Intraseasonal and Interdecadal Jet Shifts in the Northern Hemisphere: The Role of Warm Pool Tropical Convection and Sea Ice

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ABSTRACT

This study uses cluster analysis to investigate the interdecadal poleward shift of the subtropical and eddy-driven jets and its relationship to intraseasonal teleconnections. For this purpose, self-organizing map (SOM) analysis is applied to the ECMWF Interim Re-Analysis (ERA-Interim) zonal-mean zonal wind. The resulting SOM patterns have time scales of 4.8–5.7 days and undergo notable interdecadal trends in their frequency of occurrence. The sum of these trends closely resembles the observed interdecadal trend of the subtropical and eddy-driven jets, indicating that much of the interdecadal climate forcing is manifested through changes in the frequency of intraseasonal teleconnection patterns.

Two classes of jet cluster patterns are identified. The first class of SOM pattern is preceded by anomalies in convection over the warm pool followed by changes in the poleward wave activity flux. The second class of patterns is preceded by sea ice and stratospheric polar vortex anomalies; when the Arctic sea ice area is reduced, the subsequent planetary wave anomalies destructively interfere with the climatological stationary waves. This is followed by a decrease in the vertical wave activity flux and a strengthening of the stratospheric polar vortex. An increase in sea ice area leads to the opposite chain of events. Analysis suggests that the positive trend in the Arctic Oscillation (AO) up until the early 1990s might be attributed to increased warm pool tropical convection, while the subsequent reversal in its trend may be due to the influence of tropical convection being overshadowed by the accelerated loss of Arctic sea ice.

1. Introduction

The jet streams in the Northern Hemisphere (NH) exhibit fluctuations in both their strength and latitude. On the interdecadal time scale, previous studies have found that increased greenhouse gas (GHG) driving coincides with a poleward shift of both the subtropical jet (e.g., Archer and Caldeira 2008; Chen et al. 2008; Fu et al. 2006; Fu and Lin 2011; Hu and Fu 2007; Lu et al. 2008; Seager et al. 2003) and the midlatitude eddy-driven jet (e.g., Yin 2005; Lorenz and DeWeaver 2007; Lu et al. 2008; Kidston et al. 2011). This poleward shift of the eddy-driven jet also corresponds to a trend toward the positive phase of the North Atlantic Oscillation (NAO) and Arctic Oscillation (AO) teleconnection patterns1 (Thompson and Wallace 2000).

Beginning in the early 1990s, after a 20-yr upward trend toward its largest positive value, the 5-yr running mean winter NAO/AO index began to decline, becoming negative at about 2010 (http://www.cpc.ncep.noaa.gov/products/precip/CWlink/daily_ao_index/JFM_season_ao_index.shtml). During the same time period, Arctic sea ice area was observed to undergo a steep decline in the summer and autumn (Comiso et al. 2008). Anomalously low sea ice area has been linked to the negative phase of the NAO/AO (Seierstad and Bader 2009; Deser et al. 2010;

1 As discussed by Thompson and Wallace (2000), the NAO can be regarded as the North Atlantic regional contribution to the hemispheric-scale AO.
Francis et al. 2009; Jaiser et al. 2012; Liu et al. 2012), al-
cluding to the possibility that the recent trend toward the
negative phase of the NAO/AO is driven by the declining
sea ice (Jaiser et al. 2012).

As will be shown in this study, the interdecadal pole-
ward jet shift can be expressed in terms of the change in the
frequency of occurrence of intraseasonal time-scale
teleconnection patterns, as identified with cluster analysis
(self-organizing map analysis; see section 2). Mathem-
tically, this can be written as

$$\Delta U(\theta, p) \approx \sum_{i=1}^{K} \Delta f_i U_i(\theta, p),$$  \hspace{1cm} (1)

where $\Delta U(\theta, p)$ is the trend of the zonal-mean zonal wind
at latitude $\theta$ and pressure $p$; $U_i(\theta, p)$ are the $i = 1, \ldots, K$
cluster patterns; and $\Delta f_i$ are the trends in the frequency of
the cluster patterns. (For brevity, rather than frequency
of occurrence, we use the shorter-term frequency.) For
the data to be examined in this study, the cluster pat-
terns can be interpreted as varying at the intraseasonal
time scale, since as we will see, each of the $U_i$ has an $e$
-folding time scale between 4.8 and 5.7 days. The in-
terdecadal trend in variables such as the zonal-mean
zonal wind $\Delta U$ typically has contributions from inter-
decadal driving, possibly from GHG driving and changes
in Arctic sea ice, and from climate noise (Feldstein 2002).
Thus, to the extent that (1) accurately describes the
interdecadal trend $\Delta U$ and if climate noise makes a rela-
tively small contribution to $\Delta U$, the form of (1) suggests
that the interdecadal driving of $\Delta U$ is primarily man-
ifested through interdecadal changes in the frequency of
the intraseasonal cluster patterns.

The choice to use cluster patterns for this attribution
study is motivated by the continuum property of atmo-
spheric teleconnection patterns (Franzke and Feldstein
This cluster analysis approach was used by Lee and
Feldstein (2013) to examine the interdecadal poleward
shift of the midlatitude jet in the Southern Hemisphere
(SH). They showed that the SH interdecadal jet shift can be
accurately represented by an interdecadal trend in the
frequency of intraseasonal time-scale cluster patterns as
described by (1). This linkage between interdecadal vari-
bility and intraseasonal teleconnection patterns suggests
that it can often be helpful to decompose interdecadal
variability into two separate but related questions. These
are as follows: 1) What dynamical processes drive the
intraseasonal teleconnection patterns that contribute to
the interdecadal variability? 2) What are the interdecadal
processes that can account for the changes in the fre-
quency of the intraseasonal teleconnection patterns? In
this study, we focus on the former question. After showing
that the interdecadal poleward jet shifts can be accurately
described by changes in the frequency of four intraseasonal
cluster patterns, we proceed to examine the intraseasonal
dynamical processes that drive these four cluster patterns.
One advantage to this approach is that, for the relatively
short intraseasonal time scale, causal relationships can be
revealed more readily than with steady-state or long-term
averaged responses. While it may be counterintuitive that
interdecadal variability has such short time-scale linkages,
as will be discussed later, there are physical arguments as to
how intraseasonal time-scale processes can play an im-
portant role in the interdecadal time-scale changes of the
atmospheric circulation.

2. Data and methodology

For this study, the European Center for Medium-Range
Weather Forecasts (ECMWF) Interim Re-Analysis (ERA-
Interim) dataset (Dee et al. 2011) for the years 1979–2008
is used. We also use daily National Oceanic and At-
mospheric Administration outgoing longwave radiation
(OLR) data as a proxy for tropical convection and daily
and monthly Arctic sea ice data obtained from the Na-
tional Snow and Ice Data Center (http://nsidc.org/).

We apply the method of self-organizing maps (SOMs)
(Kohonen 2001; Johnson et al. 2008) to the zonal-mean
zonal wind during the NH winter season [December–
February (DJF)]. Prior to calculating the SOMs, the
zonal-mean zonal winds are multiplied by the cosine of
latitude and mass weighted in the vertical direction. The
primary motivation for examining the zonal-mean zonal
wind, rather than a zonally varying quantity, is to facilitate
comparison with previous studies, such as many of the
papers listed in section 1. The SOM calculation organizes
the daily data into a much smaller number of $m \times n$
cluster patterns, where each cluster represents a large
number of similar daily patterns. The SOM patterns are
displayed on a two-dimensional $m \times n$ grid, where $m$
is the number of rows and $n$ is the number of columns. The
SOM analysis organizes the data so that similar patterns
are assigned to a nearby location and dissimilar patterns
to a distant location on the grid. The SOM patterns are
obtained by minimizing the Euclidean distance between
each of the SOM patterns and the observed daily field in
an $N$-dimensional phase space, where $N$ is the number of
grid points within the domain.

The choice of the size of the SOM grid is motivated by
two criteria: that the number of SOM patterns is not in-
conveniently large and that the SOM patterns are similar
to the observed daily fields. These criteria are evaluated
for three different SOM grid sizes, $4 \times 1, 6 \times 1$, and $8 \times 1$
(see Table 1). Columns 2–5 of Table 1 show, for each
SOM pattern, the mean pattern correlation between the
daily field and the representative SOM pattern for that day (i.e., the SOM pattern with the smallest Euclidean distance on that day). The first column of Table 1 shows the weighted-mean pattern correlation over all four SOMs, where the weighting is based on the different SOM frequencies. As can be seen, the mean pattern correlation increases modestly from 0.50 to 0.58 when the size of the SOM grid doubles. These results suggest that a $4 \times 1$ grid is sufficient for the goals of this study.

To explore the physical processes associated with the SOM patterns, composites and correlations are utilized that are based on either the representative SOM pattern or the SOM frequency. In most evaluations of statistical significance, a two-sided Student’s $t$ test is performed. For the composites, the calculation is based on those days when a particular SOM pattern has the smallest Euclidean distance. If two days occur within 15 days of each other, then the day with the larger Euclidean distance is discarded. This procedure finds 68, 70, 65, and 66 days for SOM1, SOM2, SOM3, and SOM4, respectively. These values are used for the number of degrees of freedom in the test of statistical significance. For the correlations, all quantities are averaged over the DJF months. With the exception of correlations involving the global-mean surface air temperature (SAT), where the number of degrees of freedom is determined by the method of Oort and Yienger (1996), for all other correlations the number of degrees of freedom is specified to equal 29, corresponding to the number of winter seasons in this study. In contrast to the above tests of statistical significance, for OLR, to which a nine-point local spatial smoothing is applied, a Monte Carlo approach is used. For this calculation, to determine the probability distribution, we performed 1000 sets of composite calculations with randomly chosen days and sample sizes that are the same as those in the SOM composites.

3. SOM patterns and interdecadal variability

The linear trend in the zonal-mean zonal wind (Fig. 1a) indicates that there has been a poleward shift in the latitude of both the subtropical jet and the midlatitude eddy-driven jet, as found in many previous studies. (In Fig. 1a, the climatological subtropical jet can be identified by the shallow zonal wind maximum near 30°N and the climatological midlatitude eddy-driven jet can be identified by its deep vertical structure and local maximum in the lower troposphere near 42°N.) As will be seen below, this trend is reasonably well captured by the sum of the trends from all SOM patterns. The first SOM pattern (SOM1; Fig. 2, first row) corresponds primarily to an equatorward shift of the subtropical jet and a poleward shift of the eddy-driven jet. The second SOM pattern (SOM2; Fig. 2, second row) indicates a poleward displacement of both the subtropical and eddy-driven jets, along with a strengthening of the latter jet. The third SOM pattern (SOM3; Fig. 2, third row) resembles SOM1 but with the opposite sign for each anomaly, thus corresponding to a poleward shift of the subtropical jet and an equatorward shift of the eddy-driven jet. The fourth SOM pattern (SOM4; Fig. 2, fourth row) has a structure that is close to being opposite to that of SOM2, thus coinciding with an equatorward movement of both jets and a weakening of the eddy-driven jet.

Even though all four SOM patterns exhibit distinct properties, all four SOM patterns are associated with a large, statistically significant ($p < 0.05$) composite AO index [from the National Oceanic and Atmospheric Administration/Climate Prediction Center (NOAA/CPC)], with maximum standardized values of 1.18, 0.82, $-1.10$, and $-0.64$, for SOM1, SOM2, SOM3, and SOM4, respectively. These results indicate that the SOM patterns can identify different AO-like patterns that occur in nature.

The DJF-mean frequency time series (the number of days in each winter season for which a particular SOM pattern is the representative pattern) is shown for each SOM pattern in the right panels of Fig. 2. Linear regression is used to determine the trend for the frequency time series for each SOM pattern. The corresponding trends associated with each SOM are calculated by multiplying the frequency trend of each SOM pattern by the corresponding SOM spatial pattern, as in (1), and then dividing by 90 (DJF) days. The results (shown in the two middle columns of Fig. 2) indicate that the SOM2 and SOM4 frequency trends dominate the 1979–2008 time period but that during 1990–2008 all four patterns have a similar amplitude in their trends. The latter time period was chosen because the rapid decline in summer and autumn Arctic sea ice began in the early 1990s. (Note that the sign of the trend patterns for SOM1 and SOM4 is opposite to that of the corresponding SOM patterns because of the downward trend in the SOM1 and SOM4 frequencies.) The sum of the trends from all four SOM patterns for the 1979–2008 time period is shown in Fig. 1b. A comparison with the full observed trend (Fig. 1a) shows that a large

<table>
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<th>SOM grid</th>
<th>All SOMs</th>
<th>SOM1</th>
<th>SOM2</th>
<th>SOM3</th>
<th>SOM4</th>
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<td>4 $\times$ 1</td>
<td>0.50</td>
<td>0.50</td>
<td>0.52</td>
<td>0.42</td>
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<td>6 $\times$ 1</td>
<td>0.54</td>
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<td>0.64</td>
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<td>8 $\times$ 1</td>
<td>0.58</td>
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<td>0.62 and 0.70</td>
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fraction of the observed trend in the zonal-mean zonal wind can be expressed by the sum of the individual trends of the four SOM patterns. Analogous results were obtained by Lee and Feldstein (2013) for the SH.

The statistical significance of the linear trends in the SOM frequency is also indicated in the right column of Fig. 2. It found that the SOM2 and SOM4 frequency trends are statistically significant at \( p < 0.10 \) for the full 1979–2008 time period and the SOM1 frequency trend is statistically significant at \( p < 0.05 \) for the shorter 1990–2008 time period. The SOM3 frequency is found to be marginally significant at \( p < 0.10 \) for the shorter time period, as its trend was significant at \( p < 0.05 \) (\( p < 0.10 \)) for initial years of 1988 (1989), slightly misses \( p < 0.10 \) for 1990, and is further from the \( p < 0.10 \) threshold for the initial years of 1991 and 1992.

We next discuss the time scale for each of the SOM patterns. The time scales are determined by calculating lagged autocorrelations of time series that are obtained by projecting the daily zonal-mean zonal wind field onto each SOM pattern. The corresponding \( e \)-folding time scales are 4.8, 5.7, 5.3, and 5.2 days for SOM1, SOM2, SOM3, and SOM4, respectively (see Fig. 3). The time over which the lagged autocorrelations decay to zero varies from 14 to 19 days for each SOM pattern. These findings provide the support for the statement in the introduction [see also (1)] that the interdecadal trend of the NH zonal-mean zonal wind can be expressed in terms of the interdecadal trend in the frequency of the intraseasonal (4.8–5.7 days) time-scale SOM patterns. The implication of this result is that interdecadal driving is manifested mostly through its impact on the frequency of the SOM patterns and to a lesser extent through a slow interdecadal modulation of the background flow by the forcing. Otherwise, the \( e \)-folding time scales of the SOM patterns would have to be much longer, with
some SOM patterns persisting for an entire winter season. It is also important to state that (1) does not imply causality in the sense of the intraseasonal SOM patterns driving the interdecadal trend or vice versa. Equation (1) merely indicates the relationship between the interdecadal trend and the intraseasonal SOM patterns, with the linkage between the interdecadal trend and intraseasonal SOM patterns being realized via the trend in the SOM frequencies.

One may question why the primary impact of the interdecadal forcing is to alter the frequency of the much shorter time-scale SOM patterns rather than to drive slow interdecadal changes to the background flow. We provide an example of this type of relationship by considering the question addressed by Gong et al. (2010) in the context of the impact of the El Niño–Southern Oscillation (ENSO) on the frequency of positive and negative phase events of the intraseasonal southern annular mode (SAM), the most prominent teleconnection pattern in the Southern Hemisphere. They found that during La Niña the subtropical jet becomes weaker and vice versa during El Niño, with the zonal wind difference between La Niña and El Niño (excluding the SAM events; i.e., corresponding to the slow changes of the background state) being 4 times smaller than the SAM zonal-mean zonal wind anomalies driven by the short time-scale eddy momentum fluxes. The reason that the difference in jet strength between La Niña and El Niño is relatively small, being about 10% of the climatological
value, lies in the changes to the meridional potential vorticity (PV) gradient, which are much greater, corresponding to a 75% decline on the equatorward side of the midlatitude jet. These changes in the PV gradient result in a much greater frequency of anticyclonic wave breaking\(^2\) (for La Niña) together with the concomitant increase in the strength of the poleward eddy momentum fluxes, followed by the excitation of the positive phase of the SAM. Opposite features were obtained for El Niño. As a result, during La Niña, the positive phase of the SAM is much more frequent than the negative phase and vice versa for El Niño. These SAM events were found to have an e-folding time scale of between 12 and 27 days, depending upon the particular season and whether ENSO is active. Similar time scales for the SAM and the closely related SH zonal index have been found by Feldstein and Lee (1998), Lorenz and Hartmann (2001), and Gerber et al. (2008). In the NH, the analogous northern annular mode (NAM) has been shown to exhibit an e-folding time scale of 7–15 days (e.g., Feldstein and Lee 1998; Lorenz and Hartmann 2003; Gerber et al. 2008).

In other words, the slow forcing indeed changes the background state, but the lion’s share of the actual circulation change is realized through intraseasonal time-scale processes. This perspective can be summarized as follows: 1) interdecadal forcing \(\rightarrow\) 2) alters the low-frequency background flow \(\rightarrow\) 3) influences high-frequency eddy driving of the background flow \(\rightarrow\) 4) changes the frequency of the intraseasonal SOM patterns, which corresponds to changes in the low-frequency background flow. As indicated in the above ENSO/SAM example, the third step, 3) \(\rightarrow\) 4), can be very large. It is beyond the scope of our manuscript to determine whether this type of process is taking place in response to GHG or sea ice driving, but in our view this picture presents a plausible explanation for why the trend in the SOM frequencies is able to account for most of the long-term trend (Fig. 1).

This type of relationship between intraseasonal processes and interdecadal variability is not restricted to zonal-mean wind. It has been shown that a substantial fraction of the interdecadal variability in the NH atmospheric sea level pressure (Johnson et al. 2008; Johnson and Feldstein 2010) and 250-hPa geopotential height (Lee et al. 2011) can be expressed in terms of interdecadal fluctuations in the frequency of a relatively small number of SOM patterns that fluctuate on a 5–10-day time scale.

### 4. Exploring the possible GHG-driven response

The spatial structure of the SOM patterns and the trends in their frequencies allude to the possibility that SOM2 and SOM4 are linked to increased GHG driving, whereas SOM1 and SOM3 are linked to the decline in Arctic sea ice. This is because the trends associated with SOM2 and SOM4 correspond to a poleward shift of both jets and those for SOM1 and SOM3 correspond to an equatorward (poleward) shift of the eddy-driven (subtropical) jet (see Fig. 2), which match the trends associated with GHG driving and sea ice loss in modeling and observational studies, as discussed in the introduction. To explore this possible relationship with GHG driving, we correlate the DJF-mean SOM frequency time series with the DJF NOAA/CPC global-mean SAT, defined as the deviation from the 1901–2000 time average. We use global-mean SAT as an indicator of the thermodynamic response of the atmosphere to GHG loading. However, since the global-mean SAT is also modulated by internal variability, most notably that by ENSO, we apply 7-yr low- and high-pass filters to the global-mean SAT and SOM frequency time series in order to evaluate the relationship with ENSO.

Among the unfiltered, low-pass, and high-pass correlations between the global-mean SAT and the SOM frequencies, the only statistically significant \((p < 0.10)\) correlation is found for SOM4 at high frequencies with a value of 0.45. The unfiltered correlations were all very small, with absolute values less than 0.15, with the signs of the high- and low-frequency correlations opposing each other for each SOM pattern. The signs of these correlations

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2 Following Thornicroft et al. (1993), anticyclonic (cyclonic) wave breaking corresponds to Rossby waves that break on the anticyclonically (cyclonically) sheared side of the jet.
are positive for SOM2 and negative for SOM4, with values of 0.48 (statistically significant at $p < 0.15$) and $-0.28$, respectively. Perhaps in the future, when the data period becomes longer, which would increase the number of degrees of freedom, the correlations between patterns resembling SOM2 and SOM4 and the global-mean SAT may be statistically significant. However, even if the correlations are statistically significant, it would be only one piece of evidence, and a more conclusive attribution requires identification of the mechanism that links the GHG warming and the poleward jet shift.

The correlations between the high-pass SOM frequencies and the Niño-3.4 index [the sea surface temperature (SST) averaged over 5°N–5°S, 120°–170°W] yield statistically significant ($p < 0.05$) correlations only for SOM2 and SOM4, with values of $-0.57$ and 0.43, respectively. Furthermore, the linear correlation between the high-frequency global-mean SAT and the Niño-3.4 index is found to be 0.64, also a statistically significant ($p < 0.05$) value. These results suggest that El Niño raises the global-mean SAT and is responsible for a decrease (increase) in the frequency of SOM2 (SOM4) and vice versa during La Niña. Since the signs of the high- and low-frequency correlations (between the SOMs and SAT) are opposite, this finding also suggests that, in the long term, higher values of the global-mean SAT may be associated with La Niña–like atmospheric conditions. As mentioned above, however, a concrete evaluation of this possibility will have to wait for a longer data period in the future.

5. A possible mechanism for the simultaneous poleward shift of both jets

We next examine possible driving mechanisms for SOM2 and SOM4. The link between ENSO and the SOM2 and SOM4 frequencies suggests that these patterns are driven in part by anomalies in tropical convection. Since the extratropical response to tropical convection takes place on a time scale of 5–10 days (Hoskins and Karoly 1981), we examine daily lagged composites of anomalous OLR for the SOM patterns (see Fig. 4). For this calculation, the OLR anomalies are shown as a function of longitude averaged between 10°S and 10°N. Consistent with the above relationship with ENSO, SOM2 is associated with enhanced (reduced) convection over the Maritime Continent (central tropical Pacific Ocean) (i.e., a La Niña–like OLR pattern), with SOM4 showing the opposite relationship. Although not related to ENSO, the pattern of the OLR anomalies associated with SOM1 resemble those for SOM2, except that they are opposite in sign. Moreover, the anomalies in tropical convection are seen to lead the SOM patterns (except for SOM3), suggesting that intraseasonal tropical convection may excite the SOM patterns and perhaps plays an important role in the interdecadal modulation of the frequencies of the SOM patterns.

This connection between intraseasonal time-scale tropical convection and the SOM patterns may be understood from the findings of Moon and Feldstein (2009) and Park and Lee (2013). These studies showed that poleward-propagating wave trains excited by tropical convection can have a large impact on the zonal-mean flow, as the eddy momentum flux associated with the wave trains alters the zonal-mean flow in a manner that the ensuing synoptic-scale eddy momentum fluxes drive the midlatitude jet poleward.

We investigate if this mechanism operates in the excitation of SOM2 by examining wave activity [Eliassen–Palm (EP)] fluxes. Figure 5 presents lagged composites of the anomalous EP flux vectors and their divergence, with the planetary-scale (zonal wavenumbers 1 and 2; Fig. 5, right) and synoptic-scale3 (zonal wavenumbers greater than or equal to 3; Fig. 5, left) contributions shown separately. (We will not show lagged EP flux composites for SOM4, as the EP flux vectors and their divergence are found to be opposite in sign to those for SOM2.) As can be seen, between lag $-30$ and $-20$ days, there is poleward planetary-scale wave activity propagation in the upper troposphere from the deep tropics to about 25°N, which is consistent with the expected response from the enhanced warm pool tropical convection associated with SOM2. During the same time interval, confined mostly to mid-latitudes, there is equatorward synoptic-scale wave activity propagation.

As evidenced from the EP flux divergence for SOM2 (Fig. 5), the primary impact of this wave activity propagation from both the planetary and synoptic scales is a deceleration of the zonal-mean zonal wind throughout much of the troposphere at lag $-20$ days near 20°N (Fig. 6). Between lag $-20$ and $-10$ days, the planetary-scale EP flux (Fig. 5, right) is dominant in the subtropics while the synoptic-scale EP flux (Fig. 5, left) is stronger in mid-latitudes, resulting in a weakening of the subtropical jet and a slight strengthening of the midlatitude jet at lag $-10$ days. These changes to the zonal wind structure correspond to an increase in the anticyclonic shear between the two jets. Such changes to the subtropical and midlatitude jet structure typically results in a strengthened equatorward

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3Synoptic scale is typically regarded as excluding zonal wavenumbers 3 and perhaps 4. However, since the EP fluxes associated with all zonal wavenumbers greater than or equal to 3 is dominated by its synoptic-scale contribution, for brevity, we use the term synoptic scale to refer to all wavenumbers excluding 1 and 2.
synoptic-scale wave activity flux and a poleward shift of the midlatitude jet (Gong et al. 2010, and references therein). Both of these features are seen between lag $-10$ and $-5$ days and again between lag $-5$ and 0 days. The short 5.7-day $e$-folding time scale for SOM2 is consistent with the eddy driving and zonal-mean flow changes being largest in the latter time interval. These results are consistent with those of Moon and Feldstein (2009) and Park and Lee (2013) and therefore suggest that increased warm pool tropical convection and the subsequent poleward Rossby wave propagation alters the zonal-mean flow in a manner that leads to a strong equatorward synoptic-scale wave activity flux that excites the SOM2 pattern.

6. Identifying the Arctic sea ice–driven response

Recent studies have shown that anomalies in summer and autumn Arctic sea ice area$^4$ were followed in the winter by anomalies in the strength and latitude of the eddy-driven jet that project onto the NAO/AO spatial pattern, along with widespread changes in midlatitude

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$^4$Sea ice area is defined as the area of the region that is covered by sea ice. The contribution to the sea ice area from an individual grid cell comes from the portion of the grid cell that is covered in sea ice (for more information, see http://nsidc.org/cryosphere/seacie/data/terminology.html).
Surface air temperature, storm-track location, stationary wave location, blocking frequency, cold-air outbreaks, and snow cover (e.g., Singarayer et al. 2006; Honda et al. 2009; Petoukhov and Semenov 2010; Deser et al. 2010; Overland and Wang 2010; Liu et al. 2012). This relationship has been most noticeable over the past several years, when anomalously low sea ice area in the autumn has been followed by severely cold winters over large parts of the midlatitudes in the NH (e.g., Petoukhov and Semenov 2010).

We examine lagged correlations between the detrended DJF SOM frequencies and anomalous Arctic sea ice area, with the months chosen for the sea ice anomalies ranging from 12 months prior to 12 months following the SOM frequency anomalies. Unlike for the global-mean SAT, the Arctic sea ice correlations are not split into low- and high-frequency components because Arctic sea ice is not correlated with ENSO (Fig. 7f). As can be seen in Figs. 7a,c, the frequencies of SOM1 and SOM3 are strongly linked to Arctic sea ice area over a wide range of negative lags. These correlations imply that SOM1 occurs more frequently during the DJF winter following a period of enhanced sea ice and vice versa for SOM3. Most strikingly, for SOM1 and SOM3, statistically significant correlations ($p < 0.05$) are found as far back as 12 months before the start of the winter season. Also, because SOM1 (SOM3) has a large positive (negative) projection onto the AO, not surprisingly, positive (negative) AO DJF winters are preceded by positive (negative) sea ice area anomalies (Fig. 7e).

The above correlations are also consistent with the interdecadal decline in Arctic sea ice, because the SOM1 (SOM3) frequency shows a downward (upward) trend (see Fig. 2). In contrast, there is no indication that SOM2 and SOM4 are preceded by anomalies in the sea ice area during the preceding months (Figs. 7b,d).

7. Possible mechanism for the impact of Arctic sea ice on meridional jet displacements

In the previous section, a linkage between Arctic sea ice area and the SOM1 and SOM3 frequencies on interannual time scales was found. That result motivates us to examine whether anomalies in Arctic sea ice contribute toward the driving of these two SOM patterns on the intraseasonal time scale. To investigate this possible relationship, we first perform lagged composites of the 15-day running mean Arctic sea ice area, where the sea ice is averaged over $60^\circ$–$90^\circ$N, $30^\circ$–$120^\circ$E, a region that is centered on the Barents and Kara Seas (Fig. 8). (This domain is chosen because, as we will see, SOM1 and SOM3 are most closely linked to sea ice concentration anomalies in this region.) As can be seen, SOM1 is preceded by positive statistically significant sea ice area anomalies ($p < 0.05$) over a wide range of negative lags. (Lagged composites for the entire Arctic Ocean showed similar results, except that the number of days with statistically significant values was slightly less.) Although the other SOM pattern composites are not found to be statistically significant ($p < 0.05$), SOM3 is an exception, as it exhibits statistically significant ($p < 0.15$) negative composite values at negative lags. Because SOM1 and SOM3 both show a strong connection to Arctic sea ice area at interannual time scales (Fig. 7), in spite of the connection between Arctic sea ice and SOM3 being weak at intraseasonal time scales, in this section we examine the possible relationship between Arctic sea ice and both SOM1 and SOM3 at intraseasonal time scales.

To investigate the possible mechanisms by which Arctic sea ice drives SOM1 and SOM3, we again examine the anomalous EP flux vectors and zonal-mean zonal wind anomalies (Figs. 9–11). Figure 11 indicates that SOM1 is associated with a persistent and strengthened stratospheric polar vortex and vice versa for SOM3. For both SOM1 and SOM3, the presence of large amplitude stratospheric polar vortex anomalies at negative lags suggests that the atmospheric response to sea ice is first imprinted upon the stratosphere, which in turn alters the frequency of occurrence of these two SOM patterns. In contrast, the stratospheric anomalies associated with SOM2 and SOM4 are much weaker and shorter lived (see Fig. 6 for SOM2; not shown for SOM4).

a. Stratospheric polar vortex and wave activity flux

We first investigate the relationship between the strength of the stratospheric polar vortex and the occurrence of SOM1 and SOM3. Beginning with SOM1, it can be seen that the zonal-mean zonal wind anomalies are relatively small at lag $-60$ days (Fig. 11). An anomalous downward planetary-scale EP flux in the lower and middle stratosphere near $60^\circ$N results in an EP flux divergence (Fig. 9) and corresponding strengthening of the stratospheric polar vortex by lag $-45$ days. Continuation of this downward (and also equatorward) anomalous planetary-scale EP flux further accelerates the stratospheric polar vortex within the lag $-45$ to $-30$ day and lag $-30$ to $-10$ day periods. During this time interval, an anomalous equatorward synoptic-scale EP flux leads to the acceleration of the midlatitude jet and deceleration of the subtropical jet in the troposphere. The occurrence of these particular synoptic-scale EP fluxes is consistent with the findings of Garfinkel et al. (2013), who show with a general circulation model that an accelerated stratospheric polar vortex coincides with a synoptic-scale eddy feedback in the troposphere that shifts the midlatitude
tropospheric jet poleward. [Consistently, modeling studies such as Polvani and Kushner (2002), Kushner and Polvani (2004), Song and Robinson (2004), and Simpson et al. (2009) show that changes in the stratosphere influence the troposphere through downward control and an eddy feedback.] From lag $-10$ to 0 days, both the planetary-scale and synoptic-scale eddies in the troposphere undergo a substantial strengthening that drives

Fig. 5. The composite SOM2 EP flux vectors and their divergence (shading) for (right) planetary-scale (zonal wavenumbers 1 and 2) and (left) synoptic-scale (zonal wavenumbers greater than or equal to 3) eddies. Each panel shows a time average for the interval indicated. Vectors shown correspond to those that have a least one statistically significant ($p < 0.05$) component.
the anomalous zonal winds toward a pattern that closely resembles SOM1 both in spatial structure and amplitude (see Fig. 2). This rapid acceleration of the zonal wind field is consistent with the short 4.8-day e-folding time scale of SOM1.

SOM3 exhibits similar features in its anomalous EP flux and zonal-mean zonal wind (Figs. 10 and 11) as that for SOM1, except for a change in sign. For example, from lag −60 to −10 days, the anomalous planetary-scale EP fluxes are upward and poleward. This results in a substantial weakening of the stratospheric polar vortex by lag −10 days. Also, the anomalous synoptic-scale EP fluxes are poleward in the troposphere [again consistent with the synoptic-scale eddy feedback discussed in Garfinkel et al. (2013)], which results in a deceleration and equatorward shift of the midlatitude jet and acceleration and poleward shift of the subtropical jet. Between lag −45 and 0 days, the anomalous tropospheric fluxes attain their largest amplitude and the SOM3 pattern is excited. Again, the rapid increase in the strength of the EP fluxes during this time interval is consistent with the short 5.3-day e-folding time scale of SOM3.

b. Planetary wave interference and the stratospheric polar vortex

For SOM1 and SOM3, the question remains as to what process can change the strength of the stratospheric polar vortex. As shown in Garfinkel et al. (2010), the strength of the stratospheric polar vortex can be altered by interference between the planetary wave contributions to the anomalous circulation and the climatological stationary eddy fields in the troposphere. Their study showed that constructive interference leads to a strengthening of planetary-scale vertical wave activity propagation hence to a deceleration of the stratospheric polar vortex and vice versa for destructive interference. Other studies that have linked interference in the troposphere with changes in the strength of the stratospheric polar vortex include Smith et al. (2011), Garfinkel et al. (2012), and Jiang et al. (2014, manuscript submitted to J. Atmos. Sci.).

To investigate whether interference is playing a role in the changes to the strength of the stratospheric polar vortex, we show in Fig. 12 the anomalous (shading) and the climatological (contours) planetary-scale streamfunction at 300 hPa, both with their zonal-mean values subtracted. As can be seen for SOM1 during the lag −60 to −45 day and lag −30 to −10 day intervals, the positive anomaly overlaps with the negative climatological eddy streamfunction over eastern Asia and the North Pacific and vice versa over northern Europe. From lag −45 to −30 days and from lag −10 to 0 days, this overlap occurs primarily over eastern Asia and the North Pacific. Therefore, destructive interference may indeed account
Fig. 7. Time-lag correlations between monthly-mean Arctic sea ice area and the DJF-mean frequency of occurrences for (a) SOM1, (b) SOM2, (c) SOM3, and (d) SOM4. Also shown are correlations between the monthly-mean Arctic sea ice area and (e) the DJF-mean AO index and (f) the DJF-mean Niño-3.4 index. The correlation values shown with black dots indicate statistical significance ($p < 0.05$) for a two-sided Student’s $t$ test. Lag 0 corresponds to December, and negative (positive) lags correspond to sea ice leading (lagging) the SOM frequency.
for the acceleration of the stratospheric polar vortex for SOM1. Analogous interference features (but of opposite sign) can be seen for SOM3 (Fig. 12). Unlike for SOM1, however, for the lag 260 to 245 day and lag 245 to 230 day intervals, there is no clear pattern of interference, and constructive interference becomes apparent only at later periods (from lag 230 to 210 days and from lag 210 to 0 days) over the North Pacific/eastern Asia and northern Europe.

In a recent modeling study with the Community Atmospheric Model, version 5 (CAM5), Peings and Magnusdottir (2014) examined the response of the atmospheric circulation to sea ice anomalies. For the years 2007–12, they found interference within the troposphere to be followed by a weakening of the stratospheric polar vortex and the excitation of the negative NAM, as in the present study.

c. Sea ice anomalies and planetary-scale waves

To examine the plausibility of the anomalous 300-hPa streamfunction field in Fig. 12 arising as a response to diabatic heating anomalies associated with changes in Arctic sea ice, we compare the anomalous 300-hPa streamfunction field in Fig. 12 with the atmospheric response to sea ice concentration anomalies in climate models. From this perspective, an anomaly in sea ice concentration may excite a particular wave field that either constructively (SOM1) or destructively (SOM3) interferes with the climatological stationary eddy field, which leads to the more frequent excitation of SOM1 or SOM3.

We first compare the results in Fig. 12 with the findings of Deser et al. (2007), who examine the transient atmospheric response to sea ice concentration anomalies in the National Center for Atmospheric Research (NCAR) Community Climate Model, version 3 (CCM3) for the winter season. In their model calculation, they imposed a negative sea ice concentration anomaly over the Barents and Greenland Seas and a positive sea ice concentration anomaly over the Labrador Sea. They found that the initial adjustment exhibits a baroclinic vertical structure with an anomalous low at 1000 hPa and an anomalous high at 300 hPa over the region where sea ice has been removed and vice versa over the region where sea ice has been added. This baroclinic local response persisted for 2–3 weeks. Throughout the following 2 months, the vertical structure of the atmospheric circulation became increasingly barotropic and increasingly resembled the negative phase of the NAM (i.e., SOM3).

They also showed with a linear primitive equation model that the initial baroclinic circulation can be understood as the forced response to diabatic heating in the lower troposphere and the equivalent barotropic circulation 2 months later as being due to driving by transient eddy vorticity and heat fluxes. Similar findings were obtained by Deser et al. (2004) by separating the response to Arctic sea ice forcing into indirect responses (projection onto the leading EOF) and a direct response (obtained as a residual).

In another more recent modeling study of the atmospheric response to changes in sea ice over the Barents Sea, Liptak and Strong (2014) found a similar initial baroclinic response over the Arctic Ocean that was followed 2 months later by an equivalent barotropic response that extended into midlatitudes. When sea ice was reduced over the Barents Sea, the initial baroclinic resembled that of Deser et al. (2007); when the sea ice was increased, the sign of the baroclinic response was reversed.

For SOM3, inspection of Fig. 13 shows a reduction in sea ice over the Greenland and Barents Seas with the largest amplitude anomalies occurring from lag −60 to −30 days. (Note that the sea ice concentration anomalies in the Barents and Kara Seas in Fig. 13 have opposite signs, with those in the Barents Sea being stronger and more widespread. This contrasts many studies where the sea ice concentration anomalies in both seas have the same sign.) In the upper troposphere, at 300 hPa, an anomalous high can be seen over the Greenland, Barents, and Kara Seas at lag −45 to −10 days. This is followed by the equivalent barotropic negative NAM (Fig. 11) almost 2 months later. These results are consistent with the findings of Deser et al. (2007) discussed
above. For SOM1, there is an increase in sea ice mostly over the Kara Sea that is largest from lag $-60$ to $-45$ days, which then slowly declines for the rest of the time period (Fig. 13). An anomalous low at 300 hPa is observed over the same region and time period (Fig. 12), which is followed about 1 month later by the equivalent barotropic positive NAM (Fig. 11). Thus, the sea ice concentration, 300-hPa streamfunction, and zonal-mean zonal wind anomalies are consistent with sea ice contributing to the excitation of SOM1 and SOM3 via the wave response to the diabatic heating associated with the sea ice changes, followed by interference, changes to the strength of the stratospheric polar vortex, and the subsequent excitation of SOM1 and SOM3.

Fig. 9. As in Fig. 5, but for SOM1.
8. Discussion and conclusions

This study uses SOM analysis to examine the interdecadal poleward shift of the subtropical and eddy-driven jets and its relationship to intraseasonal teleconnections as determined from SOM analysis. It is found that these jet shifts can be expressed in terms of an interdecadal trend in the frequency of four SOM patterns, each of which has an e-folding time scale of between 4.8 and 5.7 days. The SOM analysis finds two classes of zonal-mean zonal wind patterns. One class is associated with simultaneous shifts of the subtropical and eddy-driven jets in the same direction and the second class with jet shifts in opposite directions. The interdecadal trend in the frequency of the...
The first class of SOM patterns corresponds to a poleward shift of both jets and that for the second class to a poleward (equatorward) shift of the subtropical (eddy-driven jet).

In this study we exploit the short time-scale characteristics to explore the physical mechanisms that drive the jet variability represented by the SOM patterns. Among the first class, it was found for SOM2 that the jet shifts were preceded by enhanced warm pool tropical convection and then by poleward wave activity propagation (identified with EP fluxes). This wave activity...
FIG. 12. Lagged composites of the planetary-scale (zonal wavenumbers 1 and 2) (left) SOM1 and (right) SOM3 anomalous 300-hPa streamfunction (shading) and planetary-scale 300-hPa climatological streamfunction (contours) for the averaging time period indicated.
Fig. 13. Lagged composites of the (left) SOM1 and (right) SOM3 anomalous sea ice concentration for the time interval indicated. Note that the shading level is reversed with that for the previous figures.
propagation with its attendant equatorward eddy momentum flux weakens the subtropical jet and is followed by the excitement of a synoptic-scale eddy momentum flux that drives both jets poleward. The excitement of SOM4 was found to exhibit the same features but opposite in sign.

For the second class, the driving mechanism involves sea ice concentration anomalies over the Greenland, Barents, and Kara Seas. For SOM1, which is preceded by an increase in sea ice over the Kara Sea, an anomalous low develops in the upper troposphere over the same

FIG. 14. A schematic depiction of the mechanism proposed by this study for the linkage between Arctic sea ice anomalies and changes in the strength of the stratospheric polar vortex for (a) SOM1 and (b) SOM3. For SOM1, the phase of the anomaly coincides with destructive interference, weaker vertical wave activity propagation, and an acceleration of the stratospheric polar vortex. SOM3 shows the opposite features.
region. (This relationship is consistent with the findings based on climate model experiments when sea ice is added.) This is followed by destructive interference between the anomalous wave field and the climatological stationary eddies. These changes result in a weakening of the vertical flux of wave activity into the stratosphere and thus an acceleration of the stratospheric polar vortex. Each of these steps is illustrated in Fig. 14a. The strengthened stratospheric polar vortex is then followed by the excitation of SOM1, presumably through a positive synoptic-scale eddy feedback process (Garfinkel et al. 2013).

SOM3 was found to exhibit the same features but opposite in sign. The excitation of this SOM pattern was preceded by a loss of sea ice over the Greenland and Barents Seas. This was followed by the formation of an anomalous high over these two seas (again consistent with climate model experiments), constructive interference, a strengthening of the vertical flux of wave activity into the stratosphere, and a weakening of the stratospheric polar vortex (see Fig. 14b). This is followed by the excitation of SOM3, again most likely through a synoptic-scale feedback process.

From the findings above, what can we learn about the interdecadal jet shifts? For the first class, to the extent that GHG driving enhances warm pool convection (Lee et al. 2011; L’Heureux et al. 2013), one expects a trend toward an increased frequency of convection on intra-seasonal time scales that has a La Niña–like spatial structure. This would lead to more frequent excitation of planetary-scale poleward-propagating wave trains and the subsequent poleward jet shifts represented by SOM2 and SOM4. This study cannot be conclusive about this linkage (between the frequency trends of SOM2 and SOM4 and the global-mean SAT, which is treated as a proxy for GHG driving) perhaps because the data period is too short; at time scales longer than that of ENSO, the correlations were statistically significant only at the $p < 0.15$ level. In the future, as the dataset becomes longer, the relationship between this class of patterns and the global-mean SAT may exhibit a higher degree of statistical significance, which would be consistent with results from climate model runs with increased CO$_2$. On the other hand, it is also possible that these interdecadal jet shifts can be explained by climate noise (Feldstein 2002). For the second class, its interdecadal trend was associated with the decline in Arctic sea ice. Since the poleward (equatorward) shift of the eddy-driven jet corresponds to the positive (negative) phase of the NAO/AO, these findings suggest that the positive trend in the Arctic Oscillation (AO) from the late 1960s through the early 1990s might possibly be attributed to the influence of GHG warming on tropical convection, while the subsequent reversal in its trend since that time period is likely due to the loss of Arctic sea ice, which may also be influenced by the GHG warming.

Returning to the question of the mechanism that drives latitudinal jet shifts, there is a number of other mechanisms that have been proposed for poleward jet shifts. These include an increase in latent heat release by midlatitude storms (Son and Lee 2005); a higher tropopause (Lorenz and DeWeaver 2007); an enhancement in subtropical and midlatitude static stability and thus a poleward shift of both the subtropical jet and latitude of strongest baroclinic instability (Lu et al. 2008); higher-latitude wave breaking caused by an increased midlatitude eddy phase speed, poleward-displaced critical latitudes, and an increased eddy length scale (Chen et al. 2008; Lu et al. 2008; Kidston et al. 2010, 2011); an increase in the frequency of anticyclonic wave breaking of longer waves associated with greater upper-tropospheric baroclinicity (Riviere 2011); and a poleward shift in the midlatitude storm track resulting from both tropical upper tropospheric warming and a cooling of the polar stratosphere (Butler et al. 2010). The finding in this study, that the poleward trend in the NH jets can be expressed through changes in the frequency of occurrence of teleconnection patterns with an $e$-folding time scale of 4.8–5.7 days, poses the constraint that these possible mechanisms also occur on a similarly short time scale.

Finally, it was found that the frequencies of two SOM patterns were preceded by statistically significant ($p < 0.05$) anomalies in Arctic sea ice area that extend back as far as 12 months. These findings suggest that Arctic sea ice may serve as an important source of predictability for the NH with lead times of many months.

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