2-CC	144×192	60 (39.25 km)	216×360	MOHC
IPSL-CM5A-LR	96×96	39 (0.04 hPa)	149×149	LIM2
IPSL-CM5A-MR*	143×144	39 (0.04 hPa)	149×149	IPSL
MIROC-ESM	64×128	80 (0.003 hPa)	192×256	MIROC
MIROC-ESM-CHEM	64×128	80 (0.003 hPa)	192×256	MIROC
MPI-ESM-MR	96×192	95 (0.01 hPa)	404×404	MPI

be virtually identical to the integrated potential vorticity above (Baldwin et al. 2019). In addition, the 100-hPa level is close to the tropopause, which is desirable for stratosphere–troposphere coupling analyses. The use of this index also ensures the consideration of full data of the reanalysis period.

b. CMIP5 model data

Preindustrial control runs from the CMIP5 archive (Taylor et al. 2012) were used to create composites relative to the SFW date. A subset of the CMIP archive was selected to include only CMIP5 high-top models (top higher than 1 hPa). Table 2 shows all the high-top models included, with their resolutions, length of preindustrial control run, and sea ice model. Note that sea ice thickness (SIT) in these models represents “the mean thickness of sea ice in the ocean portion of the grid cell (averaging over the entire ocean portion, including the ice-free fraction). Reported as 0.0 in regions free of sea ice” (Taylor 2013).

As just mentioned, composites were constructed for each model based on the final warming date of each year. However, unless otherwise stated, the results displayed are for the multimodel mean difference between late and early composites. This was computed by first differencing late and early final warming composites for each model, then each is regridded to a common grid of 64×128 (latitude–longitude) for atmospheric data and 149×182 (latitude–longitude) for sea ice data. Finally, the regridded composites of all models are averaged to obtain the multimodel mean. Statistical significance of results was computed by combining p values from a Student’s t test for the difference in composites (late minus early) for each of the 10 models using Fisher’s method (Kost and McDermott 2002). This method combines p values for independent observations with a common null hypothesis. A chi-squared test was performed on this combined value to obtain a confidence bound on the difference between composite late and

early final warming events being different from zero. The results were identified as robust when at least 60% of model composite differences match the sign of the mean composite difference.

FIG. 2. (a) Multimodel-mean standardized anomalies of geopotential height differences averaged over the polar cap (66° – 90° N) (contour interval: 0.25) for late SFW years relative to early SFW years. (b) As in (a), but for early SFW years. (c) As in (a), but for late SFW years. Shading is plotted where anomalies are statistically significant at the 95% confidence level and greater than 60% (6 of 10) models agree on the sign of the difference.

FIG. 3. Multimodel-mean composite difference, for late SFW years relative to early SFW years of sea ice thickness (m; shaded) and sea level pressure (hPa; contours); negative lag months are prior to mean SFW month. Hatching indicates anomalies do not meet statistical significance and model agreement set in Fig. 2.

c. Reanalysis data

Monthly reanalysis data were used in the second part of the analysis to compare with and confirm the model results. More specifically, atmospheric fields (sea level pressure, zonal wind, and surface temperature) were taken from JRA-55 (Kobayashi et al. 2015) in a $2.5^\circ \times 2.5^\circ$ (latitude–longitude) grid. Sea ice variables (sea ice thickness and sea ice velocity), are taken from the sea ice reanalysis Pan-Arctic Ice-Ocean Modeling and Assimilation System (PIOMAS; Zhang and Rothrock 2003; Schweiger et al. 2011), on a generalized curvilinear coordinate system.

In all cases, we considered a common period for all datasets, extending from 1979 to 2016. In this period, variables were linearly detrended to remove an estimate of the anthropogenic influence.

3. Results

a. CMIP5 models

A time–height plot of the geopotential height anomaly over the polar cap (66° – 90° N) can be used to measure the strength of the polar stratospheric vortex and the downward propagation of this signal to the troposphere in a similar way to the northern annular mode. In the stratosphere positive anomalies represent a weaker vortex, while negative anomalies indicate a stronger, more cyclonic vortex. Figure 2 shows the composite difference of polar cap geopotential height anomalies for late and early SFW, and their difference, standardized by removing the monthly long-term mean and dividing by the monthly long-term standard deviation. Early

FIG. 4. Multimodel-mean composite difference, for late SFW years relative to early SFW years of sea ice concentration (%; shaded) and sea level pressure (hPa; contours); negative lag months are prior to mean SFW month. Hatching indicates anomalies do not meet statistical significance and model agreement set in Fig. 2.

SFW years have significantly positive stratospheric and upper-tropospheric height anomalies in spring, while late SFW years have significantly negative geopotential height anomalies in May–June. This is a consequence of the anomalously persistent cyclonic circulation in late SFW years, whereas the summertime anticyclonic circulation develops sooner in early SFW years. The vortex is weaker in March preceding a late final warming, though the magnitude of this anomaly is smaller than the strong vortex anomaly in the subsequent May and June following a delayed final warming. This might seem contradictory to [Ayarzagüena and Serrano \(2009\)](#), who found a weaker polar vortex in March in early SFW years compared to late SFW years. However, the same authors and more recent studies ([Hu et al. 2014](#); [Thièblemont et al.](#)

[2019](#)) have indicated that late SFWs are often preceded by SSWs in February/March in reanalysis. We speculate that the weaker polar vortex in March during late SFW years could reflect the increased occurrence of late SFWs following SSWs in late February/March, but confirmation is not possible due to the lack of daily output of many models for such long time periods.

The stratospheric polar height anomalies are followed by tropospheric anomalies of the same sign about 1 month later, and the tropospheric anomalies persist until June (Fig. 2). Given the negative polar anomalies of geopotential height in Fig. 2a, one would expect a significant cyclonic pattern in sea level pressure (SLP) over the pole from April in late SFW years relative to early SFW years. This pattern could in turn induce changes in sea

ice concentration and thickness (e.g., Kwok 2000; Ogi et al. 2010; Smith et al. 2018). These SLP anomalies are seen in the contours of Fig. 3. Robust SLP differences are limited to April–June period as is also the case for the midtroposphere and stratosphere. However, there are significant anomalies in SIT that persist until the following autumn (shading of Fig. 3). There is anomalously low SIT in the East Siberian Sea, concurrent with anomalously high SIT in the Beaufort Sea in late SFW years with respect to early SFW years. Weak positive SIT anomalies are also evident in the Greenland and Lincoln Seas, while weak negative SIT anomalies are present in the Laptev Sea. The months not plotted in Fig. 3 (December–February) lack either statistically significant sea ice or circulation anomalies, or both.

To further examine Arctic sea ice cover variability in response to the anomalous SFWs, Fig. 4 shows the composite differences in sea ice concentration (SIC), which are qualitatively similar to those of SIT differences. Additionally, SIC anomalies are also present prior to the SFW, including increased concentration in the Barents Sea and lower concentrations in Baffin Bay in late SFW years relative to early SFW years. These anomalies, however, are spatially and temporally distinct from the SIC and SIT dipole between the Laptev/East Siberian and Beaufort Seas that appears in the months following the SFW.

As previously discussed, sea ice responds primarily to atmospheric winds on the time scales of interest here. While surface wind anomalies can be implied from the SLP anomalies, surface wind and sea ice motion composites were constructed for a smaller subset of models with available data in order to compare these fields directly (Fig. 5). Only three models of the 10 previously discussed have sufficient sea ice motion data available. Thus, for a fairer comparison, Fig. 5 uses surface winds, sea ice motion, and sea ice thickness from CanESM2, CMCC-CESM, and IPSL-CM5A-MR only. Note that while statistical significance in SIT differences is reduced, the pattern is qualitatively similar to that of the full subset in Fig. 3. Here, only May is shown as it has the most widespread statistically significant anomalies in both sea ice motion and surface winds. From previous studies (e.g., Thorndike and Colony 1982) it was observed that sea ice motion typically is of smaller magnitude and to the right of the geostrophic wind, and in general, the composite differences of Fig. 5 follow this relationship. The significant differences in sea ice motion are limited to the months in which there are significant differences in composite sea level pressure, and thus significant differences in surface wind.

These anomalies in atmospheric circulation, in the models with sea ice motion available, coincide with sea

FIG. 5. Multimodel-mean composite difference, for late SFW years relative to early SFW years of sea ice thickness (m; shaded), composite difference surface wind (purple vectors; m s^{-1}), and composite difference sea ice motion vectors (black vectors; m s^{-1}) for month +1 (May) for models with sea ice motion available. Shading and vectors are plotted where statistically significant at the 95% confidence level.

ice motion anomalies. The sea ice motion anomaly vectors, in late SFW years relative to early SFW years, are directed from negative SIT/SIC anomalies toward positive SIT/SIC anomalies, suggesting that the changes in SIT and SIC are driven by atmospheric winds moving ice. There is a dipole of SIT between the East Siberian Sea and the Beaufort Sea, where both surface wind and sea ice motion anomalies are directed. Under the circulation pattern seen in Fig. 3, the east-to-west sea ice motion results in an accumulation of ice in the Beaufort Sea and a divergence of ice in the East Siberian Sea, in late SFW years relative to early SFW years. This anomaly dipole in SIT and SIC then persists for several months longer than those of the atmospheric circulation and wind-driven sea ice motion, due to the greater thermal inertia of the ocean/sea ice than the atmosphere. This persistence of sea ice anomalies from spring into summer is consistent with observational studies showing that spring atmospheric and sea ice conditions are a skillful predictor of late summer sea ice conditions (Kapsch et al. 2014; Schröder et al. 2014; Petty et al. 2017).

Figure 6 shows the subsequent autumn (SON) surface temperature (TAS), in late SFW years compared to

FIG. 6. Multimodel-mean composite difference, for late SFW years relative to early SFW years of surface air temperature (K; shaded) for autumn months +5 (September) through +7 (November). Hatching of TAS is as in Fig. 3.

early SFW years. Statistically significant warm anomalies are seen over the East Siberian Sea, spreading over eastern Siberia in September and October. As previously shown in Fig. 3, there are no concurrent significant

circulation anomalies. However, in the months with statistically significant TAS anomalies, there are corresponding anomalies in SIT (Fig. 3) and SIC (Fig. 4). Thus, it would appear that the anomalies in sea ice thickness and

FIG. 7. Regression coefficient (shaded) of polar cap geopotential height on April ST100 index for (a) JRA-55 and (b) the multimodel mean where each model is regressed on its own ST100 index, then an intermodel mean is taken. Areas that are not significant at the 95% confidence level are hatched. Regressions are multiplied by -1 to emphasize the relationships with a colder/stronger polar vortex.

concentration are responsible for the TAS anomalies seen here. Therefore, our results indicate the potential for interseasonal prediction of surface temperatures in selected Arctic regions based on the state of the springtime polar vortex.

b. Regression on ST100

We will soon turn to results using JRA-55, for which we use the ST100 index rather than compositing based in the SFW. First, however, in order to bridge the gap between two different methods, we performed regressions on the ST100 index using the CMIP5 models. Figure 7 shows the regression of polar cap geopotential height anomaly on the ST100 index, which matches well between the model ensemble mean and the reanalysis. The regression coefficients are multiplied by -1 , to aid comparison with the composite differences and to highlight the effects of a stronger spring stratospheric polar vortex. We note that although the ST100 index is not a direct measure of the SFW timing and therefore, there are differences between the polar cap height evolution between composite (Fig. 2) and regression (Fig. 7) analyses, both methods identify anomalies in the strength of the stratospheric polar vortex and the downward propagation of this signal. Despite the two methods being nonidentical, the SIT and SLP anomalies in the models when regressed onto the April ST100 index are highly consistent with the composite anomalies for late minus early SFW years. Figure 8 shows the results for three representative months. The dipole between the East Siberian/Laptev Seas and the Beaufort Sea seen in Fig. 3 appears here as well. Though the sea level pressure pattern is not broadly statistically significant, it is present. Having established that the ST100 index yields similar patterns of SIT and SLP anomalies in the models as does compositing by the SFW date, Fig. 9 displays regression maps of SLP and SIT onto the ST100 index for JRA-55 data. The SLP and SIT patterns closely resemble those in the models, shown in Fig. 3. The similarity in results between reanalysis and models is also detected in the patterns of SIC anomalies (Fig. 11). Negative anomalies of SLP in April are still detected over the pole, consistent with an anomalously strong polar stratospheric vortex in the same month. These anomalies are almost undetectable one month later, and so, disappear more quickly in reanalysis than in models. This could be explained by the earlier average occurrence of SFW in the former (Table 1). However, statistical significance of results is much lower in JRA-55 than in model results, probably due to the smaller data sample. Nevertheless, although SLP results are not statistically significant in JRA-55 data, we do detect a significant dipole of SIT anomalies between the eastern and western Arctic in its regression map on the ST100 index

FIG. 8. Multimodel-mean regression coefficient of sea ice thickness (shaded) and sea level pressure (contours), against April (month 0) ST100 index for (a) April, (b) July, and (c) October (months 0, +3, and +6, respectively). SLP contours are thicker, and SIT shading is unhatched where each meet the significance thresholds described in Fig. 2. Regressions are multiplied by -1 to emphasize the relationships with a colder/stronger polar vortex.

FIG. 9. JRA-55 regression of sea ice thickness (shaded; m) and sea level pressure (contours; hPa) against April (month 0) ST100 index in JRA-55 for March (−1) to November (+7). SLP contours are plotted from −4 to +4 hPa every 0.75 hPa. Areas that are hatched indicate that sea ice thickness anomalies are not statistically significant at the 90% confidence level. Thicker contours indicate areas where SLP is statistically significant at the 90% confidence level. Regressions are multiplied by −1 to emphasize the relationships with a colder/stronger polar vortex.

(shading of Fig. 9). This dipole persists until the subsequent autumn in good agreement with CMIP5 results.

The regression map of surface wind on April ST100 index (Fig. 10) in JRA-55 data shows cyclonic wind anomalies that could force anomalous sea ice movement from the eastern to the western Arctic, as was also found for models in Fig. 5. The area of significant values of surface wind (i.e., approximately from 90° to 270°E) coincides with that of SIT anomalies (Fig. 10) and SIC anomalies (Fig. 11), suggesting the sea ice cover and thickness anomalies are wind driven. The mentioned agreement in sea ice and near-surface atmospheric impacts would indicate that the wind drives the SIT and SIC anomalies. In short, results from the reanalysis

support the findings with model data, though with some differences, including that the subsequent autumn TAS anomalies are statistically insignificant and inconclusive in JRA-55 (not shown).

4. Discussion

We are cognizant that the apparent effects of the SFW timing shown here could be related to a hidden common variable that impacts both Arctic sea ice cover and the timing of the SFW, for example, El Niño–Southern Oscillation (ENSO). ENSO is known to impact stratospheric variability (Manzini et al. 2006; Butler et al. 2014; Polvani et al. 2017) and potentially, also Arctic

FIG. 10. Negative regression of sea ice thickness (shaded; m), surface wind (blue vectors; m s^{-1}), and sea ice motion (black vectors; m s^{-1}) against April (month 0) ST100 index in JRA-55 for May (+1). Shading and vectors are plotted only where are significant at the 90% confidence level; for vectors both components must meet this criterion. Regressions are multiplied by -1 to emphasize the relationships with a colder/stronger polar vortex.

FIG. 11. Negative regression of sea ice concentration (shaded; %) and sea level pressure (contours; hPa) against April (month 0) ST100 index in JRA-55 for May (+1). Shading is plotted as in Fig. 9. Regressions are multiplied by -1 to emphasize the relationships with a colder/stronger polar vortex.

surface temperature and sea ice (Lee 2012; Li et al. 2019). To remove a possible ENSO influence on our results, we have repeated our analyses but for only neutral ENSO conditions, defined using the Niño-3.4 index (Trenberth and Stepaniak 2001). El Niño or La Niña events had to last 6 months above or below a threshold of $\pm 0.4\text{ K}$, computed from a 5-month rolling mean. All other months were defined as neutral, and SFWs that occurred during or within 1 month of a nonneutral month were not included in this composite. Figure 12 shows composite SIT and SLP differences, for late SFW years compared to early SFW years, for neutral ENSO years only, and matches qualitatively well with Fig. 3. Thus, our results appear to be independent of ENSO. We noted that controlling for every external factor is not possible, and ENSO represents only the most likely confounding variable.

Our results agree with previous studies on several aspects. First, we have shown that the interannual variability of SFWs, and in particular their timing, in CMIP5 models has an impact on the tropospheric circulation in springtime in agreement with results from reanalyses shown by Wei et al. (2007), Ayarzagüena and Serrano (2009), and Hardiman et al. (2011). Second, the mentioned

SFW-induced tropospheric circulation anomalies are centered over the polar cap, which in turn induces wind anomalies, which drive sea ice motion anomalies similar to a negative AO pattern for years with late SFW and/or stronger April stratospheric polar vortex relative to years with an early SFW and/or weaker April polar vortex. As a result, sea ice thickness and concentration changes in agreement with previous studies that have related variability of the AO to changes in Arctic sea ice on different time scales (e.g., Kwok 2000; Rigor et al. 2002; Park et al. 2018). Despite these common results with previous work, the present study constitutes a novel work as it combines all these separate aspects, connecting changes in the stratosphere in late winter and early spring to Arctic sea ice state in the following summer and autumn.

Recently, Smith et al. (2018) linked winter polar stratospheric variability to Arctic sea ice cover in the CESM1(WACCM) model. They found that midwinter strong polar vortex events are associated with a dipole of sea ice cover, similar to our findings, although with some important differences. In Smith et al.'s (2018) case, they focused on midwinter stratospheric changes whereas we have focused on springtime stratospheric variability. This different timing might be extremely important for

FIG. 12. Multimodel-mean composite difference, for late SFW years relative to early SFW years of sea ice thickness (m; shaded) and sea level pressure (hPa; contours), for anomalous final warming years in ENSO neutral years. Negative lag months are prior to mean SFW month. Hatching indicates anomalies do not meet statistical significance and model agreement set in Fig. 2.

the resulting surface circulation and sea ice. Although a midwinter strong polar vortex could theoretically result in a very persistent spring vortex, some previous studies and ours have shown that delayed SFWs may be preceded by a midwinter SSW. It should also be noted that our results represent an ensemble of multiple models, whereas Smith et al. (2018) used a single model and so, we have demonstrated that the results are robust across a variety of high-top models.

5. Conclusions

We have presented results from analysis of CMIP5 models and reanalysis data on the effects of the timing of SFWs on Arctic sea ice distribution. Composites were

constructed in a subset of high-top CMIP5 models based on the date of the stratospheric final warming relative to the mean final warming date in each model. Differences between anomalously late and early years were examined for various properties of the atmosphere and sea ice.

The primary conclusions of this study are as follows:

- Selected CMIP5 high-top models represent interannual variability of SFW date reasonably well, although the mean date is generally later than in reanalysis.
- The timing of SFW affects springtime surface circulation in the Northern Hemisphere.
- The induced circulation anomalies generate a dipole of anomalies in sea ice thickness and concentration that persists until autumn (up to 7 months).

- These ice anomalies are associated with surface wind-driven sea ice motion from the East Siberian Sea toward the Beaufort Sea.
- Changes in sea ice thickness and concentration induce coincident surface temperature anomalies in early autumn, without the presence of circulation anomalies that dissipate in June.
- Closely comparable results are found in the models when regressing against lower-stratospheric temperature as an alternative measure of the polar vortex strength rather than the SFW timing.
- Qualitatively similar results are also found when the regression method is applied to reanalysis data, but with reduced statistical significance due to the time limited data record.

From our results, a consistent picture emerges, whereby the spring and more importantly, summer and autumn sea ice distribution is impacted by changes in the stratospheric circulation. Previous studies have already shown that the stratosphere plays an important role in monthly to seasonal predictability of the troposphere (e.g., Baldwin et al. 2003; Sigmond et al. 2013), and our results provide evidence that stratospheric variability also affects Arctic sea ice, supporting previous work (e.g., Smith et al. 2018) that it also affects Arctic sea ice. Moreover, unlike other polar stratospheric extreme events (SSWs or strong polar vortex), SFWs take place every year and so, our study suggests that its date could be included as one of the predictors of the Arctic sea ice state in summer and autumn. A similar analysis with hindcast simulations from seasonal prediction systems might gain further insight here.

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