On the Possible Link between Tropical Convection and the Northern Hemisphere Arctic Surface Air Temperature Change between 1958 and 2001

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ABSTRACT

This study presents mechanisms for the polar amplification of surface air temperature that occurred in the Northern Hemisphere (NH) between the periods of 1958–77 (P1) and 1982–2001 (P2). Using European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40) reanalysis data, it is found that over the ice-covered Arctic Ocean, the winter surface warming arises from dynamic warming (stationary eddy heat flux and adiabatic warming). Over the ice-free Arctic Ocean between the Greenland and the Barents Seas, downward infrared radiative (IR) flux is found to dominate the warming.

To investigate whether the difference in the flow between P1 and P2 is due to changes in the frequency of occurrence of a small number of teleconnection patterns, a coupled self-organizing map (SOM) analysis of the 250-hPa streamfunction and tropical convective precipitation is performed. The latter field was specified to lead the former by 5 days. The results of the analysis showed that the P2 − P1 trend arises from a decrease in the frequency of negative phase PNA-like and circumglobal streamfunction patterns and a corresponding increase in the frequency of positive PNA-like and circumglobal streamfunction patterns. The occurrence of the two strong 1982–83 and 1997–98 El Niño events also contributes toward this trend. The corresponding trend in the convective precipitation is from below average to above average values in the tropical Indo-western Pacific region. Each of the above patterns was found to have an e-folding time scale from 6 to 8 days, which implies that the P2 − P1 trend can be understood as arising from the change in the frequency of occurrence of teleconnection patterns that fluctuate on intraseasonal time scales.

The link between intraseasonal and interannual variability was further examined by linearly regressing various quantities against trend patterns with interannual variability subtracted. It was found that enhanced convective precipitation is followed 3–6 days later by the occurrence of the P2 − P1 circulation trend pattern, and then 1–2 days later by the corresponding trend pattern in the downward IR flux. This finding suggests that an increased frequency of the above sequence of events, which occurs on intraseasonal time scales, can account for the NH winter polar amplification of the surface air temperature via increased dynamic warming and downward IR flux.

1. Introduction

Geological evidence shows that a warmer climate is characterized by so-called polar amplification (e.g., Manabe and Wetherald 1975a,b; Budyko and Izrael...
1991; Rigor et al. 2000; Johannessen et al. 2004), which refers to the surface warming being more pronounced in polar regions than in lower latitudes. In fact, an explanation for possible mechanisms that can account for polar amplification remains as one of the long-standing questions in the paleoclimate research community (Huber 2008; Spicer et al. 2008). Since the 1970s, the most pronounced increase in Northern Hemisphere (NH) surface air temperature occurred over the high-latitude continents during the cold season (Serreze et al. 2000). Recently, it was shown that the Southern Hemisphere has also been experiencing winter season polar amplification (Steig et al. 2009).

The most prominent explanation for this phenomenon is the surface albedo feedback (SAF) mechanism (Budyko 1969; Sellers 1969). Evidence for this theory is mixed. With a coupled global climate model of the atmosphere, ocean, and land surface, Hall (2004) found that the albedo feedback plays the major role. He compared the responses to CO₂ doubling in two different models: one with the climatological albedo and the other with a parameterized variable albedo that is a function of the surface temperature and the thickness of the sea ice, land ice, and land snow. The results showed that the polar warming was substantially greater in the variable albedo model than in the fixed albedo model. In the zonal mean, the ratio of the surface air temperature increase in the variable and fixed albedo runs was 1.5–2.0 in the polar regions. However, this conclusion is not shared by other GCM studies (Winton 2006; Graversen and Wang 2009). With a zero-dimensional energy balance model, Winton (2006) quantified the role of the SAF for the temperature sensitivity to a CO₂ increase in a group of 12 Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) climate models. He found that, while the SAF contributes to the amplification of the model’s Arctic surface warming, it is not the dominating factor. Instead, longwave feedback is found to be far more influential for the model’s polar amplification.

An alternative viable candidate for the cold season polar amplification is an increase in poleward heat flux. Indeed, Graversen (2006) has shown that the column-integrated moist static energy transport across 60°N has increased during the period of 1979–2001. Consistently, Lu and Cai (2008) found with a GCM that an increase in the poleward sensible heat flux plays the major role in generating the model’s polar amplification. However, the mechanism for this increase in the sensible heat flux and surface warming is still unknown. Another feasible candidate is the winter convective cloud feedback mechanism proposed by Abbot and Tziperman (2008a,b). In this mechanism, sea ice reduction led by the CO₂-initiated warming destabilizes the atmosphere, which results in a moistening of the atmosphere and a generation of convective clouds. Because these two processes reduce outgoing longwave radiation, the surface further warms. Abbot et al. (2009) analyzed IPCC AR4 models and found that the winter Arctic cloud feedback is active in 4 × CO₂ runs. In their GCM study of equable climates of the Cretaceous and early Cenozoic, Lee et al. (2011) found that both the dynamic poleward heat flux and the convective cloud feedback contribute, as a duet, toward the GCM’s Arctic amplification.

The generally accepted picture for the climatology is that extratropical baroclinic eddies are the main agent for the poleward heat flux (see Peixoto and Oort 1992). While we concur with this picture, in light of the fact that the strength of the eddy heat flux depends on a suitably averaged baroclinicity, which in turn is proportional to the meridional temperature gradient, it is difficult to see how a stronger poleward transient eddy heat flux can be maintained in the face of a weaker meridional temperature gradient. Instead, in this study, we present observational evidence that horizontal temperature advection and adiabatic warming (downward motion) associated with stationary Rossby waves can play an important role in the cold season polar amplification. [Lee et al. (2011) showed in a GCM study of equable climates that the horizontal momentum flux associated with stationary Rossby waves can drive adiabatic warming in the Arctic.] We refer to the sum of the horizontal advection and the adiabatic warming as dynamic warming.

There are also theoretical reasons why the poleward heat transport may be triggered by tropical convection. The rationale is that: 1) if the tropical sea surface temperature (SST) were to increase uniformly everywhere, since saturation vapor pressure has an exponential dependency on temperature (the Clausius–Clapeyron relationship), tropical convection over the Pacific warm pool region will intensify more rapidly than elsewhere; 2) this localization of convection will generate poleward-propagating Rossby waves (Hoskins and Karoly 1981); and 3) that, unlike midlatitude synoptic-scale baroclinic waves, the strength of the tropically-forced Rossby wave heat flux is expected to be at best only weakly dependent upon the equator-to-pole temperature gradient. Regarding 1), there is a debate as to whether the SST change associated with increased greenhouse gas forcing is more El Niño–like or La Niña–like. Vecchi and Soden (2007) found that in the A1B scenario runs of the IPCC AR4, the SST becomes more “El Niño–like.” On the other hand, with historical SSTs dating back to 1900, Cane et al. (1997) found that the eastern equatorial Pacific cooled and, as a result, the zonal sea surface temperature gradient had increased. Kumar et al. (2010) showed that
between 1950 and 2008 the greatest tropical SST increase occurred in the tropical Indian and western Pacific Oceans. These findings further strengthen the first rationale.

The main goal of this study is to investigate whether there is observational evidence that the late twentieth century, cold-season, polar surface warming is caused by dynamic warming associated with large-scale circulation changes and whether these changes, if they occur, are driven by a localization of tropical convective heating. For this purpose, we analyze the global 40-yr European Centre for Medium-Range Weather Forecasts Re-Analysis (ERA-40), focusing on the contrast between the periods of 1958–77 (P1) and 1982–2001 (P2). The data from 1978 to 1981 is not included in either of these periods because of the marked shift in the North Pacific circulation in the late 1970s, as shown in Trenberth (1990) and subsequent studies. However, at least for surface air temperature, upper-tropospheric circulation, and tropical convective precipitation, the P2–P1 fields are found to be insensitive to small changes in the two periods. Specifically, two additional sets of calculations with slightly different time periods (P1 = 1958–79, P2 = 1980–2001, P1 = 1958–74, P2 = 1985–2001) yielded results that were nearly identical to those for the P1 and P2 periods chosen for this study (not shown).

There are quality issues with the ERA-40 tropical precipitation (Uppala et al. 2005) and surface temperature (Simmons et al. 2004; Karl et al. 2006). To address this issue for the tropical precipitation, we illustrate in Fig. 1 the 1979–20021 December–February (DJF) trend in the ERA-40, Global Precipitation Climatology Project (GPCP), and Climate Prediction Center Merged Analysis of Precipitation (CMAP) datasets. Although considerable variation can be seen in the sign and magnitude of the precipitation trend among the datasets, it is reassuring that within the Indo-western Pacific warm pool region all three datasets show a positive trend. This suggests that an interdecadal trend in warm pool convective precipitation is likely a robust feature of the atmosphere over the past few decades.

Reanalysis products prior to 1979, being of the pre-satellite era, have additional limitations. However, as we will see, the spatial pattern of the P2–P1 interdecadal trend in tropical convective precipitation obtained from the ERA-40 dataset is consistent with the 1979–2002 trends shown in Fig. 1. Moreover, the results to be presented are consistent with theoretical expectations. It is

Fig. 1. 1979–2002 linear precipitation trend (mm day$^{-1}$ century$^{-1}$) from (a) ERA-40, (b) GPCP, and (c) CMAP.

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1 This trend calculations begin with 1979 because the latter two datasets are first available for that year.
thus reasonable to take the findings to be described below as a starting point for future investigations.

In this study, we also address the question of how tropically forced Rossby waves, which propagate from the tropics to high latitudes within 10 days (Hoskins and Karoly 1981), are related to the high-latitude warming that occurs over interdecadal time scales (Serreze et al. 2000). Following the lead of Feldstein (2000) and Johnson and Feldstein (2010), we will show evidence that the interdecadal time-scale change is a result of a change in the frequency of occurrence of intraseasonal time scale teleconnection patterns.

2. Changes in the surface temperature and the attributes

Throughout this study, for a given time-mean field, $\overline{F}$, the difference between the two time periods, 1982–2001 and 1958–1977, will be denoted as $\delta F$. For this calculation, we used daily ERA-40 data archived at $2.5^\circ \times 2.5^\circ$ horizontal resolution. All data to be presented in this section are averaged over DJF.

The $P2 - P1$ difference in the surface (2 m AGL) air temperature, $\delta T_s$ (Fig. 2a), shows an overall warming in the Arctic except over the Greenland–Labrador Sea, a small region at the eastern tip of Siberia, and the Beaufort and Chukchi Seas. Furthermore, an additional warming can be seen over the North American and Eurasian continents and a cooling over the North Pacific and North Atlantic Oceans. This pattern is reminiscent of the cold ocean–warm land (COWL) teleconnection pattern (Wallace et al. 1996), a spatial pattern that refers to the change in the NH surface air temperature during the late-twentieth-century NH winter.

The $P2 - P1$ 250-hPa streamfunction field, $\delta \psi_{250}$ (Fig. 2b), displays mostly positive values in midlatitudes and negative values at higher latitudes, indicating a poleward shift of the upper-tropospheric jet. Other key features in Fig. 2b include a strong zonal mean and zonal wavenumber 5 contribution to $\delta \psi_{250}$. The 250-hPa eddy streamfunction field, $\delta \psi_{250}$, is shown in Fig. 2c, where the asterisk denotes a deviation from the zonal mean. Over the North Pacific and northwestern Canada a wave train can be seen, which resembles the positive phase of the Pacific–North American (PNA) teleconnection pattern (Wallace and Gutzler 1981; Barnston and Livezey 1987), and, over the North Atlantic sector, $\delta \psi_{250}$ resembles the positive phase of the North Atlantic Oscillation (NAO) (Barnston and Livezey 1987). The change in the convective precipitation, $\delta P$ (Fig. 2d), indicates an increase over the Indo-western Pacific warm pool region, as discussed in the introduction.

2a. Dynamic warming

We next evaluate whether the proposed mechanism, that is, the dynamical warming, can explain the observed near-surface polar amplification. The thermodynamic energy equation in pressure coordinates can be written as

$$0 = \frac{\partial \overline{T}}{\partial t} = -\nabla_h \cdot \overline{\nabla h T} + \overline{\nabla_p \omega} - \gamma T + \mathcal{Q},$$

(1)

where the overbar denotes the time mean for each period (P1 or P2). Other notations are standard: $T$ is
temperature, \( \mathbf{v}_h \), the horizontal velocity, and \( \omega = dp/dt \) in which \( p \) is pressure. The first term on the right-hand side (rhs) is the horizontal temperature advection; the second term accounts for adiabatic warming (downward motion) or cooling (upward motion);

\[
S_p = -\frac{T}{\theta} \frac{\partial \theta}{\partial p}
\]

is a static stability parameter for which \( \theta \) is potential temperature. The third term, \( \gamma T \), approximates clear-sky radiative cooling, where \( \gamma^{-1} \) is the radiative relaxation time scale. Finally, \( \overline{Q} \) represents the remaining diabatic heating contribution including latent heating, infrared radiation (IR) warming by clouds, and surface heat fluxes. Rewriting (1) in terms of \( \delta T \), we have

\[
\delta T = \gamma^{-1} \left[ -\delta(\mathbf{v}_h \cdot \nabla_h T) + \delta(\omega S_p) + \delta \overline{Q} \right]. \quad (2)
\]

This equation states that \( \delta T > 0 \) if the sum of the rhs of (2) is positive in that region. In this section, we investigate to what extent the dynamic warming, \( -\delta(\mathbf{v}_h \cdot \nabla_h T) + \delta(\omega S_p) \), can account for \( \delta T \).

Because the ERA-40 reanalysis data are available on pressure levels (1000, 925, 850, 775, 700 hPa, etc.), while surface pressure at different grid points ranges over these pressure levels, to evaluate the surface dynamic warming represented by the first two terms on the rhs of (2), we make the following assumption: for the \( i \)th grid point, which satisfies \( P_{n-1} \leq P_i < P_n \), where \( P_n \) is the pressure at the \( n \)th pressure level and \( P_i \) is the surface pressure, \( P_{n-1} \) is treated as the surface pressure. For example, for grid points where the surface pressure is less than 1000 hPa but greater than or equal to 925 hPa, data at 925 hPa are regarded as surface data. Accordingly, we adopt the following analysis procedure: for those grid points where \( P_i \geq 1000 \) hPa, assign the estimated surface temperature \( T_s = T_{925} \) while excluding those values at grid points where the surface pressure is less than 1000 hPa; then for the empty grid points, assign \( T_s = T_{925} \) provided that the surface pressure at those points is greater than 925 hPa; repeat the same procedure for the next three levels: 850, 775, and 700 hPa. The P2 – P1 surface temperature as estimated with this “patching” procedure is shown in Fig. 3a. The same approach is adopted to estimate the horizontal temperature advection and adiabatic warming terms.

This patched \( \delta T_s \) (Fig. 3a) closely resembles \( \delta T_i \) (Fig. 2a). Pattern correlations between these two fields are 0.79 and 0.85 over the domain 0°–90°N and 60°–90°N, respectively. (If area means are not subtracted, the pattern correlations are 0.82 and 0.88, respectively.) All longitudes are included in the pattern correlations. This resemblance indicates that the patching procedure is suitable for our purpose. Figure 3b shows the patched \( -\delta(\mathbf{v}_h \cdot \nabla_h T) \) with small-scale features filtered out. Specifically, for all of the dynamic warming fields (Figs. 3b–e), we applied a spatial smoothing procedure that filters scales greater than zonal wavenumber 5 and meridional wavenumber 10. By comparing Figs. 3a to 3b, it can be seen that horizontal temperature advection can explain a substantial portion of the warming over the Arctic as well as over Eurasia and North America. To compare the stationary and transient contributions to the horizontal advection term, we calculate the patched stationary component, \( -\delta(\mathbf{v}_h \cdot \nabla_h T) \) (Fig. 3c). This similarity between Figs. 3b and 3c indicates that the transient eddies play a minor role, although their impact on the stationary eddies may be substantial.

Over the Arctic Ocean between 150°E and 90°W, the horizontal temperature advection (Fig. 3b) is slightly negative. Instead, adiabatic warming, that is, the patched \( \delta(\omega S_p) \), dominates in this region (Fig. 3d). For the remainder of the Arctic Ocean, with the exception of the Greenland and the Barents Seas, the atmosphere is warmed by both horizontal temperature advection and adiabatic warming. As such, the sum of the temperature advection and adiabatic terms yields a pattern that is overall similar to \( \delta T \) (cf. Figs. 3a to 3e). In contrast, over the Greenland and Barents Seas, where the ocean is ice free, there is a distinct net dynamic cooling. Since \( \delta T_s \) is positive in these regions (Figs. 2a and 3a), the observed warming must be due to the diabatic heating term \( \delta \overline{Q} \). This term is examined in section 2b. Overall, the dynamical processes warm (cool) the surface air over the ice-covered (ice free) regions of the Arctic Ocean: the pattern correlation between Figs. 3a and 3e is 0.31 over the domain 60°–90°N, 90°E–20°W (mostly ice covered), while it is −0.45 over 60°–90°N, 20°E–90°W (mostly ice free). If the area means are not subtracted, the correlations are 0.26 and −0.39, respectively. These correlations are rather moderate. However, given the inaccuracies introduced through the patching method and the assumption of linear Newtonian cooling [Eq. (2)], we believe that these correlations indeed support the above interpretations. Moreover, it is important to keep in mind that the focus of this study is to address the question of whether an interdecadal trend in dynamic warming contributes to an increase in temperature over most of the Arctic, not to whether this process accounts for the particular spatial pattern of the warming trend.

### b. Downward IR radiation and surface heat fluxes

The role of \( \delta \overline{Q} \) in warming the ice-free Arctic Ocean can be seen in Fig. 3f, which reveals that the surface warming is caused by an increase in downward infrared radiation. The cloud fraction change between P1 and P2...
shows that over the Greenland Sea the IR warming can be explained in part by an increase in both midlevel \((0.45 < p/p_s \leq 0.8)\) and low-level \((0.8 < p/p_s \leq 1.0)\) cloud cover (see Figs. 3h and 3i, respectively). Over the Barents Sea, neither cloud cover change matches the downward IR increase, suggesting that other factors such as increased specific humidity may contribute to the warming over this region. It is interesting that the low-level cloud cover change also agrees with the IR warming over the Eurasian continent and North America and the IR cooling over the North Pacific. These changes project onto the COWL pattern.

Although cloud cover is known to be notoriously difficult to simulate, Liu et al. (2007) showed that the main pattern of the 1982–2000 Arctic winter cloud cover trend is similar between the ERA-40 data and the extended Advanced Very High Resolution Radiometer (AVHRR) Polar Pathfinder product. Overall, for the ice-covered Arctic Ocean between 150°E and 60°W where adiabatic warming dominates the horizontal temperature advection, the change in IR acts to cool the atmosphere. In contrast, over the ice-free Arctic Ocean where cold advection prevails over the adiabatic warming, the IR change warms the atmosphere.
Above the ice-free open ocean, between the Greenland and Barents Seas (80°N, 20°W–60°E), the change in the surface heat flux warms the atmosphere [Fig. 3g, red (blue) denotes an upward (downward) heat flux from (toward) the surface]. Together with the IR warming, these two processes oppose the cooling caused by horizontal temperature advection (Fig. 3b). At most other locations over the Arctic Ocean, the surface heat flux change cools the atmosphere.

To summarize the budget analysis, except over the Greenland and Barents Seas, dynamic warming plays the major role in explaining the observed temperature change over the Arctic. This dynamic warming is partially compensated by cooling via both the surface heat flux and the downward IR flux. This reduction in the downward IR flux is consistent with the decreased lower- and midlevel cloud cover. Because this reduction in cloud cover is likely to be a result of anomalous downward motion (cf. Figs. 3d,f), the adiabatic warming and the cooling through the reduced downward IR flux are unlikely to be mutually independent. This will be revisited in section 5. Over the Greenland and Barents Seas, where the Arctic Ocean is free of sea ice, the anomalous downward IR warming dominates the dynamic cooling. Because the surface heat flux change over most of the Arctic Ocean is found to be downward (i.e., a net heat loss by the atmosphere), the ERA-40 data does not support the idea that delayed summer ice melting/ocean heating accounts for the winter warming of the surface air over the Arctic Ocean (Serreze and Francis 2006).

Given that the Pacific decadal oscillation2 (PDO) (Mantua et al. 1997; Zhang et al. 1997) has undergone an upward swing from the late 1950s to the mid-1990s, one may wonder whether the dynamic warming described above is driven by the PDO. To gauge the impact of the PDO, for each of the variables shown in Fig. 2, we calculated its PDO component by linearly regressing these variables against the normalized PDO amplitude time series. The regressed pattern was then subtracted from the corresponding pattern shown in Fig. 2 using the methodology3 of Feldstein (2003). Since the PDO regressed patterns are similar to but of smaller amplitude than the full P2–P1 patterns, the resulting PDO-removed P2–P1 patterns (not shown) and full P2–P1 patterns strongly resemble each other. This finding indicates that while the PDO does contribute to the P2–P1 change, the PDO does not play a crucial role. In fact, the similarity between the P2–P1 change patterns (Fig. 2) and the corresponding PDO regressed patterns (not shown) suggests that the mechanism that drives the P2 – P1 change may also be involved in driving the PDO. This is consistent with the finding by Schneider and Cornuelle (2005) that the observed PDO can be reasonably well simulated by a first-order autoregressive model forced by sea level pressure.

3. The relationship between interdecadal variability and intraseasonal teleconnection patterns

According to our hypothesis, the increase in the surface air temperature over the Arctic is caused by an atmospheric circulation that is driven by tropical convection. Here, we will focus on the upper-troposphere streamfunction, that is, $\psi_{250}$, because the strongest Rossby wave response to tropical convection takes place in the upper troposphere. To examine the mechanisms that drive the $\delta\psi_{250}$ pattern (Fig. 2b), we next address the question of whether $\delta\psi_{250}$ corresponds to a change in the frequency of occurrence of a small number of teleconnection patterns whose amplitude fluctuates on intraseasonal time scales. This question is motivated by the findings of Johnson et al. (2008) and Johnson and Feldstein (2010) that much of the NH interdecadal variability arises from changes in the frequency of the intraseasonal teleconnection patterns.

To address this question, we use the approach of Johnson et al. (2008) and Johnson and Feldstein (2010) who used the method of self-organizing maps (SOMs) (Kohonen 2001; Hewitson and Crane 2002) to investigate teleconnection patterns from a continuum perspective. We refer the reader to the appendix in Johnson et al. (2008) for a detailed discussion of SOM analysis and its application to the continuum of atmospheric teleconnection patterns, but provide a brief description here. SOM analysis is a type of cluster analysis that partitions data into a specified number of representative patterns that lie on a one- or two-dimensional grid. The patterns are organized such that similar patterns lie close together and dissimilar patterns lie farther apart within this one-or two-dimensional grid. The SOM methodology has the advantage over the more commonly used EOF analysis in that the SOM patterns correspond approximately to a minimization in the Euclidean distance (in a $N$-dimensional phase space where $N$ equals the number of grid points) between the observed patterns and the SOM patterns. As a result, SOM analysis often

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2 The PDO is defined as the leading empirical orthogonal function (EOF) of monthly SST anomalies in the North Pacific Ocean poleward of 20°N (Mantua et al. 1997).

3 The methodology is briefly described for the 250-hPa streamfunction. We first divide $\psi_{250}$ into the PDO component and a residual where the PDO component is calculated by linearly regressing $\psi_{250}$ against the SST-based PDO index and then projecting daily $\psi_{250}$ onto this $\hat{\psi}_{250}$ PDO pattern. In doing so, the PDO and the residual components are constructed to be orthogonal to each other. The same approach is used for the other variables in Fig. 2. For more detail, see Eqs. (2)–(5) of Feldstein (2003).
outperforms EOF analysis in the extraction of patterns from datasets (Reusch et al. 2005; Liu et al. 2006).

For this study, in analogy with Johnson and Feldstein (2010), we perform a coupled SOM analysis with two fields: NH 250-hPa streamfunction and tropical convective precipitation. The data comprise 5-day mean (pentad) anomalies, where the convective precipitation leads the 250-hPa streamfunction by one pentad. The anomalies are defined as deviations from the seasonal cycle. Because the streamfunction and convective precipitation fields are on a 2.5° latitude–longitude grid, we weight the data by the square root of cosine of latitude in the SOM analysis to account for the dependence of grid point density on latitude. This weighting was applied because the Euclidean distance in the SOM analysis incorporates the squares of distance. To ensure equal weighting to the streamfunction and convective precipitation anomalies in the SOM analysis, both of these quantities are standardized. We have chosen to use a one-dimensional 20 × 1 grid (Fig. 4), as in Johnson and Feldstein (2010), so that the temporal evolution of the pattern frequency distribution can be clearly illustrated (see Fig. 5). In Fig. 4, each coupled SOM pattern is illustrated with two panels, the top panel corresponding to the streamfunction and the bottom panel to the convective precipitation field.

Each (weighted) coupled streamfunction–convective precipitation pentad field is assigned to the best-matching SOM pattern on the basis of minimum Euclidean distance. Based on these assignments, we calculate the frequency of occurrence of each coupled SOM pattern for the 45-yr period (Fig. 5). As will be discussed later, this calculation shows the temporal evolution of the SOM pattern frequency of occurrence. The mean pattern correlation between the pentad 250-hPa streamfunction (convective precipitation) field and the best-matching SOM pattern is 0.48 (0.29); both values are statistically significant at the 95% confidence level based on a one-sided t-test. Thus, we conclude that 20 SOM patterns are sufficient to capture the large-scale streamfunction and convective precipitation features in the pentad fields and that the patterns in Fig. 4 are real physical patterns, not artifacts of the statistical analysis. Although the number is somewhat arbitrary, we have chosen 20 patterns because the number is large enough to ensure that each pattern bears a reasonably strong resemblance to corresponding pentad fields, but small enough to provide insight into the 250-hPa streamfunction and convective precipitation changes between P1 and P2, as discussed below.

As can be seen from Fig. 4, although the 20 SOM patterns exhibit much variability, most of the patterns resemble at least one or two other patterns. Thus, the 20 SOM patterns can be understood as corresponding to a finite representation of the full continuum. Moreover, the 20 SOM patterns resemble a number of the well-known NH winter teleconnections. For example, streamfunction SOM patterns 6–8 are similar to the positive PNA phase and patterns 12–15 to the negative PNA phase. We collectively refer to these as PNA-like patterns. The positive phase PNA-like patterns are preceded by positive convective precipitation anomalies over the central tropical Pacific and warm pool regions. The negative phase PNA-like patterns exhibit opposite features. Among these PNA-like patterns, pattern 8 also shows a strong projection onto the positive NAO phase and patterns 12–14 onto the negative NAO phase. Pattern 1 exhibits strong El Niño features as evidenced by the dipole streamfunction anomaly over the North Pacific, the strengthened and zonally extended subtropical jet, and the largest increase in convective precipitation taking place over the central Pacific Ocean.

There are also global patterns in Fig. 4 that include strong zonal mean and zonal wavenumber 5 contributions, which are dominated by one particular sign and consist of mostly positive anomalies (patterns 17–19) and negative anomalies (patterns 4 and 5). Some of these patterns also exhibit strong projections onto the NAO. We refer to these patterns as the positive and negative phases of the circumboreal streamfunction pattern, respectively. The circumboreal streamfunction anomalies are preceded by zonally oriented dipole anomalies in the convective precipitation field, with positive (negative) anomalies over the Indian Ocean (western tropical Pacific Ocean) for the positive phase, and vice versa for the negative phase.

The e-folding time scales of the SOM patterns were also calculated with time series generated by projecting the daily 250-hPa streamfunction field onto each of the SOM patterns. Similar to Feldstein (2000) and Johnson and Feldstein (2010), the amplitude time series for all of the SOM patterns have e-folding time scales that vary between 6 and 8 days, indicating that each of SOM patterns fluctuate on intraseasonal time scales.

The interannual variability of the SOM patterns is illustrated in Fig. 5a, where the 5-yr moving average of the frequency of occurrence for each pattern is shown. As can be seen, there is a transition from patterns 4–5 and 11–14 dominating during P1 to patterns 1, 7–8, and 17–19 occurring most frequently in P2. This suggests that the interdecadal trend in the Ψ250 field can be understood

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4 For this calculation, we estimate the number of effective spatial degrees of freedom to be 18 and 86 for the streamfunction and convective precipitation fields, respectively, based on the eigenvalue formula estimate (Bretherton et al. 1999).
FIG. 4. The joint 250-hPa and convective precipitation SOM patterns. The pattern number is indicated for each set. The percentage below each set indicates the frequency of occurrence of the pattern for the period from 1958 to 2002. The contour interval is $2 \times 10^6$ m$^2$ s$^{-1}$ for the streamfunction and 1 mm day$^{-1}$ for the precipitation.
as arising from a shift in the frequency of occurrence of the above patterns, with P1 being dominated by negative PNA-like and negative circumglobal streamfunction patterns, and P2 by positive PNA-like and positive circumglobal streamfunction patterns together with the strong 1982–83 and 1997–98 El Niño events. For the convection precipitation, this transition corresponds to a trend from below average to above average convective precipitation in the tropical Indo-western Pacific region. To evaluate the sensitivity of these findings to the number of SOM patterns, we also show the frequency of occurrence as a function of the year for a choice of 35 patterns (Fig. 5b). As can be seen, there is a noticeable decline in the frequency of occurrence of patterns 6–9 and 19–24, and marked increase for patterns 1, 2, 13–17, and 28–33. The pattern of these changes clearly resembles that shown in Fig. 5a. These SOM patterns are found to correspond to positive and negative PNA-like and circumglobal streamfunction patterns along with the El Niño pattern (not shown). Therefore, these results do indicate that our results are not sensitive to the size of the SOM array.

To quantitatively evaluate the extent to which the SOM analysis captures the interdecadal variability of the streamfunction and convective precipitation fields, the following calculation is performed. For P1 and P2, we estimate the time-mean streamfunction and convective precipitation anomalies as

\[ \tau_i = \sum_{c=1}^{20} m_c(x,y) f_c(m_c), \quad i = 1, 2, \]

where \( \tau_i \) corresponds to the P1 \((i = 1)\) or P2 \((i = 2)\) time-mean anomaly, \( m_c(x,y) \) is SOM pattern \( c \), and \( f_c(m_c) \) is the frequency of occurrence of \( m_c(x,y) \). Figure 6 shows the SOM-derived P1 and P2 time-mean streamfunction and convective precipitation fields along with the actual fields.

It is found that the pattern correlations between Figs. 6a and 6b are 0.83 for the streamfunction (over the entire NH) and 0.87 for the convective precipitation (between 30°S and 30°N). The pattern correlations between Figs. 6c and 6d are 0.84 and 0.83 for the streamfunction and convective precipitation fields, respectively. If the area means are included, the pattern correlations are somewhat higher: 0.85 (for streamfunction) and 0.89 (for precipitation) between Figs. 6a and 6b; 0.85 (for streamfunction) and 0.86 (for precipitation) between Figs. 6c and 6d. As such, the SOM-based patterns capture a substantial fraction of the interdecadal variability between P1 and P2. Therefore, from Figs. 5 and 6, we can conclude that the interdecadal changes between P1 and P2 can be understood as arising from interdecadal changes in the frequency of occurrence of the patterns in Fig. 4, which fluctuate on intraseasonal time scales.

The above results show that from P1 to P2 there is a trend in the frequency of occurrence of the SOM patterns, namely from negative PNA-like and circumglobal streamfunction patterns to positive PNA-like and circumglobal streamfunction patterns, plus a trend toward more frequent and stronger El Niño events. To examine if these trends are born out in the \( \delta \psi_{250} \) pattern, we calculate pattern correlations between each of the SOM streamfunction patterns and the \( \delta \psi_{250} \) pattern. As can be seen from Fig. 7a, the largest positive correlations correspond to SOM patterns 8 and 17–19 and the largest negative correlation for SOM patterns 4, 5, and 12–14. An analogous calculation of the pattern correlations between the SOM convective precipitation patterns and \( \delta P \) shows relatively large positive values for SOM patterns 7–9 and 16–19 and negative values for SOM patterns 4, 5, and 11–14 (see Fig. 7b). These results provide additional support for the P2–P1 trend arising from a shift in the frequency of occurrence of SOM patterns.
from negative to positive phase PNA-like and circum-global streamfunction patterns.

The pattern correlation approach in Figs. 7a,b is useful for evaluating the hemispheric relationship between the SOM patterns and the P2–P1 trend. Such a calculation should not be applicable to SOM patterns whose contribution to the P2–P1 trend is local. If we compare SOM pattern 1 with the \( d_{250} \) and \( d_P \) patterns (Figs. 2b and 2d), it does indeed appear that the El Niño contribution to the P2–P1 trend is more local than for the other SOM patterns. To examine the role of El Niño in the trend, and also to succinctly summarize the results shown in Fig. 5, we calculate the P2–P1 frequency of occurrence (Fig. 7c). As can be seen, SOM patterns 4–6 and 11–14 undergo a marked decline in their frequency, while SOM patterns 1, 7–9, and 15–19 show a large increase in their frequency, with pattern 1 showing the largest gain. This calculation confirms the visual impression from Fig. 5 that El Niño does indeed make an important contribution to the local P2 – P1 trend.

4. Intraseasonal variability of the P2 – P1 patterns

The findings of this study suggest that polar amplification during the NH cold season can be largely accounted for by dynamic warming associated with the circulation pattern in Fig. 2b. It was shown, moreover, that the circulation and convective precipitation change can be interpreted as being a manifestation of the changes in the frequency occurrence of a few SOM patterns. Because the time scale of these SOM patterns varies between 6 and 8 days, to understand the mechanistic relationship among the precipitation, circulation, and the polar amplification, we examine the intraseasonal variability of these patterns.

For this purpose, we first construct indices that can represent the intraseasonal variability of the circulation change pattern. To do so, we first remove interannual variability by subtracting the DJF average of \( d_{250} \), for each year. The resulting daily fields are then projected onto the \( d_{250} \) field (Fig. 2b) for a domain that extends over the entire NH poleward of 20°N. The resulting daily time series, normalized by its one standard deviation value, is shown in Fig. 8a and is referred to as the circulation change (CC) index. For the purpose of comparison, Fig. 8a also displays the projection time series with the DJF averages retained for each year. With the interannual variability intact, an upward trend is present in this latter time series, but embedded in this interdecadal time-scale trend are intraseasonal time-scale fluctuations. As is expected from this construction, this
same intraseasonal variability is evident in the detrended CC index.

In a similar manner, we also construct an index that measures the amplitude of the convective precipitation change pattern. This index, which is referred to as the convective precipitation change (CPC) index, was constructed by projecting the daily convection precipitation (with each winter mean subtracted) onto the CPC pattern (Fig. 2d). In this case, the range of the projection domain is $20^\circ$S–$20^\circ$N and it spans all longitudes. The resulting index (Fig. 8b), as for the CC index, shows that there are ample intraseasonal fluctuations. Once again, an upward trend is evident in the corresponding projection time series when the interannual variability is retained (see the thin line in Fig. 8b). To quantify the time scale of these intraseasonal fluctuations, the autocorrelations of both CC and CPC indices are calculated. It is found that the $e$-folding time scales for the
CC and CPC time series are 5 and 3 days, respectively. Clearly these fall into intraseasonal time scales.

With the CC index, we next use time-lagged linear regression analysis to explore temporal relationships between the CC pattern (δϕ250) and each of the following quantities: convective precipitation, ϕ250, downward IR, and T_s. For this analysis, in the same manner as described above, interannual variability is first removed from these four variables, and then the resulting daily fields are linearly regressed against the CC index. The time-lagged linear regression coefficients are first calculated for each DJF, and then averaged for 1958–2001. The resulting regression fields are shown in Fig. 9 for selected time lags. Positive (negative) lags indicate that the pattern in question lags (leads) the CC pattern. At lag −10 days, convective precipitation over the Indian Ocean is positive, projecting onto the CPC pattern, but there is still no hint of a P2 − P1 pattern in the other fields. At lag 0 days, the P2 − P1 patterns are evident in all four fields, except for the convective precipitation over the western Pacific and the surface temperature over the ice-covered Arctic Ocean. These patterns then steadily weaken except for the ice-free part of the Arctic Ocean, which undergoes a warming, attaining its maximum temperature at lags +13 days.

The above temporal evolution can be concisely summarized by spatially correlating the regressed fields onto the corresponding P2 − P1 patterns. Figure 10a shows the resulting projection time series where the domain of the projection is 20°–20°N for convective precipitation and 20°–90°N for all other fields. For the convective precipitation, we also show projection values obtained separately for the Indian Ocean and the western Pacific Ocean. For these two oceans, the domain is 20°S–10°N, 45°–100°E and 20°S–20°N, 120°E–180°, respectively. The values that exceed the 90% confidence level5 are indicated by dots. It can be seen from this figure that the growth and decay of the P2 − P1

\[ \text{FIG. 9. (from left to right) Time-lagged patterns of convective precipitation, } \phi_{250}, \text{ downward IR, and } T_s, \text{ regressed against the CC index. The contour interval is } 1 \times 10^{-2} \text{ mm day}^{-1} \text{ for the precipitation, } 5 \times 10^5 \text{ m}^2 \text{ s}^{-1} \text{ for } \phi_{250}, 1.0 \text{ W m}^{-2} \text{ for the downward IR, and } 0.1 \text{ K for } T_s. \text{ Shading denotes values that exceed the 90% confidence level.} \]

5 For the statistical significance test, we used the Fisher’s Z transformation (Spiegel 1975).
patterns take place over a period of about 10–15 days, the convective precipitation over the Indian Ocean leads $\psi_{250}$ by about 3–6 days, and $\psi_{250}$ leads the downward IR flux, heat flux, and $T_s$ by 1–2 days.

Given that the Indian Ocean precipitation leads $\psi_{250}$, downward IR, and $T_s$, we ask whether the same temporal relationships are present if the time lagged regression is performed against the CPC index. Figure 10b summarizes the results. The peak pattern correlation for $\psi_{250}$ occurs at lag +3 days, followed by that for the downward IR flux and $T_s$ at lag +4 days. These timings are consistent with the CC-based regression (Fig. 10a), suggesting that
anomalous tropical convection excites large-scale circulation patterns that resemble $\delta\psi_{250}$, and that this in turn leads to downward IR flux and $T_s$ anomalies, which again project onto the corresponding $P_2 - P_1$ patterns. However, for all three variables, since the pattern correlations are relatively small, although statistically significant, there must be many convective precipitation events that take on the CPC spatial structure yet do not excite the CC pattern.

Finally, we ask how the CC or CPC patterns evolve in the SOM space on intraseasonal time scales. This can be examined objectively by calculating the pattern correlations between the above regression fields and the SOMs. Figure 11a shows the pattern correlations between the $\psi_{250}$ SOMs and the $\psi_{250}$ regressed against the CC index. Consistent with the result shown in Fig. 8a, at the lag 0 day, relatively large positive pattern correlations occur for SOM patterns 1, 2, 7–11, and 16–19. However, prior to lag −10 days, patterns 1, 2, and 7–11 show relatively large negative pattern correlations. Similarly, prior to lag −25 days or so, patterns 16–19 also exhibit negative correlations. Likewise, the pattern correlations between the precipitation SOMs and the precipitation regressed against the CPC index (Fig. 11b) show that during the lag −6 to lag +6 day interval, the largest positive correlations occur for SOM patterns 7–10 and 16–19. These correlations show that the intraseasonal variation of the $P_2 - P_1$ trend patterns are dominated by the same SOM patterns as those for the interdecadal trend (Figs. 5 and 7). Therefore, these findings suggest that the $P_2 - P_1$ trend can be understood as arising from an increase in the frequency of occurrence of the intraseasonal time-scale process represented by Fig. 10.

5. Summary and conclusions

In this study, the ERA-40 dataset is used to study the causes for the polar amplification of the surface air temperature, which was observed to occur during the Northern Hemisphere cold season in the 1958–2001 time period. For this purpose, the data examined were separated into two periods: 1958–77 (P1) and 1982–2001 (P2). The surface temperature budget analysis indicates that over the ice-covered parts of the Arctic Ocean, the warming is caused by the change in horizontal temperature advection and adiabatic warming between the two time periods. The impact of the transient eddy heat flux was found to be minor. Over the ice-free regions of the Arctic Ocean, between the Greenland and Barents Seas, increased downward IR flux was found to be crucial for the warming. This downward IR warming is consistent with an increase in low- to midtropospheric cloud cover. The finding that the downward IR warming is associated with an increase in cloud cover over the open Arctic Ocean is consistent with the mechanism proposed by Abbot and Tziperman (2008a,b). The change in the surface heat fluxes generally act to damp the surface air temperature change. This suggests that at least in the ERA-40 dataset the delayed effect of the summer ice melting/ocean warming plays a minor role.

To investigate whether the difference in the flow between P1 and P2 is associated with the change in the frequency of occurrence of a small number of teleconnection patterns, we perform a coupled SOM analysis of the NH 250-hPa streamfunction and tropical convective precipitation fields. A similar methodology was used by Johnson et al. (2008) and Johnson and Feldstein (2010) to investigate teleconnection patterns from a continuum perspective. For this calculation, 20 SOM patterns were used to approximate the continuum. The convective precipitation field was specified to lead the 250-hPa streamfunction field by 5 days. It was indeed found that the $P_2 - P_1$ trend can be explained as arising from a decrease in the frequency of
negative phase PNA-like and circumglobal streamfunction patterns together with an increase in the frequency of positive phase PNA-like and circumglobal streamfunction patterns, along with the occurrence of the two strong 1982–83 and 1997–98 events El Niño events. (These circumglobal streamfunctions are dominated by anomalies of one sign and are characterized by large zonal mean and zonal wavenumber 5 contributions.) Although not previously discussed in this study, the NAO also adds to the P2 − P1 trend. This is because the NAO pattern projects onto several of the PNA-like and circumglobal streamfunction patterns over the North Atlantic. The trend in convective precipitation associated with these patterns was found to correspond to a transition from below average to above average values in the tropical Indo-western Pacific region. Since each of the SOM patterns was found to have an e-folding time scale of 6–8 days, these findings also imply that the P2 − P1 trend can be understood as arising from shifts in the frequency of occurrence of spatial patterns that vary on intraseasonal time scales.

To explore the above possible linkage between the interdecadal change (from P1 to P2) and intraseasonal variability, we generated indices that represent the intraseasonal variation of the 250-hPa streamfunction and convective precipitation trend patterns. A lagged regression of various fields onto these indices shows that the intraseasonal variation of the 250-hPa streamfunction trend pattern is first excited by enhanced convection over the tropical Indian Ocean, which projects onto the P2 − P1 precipitation pattern. The results of the calculation indicate the following temporal relationships among the various P2 − P1 patterns: enhanced Indian Ocean precipitation is followed 3–6 days later by the P2 − P1 circulation trend pattern, which is then followed within 1–2 days by the P2 − P1 downward IR flux. Because the strongest Madden–Julian oscillation (MJO) convection occurs over the tropical Indian and western Pacific Oceans, a similar mechanism may be operating during MJO events. Vecchi and Bond (2004) showed that the MJO can influence high-latitude surface temperatures. Based on composites of geopotential height and specific humidity, they suggested that both radiative and advection effects are responsible for this connection.

The finding that the interdecadal trend can be realized through changes in the frequency of occurrence of intraseasonal time-scale processes is consistent with the idea of Corti et al. (1999). Based on findings with an idealized model and observational data, they suggested that an external forcing can change the frequency of occurrence of particular circulation patterns. The results of our study further identify what form the external forcing can take, that is, tropical convective heating over the Indian and western Pacific Oceans. Moreover, we show how the circulation that responds to the external forcing can bring about the NH winter Arctic warming, both by dynamical processes and by the subsequent triggering of radiative warming.

In a calculation of the interdecadal trend in the Community Climate System Model (CCSM) of the National Center for Atmospheric Research, with greenhouse gas and sulfate aerosol forcing included, Selten et al. (2004) found a circumglobal 300-hPa streamfunction trend pattern (Fig. 4a of that study) that has a strong zonal wavenumber 5 component as in Fig. 2b, but which lacks the notable zonal mean contribution of the P2 − P1 trend. Their results were presented as the 2051–80 minus 1951–80 300-hPa streamfunction field, which corresponds to a time interval that only partially overlaps with that in our study. To quantify the difference between the trend pattern of Selten et al. (2004) and that of our study, we performed an EOF analysis of the 250-hPa streamfunction field (Fig. 12). It was found that our EOF3 exhibits much overlap with the interdecadal trend pattern of Selten et al. A calculation of the pattern correlation between our trend pattern (Fig. 2b) and the first three 250-hPa streamfunction EOFs yielded values of 0.321, 0.350, and −0.005, respectively. In addition, the linear correlations between the CC index and the principal component time series of the first three 250-hPa streamfunction EOFs are 0.002, 0.230, and 0.172.
respectively. These results further indicate that the trend pattern of Selten et al. (2004) is dissimilar to that which we find. While this difference may be caused by our analysis period being too short, or by deficiencies in the CCSM, given the resemblance between the trend pattern in Selten et al. and our streamfunction EOF3, it is also plausible that this particular pattern may play an increasingly important role throughout the twenty-first century.

Branstator (2002) also defines a circumglobal teleconnection pattern (CTP) based on the first two EOFs of the nondivergent meridional wind. He found that the first two meridional wind EOFs are dominated by zonal wavenumber 5 with very little zonal mean contribution. Because our streamfunction trend and SOM patterns 16–19 are all circumglobal, to determine if the CTP is associated with the trend, we performed an EOF analysis of the 250-hPa nondivergent meridional wind. Our EOF1 and EOF2 patterns were found to closely resemble those of Branstator (not shown). The linear correlations between the CC index and the principal component time series of the first three 250-hPa meridional wind EOFs are 0.166, 0.004, and −0.319, respectively. This lack of a relationship between the CTP and the CC indicates that the CTP is most likely not related to the observed P2 – P1 trend. Returning to the finding by Selten et al. (2004), our calculation showed that the linear correlations between the principal component time series of the meridional wind EOF1 (i.e., one of the CTP patterns) with those of the first three 250-hPa streamfunction EOFs are 0.106, 0.301, and 0.659, respectively. Because the streamfunction trend pattern of Selten et al. resembles our streamfunction EOF3, this suggests that the CCSM trend is associated with the CTP.

Finally, by showing that dynamic warming plays a crucial role for polar amplification, the results of this study contrast the long-held view that the ice albedo feedback is responsible for polar amplification. This study supports the perspective that poleward heat flux associated with changes in the atmospheric circulation plays a critical role in the polar amplification (Graversen 2006; Lu and Cai 2008). Our finding that these circulation changes are excited by enhanced convection over the tropical Indian and western Pacific Oceans is also consistent with the finding of Blessing et al. (2008) who studied the forcing mechanisms that drive the cold ocean–warm land pattern (Wallace et al. 1996). In another study, Compo and Sardeshmukh (2008) showed with a GCM that the impact of increasing CO2 on land warming is not realized directly through radiative processes, but instead through a warming of the sea surface. Although the focus of our study is not the land warming, the results allude to the possibility that the warmer sea surface impacts the NH land area by the triggering of additional convective precipitation followed by dynamic warming and the associated downward IR flux. As discussed in this study, it may be that the land warming is also realized through the increased frequency of these processes. These ideas are presently being investigated with numerical model experiments.

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