

Interdependent Extratropical Atmospheric Responses to Arctic Sea Ice Loss, QBO, and ENSO

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ABSTRACT: The quasi-biennial oscillation (QBO) and El Niño–Southern Oscillation (ENSO) both influence the strength of the stratospheric polar vortex, with weakening of the polar vortex during the easterly QBO and El Niño. The polar vortex is also thought to weaken in response to Arctic sea ice loss, related to an equatorward shift of the midlatitude jet stream. However, it is unclear how these factors impact the polar vortex in combination and whether their influences are linearly additive. Here, using atmospheric general circulation model experiments, we show that in our model, in the presence of Arctic sea ice loss, the polar vortex weakens significantly during easterly QBO, but not during westerly QBO and weakens significantly during neutral ENSO, but not during El Niño. This suggests that the polar vortex responses to sea ice loss, QBO, and ENSO are interdependent and nonadditive. This interdependence cannot be explained by the polar vortex background state, since the easterly QBO and El Niño both separately weaken the polar vortex but have an opposite-signed influence on the response to sea ice loss. Instead, we surmise the modulation is governed by how QBO and ENSO affect anomalous wave propagation into the stratosphere in response to sea ice loss. At the surface, sea ice loss enhances the Siberian high and causes cooling over Eurasia in QBO easterly phase (QBO-E) but not QBO westerly phase (QBO-W). Both QBO-W and El Niño damp the equatorward tropospheric jet shift in response to sea ice loss, relative to QBO-E or neutral ENSO, respectively. The existence of these nonlinearities has implications for subseasonal-to-interannual climate prediction.

SIGNIFICANCE STATEMENT: This work aims to understand how the atmospheric changes triggered by sea ice loss in the Arctic will differ depending on the background state of the atmosphere. We find that the response to sea ice loss is stronger under certain conditions, meaning sea ice loss is more impactful to humans in years where these conditions are present. The results of this paper have implications for accurate climate prediction, particularly on the expected impacts of climate change.

KEYWORDS: Sea ice; ENSO; Quasi-biennial oscillation; Stratospheric circulation; Teleconnections

1. Introduction

The Arctic is warming almost 4 times faster than the global average (Rantanen et al. 2022). This warming has been accompanied by significant sea ice decline, with both ecological and climatic implications (Stroeve and Notz 2018). The greater warming of the Arctic than the global average—termed Arctic amplification (AA)—is driven by a combination of feedbacks including the ice–albedo, Planck, and lapse-rate feedbacks; the reduction in the insulating effect of sea ice due to its decline; and changes in poleward heat transport (Previdi et al. 2021). Sea ice decline has accelerated lower-tropospheric warming in the Arctic (Screen and Simmonds 2010), but its influence is not limited to the Arctic. There is also growing evidence for

remote effects on the atmosphere and ocean from the sub-Arctic to the tropics. However, the quantification and understanding of these remote impacts of Arctic sea ice loss have been hampered by disagreements between modeling and observational studies and the diverse responses simulated by different models (Screen et al. 2018; Smith et al. 2019; Cohen et al. 2020).

Models simulate a weakening and equatorial shift of the Northern Hemisphere midlatitude jet stream in response to future Arctic sea ice loss (Smith et al. 2022). Arctic heating induced by the imposed sea ice loss reduces the equator-to-pole temperature gradient, weakening the jet predominantly on its poleward flank, and positive eddy feedback enhances the equatorward shift. Although robust in sign across models, the magnitude of this response varies between models and may be related to the strength of simulated eddy feedback (Smith et al. 2022; Screen et al. 2022). Another possible response to Arctic sea ice loss is a weakening of the stratospheric polar vortex (Kim et al. 2014), through enhancing wave propagation into the stratosphere. However, this sign and magnitude of the stratospheric response differs between models for reasons that are not fully clear. Possible reasons for these differences include internal variability (Peings et al. 2021; Sun et al. 2022), different model physics, model resolution (Steffing et al. 2021), and dependence on the background state. Regarding the latter, numerous

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studies have suggested that the response to sea ice loss is sensitive to the atmospheric and/or oceanic background state (e.g., Screen and Francis 2016; Smith et al. 2017; Osborne et al. 2017; Deser et al. 2015; Labe et al. 2019; England et al. 2020; Simon et al. 2022; Sigmond and Sun 2024).

One example of sensitivity to the background mean state is that the stratospheric mean state may affect the response to sea ice decline. Labe et al. (2019) proposed that in response to a reduction in sea ice, the stratospheric polar vortex weakens during the easterly phase of the quasi-biennial oscillation (QBO) but strengthens in response to a reduction in sea ice during westerly QBO. The QBO is an important mode of climate variability in the tropical stratosphere and consists of downward-propagating easterly and westerly wind regimes, which oscillate with a period of approximately 28 months (Baldwin et al. 2001). Although the QBO originates in the tropics, its teleconnections affect the mid-high latitudes in winter (Anstey and Shepherd 2013; Scaife et al. 2014; Andrews et al. 2019). It is observed that the QBO easterly phase (QBO-E) weakens the polar vortex and the QBO westerly phase (QBO-W) strengthens it (Ebdon 1975; Holton and Tan 1980). Holton and Tan (1980) suggest that this teleconnection works by changing the location of the subtropical critical line. During QBO-E, this critical line shifts toward the pole and planetary waves are increasingly reflected toward the polar vortex, where they break and exert a decelerating force on the vortex, weakening it (Baldwin et al. 2001; Anstey and Shepherd 2013). This is known as the Holton–Tan mechanism, although other work has suggested that a shift of the critical line is not the most important factor in driving it (e.g., Garfinkel et al. 2012; Lu et al. 2014). Through its effect on the polar vortex mean state, the QBO may modulate the stratospheric response due to Arctic sea ice loss. Labe et al. (2019) investigated this relationship in one model [Specified Chemistry-WACCM4 (SC-WACCM4)]. Here, we identify a similar QBO modulation of the response to sea ice loss in an independent model (HadGEM3) and further elucidate its mechanism. We do this using a higher-resolution model than in Labe et al. (2019), which improves the representation of eddy–mean flow interactions (Smith et al. 2022), and with more vertical levels, which dynamically produces a realistic QBO (Menary et al. 2018).

We hypothesize that other climate factors that can influence the stratospheric polar vortex, in addition to the QBO, may also modulate the polar vortex response to sea ice loss. El Niño–Southern Oscillation (ENSO) weakens the polar vortex during El Niño (Hamilton 1993; Brönnimann et al. 2004; Taguchi and Hartmann 2006; Manzini et al. 2006; Ineson and Scaife 2009). Positive sea surface temperature (SST) anomalies in the tropical east Pacific during El Niño generate a Rossby wave train that deepens the Aleutian low (Domeisen et al. 2019; Bell et al. 2009) and increases wave activity entering the stratosphere, weakening the polar vortex (Manzini et al. 2006; Ineson and Scaife 2009). There is evidence that the Pacific decadal oscillation (PDO), which is analogous to ENSO but on a decadal rather than interannual time scale, also moderates the polar vortex response to sea ice loss. Simon et al. (2022) find that the warm PDO phase (equivalent to prolonged El Niño) and sea ice loss both separately deepen the Aleutian low, increase in wave activity into stratosphere, and subsequently weaken the polar vortex,

but when both effects are combined, the polar vortex weakens less than the sum of their individual effects. Here, for the first time, we identify a modulation of the response to sea ice loss by El Niño and contrast its mechanisms to the modulation of the polar vortex by the QBO.

2. Data and methods

a. Model experiments

This study presents results from HadGEM3 at N216 resolution (~60 km) configured with different sea ice, QBO, and ENSO states. The model setup is based on two experiments from the Polar Amplification Model Intercomparison Project (PAMIP): one with present-day SST (pdSST) and present-day sea ice concentration (SIC) (pdSIC) ocean boundary conditions (pdSST–pdSIC; Fig. 1a) and the other with pdSST with future Arctic SIC (futArcSIC) (pdSST–futArcSIC; Fig. 1b). The prescribed loss of SIC corresponds to that simulated at 2°C of global warming [from CMIP5; see Smith et al. (2019) for further details]. Relative to the present-day sea ice state, there is a marked reduction in winter SIC, most pronounced (40% or above) in the Barents–Kara Sea, Bering Sea, Sea of Okhotsk, and Hudson Bay (Fig. 1c).

Following the PAMIP protocol, all our simulations start on 1 April and were run for 14 months, with the first 2 months being discarded for model spinup. To test the dependence on the QBO, two ensembles were conducted for each sea ice and ENSO state, with atmospheric initial conditions chosen to give a strong QBO-W or QBO-E in the following winter (December–February). Each ensemble member, for a given QBO polarity, was initialized from essentially the same atmospheric state, with ensemble spread achieved through small perturbations to atmospheric initial conditions. As the QBO evolves relatively slowly, its sign and magnitude do not change drastically between 1 April and the following winter, and thus, the ensembles have distinct and opposite QBO phases (Fig. 1d). HadGEM3-GC3.1 adequately represents the QBO by having a high-top and nonorographic gravity wave drag parameterization (Scaife et al. 2002). The model hence features a realistic QBO, albeit with a period 3–4 months longer than observed and a weaker amplitude than observations, and reproduces the Holton–Tan effect (Menary et al. 2018). Here and elsewhere, the QBO is defined as the zonal-mean westerly wind at 30 hPa from 5°N to 5°S in DJF (unless otherwise stated).

To investigate the potential modulation by ENSO, we adapted the PAMIP boundary conditions to include an El Niño signal in the SST. The standard PAMIP SST uses the 1979–2008 climatology from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST). To define the El Niño SST anomaly, we regressed the Niño-3.4 index against global SST from the HadISST observational dataset (Rayner et al. 2003), for the years 1950–2019. The Niño-3.4 index was defined as the area-averaged SST anomaly (with the 1950–2019 climatology removed at each grid point) in the region 5°N–5°S and 120°–170°W (Trenberth and Stepaniak 2001). The resultant regressions were doubled to correspond to a strong but realistic El Niño event (i.e., 2°C). El Niño

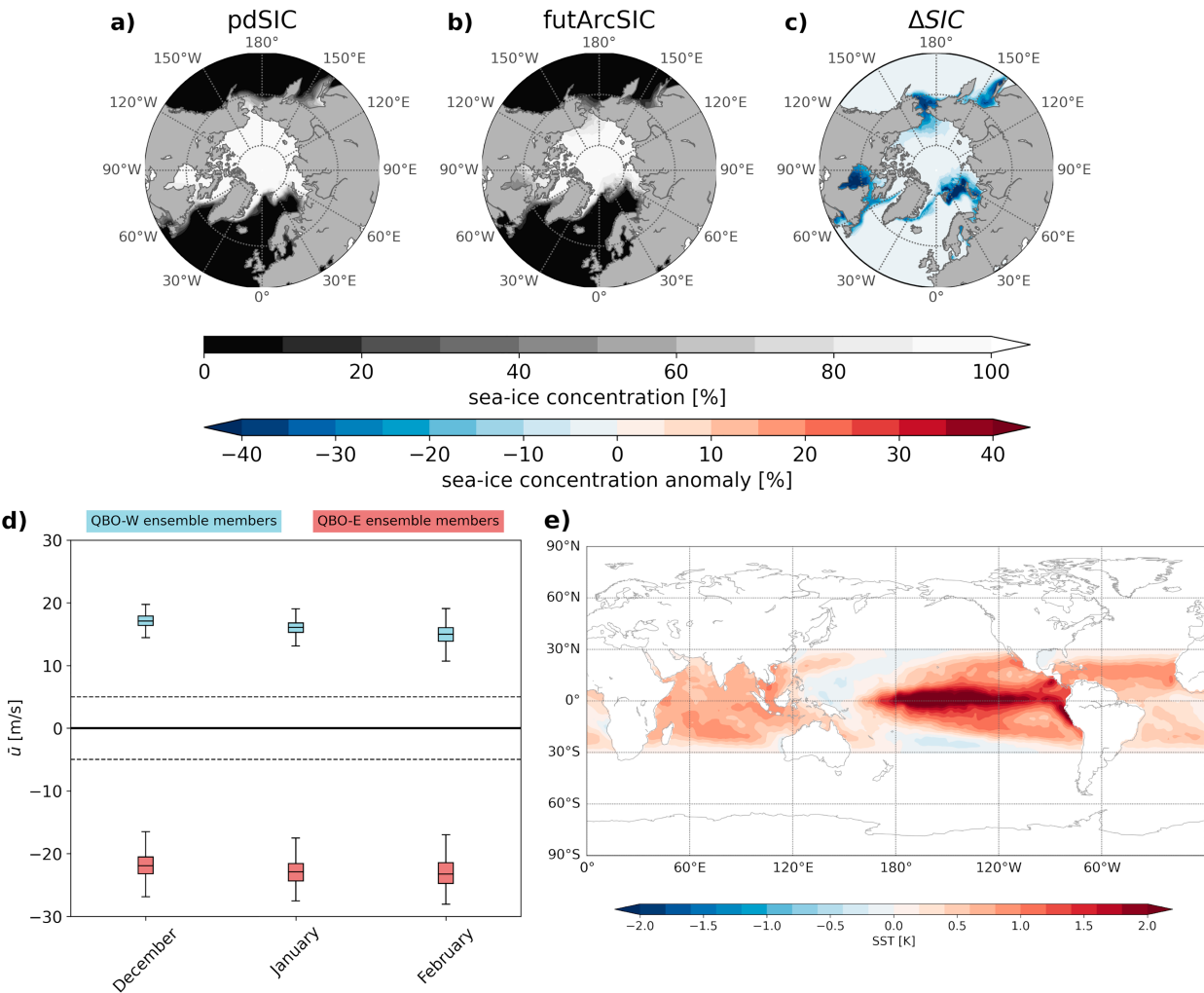


FIG. 1. (a) Winter Arctic SIC prescribed in the PAMIP pdSIC and (b) future experiments (futArcSIC), and (c) their difference (futArcSIC minus pdSIC). (d) Ensemble spread of monthly QBO indices in winter for the QBO-W and QBO-E experiments. The dashed lines at $\pm 5 \text{ m s}^{-1}$ denote neutral QBO conditions. (e) The DJF SST anomalies added to the El Niño experiments.

anomalies have been observed to reach as high as 3°C , but 2°C is more typical of a strong El Niño event. The derived SST anomalies were added to the baseline PAMIP boundary conditions separately for each month but only within the tropics, 20°N – 20°S , and a linear tapering between 20° and 30°

(Fig. 1e). The standard PAMIP experiments have neutral ENSO conditions and were used as a reference.

In summary, we analyze eight experiments with different combinations of Arctic sea ice, QBO, and ENSO states (Table 1).

TABLE 1. Summary of HadGEM3 experiments, including the PAMIP experiment number (column two) that most closely corresponds with them. See Smith et al. (2019) for further information.

Experiment name	PAMIP experiment number	Arctic sea ice	QBO	ENSO	Members
pdSST-pdSIC-qboe	1.1	Present day	QBO-E	Neutral	150
pdSST-futArcSIC-qboe	1.6	Future	QBO-E	Neutral	150
pdSST-pdSIC-qbow	1.1	Present day	QBO-W	Neutral	150
pdSST-futArcSIC-qbow	1.6	Future	QBO-W	Neutral	150
niñoSST-pdSIC-qboe	1.1 (modified SST)	Present day	QBO-E	El Niño	150
niñoSST-futArcSIC-qboe	1.6 (modified SST)	Future	QBO-E	El Niño	150
niñoSST-pdSIC-qbow	1.1 (modified SST)	Present day	QBO-W	El Niño	150
niñoSST-futArcSIC-qbow	1.6 (modified SST)	Future	QBO-W	El Niño	150

b. Response calculation and notation

The response to sea ice loss, conditional on another factor such as the QBO or ENSO phase, is denoted as, for example, $\Delta I | \text{QBO-W}$, for the response to sea ice loss, given QBO-W conditions:

$$\Delta I | \text{QBO-W} = [\text{pdSST-futArcSIC-qbow, ni\~{n}oSST-futArcSIC-qbow}] - [\text{pdSST-pdSIC-qbow, ni\~{n}oSST-pdSIC-qbow}], \quad (1)$$

where the square brackets denote the pooling of experiments and each experiment name denotes an ensemble mean from that experiment. Experiments are pooled to maximize the ensemble sizes. When testing the dependence on the QBO, we pool experiments from both ENSO states, and when testing the dependence on ENSO, we pool experiments from both QBO states. For completeness, the other combinations are

$$\Delta I | \text{QBO-E} = [\text{pdSST-futArcSIC-qboe, ni\~{n}oSST-futArcSIC-qboe}] - [\text{pdSST-pdSIC-qboe, ni\~{n}oSST-pdSIC-qboe}], \quad (2)$$

$$\Delta I | \text{El Ni\~{n}o} = [\text{ni\~{n}oSST-futArcSIC-qboe, ni\~{n}oSST-futArcSIC-qbow}] - [\text{ni\~{n}oSST-pdSIC-qbow, ni\~{n}oSST-pdSIC-qboe}], \quad (3)$$

$$\Delta I | \text{neutral ENSO} = [\text{pdSST-futArcSIC-qbow, pdSST-futArcSIC-qboe}] - [\text{pdSST-pdSIC-qbow, pdSST-pdSIC-qboe}]. \quad (4)$$

For all cases above, the difference is computed between two 300-member pooled ensembles. Additionally, the difference in the responses to sea ice loss between QBO-E and QBO-W is written as $\Delta I | \Delta \text{QBO}$:

$$\Delta I | \Delta \text{QBO} = \Delta I | \text{QBO-E} - \Delta I | \text{QBO-W}. \quad (5)$$

Similarly, for the difference in the responses to sea ice loss between El Ni\~{n}o and neutral ENSO, the notation is $\Delta I | \Delta \text{ENSO}$:

$$\Delta I | \Delta \text{ENSO} = \Delta I | \text{El Ni\~{n}o} - \Delta I | \text{neutral}. \quad (6)$$

We note that in this framework,

$$\Delta I | \Delta \text{QBO} = \Delta \text{QBO} | \Delta I \quad (7)$$

and

$$\Delta I | \Delta \text{ENSO} = \Delta \text{ENSO} | \Delta I. \quad (8)$$

So, for the QBO, the response differences can equally be interpreted as the QBO modulation of the response to sea ice loss or as the modulation by sea ice of the response to QBO. Similarly for ENSO, the differences can be interpreted as ENSO modulation of the response to sea ice loss or as the sea ice loss modulation of the response to ENSO.

The computed differences are tested for statistical significance using the Student's t test with a threshold of $p < 0.05$.

c. Diagnostics

The Eliassen–Palm (EP) flux is a zonal-mean metric of wave activity, defined as

$$\mathbf{F} = [F_\phi, F_p] \\ = a \cos \phi \left\{ -\overline{u'v'} + \psi \frac{\partial \bar{u}}{\partial p}, -\overline{u'\omega'}, -\psi \left[\frac{\partial(\bar{u} \cos \phi)}{a \cos \phi \partial \phi} - f \right] \right\}, \quad (9)$$

where

$$\psi = \overline{v'\theta'} \left/ \frac{\partial \bar{\theta}}{\partial p} \right. = -\overline{v'T'} \left/ \left(\frac{\kappa \bar{T}}{p} - \frac{\partial \bar{T}}{\partial p} \right) \right., \quad (10)$$

and u' , v' , and ω' are the zonal, meridional, and pressure perturbations of zonal-mean wind, respectively; θ is the potential temperature; T is the temperature; f is the Coriolis parameter; ϕ is the latitude; a is the radius of Earth; and κ is the ratio of the gas constant and specific heat capacity of air (Eliassen and Palm 1961; Andrews et al. 1983). Primes denote the perturbations and overbars denote the zonal means. The divergence of EP flux is defined as

$$\nabla \cdot \mathbf{F} = \frac{\partial(F_\phi \cos \phi)}{a \cos \phi \partial \phi} + \frac{\partial F_p}{\partial p}. \quad (11)$$

Physically, the EP flux divergence corresponds to a zonal force on the flow from eddies that can accelerate or decelerate the zonal-mean wind.

The quasigeostrophic refractive index, n_k , is a diagnostic of the propagation of planetary waves. It can be derived from the quasigeostrophic potential vorticity, q , and takes the form

$$n_k^2 = \frac{\bar{q}}{\bar{u}} - \left(\frac{k}{a \cos \phi} \right)^2 - \left(\frac{f}{2NH} \right)^2, \quad (12)$$

where

$$\bar{q} = \frac{2\Omega}{a} \cos \phi - \frac{1}{a^2} \left[\frac{(\bar{u} \cos \phi)_\phi}{\cos \phi} \right]_\phi - \frac{f^2}{\rho_0} \left(\rho_0 \frac{\bar{u}}{N^2} \right), \quad (13)$$

and u , k , N , H , a , ρ_0 , and Ω are the zonal wind, wavenumber, buoyancy frequency, scale height, radius of Earth, air density, and Earth's rotation rate, respectively (Hu and Tung 2002). In the case where $n_k^2 = 0$, this results in a turning surface where waves are reflected away (Vallis 2017). Hence, the refractive index gives important information on how planetary waves propagate in the atmosphere. To compute the refractive index using model data, some further calculations and assumptions need to be made. The buoyancy frequency squared is defined as

$$N^2 = \frac{g}{\theta} \left(\frac{\partial \theta}{\partial z} \right), \quad (14)$$

where g is the acceleration due to gravity and θ is the potential temperature which itself is defined as

$$\theta = T \left(\frac{P_0}{P} \right)^{R/c_p}. \quad (15)$$

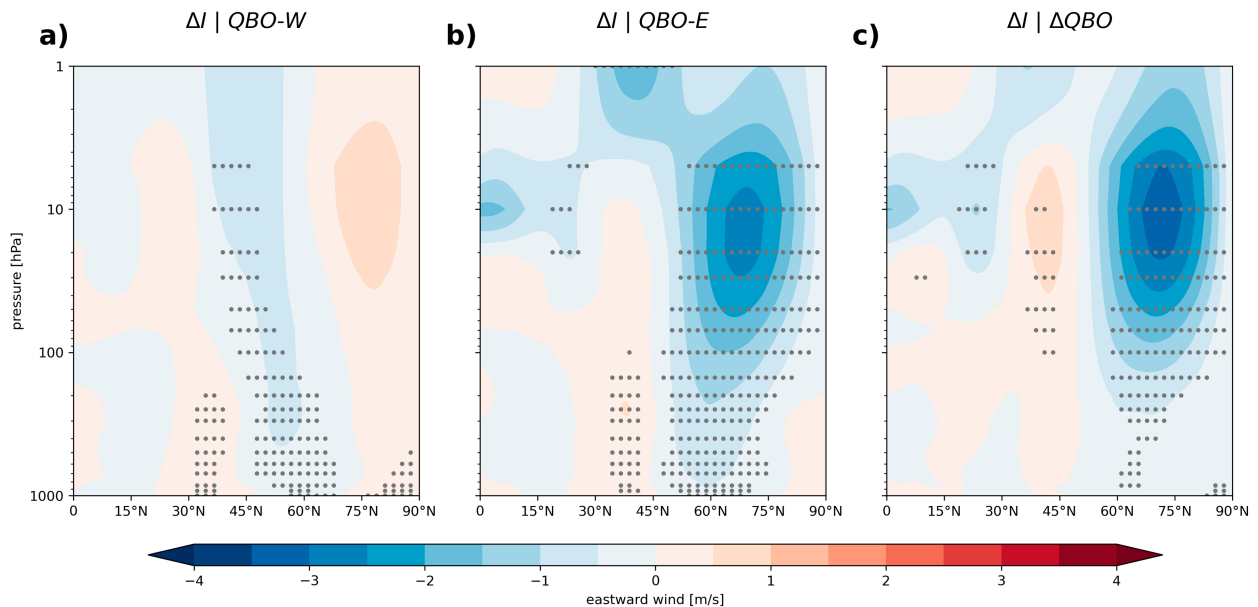


FIG. 2. Winter zonal-mean zonal wind response to sea ice loss during (a) QBO-W and (b) QBO-E and (c) their difference. Stippling denotes (a),(b) responses or (c) response differences that are significant at the 95% level.

Here, T is the temperature, P is the pressure, P_0 is the surface pressure, R is the specific gas constant for air, and c_p is the specific heat capacity. Hence, buoyancy frequency can be calculated just from temperature as well as various constants. Scale height is set to a constant value of 7 km, which corresponds to the value at approximately 240 K, which is close to the average temperature of the atmosphere.

The stratospheric polar vortex strength is defined as the zonal-mean zonal wind at 60°N and 10 hPa. Sudden stratospheric warmings are defined as occurrences when the polar vortex strength becomes negative for at least five consecutive days. Final warmings, that is when polar winds turn easterly each year in spring and remain so until autumn, are excluded from the sudden stratospheric warming (SSW) count.

The fast Fourier transform is used here to decompose atmospheric circulation fields into their constituent zonal wavenumbers, for the purpose of examining linear interference (Smith and Kushner 2012; Fletcher and Kushner 2011) between the circulation response and climatology, specifically for zonal wavenumbers 1 and 2, which are most important for vertical wave propagation into the stratosphere. If the phase of the wave response matches that of climatology, the waves constructively interfere, and vertical wave propagation is enhanced. If the wave response is out of phase with that of climatology, the waves destructively interfere, and vertical wave propagation is reduced. The climatologies shown throughout are present-day sea ice, with the QBO states averaged (in the QBO case) or neutral ENSO (for the ENSO case).

3. Results

a. Interdependent responses to sea ice loss and QBO

The winter zonal-mean zonal wind response to sea ice loss is shown in Fig. 2 split by QBO phase. These QBO composites are comprised of 300 ensemble members per sea ice

state. The $\Delta/QBO-W$ (Fig. 2a) and $\Delta/QBO-E$ (Fig. 2b) responses show a similar tropospheric signal, with both featuring a weakening and equatorward shift of the eddy-driven jet, consistent with the PAMIP multimodel ensemble (Smith et al. 2022). However, the weakening on the poleward side of the jet is larger for $\Delta/QBO-E$ than $\Delta/QBO-W$. This difference is statistically significant (Fig. 2c) throughout the troposphere at 60°N.

The $\Delta/QBO-E$ response features a weakened polar vortex, while, in contrast, the polar vortex is strengthened in $\Delta/QBO-W$, albeit by lesser magnitude and not statistically different from zero. This suggests a nonlinear polar vortex response to the combined effects of sea ice loss and the QBO that agrees with Labe et al. (2019). However, the exact interpretation of the nonlinearity is unclear, since it is possible that either the QBO modulates the response to sea ice loss and/or that sea ice loss alters the QBO teleconnection to the polar vortex.

The enhanced weakening on the poleward side of the jet, in $\Delta/QBO-E$ relative to $\Delta/QBO-W$, is likely partly related to the differing stratospheric responses. Weakening of the polar vortex in $\Delta/QBO-E$ would be expected to reinforce the weakening of the tropospheric jet stream, whereas the stronger polar vortex in $\Delta/QBO-W$ would be expected to dampen the weakening of the tropospheric jet stream. This is due to the feedback between polar vortex weakening, which induces negative Arctic Oscillation (AO) conditions at the surface, hence weakening the jet and shifting it equatorward (Coumou et al. 2018).

The polar vortex response to sea ice loss is often understood from a linear interference perspective (Smith and Kushner 2012; Fletcher and Kushner 2011), with enhanced wave activity corresponding to a weaker polar vortex. Figure 3 shows

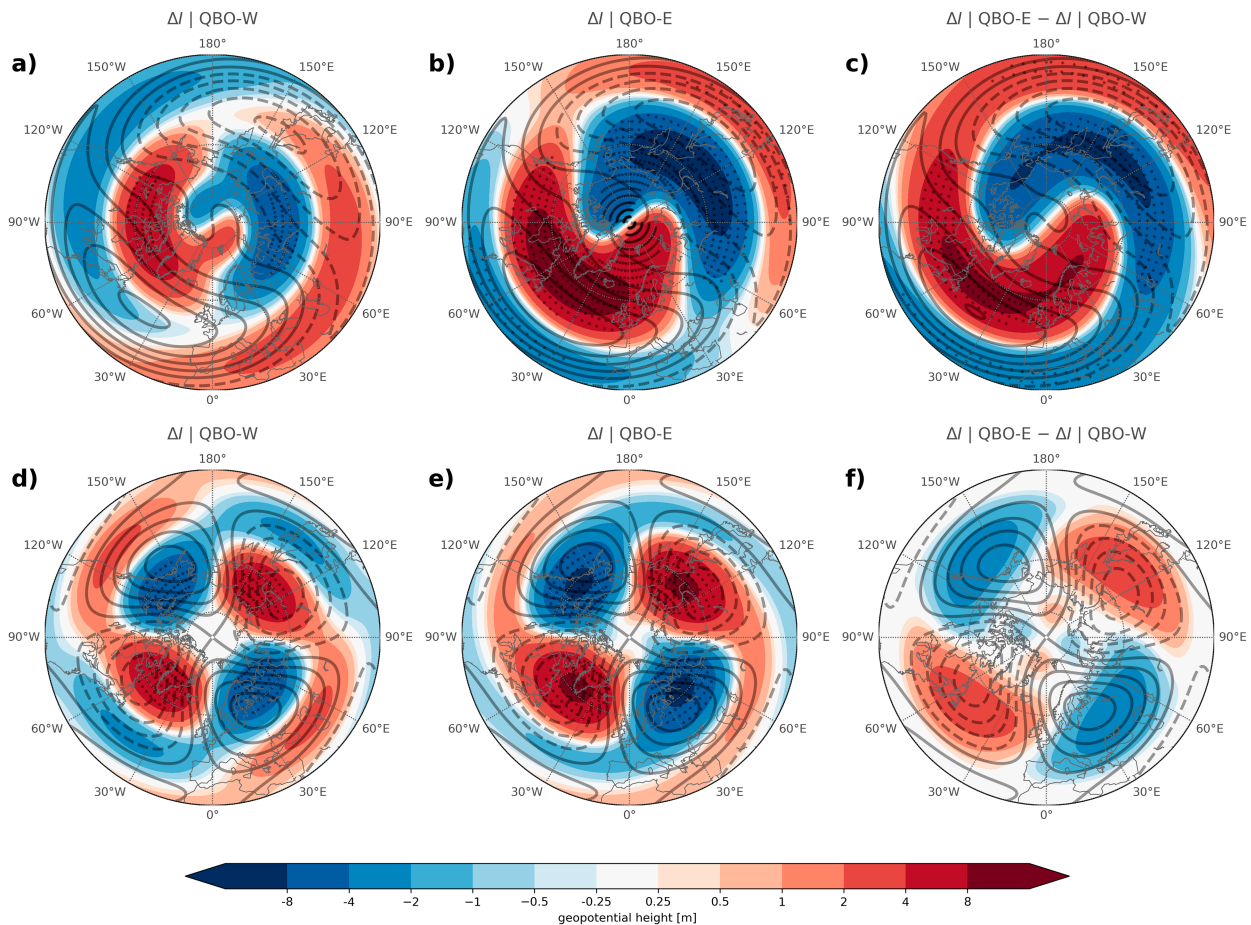


FIG. 3. Decomposition of the winter 200-hPa geopotential height response to sea ice loss into its (a)–(c) wavenumber-1 and (d)–(f) wavenumber-2 components, during (a),(d) QBO-W, (b),(e) QBO-E, and (c),(f) their difference. Climatologies are depicted by black contours, where solid and dashed lines indicate positive and negative values, respectively. Stippling denotes (a),(b),(d),(e) responses or (c),(f) response differences that are significant at the 95% level.

the geopotential height response at 200 hPa decomposed into wavenumbers 1 and 2. The 200-hPa level is chosen due to its proximity to the tropopause, since we are interested in planetary waves entering the stratosphere that can influence the polar vortex, and wavenumbers 1 and 2 because they are predominantly responsible for decelerating the winds in the polar stratosphere (Charney and Drazin 1961).

The wavenumber-1 component of geopotential height at 200-hPa response in $\Delta/QBO-E$ (Fig. 3b) aligns with the climatology, indicating constructive interference. In contrast, the wave-1 response is weaker and out of phase with climatology in $\Delta/QBO-W$ (Fig. 3a). These response differences between the two QBO phases are statistically significant (Fig. 3c) and imply that there is enhanced wave propagation from the troposphere into the stratosphere in $\Delta/QBO-E$ relative to $\Delta/QBO-W$, which helps explain the extra weakening of the polar vortex in $\Delta/QBO-E$ (Fig. 3b) than in $\Delta/QBO-W$ (Fig. 3a). The wave-2 responses to sea ice loss in both QBO states are out of phase with the climatology (Figs. 3d,e), and their differences are small and not statistically significant (Fig. 3f). Therefore, the different responses of planetary wave 1, between QBO phases, appear

more important for the diverging polar vortex responses than the wave-2 responses.

While the enhancement of climatological wave 1 in $\Delta/QBO-E$, but not in $\Delta/QBO-W$, may help to explain the differing polar vortex responses to sea ice loss, we seek further evidence by examining the EP fluxes and refractive index. The refractive index gives an indication of the regions of the atmosphere where waves will preferentially converge toward (positive anomalies) or away from (negative anomalies). The refractive index (squared) for wave 1 is inspected since this wavenumber is found to be reinforced in $\Delta/QBO-E$ (Fig. 3b) and appears to be a reason for the weakened vortex. For $\Delta/QBO-E$, there are statistically significant increases in refractive index between 30° and 50°N in the stratosphere (Fig. 4b) that are not found for $\Delta/QBO-W$ (Fig. 4a). These response differences are statistically significant (Fig. 4c). Regions where the refractive index squared is negative, and no wave propagation can occur, are shown in white in Fig. 4. In $\Delta/QBO-E$, there is a region of evanescence for both high and low sea ice states in the tropical stratosphere (whited out region) that is not found for $\Delta/QBO-W$. This is a result of the Holton–Tan effect which prevents wave propagation into the tropical

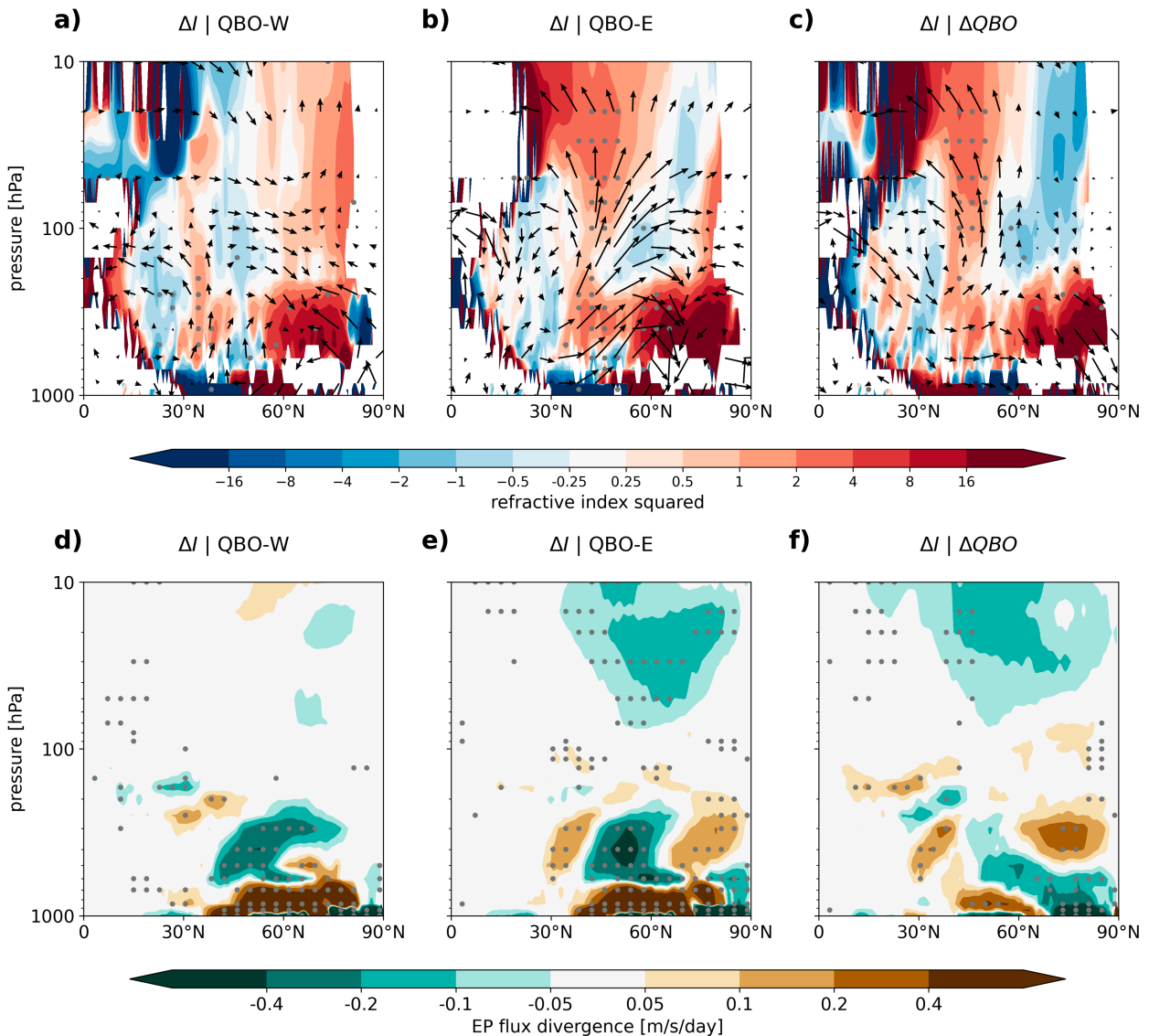


FIG. 4. Winter refractive index (squared; shading) and EP flux responses (arrows) to sea ice loss for (a) QBO-W, (b) QBO-E, and (c) their difference. EP flux divergence response for (d) QBO-W, (e) QBO-E, and (f) their difference. Stippling denotes (a), (b), (d), (e) responses or (c), (f) response differences that are significant at the 95% level.

stratosphere. The easterly winds in QBO-E phase prevent equatorward wave propagation beyond 20°N.

The region of positive refractive index response in $\Delta/QBO-E$ is collocated with an increase in upward EP flux (vectors in Fig. 4b). This is consistent with the refractive index increase and shows that additional wave activity is propagating into the stratosphere in this latitude band and ultimately is consistent with a more weakened polar vortex in response to sea ice loss during QBO-E. To understand what effect this additional wave activity into the stratosphere has on the background flow, we examine EP flux divergence response to sea ice loss. In $\Delta/QBO-W$ (Fig. 4d), there is little change in EP flux divergence in the polar stratosphere. However, this is not the case for $\Delta/QBO-E$ (Fig. 4e), where the increase in vertical EP flux

results in a notable convergence (i.e., negative divergence) of waves in the polar stratosphere. A convergence of EP flux corresponds with a deceleration of zonal winds; hence, this EP flux convergence helps explain the weakened polar vortex in $\Delta/QBO-E$ (Fig. 2b) compared to $\Delta/QBO-W$ (Fig. 2a).

The frequency of SSW events is also affected by the sea ice and QBO states (Fig. 5). For the extended winter [November–April (NDJFMA), inclusive], the experiments with future Arctic sea ice and QBO-E simulate, on average, around 10 SSWs per decade (Fig. 5a, light orange bar), which is significantly greater than other experiments that have about six SSWs per decade, close to the observed frequency. Early winter [November–January (NDJ)] has a lower SSW frequency than late winter (February–April) in all cases (Fig. 5a), but

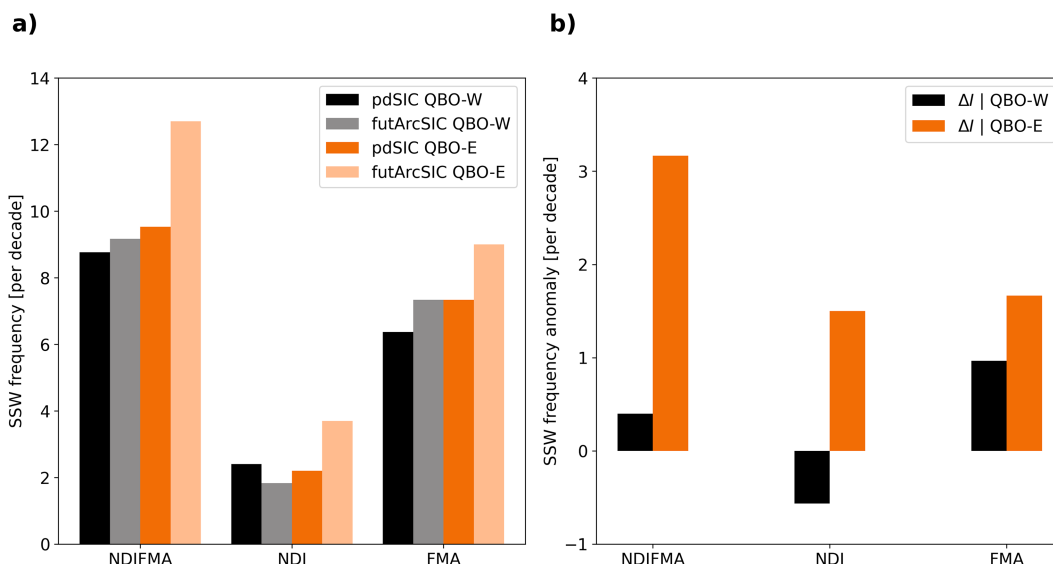


FIG. 5. (a) SSW frequency for different combinations of sea ice and QBO states and (b) SSW frequency response to sea ice loss conditioned on the QBO phase. (b) The difference between futArcSIC and pdSIC (ΔI) for both QBO-W (black bars) and QBO-E (orange bars). Frequencies are shown for NDJFMA, NDJ, and FMA.

the increase in SSW frequency in ΔI /QBO-E is found in both early and late winter (Fig. 5b). In ΔI /QBO-W, there is a small reduction in SSW frequency in early winter and an increase in late winter, albeit smaller than in ΔI /QBO-E.

Next, we consider the surface impacts of the differing dynamical responses to sea ice loss in QBO-E and QBO-W. For ΔI /QBO-W, sea level pressure (SLP) is reduced in the regions of maximum surface warming and sea ice loss (Fig. 5a). In contrast, ΔI /QBO-E additionally has a prominent increase in sea level pressure over northern Eurasia (Fig. 6b), indicating an enhancement of the climatological Siberian high. While the SLP response over the North Atlantic resembles the negative NAO in both QBO phases, the NAO-like anomaly is of significantly greater magnitude in ΔI /QBO-E. The difference between ΔI /QBO-W and ΔI /QBO-E projects on to the negative NAO/AO pattern, consistent with the larger jet shift in ΔI /QBO-E than in ΔI /QBO-W.

The surface temperature response to sea ice decline is dominated by warming in both QBO phases (Figs. 6d,e). One notable difference, however, is a cooling response over northern Eurasia seen in ΔI /QBO-E but not in ΔI /QBO-W. This response difference is found to be statistically significant (Fig. 6f) and suggests a cooling response to sea ice loss, related to the enhanced Siberian high, which is conditional on the QBO-E. The temperature response pattern in ΔI /QBO-E resembles the so-called warm Arctic, cold Eurasia (WACE) pattern discussed in past literature (e.g., Mori et al. 2014, 2019; Blackport et al. 2019; Screen and Blackport 2019; Zappa et al. 2021; Outten et al. 2023).

b. Interdependent responses to sea ice loss and El Niño

Like the QBO-E, El Niño exhibits a teleconnection to the extratropics that weakens the polar vortex (see Figs. S3c,d in

the online supplemental material). We hypothesize that any factor that modulates the polar vortex state may show evidence of a modulation of the polar vortex response to sea ice loss. We hence now look for the evidence of nonlinearity in the response to sea ice loss due to the phase of ENSO.

The familiar tropospheric response to sea ice loss of a weakened and equatorward shifted midlatitude jet is seen and is statistically significant in both neutral ENSO and El Niño cases (Figs. 7a,b). The stratospheric polar vortex is weakened in both cases, although by greater magnitude and only statistically significant in ΔI /neutral. However, there are no statistically significant differences between the zonal-mean zonal wind response to sea ice loss in neutral ENSO and El Niño (Fig. 7c), at least in winter. Larger and significant differences are found for late winter (February–April), when there is a prominent weakening of the polar vortex in ΔI /neutral but not in ΔI /El Niño. Relatedly, the response of the tropospheric-eddy-driven jet, notably an equatorward shift, is significantly larger in ΔI /neutral (Fig. 7d) than in ΔI /El Niño (Fig. 7e). This suggests that El Niño damps, or even prevents, a polar vortex response to sea ice decline. An alternative interpretation of the response difference in Fig. 7f is that reduced sea ice damps the polar vortex weakening by El Niño. It is noteworthy that the sign of this modulation is opposite to that seen for the QBO (Fig. 2) in the sense that although sea ice loss, El Niño, and QBO-E all weaken the polar vortex when acting alone, the combination of sea ice loss and QBO-E is greater than the sum of their separate effects, while the combination of sea ice loss and El Niño is less than the sum of their separate effects. Since the El Niño modulation of the response to sea ice loss is significant only in late winter, we focus on the months February–April in what follows unless stated otherwise.

Again, we inspect the wavenumber-1 and wavenumber-2 components of 200-hPa geopotential height to seek to

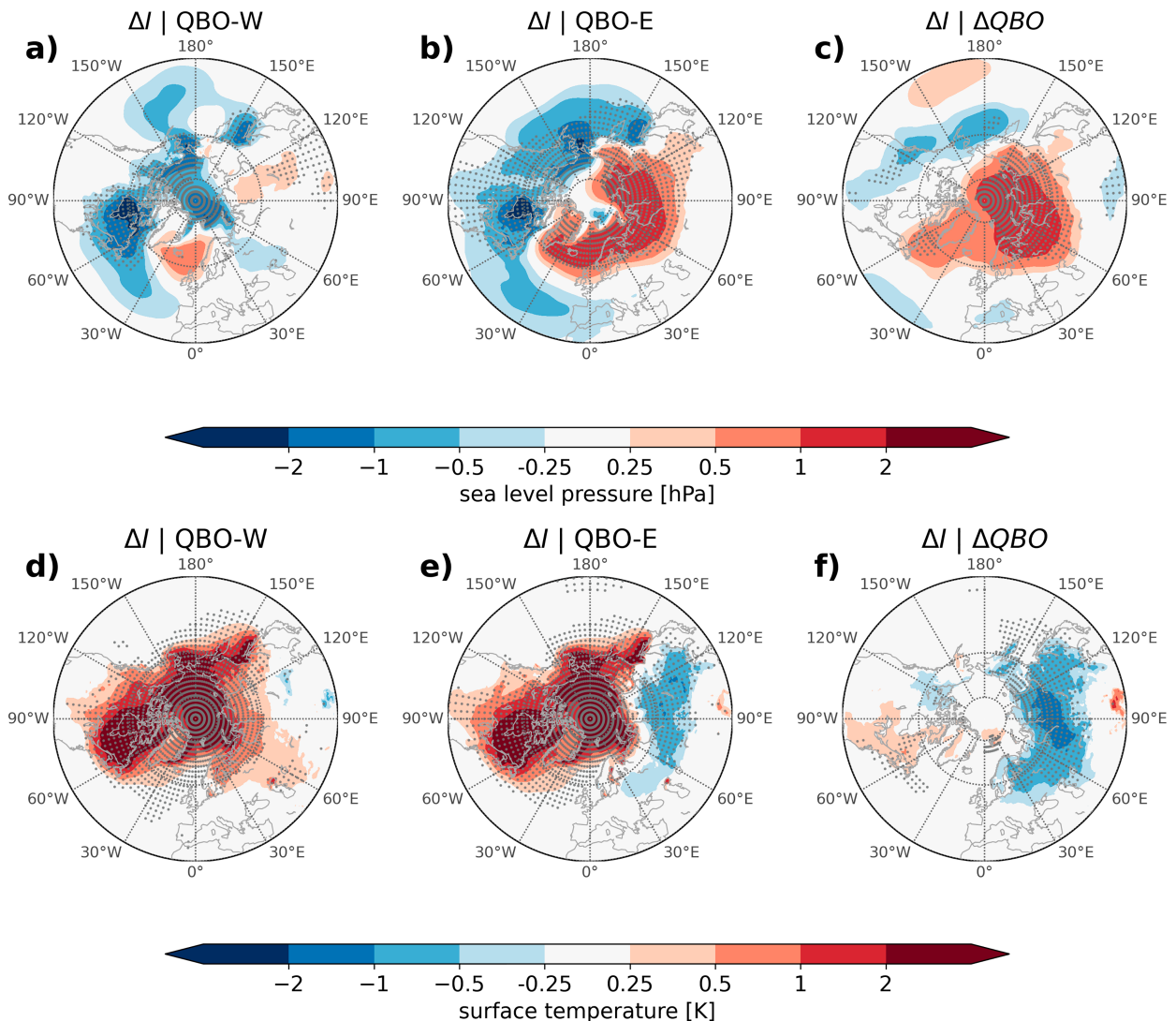


FIG. 6. Winter SLP response to sea ice loss for (a) QBO-W, (b) QBO-E, and (c) their difference. Winter near-surface air temperature response to sea ice loss for (d) QBO-W, (e) QBO-E, and (f) their difference. Stippling denotes (a),(b),(d),(e) responses or (c),(f) response differences that are significant at the 95% level.

understand the differing polar vortex responses to sea ice loss, but this time between neutral ENSO and El Niño. The wave-1 response for $\Delta I/\text{neutral}$ constructively interferes with the climatology (Fig. 8a). In $\Delta I/\text{El Niño}$, a similar response pattern is seen but of lesser magnitude (Fig. 8b). Although the wave-1 response difference is not statistically significant (Fig. 8c), it is consistent with greater wave propagation into the stratosphere in $\Delta I/\text{neutral}$ than in $\Delta I/\text{El Niño}$, which helps to explain the polar vortex weakening only in the former (Fig. 7). The wavenumber-2 responses are out of phase with the wave-2 climatology in both $\Delta I/\text{neutral}$ and $\Delta I/\text{El Niño}$ (Figs. 8d,e), and their difference corresponds to a slight eastward shift rather than a change in the magnitude of wave 2 (Fig. 8f). Again, it appears that wavenumber 1 is more important for the differing polar vortex responses in $\Delta I/\text{El Niño}$ and $\Delta I/\text{neutral}$ than is wave 2.

To determine whether $\Delta I/\text{El Niño}$ features less wave flux into the stratosphere, as might be expected from the weaker constructive interference compared to $\Delta I/\text{neutral}$, we inspect changes to the refractive index and EP flux. Both ENSO states feature that the refractive index increases between 35° and 50°N for wavenumber 1 (Figs. 9a,b), which are weaker for $\Delta I/\text{El Niño}$ although there are no significant differences between the two (Fig. 9c). This relatively lower refractive index for $\Delta I/\text{El Niño}$ may help explain the suppressed vertical flux of waves (vectors in Fig. 9a) compared to $\Delta I/\text{neutral}$ (vectors in Fig. 9b). Alternatively, these differences in wave activity into the stratosphere may relate to different tropospheric wave sources. The enhanced waves entering the stratosphere can be seen by the EP flux convergence in both $\Delta I/\text{El Niño}$ and $\Delta I/\text{neutral}$ (Figs. 9d,e), but this is of greater magnitude and statistically significant only in the neutral case. The

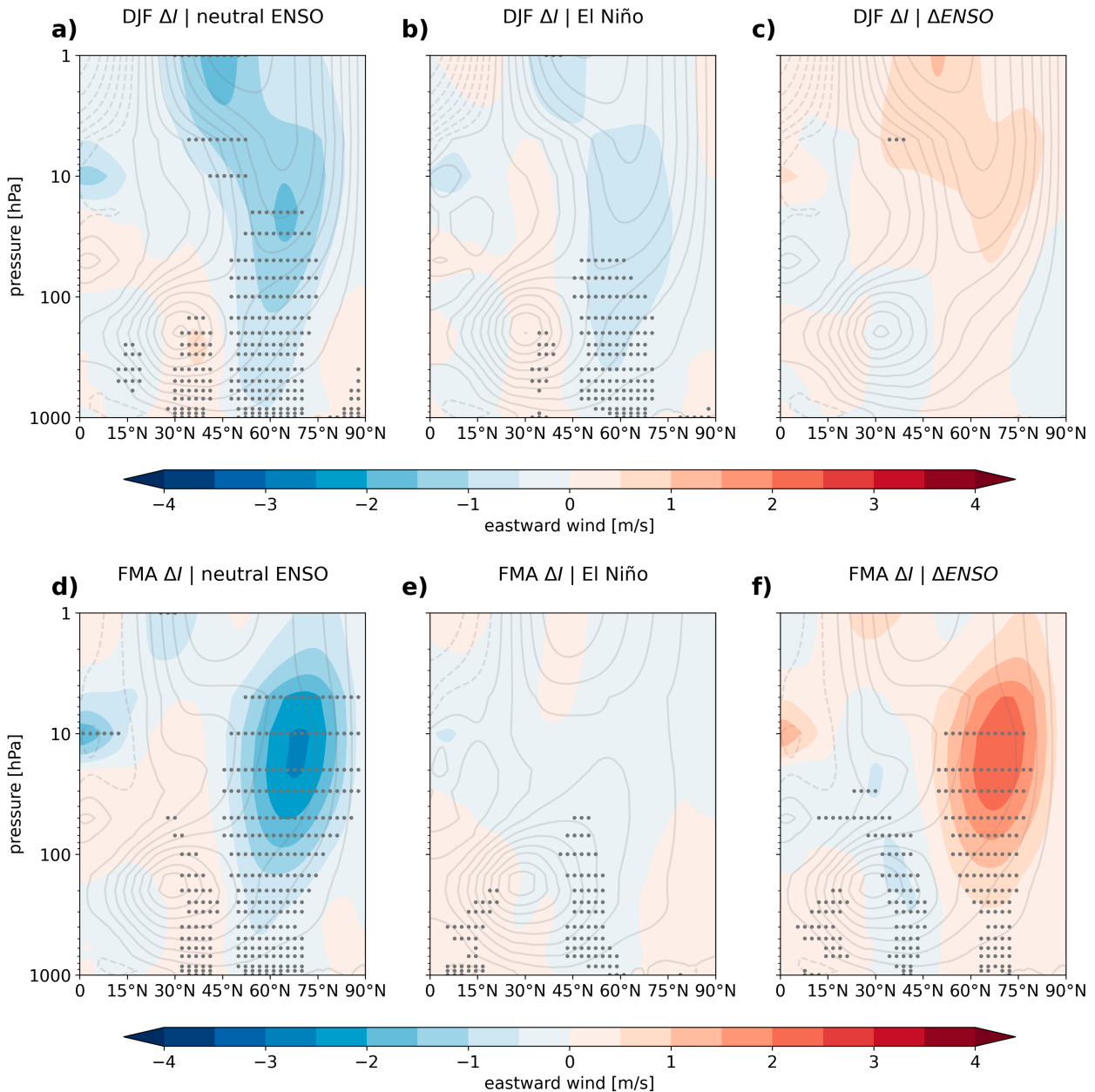


FIG. 7. Winter zonal-mean zonal wind response to sea ice loss for (a) neutral ENSO, (b) El Niño, and (c) their difference. Late winter zonal-mean zonal wind response to sea ice loss for (d) neutral ENSO, (e) El Niño, and (f) their difference. Stippling denotes (a),(b),(d),(e) responses or (c),(f) response differences that are significant at the 95% level.

difference, therefore, features a small EP flux divergence (Fig. 9f) that is consistent with the stronger polar vortex in $\Delta I/\Delta ENSO$ (Fig. 7f).

The atmospheric temperature response to sea ice loss is altered by the ENSO phase. For ΔI , warming is mostly confined to the Arctic lower troposphere which is directly heated by the areas of ice loss (Fig. S1). There is a slight warming in the Arctic stratosphere corresponding to the weakened polar vortex. The temperature response to ENSO is much more widespread, with local heating in the tropics that extends to the

stratosphere and to both hemispheres (Fig. S2b). This includes a stronger warming in the Arctic stratosphere corresponding to the ENSO–polar vortex teleconnection, as well as warming that extends down to the lower troposphere in the Arctic, but this is weaker than for ΔI . However, there are dissimilarities in the rate of Arctic amplification between ENSO phases. These differences in the meridional temperature gradient are shown in Fig. 10. There is more Arctic warming in ΔI /neutral ENSO compared with ΔI /El Niño (Fig. 10a, orange lines). Likewise, the amount of warming in the Arctic between ENSO states

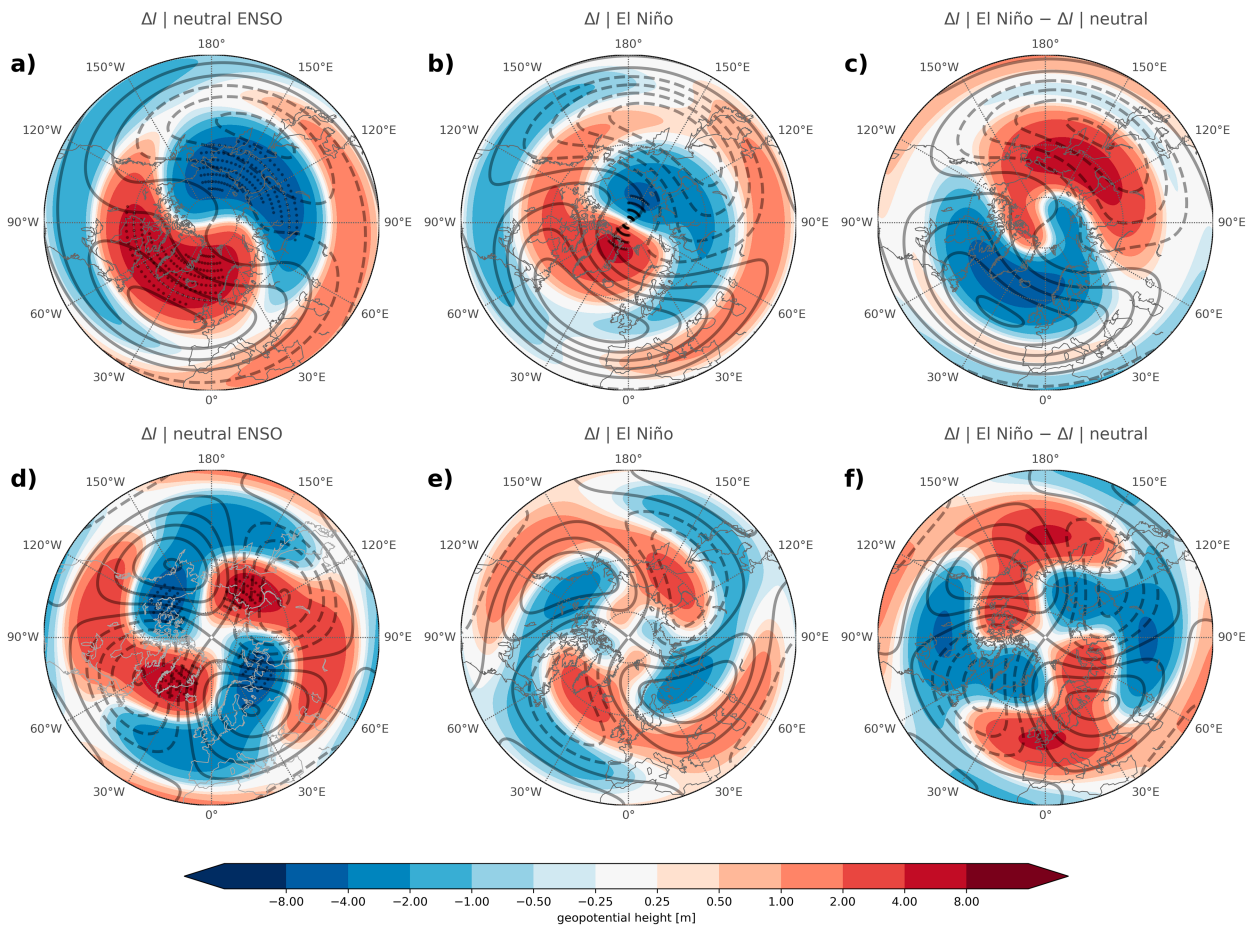


FIG. 8. Decomposition of the late winter 200-hPa geopotential height response to sea ice loss into its (a)–(c) wavenumber-1 and (d)–(f) wavenumber-2 components, during (a),(d) neutral ENSO, (b),(e) El Niño, and (c),(f) their difference. Climatologies are depicted by black contours, where solid and dashed lines indicate positive and negative values, respectively. Stippling denotes (a),(b),(d),(e) responses or (c),(f) response differences that are significant at the 95% level.

(Δ ENSO) is largest under present-day SIC rather than future Arctic SIC (Fig. 10a, green lines), indicating a modulation of the AA response by ENSO. There is a more muted response of the midlatitude jet consistent with this in Δ /El Niño compared to Δ /neutral ENSO, particularly in the Atlantic basin (Figs. 10c,d). While these jet changes do correspond in sign to the changes in temperature gradient, the Arctic warming difference is too small compared to the overall temperature gradient to explain the large difference in the jet response.

However, changes in SSW frequency between ENSO states have a much larger impact on the jet shift. In extended winter, and in early and late winter separately, SSWs are more frequent with future than present-day sea ice and in El Niño compared to neutral ENSO (Fig. 11a). These differences are larger in late than early winter. However, the SSW frequency increase in response to sea ice loss is larger under neutral ENSO than El Niño conditions (Fig. 11b). In other words, the combined effect of future sea ice and El Niño is less than the sum of their individual effects. A possible explanation for this is that the greater frequency of SSWs during El Niño limits

the ability of sea ice loss to induce further SSWs (or greater frequency of SSWs with future sea ice limits the ability of El Niño to induce further SSWs). Although there is no hard limit on the number of SSWs that can occur in each season per ensemble member, few members simulate multiple SSWs per year in any experiment (not shown), and in the El Niño experiments, almost every ensemble member features one SSW per year, but few exhibit multiple SSWs, irrespective of the sea ice state.

The smaller increase in SSW frequency in Δ /El Niño than in Δ /neutral is a possible reason for the relatively stronger mean polar vortex in Δ /ENSO. To attempt to quantify this potential relationship, we estimate the average effect on the mean polar vortex strength on an SSW by comparing the average polar vortex strength in years with one SSW compared to years without an SSW. We then scaled this difference in mean polar vortex strength by the difference in SSW frequency in Δ /ENSO, which is about 0.12 SSWs per year for late winter. We infer that the difference in SSWs could account for an increase in the late winter mean vortex response

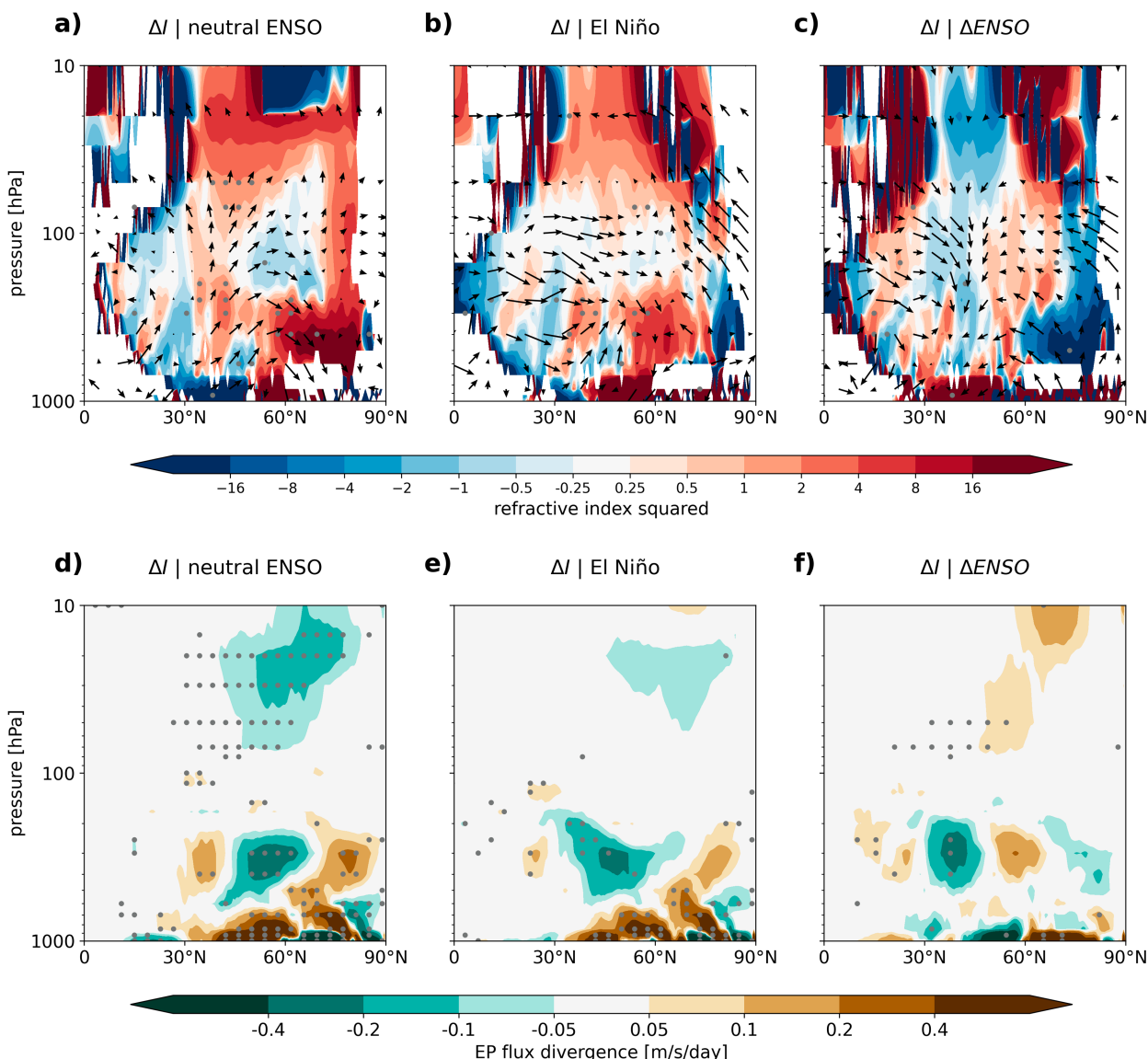


FIG. 9. Late winter refractive index (squared; shading) and EP flux responses (arrows) to sea ice loss for (a) neutral ENSO, (b) El Niño, and (c) their difference. EP flux divergence response for (d) neutral ENSO, (e) El Niño, and (f) their difference. Stippling denotes (a),(b),(d),(e) responses or (c),(f) response differences that are significant at the 95% level.

of approximately 0.5 m s^{-1} , which is only around a quarter of the response difference in $\Delta I/\Delta \text{ENSO}$. This means that the difference in SSW frequency cannot fully explain the modulation of the polar vortex by sea ice loss and El Niño combined.

Finally, we examine whether the ENSO modulation of the circulation response to sea ice loss impacts surface climate. For sea level pressure, there are notable differences in the responses between $\Delta I/\text{neutral}$ and $\Delta I/\text{El Niño}$. A negative AO pattern is featured in $\Delta I/\text{neutral}$ as there is an SLP increase over the Arctic and a decrease in the Atlantic and Pacific basins (Fig. 12a). There is also a strong negative NAO signal in the Atlantic where the largest anomalies of SLP are seen. The SLP response for $\Delta I/\text{El Niño}$ is weaker and unlike the AO (Fig. 12b). The difference resembles the positive AO, reflecting the larger negative AO

response in $\Delta I/\text{neutral}$ compared to $\Delta I/\text{El Niño}$, but is only statistically significant in a few limited regions (Fig. 12c). Over the North Atlantic, the differences in the SLP responses are related to differing precipitation responses, albeit also with limited statistical significance (Fig. 12f).

4. Discussion

We found that the QBO modulation of the response to sea ice loss, first shown by Labe et al. (2019), is reproduced in HadGEM3, an independent climate model. The mechanism proposed by Labe et al. (2019), that enhanced stationary wavenumber 1 in $\Delta I/\text{QBO-E}$ leads to a weaker polar vortex, relative to $\Delta I/\text{QBO-W}$, is also found in HadGEM3. We

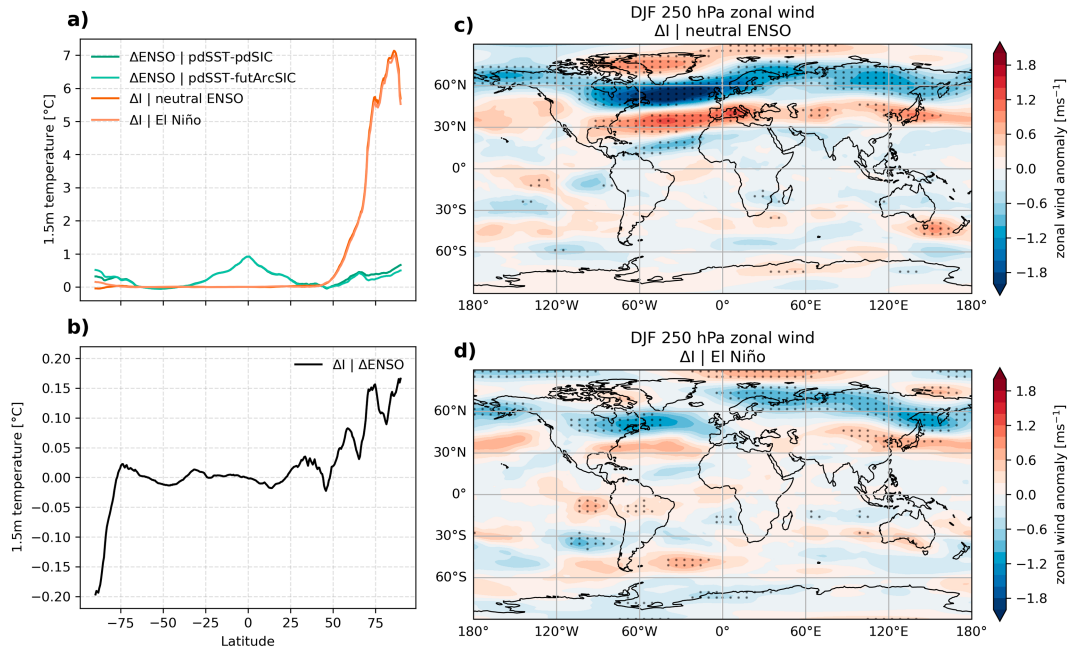


FIG. 10. (a) Change in the meridional temperature gradient for ENSO and sea ice loss and the U250 response. DJF zonal mean of 1.5-m temperature for Δ ENSO, given low and high SIC (green lines), Δ I given neutral ENSO and El Niño (orange lines) and (b) their difference Δ I/ Δ ENSO. (c) The DJF 250-hPa zonal wind for Δ I/neutral ENSO, and (d) as in (c), but for Δ I/El Niño.

further show that an increase in the refractive index in the mid-latitude stratosphere aids the propagation of waves into the polar stratosphere. We identify no notable difference in wave source, between Δ I/QBO-E and Δ I/QBO-W, and therefore, we propose the primary reason for their different polar vortex responses is that the QBO alters wave propagation and the

magnitude of anomalous wave activity entering the stratosphere in response to sea ice loss. A similar dependence of the response to sea ice loss on the QBO appears in this model when prescribed with Barents–Kara sea ice loss (Xu et al. 2024).

The differing polar vortex responses to sea ice loss correspond with differences in the troposphere for QBO-E and

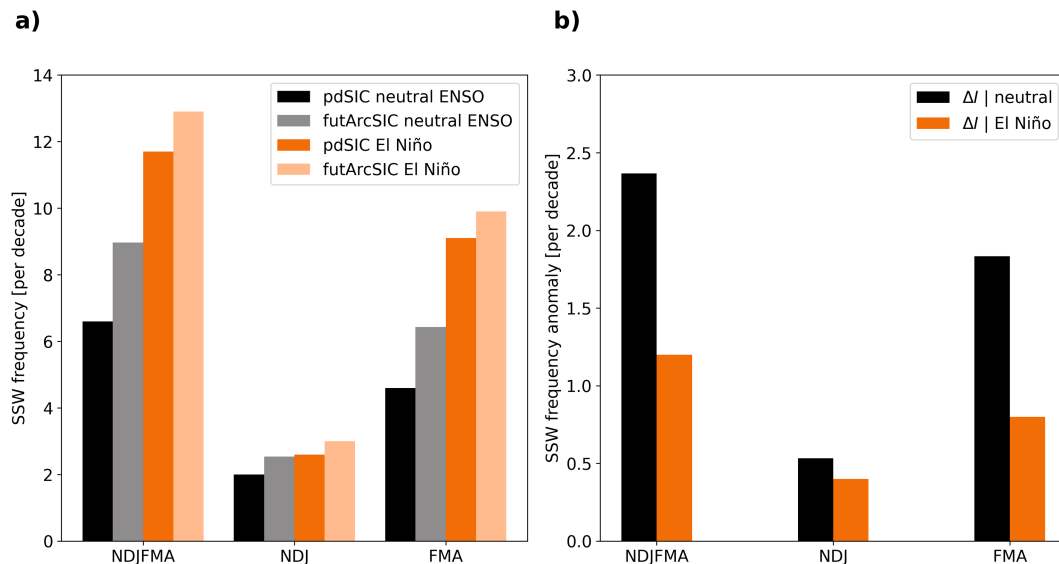


FIG. 11. (a) SSW frequency for different combinations of sea ice and ENSO states and (b) SSW frequency response to sea ice loss conditioned on the ENSO phase. (b) The difference between futArcSIC and pdSIC (Δ I) for both QBO-W (black bars) and QBO-E (orange bars). Frequencies are shown for NDJFMA, NDJ, and FMA.

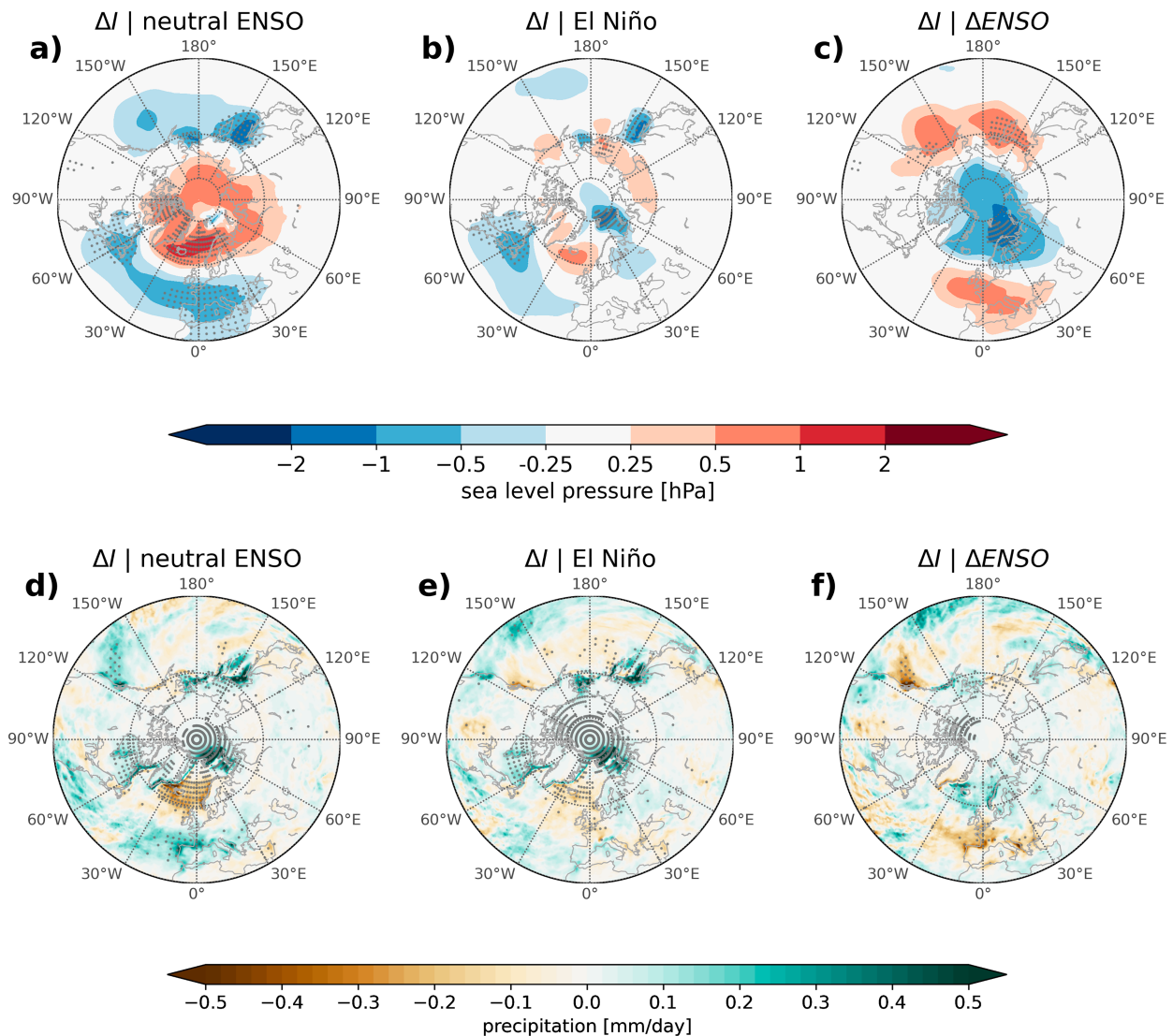


FIG. 12. Late winter SLP response to sea ice loss for (a) neutral ENSO, (b) El Niño, and (c) their difference. Late winter precipitation response to sea ice loss for (d) neutral ENSO, (e) El Niño, and (f) their difference. Stippling denotes (a),(b),(d),(e) responses or (c),(f) response differences that are significant at the 95% level.

QBO-W. Under QBO-E, but not QBO-W, there is an enhancement of the Siberian high and related cooling over Eurasia. This QBO dependence of the Eurasian temperature response to sea ice loss is also consistent with [Labe et al. \(2019\)](#). The possible Eurasian winter cooling response to sea ice loss has been widely debated and remains contentious (e.g., [Mori et al. 2014, 2019](#); [Blackport et al. 2019](#); [Screen and Blackport 2019](#); [Cohen et al. 2020](#); [Zappa et al. 2021](#); [Outten et al. 2023](#)), partly because it is not robustly simulated in models ([Screen and Blackport 2019](#); [Smith et al. 2022](#)). Our results suggest that the mean QBO state could be one source of model divergence if the full range of possible QBO states is not represented and provides a physical reason for the apparent intermittency of statistical relationships between Arctic warming and Eurasian cooling ([Overland et al. 2016, 2021](#)). This relationship is also

found when compositing ERA5 by low and high sea ice winters and QBO phase, but the difference between QBO phases is not statistically significant over most of Eurasia (not shown). It may, therefore, be important to carry out analyses under QBO-E conditions to yield more consistent results.

The negative NAO response to sea ice loss is also of greater magnitude in QBO-E than in QBO-W (although not significantly so), which we propose is partially a consequence of the weakening of the polar vortex and increased SSW frequency during QBO-E and not QBO-W. Although this is suggestive of downward influence of the stratosphere on the troposphere, causality is not clear. For instance, a shift in the jet stream equatorward can induce a weakened polar vortex, via increased vertical wave activity into the midlatitude stratosphere ([Smith et al. 2022](#)). Hence, the tropospheric jet shift can be

both a cause of and a result of a weakened polar vortex, and there likely exists feedback between the tropospheric and stratospheric responses.

We found that ENSO also modulates the polar vortex response to sea ice loss. While the polar vortex weakened in response to sea ice loss during neutral ENSO, under El Niño, this response was damped. This can be partially explained by a plateauing of the increase in SSW frequency for the low sea ice, El Niño case appearing to be the result of a saturation of SSWs as the frequency exceeds one SSW per year. This plateauing results in a smaller increase in SSW frequency for $\Delta/\text{El Niño}$ than for $\Delta/\text{neutral ENSO}$ (Fig. 11b), meaning the polar vortex will be weakened more relatively for $\Delta/\text{neutral ENSO}$. However, this does not fully explain the ENSO modulation of the polar vortex response to sea ice loss. Additionally, as with the QBO, the modulation by ENSO can be understood from the perspective of wave propagation. El Niño suppresses the stationary-wave-1 response to sea ice loss and, therefore, the vertical wave activity into the stratosphere, compared to neutral ENSO. As a result, the polar vortex weakens more in neutral ENSO than in El Niño. At the surface, the negative NAO and AO response to sea ice loss is more pronounced during neutral ENSO than El Niño.

Our original hypothesis was that the magnitude of the polar vortex response to sea ice loss is dependent on the polar vortex mean state. More specifically, climate drivers such as QBO-E and El Niño that weaken the polar vortex would promote a greater weakening of the polar vortex in the response to sea ice loss. Since $\Delta/\text{El Niño}$ does not display a weaker vortex than $\Delta/\text{neutral}$, we reject this hypothesis as it is not generally true, only true in the QBO case. We surmise the opposite modulating effects of El Niño and QBO-E arise because, although both factors weaken the polar vortex, they do so by different mechanisms. The effect of El Niño is mainly an increase in wave activity, while the effect of QBO-E is largely a confinement of wave activity [see, e.g., Walsh et al. (2022) for more discussion]. QBO-E and El Niño thus have different effects on wave propagation, despite both weakening the polar vortex. This helps explain why they have opposite modulating effects on the polar vortex response to sea ice loss, which appears governed not by the polar vortex mean state but instead by wave propagation into the stratosphere in response to sea ice loss. This conclusion is supported by two recent studies that emphasize the importance of the atmospheric mean state, particularly in the “neck region” winds, for dictating the extent to which anomalous wave activity in response to sea ice loss propagates to the polar vortex (Sigmond and Sun 2024; Mudhar et al. 2024). The timing of the modulation also differs between the QBO and ENSO. A possible reason for the ENSO modulation occurring later in winter is that the ENSO teleconnection to the polar vortex is greatest in late winter (Ayarzagüena et al. 2018), whereas the QBO influence on the polar vortex is manifest throughout winter (Zhang et al. 2019) or is possibly largest in early winter (Pascoe et al. 2006). One additional possible reason for the enhanced sea ice loss response during neutral ENSO but not El Niño is that the QBO-E phase (which makes up half of the ensemble mean) has been shown to strengthen the Walker

circulation in the western Pacific, which in turn generates a wave train that interacts at high latitudes to weaken the polar vortex (Yamazaki et al. 2020; Haynes et al. 2021; Anstey et al. 2022). There is also evidence from Honda et al. (1999) that sea ice loss in the Sea of Okhotsk specifically weakens the Aleutian low, which would reduce the influence of the El Niño response in the low sea ice case, possibly explaining the sign of the modulation we see.

A previous study examining the interaction of the PDO and sea ice loss found that the weakening of the polar vortex induced by sea ice loss and by the warm phase of the PDO, analogous to El Niño, was not additive, with their combined effects weakening the polar vortex less than would be expected from the sum of their individual contributions (Simon et al. 2022). This is consistent with our results for the interaction of El Niño and sea ice loss and suggests similar mechanisms may operate on interannual and decadal time scales. Although it has been suggested that the PDO and ENSO have opposite effects on the stratospheric polar vortex (Seabrook et al. 2023)—specifically, that the warm PDO strengthens the polar vortex on decadal time scales (due to enhanced stratospheric water vapor) while El Niño weakens it on interannual time scales (due to enhanced upward wave activity)—our results suggest that it is not the polar vortex strength that modulates its response to sea ice loss but instead how ENSO (and QBO) alters wave propagation into the stratosphere in response to sea ice loss. However, longer experiments than used here would be needed to assess whether ENSO and PDV modulate the response to sea ice in the same way or whether stratospheric water vapor increases play a role. Other modes of variability such as the Atlantic multidecadal variability (AMV) and the Atlantic meridional overturning circulation (AMOC) have also been shown to modulate the polar vortex. Liang et al. (2024) show that the response to sea ice loss varies depending on the AMV phase. For negative AMV, they find that the vortex is weakened more in late winter, with a corresponding NAO negative signal at the surface. Similarly, Li et al. (2018) show evidence that the AMV modulates the response to sea ice decline via an increase in Ural blocking in AMV negative, which leads to a weaker polar vortex and a negative AO pattern at the surface. This leads to cold temperatures over Eurasia via the stratospheric pathway (Nakamura et al. 2016). For AMOC, Dimmire-Miles et al. (2022) find that the AMOC modulates polar vortex strength with a period of about 90 years, via an enhanced QBO-W which then strengthens the polar vortex through the Holton–Tan effect. These studies show the important role that Atlantic variability plays in modulating the polar stratosphere in similar ways to what we present here.

Further support for interdependent effects of sea ice, QBO, and ENSO on the polar vortex is found in Ma et al. (2022), who explored their combined effects using observations and hindcasts. They found that the polar vortex is weaker during QBO-E than during QBO-W when combined with reduced sea ice and neutral ENSO and weaker for neutral ENSO than for El Niño when combined reduced sea ice and neutral QBO. These results are consistent with our simulations, despite different methods, and the smaller sample sizes and potential confounding factors when using observations.

Our results suggest significantly larger surface impacts for northwest Europe when reduced sea ice and QBO-E coincide than would be expected from the sum of their individual effects and weaker surface impacts when reduced sea ice and El Niño coincide than would be expected from their sum. However, we note that even during QBO-E and neutral ENSO, the NAO response to sea ice loss is relatively small, no more than one quarter of the total year-to-year standard deviation [in DJF for QBO-E and February–April (FMA) for neutral ENSO]. That said, there is evidence that the circulation response to sea ice loss and more widely, the predictable component of NAO variability, is underestimated in models (e.g., Smith et al. 2022; Screen et al. 2022; Eade et al. 2014; Dunstone et al. 2016; Scaife and Smith 2018) so this may be an underestimate. However, recent work by Sigmond and Sun (2025) find that using a new emergent constraint, the observed jet stream response lies within the model spread. A further note of caution is that experiments with a lower-resolution version of HadGEM3 (N96, ~135 km) did not reveal any significant sensitivity of the response to sea ice to the QBO or ENSO for reasons that are unclear but could include poor representation of the relevant processes at lower horizontal and vertical resolutions (Walsh 2024). The experiments at N96 resolution follow the same protocol to N216 otherwise, but using different initial conditions.

5. Conclusions

We find that both the QBO and ENSO modulate the response to sea ice loss but in opposite ways with respect to how these drivers influence the polar vortex strength. We argue that how these drivers effect wave propagation into the stratosphere is more important the mean polar vortex state for explaining their moderating influence on the response to sea ice loss. For the QBO, the stratospheric polar vortex weakens in response to sea ice loss during QBO-E but not during QBO-W. For ENSO, the stratospheric polar vortex weakens in response to sea ice loss during neutral ENSO but not during El Niño. While sea ice loss enhances the climatological stationary wave-1 and vertical wave propagation into the stratosphere, leading to a weaker polar vortex during QBO-E or neutral ENSO, these responses are suppressed during QBO-W or El Niño. These differences in the polar vortex response are related to SSW frequency, which is increased by sea ice loss during QBO-E but not QBO-W and increased more by sea ice loss during neutral ENSO than El Niño. At the surface, sea ice loss enhances the Siberian high and causes cooling over Eurasia in QBO-E but not in QBO-W. Both QBO-W and El Niño damp the negative NAO response to sea ice loss, relative to QBO-E or neutral ENSO, respectively.

Finally, while the interdependent responses to sea ice loss and QBO, and to sea ice loss and ENSO, have been presented from the perspective of the QBO or ENSO modulating the response to sea ice loss, there is an alternative interpretation. That is, that sea ice loss alters the extratropical response to the QBO or ENSO. Our results are compatible with either or both interpretations. We have provided a plausible mechanism to explain how the QBO and ENSO might affect the

response to sea ice loss, but further work is required to fully distinguish between these two possible interpretations. This may include running longer simulations to allow the QBO to oscillate between phases and hence sampling a wider set of QBO conditions and imposing a larger sea ice loss forcing to increase the significant signal.

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Data availability statement. All data used from the HadGEM3-GC31-MM sea ice, ENSO, and QBO experiments are hosted on Zenodo (<https://doi.org/10.5281/zenodo.13736537>).

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