CORRIGENDUM

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(Manuscript received 14 December 2011, in final form 19 December 2011)

Because of a production error, the figures published in Cohen and Jones (2011) were of inferior quality. To correct this, the following pages contain the full article as it should have appeared, with the figures processed properly.

The staff of the *Journal of Climate* regrets any inconvenience this error may have caused.

REFERENCE

Cohen, J., and J. Jones, 2011: Tropospheric precursors and stratospheric warmings. *J. Climate,* **24,** 6562–6572.

DOI: 10.1175/JCLI-D-11-00701.1

Tropospheric Precursors and Stratospheric Warmings

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(Manuscript received 26 October 2010, in final form 2 June 2011)

ABSTRACT

Many tropospheric Arctic Oscillation (AO) events are preceded by stratospheric AO events and even earlier in time by anomalous upward energy flux associated with Rossby waves in the troposphere. This study identifies lower-tropospheric circulation anomalies that precede large AO events in both the troposphere and stratosphere and the anomalous upward energy flux. Compositing analysis of stratospheric warming events identifies regional tropospheric precursors, which precede stratospheric warmings. The tropospheric precursor is found to vary when compositing over polar vortex displacements and splits separately. Prior to vortex displacements the main anomaly sea level pressure center of the tropospheric precursor is located across northwest Eurasia and is associated with the Siberian high. Prior to vortex splits a similar anomaly center is identified in the tropospheric precursor but is weaker and appears to be more strongly related to a shift in the storm tracks. Differences in the sea level pressure anomalies in the North Atlantic and the North Pacific are also observed when comparing the precursors prior to vortex displacements and splits. Identification of a unique tropospheric precursor to stratospheric warming and subsequent tropospheric AO events can help to improve understanding troposphere–stratosphere coupling. Furthermore, the observational evidence presented here can be compared with model simulations of winter climate variability and lead to potential model improvements.

1. Introduction

Prediction of the phase and magnitude of the dominant mode of Northern Hemisphere climate variability, referred to as the Arctic Oscillation (AO) or northern annular mode (NAM), is considered the next most important anticipated advance in seasonal winter climate prediction (Cohen 2003). Studies have shown that, on synoptic time scales, the variability in phase of the AO is due to wave breaking, for example, Feldstein and Franzke (2006). Nonetheless, most studies on longer than synoptic time scales have emphasized the lack of understanding of the underlying dynamics driving AO variability and consequently its poor predictability (Seager et al. 2010).

It has been known for a relatively long time that planetary-scale waves that propagate from the troposphere into the stratosphere control the zonal-mean stratospheric circulation and its variability (Charney and Drazin 1961; Matsuno 1970). Furthermore, later studies showed that it is variability within the troposphere that forces the changes in planetary waves that drives stratospheric variability (e.g., Ting and Held 1990; Scinocca and Haynes 1998). More recent work in turn argues that the zonal-mean stratospheric circulation can also exert a downward influence on the zonal-mean tropospheric circulation (Baldwin and Dunkerton 1999, 2001). Still, less is known about the time and meridional variation of the flow in the extratropical troposphere that triggers variability in the planetary waves, which then force stratospheric variability and eventually tropospheric variability on a hemispheric scale.

Eurasian snow cover has been shown to be a leading indicator of winter climate variability in both the troposphere and stratosphere (Cohen and Entekhabi 1999; Cohen et al. 2007; Orsolini and Kvamstø 2009; Allen and Zender 2010; Smith et al. 2010). It has been postulated that snow cover influences hemispheric climate by forcing sea level pressure (SLP) and surface temperature anomalies across northern Eurasia that, in turn, modulate the large-scale vertically propagating energy fluxes that force stratospheric variability. A six-step conceptual model has been proposed on how regional perturbations or tropospheric precursors force first

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DOI: 10.1175/2011JCLI4160.1

hemispheric-wide stratospheric variability followed by hemispheric-wide tropospheric variability (Cohen et al. 2007). There is an initial upward propagation of circulation and energy anomalies and a lagging downward propagation. The downward propagation has been well established at least statistically (Baldwin and Dunkerton 2001; Baldwin et al. 2003) if not physically (Plumb and Semeniuk 2003); however, variability in the lower troposphere that forces upward propagation of wave activity is less well understood (Cohen et al. 2007).

Anomalously extensive snow cover favors a strengthened Siberian high that acts as a tropospheric precursor to stratospheric warming and negative surface AO events (Cohen et al. 2001, 2002). Other observational and modeling studies have confirmed that tropospheric precursors to stratospheric warmings can be associated with Eurasian snow cover variability, the El Niño–Southern Oscillation, and atmospheric blocking. Martius et al. 2009 showed that atmospheric blocking up to 10 days prior to the warming event precedes stratospheric warming events. Blocking in the Atlantic basin most often precedes vortex displacements, while blocking in the Pacific basin often precedes vortex splits. Garfinkel et al. (2010) argue that both ENSO and Eurasian snow cover influence tropospheric height anomalies that force stratospheric variability. Kolstad and Charlton-Perez (2011) showed that climate models simulate tropospheric precursors, which precede stratospheric warming events similar to what is observed in the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis.

Stratospheric variability has been suggested as a long-lead predictor of winter weather (Baldwin et al. 2003). However, using stratospheric variability as a predictor in climate forecasts shows only marginal skill (e.g., Charlton et al. 2003) for three important reasons: 1) the duration and seasonality that the stratosphere and troposphere are coupled is short, 2) variability in the troposphere is more often than not independent of variability in the stratosphere, and 3) the lead time is shorter than is required for seasonal predictions. Identifying tropospheric precursors has the potential benefit of extending the lead time of weather forecasts by predicting the phase of large tropospheric AO events weeks and even months in advance. However until now, long lead tropospheric precursors have not been clearly identified.

2. Data and methods

For all atmospheric data we used the NCEP–NCAR reanalysis (Kalnay et al. 1996) 1948–2010. We repeated the analysis with the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and interim reanalyses (Uppala et al. 2005), and the results were not sensitive to the reanalysis chosen.

TABLE 1. Central date for vortex displacements and vortex splits.

Vortex displacements	Vortex splits
30 Nov 1958	30 Jan 1958
16 Jan 1960	23 Mar 1965
08 Dec 1965	24 Feb 1966
27 Nov 1968	08 Jan 1968
13 Mar 1969	17 Jan 1971
02 Jan 1970	02 Feb 1973
20 Mar 1971	22 Feb 1979
29 Feb 1980	02 Jan 1985
04 Dec 1981	08 Dec 1987
24 Feb 1984	14 Mar 1988
23 Jan 1987	22 Feb 1989
15 Dec 1998	25 Feb 1999
20 Mar 2000	11 Feb 2001
16 Dec 2000	18 Jan 2003
02 Jan 2002	21 Jan 2006
07 Jan 2004	24 Jan 2009
24 Feb 2007	09 Feb 2010
22 Feb 2008	

We defined a stratospheric warming as the reversal of the zonal-mean zonal wind from westerly to easterly at 60°N, 10 hPa to identify major stratospheric warming events (Manney et al. 2005; Charlton and Polvani 2007). The central date (day 0) of each stratospheric warming is defined as the first day that at 60°N, 10 hPa an easterly zonal-mean zonal wind is observed. All of our cases for vortex displacements and vortex splits (see Table 1) match this criterion in the NCEP–NCAR reanalysis except for the 16 December 2000 event; however, this event featured an equatorward displacement of the polar vortex, record warm polar stratospheric temperatures, and record low December AO values.

For deriving the polar cap height anomalies, we computed an area-weighted average of height anomalies poleward of 60°N at all available levels from 1000 to 10 hPa (Cohen et al. 2002). For the wave activity flux (WAF) we used the three-dimensional expansion of the two-dimensional Eliassen–Palm flux derived by Plumb (1985). Previous studies showed that stratospheric warming events are preceded by anomalously strong upward WAF originating from the troposphere (e.g., Polvani and Waugh 2004; Cohen et al. 2007). Statistical significance was computed using a two-tailed Student's t test, and in our composite analysis we removed the daily mean before averaging by day and event.

3. Results

a. Precursors

In Fig. 1, we composited SLP 45 days prior to and 45 days following a stratospheric warming event. Stratospheric

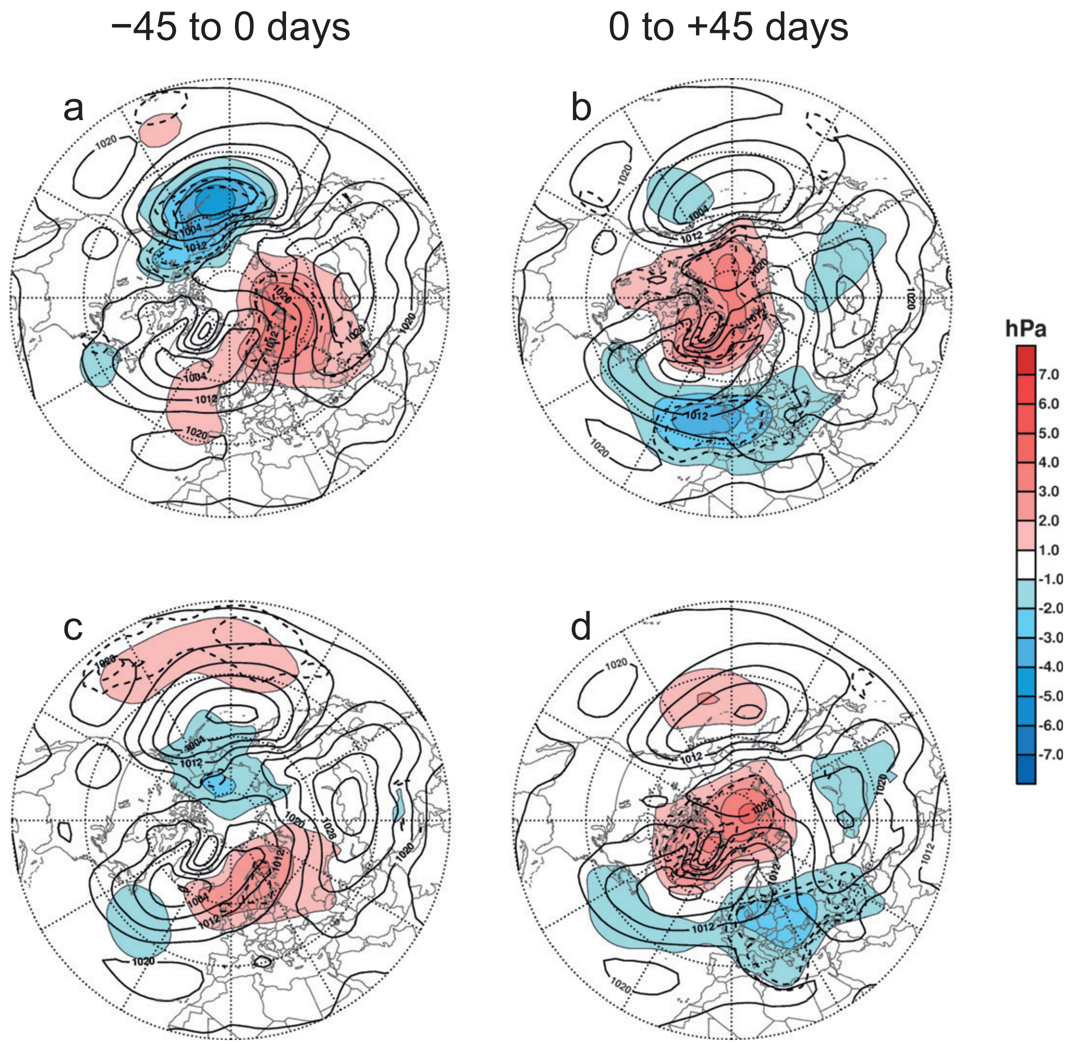

FIG. 1. Sea level pressure (SLP) anomalies (hPa) (a) averaged −45 to 0 days prior to vortex displacements, (b) averaged 0 to +45 days after vortex displacements, (c) averaged −45 to 0 days prior to vortex splits, and (d) averaged 0 to +45 days after vortex splits. Colored shading represents anomalies, solid contours show the full values of the SLP field, and dashed contours represent 90% and 95% confidence levels. Mean values are computed daily over the reanalysis period 1948–2010.

warming events can occur in two different ways, vortex displacement and splits. During vortex displacements the polar vortex is shifted off the pole and during vortex splits the polar vortex is split into two pieces of comparable size. We then separated the stratospheric warming events into vortex displacement and vortex splitting events. In defining vortex displacements and vortex splits we followed the method of Charlton and Polvani (2007). Previous studies have shown that snow cover anomalies and SLP anomalies associated with the Siberian high prior to stratospheric warmings and winter AO events have more of a wave-1 characteristic (Cohen et al. 2001, 2002), which are more closely associated with vortex displacements (Martius et al. 2009).

Following stratospheric warmings, the SLP fields are qualitatively similar, whether it is after vortex displacements or vortex splits with SLP anomalies showing positive values over the Arctic and negative values over the ocean basins (Figs. 1b and 1d). This result is consistent with the analysis of Charlton and Polvani (2007). However, the 45-day period preceding the stratospheric warming event has similarities but also some differences with respect to SLP anomalies. For vortex displacements (Fig. 1a), the Siberian high is significantly strengthened to the northwest of the climatological center; this is the dominant anomaly center in the Northern Hemisphere polar cap (see Fig. 4: prior to vortex displacements a positive height anomaly is observed in the lower

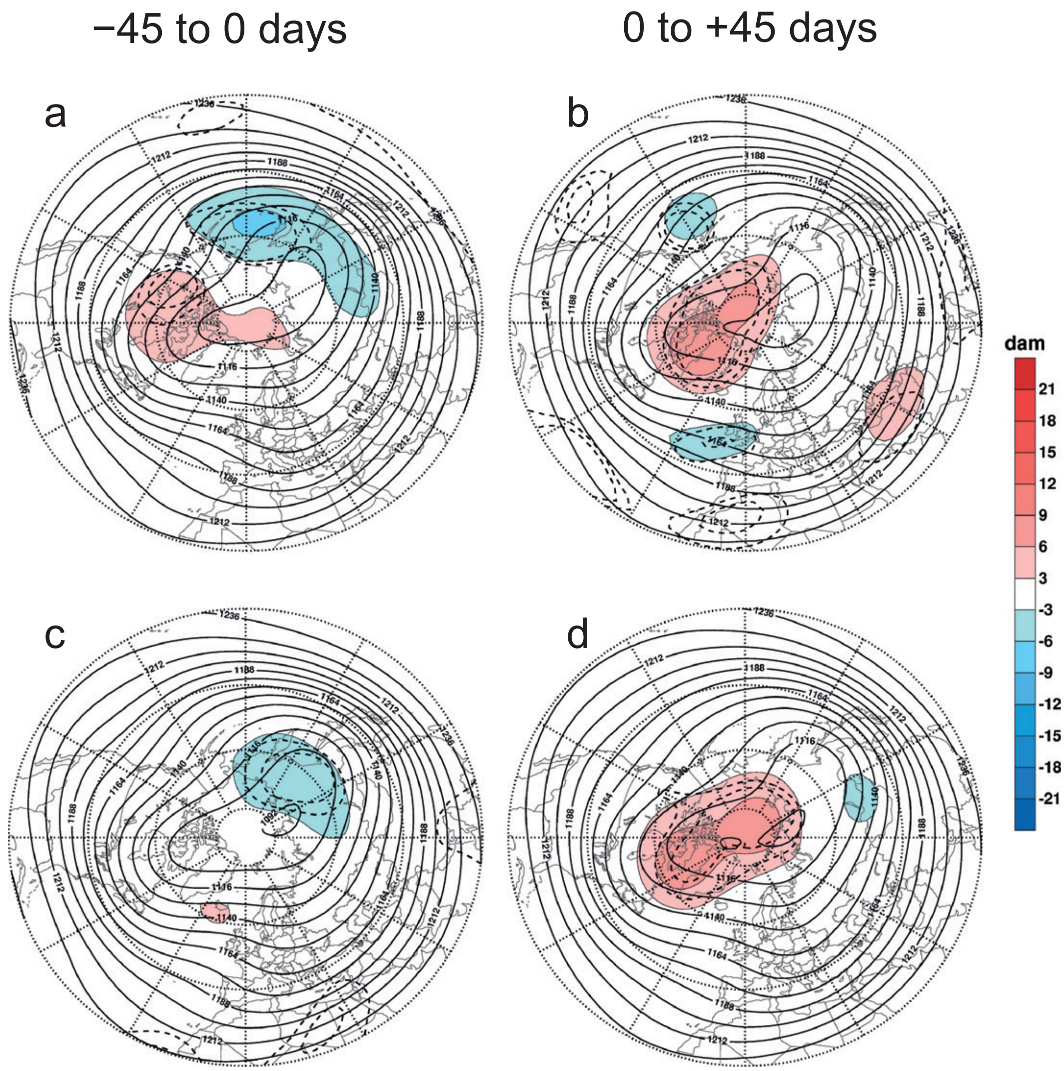

FIG. 2. As in Fig. 1, but for the 200 hPa geopotential height (dam).

troposphere) and has been argued to be associated with increased snow cover (Cohen et al. 2001; Garfinkel et al. 2010). There is also present an opposite low pressure anomaly associated with a deeper Aleutian low, which has been argued to be associated with El Niño (Ineson and Scaife 2009; Garfinkel et al. 2010). Together these two anomaly centers form a wave-1 anomaly pattern. Prior to vortex splits (Fig. 1c), both of these features are present but are weaker and, instead, the most statistically significant feature is anomalous high pressure stretched across the midlatitudes of the North Pacific. The anomaly features resemble a wave train emanating from the North Pacific and propagating over the pole or an asymmetric shift in the main storm tracks, with a northward shift in the North Pacific and a southward shift in the North Atlantic. Together these anomaly centers more closely resemble a wave-2 anomaly pattern. This is consistent with the results of Martius et al. (2009), which showed that a wave-1 pattern dominated the tropospheric height field prior to vortex displacements and a wave-2 pattern dominated the tropospheric height field prior to vortex splits.

We analyzed all of the individual cases in our composites for vortex displacements and splits, and a positive SLP anomaly is found consistently in the northwest sector of Eurasia among the individual cases; however, the location does vary from case to case and contributed to some damping of the anomaly. The primary tropospheric precursor to either vortex displacements or splits is a center of anomalously high SLP primarily to the north or northwest of the climatological position of the Siberian high. The positive SLP anomaly associated with the Siberian high tends to be coupled in time with

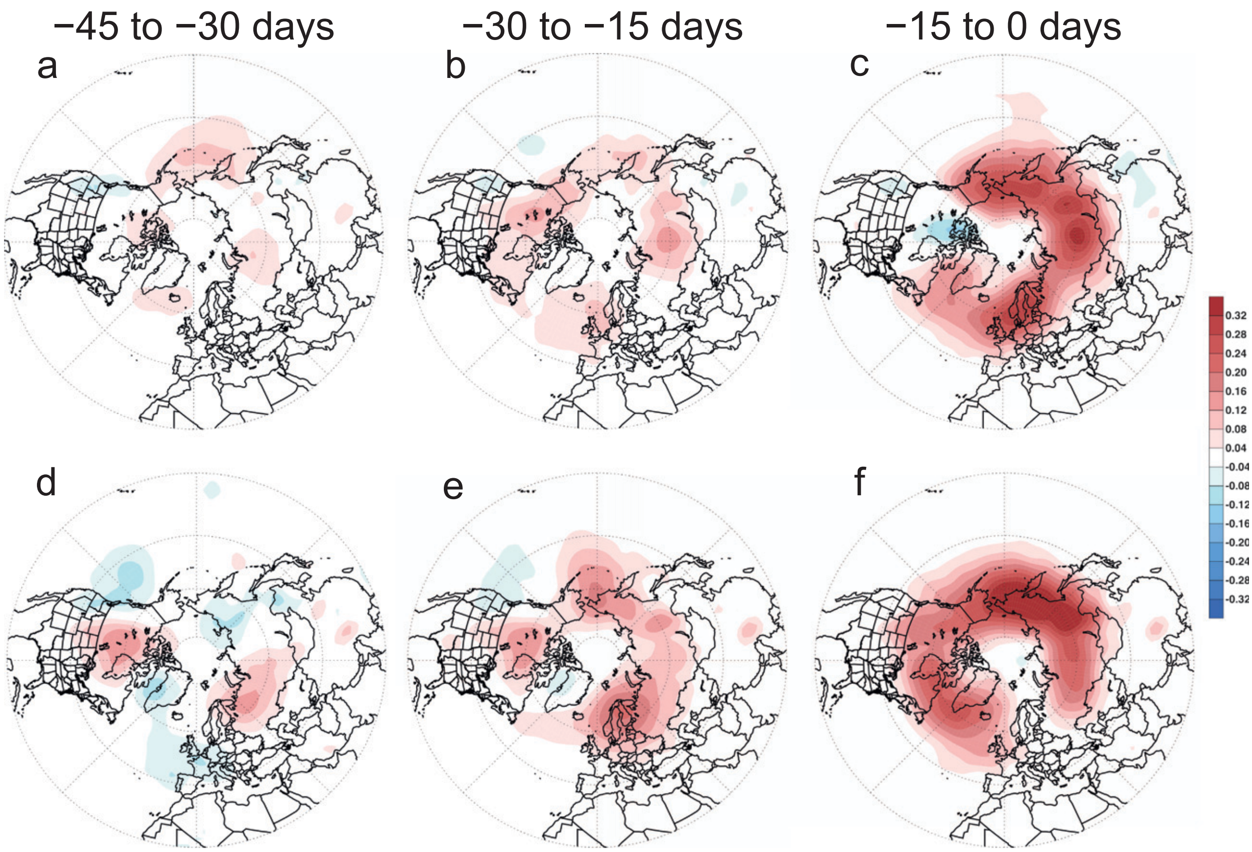

FIG. 3. Vertical wave activity flux (WAF) anomalies ($m^2\ s^{-2}$) at 100 hPa (a) averaged −45 to −30 days, (b) averaged −30 to −15 days, and (c) averaged −15 to 0 days prior to vortex displacements and (d) averaged −45 to −30 days, (e) averaged −30 to −15 days, and (f) averaged −15 to 0 days prior to vortex splits. WAF anomalies shaded.

anomalously low SLP over the Gulf of Alaska. The precursor SLP signal appears in almost every displacement case, sometime between 15 and 45 days prior to the warming. Although still common, a general area of above normal SLP in northern Eurasia was observed less frequently in vortex splits. Also, there are a number of instances in which positive SLP anomalies are observed similar to those associated with major stratospheric warmings, but are followed by either minor stratospheric warmings or have no stratospheric response 15–45 days after they were observed. These instances represent roughly half of the positive SLP anomalies of similar magnitude, which are at least one standard deviation above the area-averaged climatology.

Next, we repeat the analysis in Fig. 1 for geopotential heights at 200 hPa to compare anomalies in the upper troposphere to the lower troposphere. Polvani and Waugh (2004) argue that upward propagating WAF near the tropopause is important to stratospheric variability. Following stratospheric warming events (Figs. 2b and 2d), height anomalies are positive over the Arctic, consistent with a barotropic structure of the tropospheric response following stratospheric warmings. Prior to stratospheric warmings (Figs. 2a and 2c) the tropospheric structure is more baroclinic with negative height anomalies at 200 hPa above the positive height anomalies in the lower troposphere. The Siberian high is a shallow feature extending from the surface to as high as 500 hPa. Increased troughing and anomalously low heights over Siberia are consistent with cooling of the atmospheric column due to snow cover and stronger high pressure at the surface. Comparing Figs. 1 and 2, we note that differences are greater in the lower troposphere than in the upper troposphere prior to vortex displacements and splits.

b. *Wave activity flux and polar-cap geopotential height anomalies*

We next present in Fig. 3 the 15-day mean anomalies in the three-dimensional wave energy flux that vertically propagates from Rossby waves prior to both vortex displacements and splits at 100 hPa. The 6-week absorption of positive anomalous WAF in the stratosphere

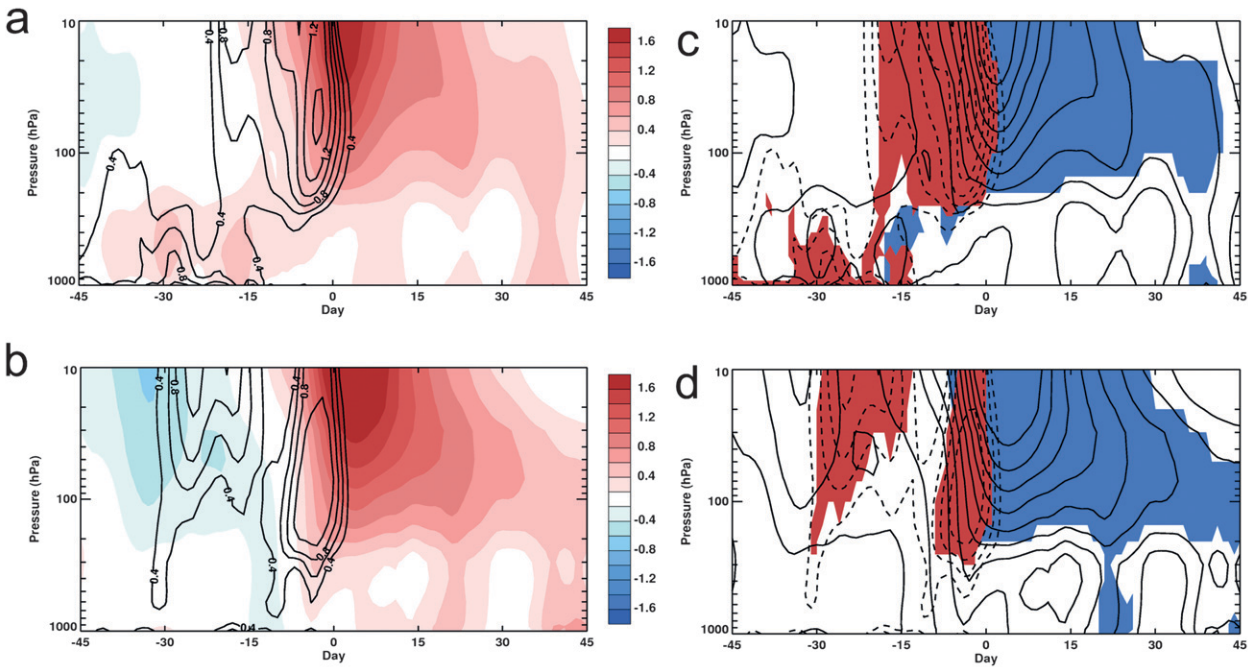

FIG. 4. Polar cap geopotential height anomalies (shading) and WAF anomalies bounded by 40°–80°N, 30°E–180° (contouring) from 45 days prior until 45 days after (a) vortex displacements and (b) vortex splits. Values shaded and contoured every 0.2 standardized anomalies. Also shown are only those values for geopotential height anomalies (shown in solid contours) found to be statistically significant at the 95% confidence level or higher (blue shading) and those WAF values (shown in dashed contouring) found to be statistically significant at the 95% confidence level or higher (red shading) for (b) vortex displacements and (d) vortex splits.

has been shown to precipitate stratospheric warmings (Polvani and Waugh 2004). Furthermore, Smith et al. (2011) show that planetary-scale waves (waves 1 and 2) contribute most to the variability in wave activity flux, while synoptic-scale waves contribute relatively much less to the variability in WAF. Given that the anomalous height field forms either a wave-1 or a wave-2 pattern prior to vortex displacements and splits, respectively, we would expect that the tropospheric precursors presented in Fig. 1 to have a strong influence on the variability in the vertical WAF.

Prior to vortex displacements, there is a slow buildup in the upward WAF starting six weeks prior to the stratospheric warming event, which peaks the last two weeks prior to the stratospheric warming event. The upward WAF anomalies are centered over Eurasia, in particular Siberia, and extend out into the North Pacific. Similarly, prior to vortex splitting events, there is also a slow buildup in upward WAF six weeks prior to the splitting event that peaks the last two weeks prior to the stratospheric warming. There is a strong focus of anomalies over Siberia; however, in contrast to the vortex displacements, the peak upward WAF anomaly is absent from western Eurasia and, instead, there exist additional anomaly centers over North America and the North Atlantic sector.

Next, we present the area-weighted polar cap geopotential height anomalies and WAF anomalies over the Eurasian sector for vortex displacements in Fig. 4a and splits in Fig. 4c throughout the atmospheric column. Also shown is the statistical significance of both variables for displacements (Fig. 4b) and splits (Fig. 4d). Prior to vortex displacements, there are two preceding events seen in the plot. The first event is a positive height anomaly averaged around the polar cap in the troposphere 30 days and even as early as 6 weeks prior to the stratospheric warming and it peaks two to three weeks prior to the stratospheric warming event. This positive polar cap anomaly is a result of the strengthened Siberian high and the increased heights northwest of the climatological center, as seen in Fig. 1. Polar cap heights are positive in the troposphere despite the deepened Aleutian low and lower heights in that region. The second event is the upward pulse of WAF that builds in the lower troposphere and eventually breaks through into the stratosphere, where it peaks just days before the climax in the stratospheric warming event. What is also noted from the plot is that the upward WAF pulse strengthens as the tropospheric precursor weakens; see the further discussion below. Following the stratospheric warming event, the upward WAF ceases and instead the downward propagation of the high heights is seen during the following 6 weeks and culminates in an annular mode pattern in the troposphere, as seen in Fig. 1 and as shown previously (Baldwin and Dunkerton 1999; Baldwin et al. 2003).

Prior to vortex splits, there is not evident in the polar cap diagnostic a long-lead coherent tropospheric precursor as seen for vortex displacements. There is some suggestion of a tropospheric precursor a week before the stratospheric warming event, but earlier the heights are mostly of the opposite sign. However, after the stratospheric warming event, the downward propagation of positive height anomalies is observed, culminating with the annular mode pattern in the lower troposphere similar to vortex displacement events.

Next, in Fig. 5, we present the time series of the SLP anomalies in the region of the strengthened Siberian high in northern Eurasia and the upward WAF anomalies at 100 hPa in the Eurasian sector for vortex displacements. The highest positive SLP anomalies are associated with the tropospheric precursor 15–30 days prior to the stratospheric warming event. Simultaneously, there is a relative minima in the upward WAF flux or a relaxation to climatological values, which is consistent with the building of lower-tropospheric heights. Then, starting three weeks prior to the warming, SLP anomalies fall precipitously while simultaneously the upward WAF turns strongly positive and peaks just prior to the stratospheric warming and then returns to normal levels after stratospheric warming. Also after the stratospheric warming, SLP anomalies slowly build, reaching a secondary maximum over a month after the stratospheric warming and a season or 90 days after the development of the tropospheric precursor. On the bottom axis we have included the average date to provide an approximate or mean timing of these events. We argue that not only does this plot demonstrate the potential application of this analysis for improved seasonal prediction, with an increased lead time of at least 30 days compared with using the stratosphere, but it provides a tangible target for modelers for improved simulation of winter climate. The inability of dynamical climate models to simulate important physical processes contributes to forecast error on seasonal time scales (Doblas-Reyes et al. 2009).

A similar plot for the SLP time series for vortex splits does not show the same features as was seen for vortex displacements (Fig. 5b). There is a building of high pressure over a month earlier than the stratospheric warming event, but it is not statistically significant. However, the strong upward WAF is seen just prior to the stratospheric warming as in the displacement events, though there are two separate large pulses in contrast to the one large pulse observed prior to vortex displacements.

c. *Differences between precursors to displacements and splits*

We performed further analysis to distinguish whether the differences seen in the tropospheric precursor in Figs. 1 and 5 are significant. We computed a tripole index where we took the difference in the area-averaged SLP in the region of the Siberian high, the Aleutian low, and the midlatitudes of the North Atlantic between vortex displacements and splits starting from 45 days before until 45 days after the stratospheric warming events, shown in Fig. 5c. The time series shows that the largest, as well as statistically significant, differences between the two time series occur between four and two weeks before the stratospheric warming. This is the critical period when the main pulses of WAF are initiated for both vortex displacements and splits. We further tested the significance of the difference in the SLP field associated with displacements and splits by taking random samples from the observed data and computed the same tripole index as shown in Fig. 5c. We found that the likelihood of SLP differences of similar magnitude to the composite of differences prior to splits and displacements to be less than 1% simply owing to chance.

In Fig. 6, we present SLP anomalies from 28 to 7 days averaged every 7 days prior to vortex displacements and splits. Prior to vortex displacements there are two dominant centers of SLP anomalies, one associated with the Siberian high and the other with the Aleutian low; together they form a clear wave-1 pattern across the mid to high latitudes. Prior to vortex splits, the SLP anomalies are more diffuse and do not form a clear dipole but, instead, may be better described as a shift in the storm tracks in the respective ocean basins. Although there is a positive SLP anomaly centered near northwest Eurasia, at least initially it does not seem to be associated with the Siberian high. However, closer in time to the stratospheric warming event this positive SLP anomaly propagates eastward and is absorbed into the Siberian high. In contrast, prior to vortex displacements, the dominant positive SLP anomaly propagates westward away from the Siberian high closer in time to the stratospheric warming. It is accepted that arctic outbreaks across Europe are associated with the Siberian high, but this is the first time that we are aware of the westward propagation of arctic high pressure associated with the Siberian high has been demonstrated for Europe. This is an active area of ongoing research.

Finally, we also analyzed the weekly accumulative snow cover extent anomalies for Eurasia. Snow cover extent is measured as a frequency of time when snow cover is observed over a given grid box. Prior to and during the building of the strong high pressure and increased upward WAF, there is a rapid increase in snow cover in the region

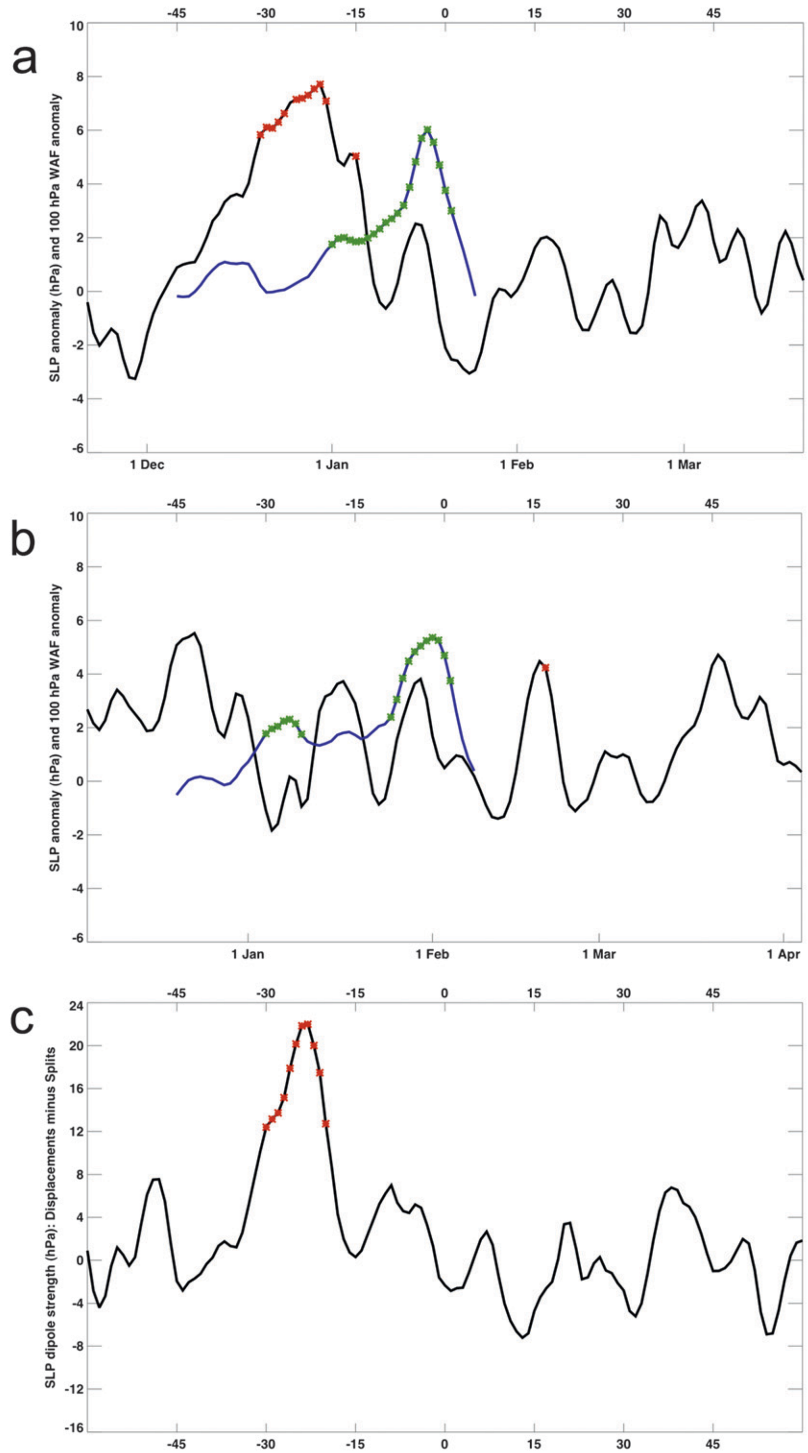

FIG. 5. Area-averaged daily SLP anomalies in the area bounded by 60°–75°N, 30–90°E (black line) and WAF anomalies over the Eurasian sector (blue line) for (a) vortex displacements and (b) vortex splits. (c) The sum of the area-averaged difference in SLP for the boxes bounded by 55°–70°N, 30°–90°E; 55°–70°N, 120°W–180°; and 40°–50°N, 30°–60°W for vortex displacements and splits. Red and green symbols represent 95% significance for SLP and WAF anomalies, respectively. SLP anomalies given in hPa, and WAF anomalies given in $m^2\ s^{-2}$ multiplied by 25. Top axis of (a)–(c) and bottom axis of (c) represent days before and after stratospheric warming event, and bottom axis in (a) and (b) shows mean calendar date.

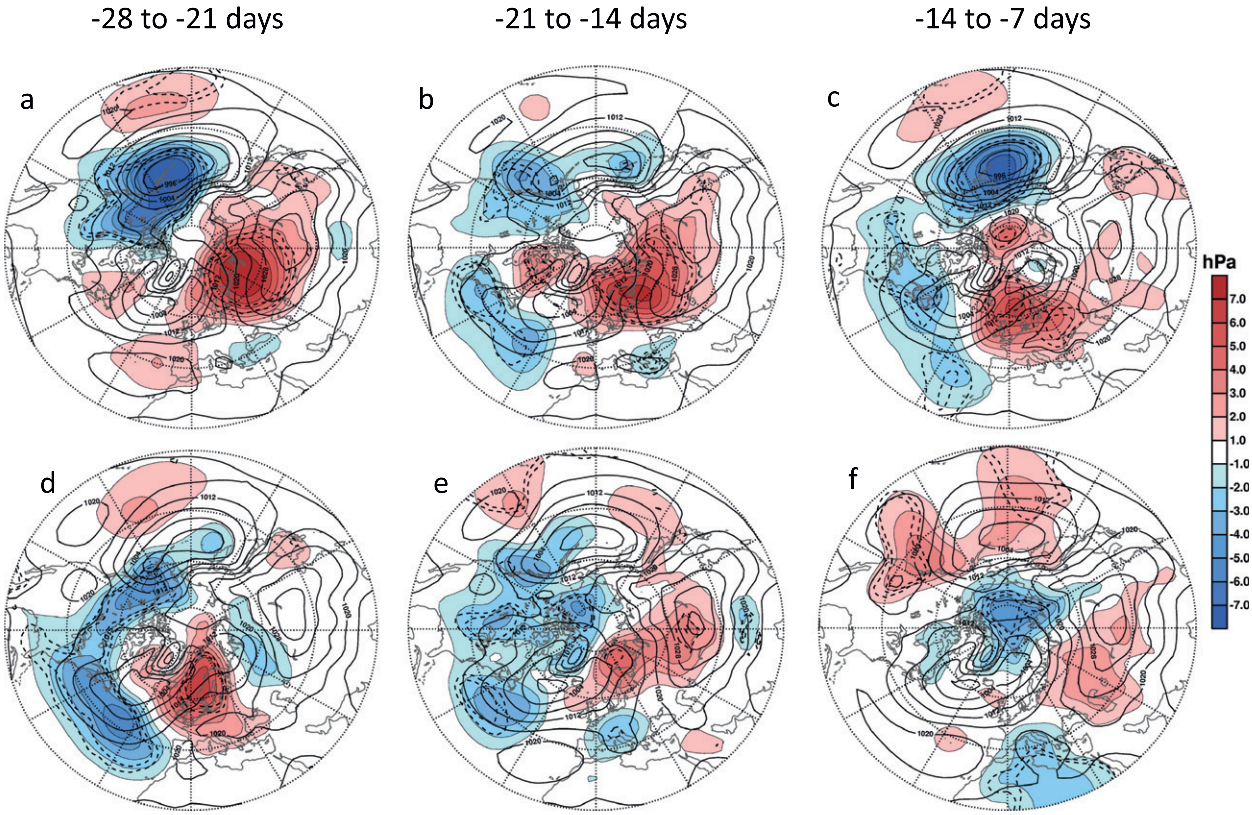

FIG. 6. SLP anomalies (hPa) (a) averaged −28 to −21 days, (b) averaged −21 to −14 days, and (c) averaged −14 to −7 days prior to vortex displacements and (d) averaged −28 to −21 days, (e) averaged −21 to −14 days, and (f) averaged −14 to −7 days prior to vortex splits. Colored shading represents anomalies; solid contours show the full values of the SLP field. Mean values are computed daily over the reanalysis period 1948–2010.

of the development of the high pressure prior to vortex displacements (Fig. 7a). The increase in snow cover extent across the Northern Hemisphere (only Eurasia is shown) is greatest close to the positive SLP anomaly center that is associated with the northwestward expansion of the Siberian high, as shown in Figs. 1a and 6a. Prior to vortex splits, there is also a rapid increase in snow cover anomalies prior to the building of high pressure and increased upward WAF as seen with displacements (Fig. 7b). However, the increase in snow cover is focused closer to Scandinavia, just as the positive SLP anomaly prior to vortex splits is focused closer to Scandinavia, as seen in Fig. 6d.

4. Conclusions

Until recently, vortex displacements and splits have been mostly grouped together, probably because, as shown by Charlton and Polvani (2007) and as we show here, following both events the tropospheric response most closely resembles a negative annular mode pattern for both events. However, prior to both events the tropospheric forcing and the upward propagation of energy emanating from the large tropospheric waves display similarities, but also important differences in the presented analysis. A distinct tropospheric precursor is identified 30 days or even 6 weeks prior to the stratospheric vortex displacement event. The precursor to a vortex displacement involves two of the three main centers of action in the Northern Hemisphere, with a strengthened Siberian high and a deepened Aleutian low. In contrast for vortex splits, the tropospheric precursor resembles a meridional shift in the major storm tracks, more so than a strengthened center of action. A second important distinction is in the upward anomalies of wave-driven energy. The upward pulse in wave activity flux (WAF) has origins in the lower troposphere prior to the main upward pulse. However, from our analysis this is more clearly seen for vortex displacements than vortex splits, and therefore vortex displacements are more easily associated with boundary forcings, such as snow cover and possibly even ENSO, when compared with vortex splits. And, as seen in Fig. 5, the SLP field and the WAF have a unique signature beginning over a month prior to the

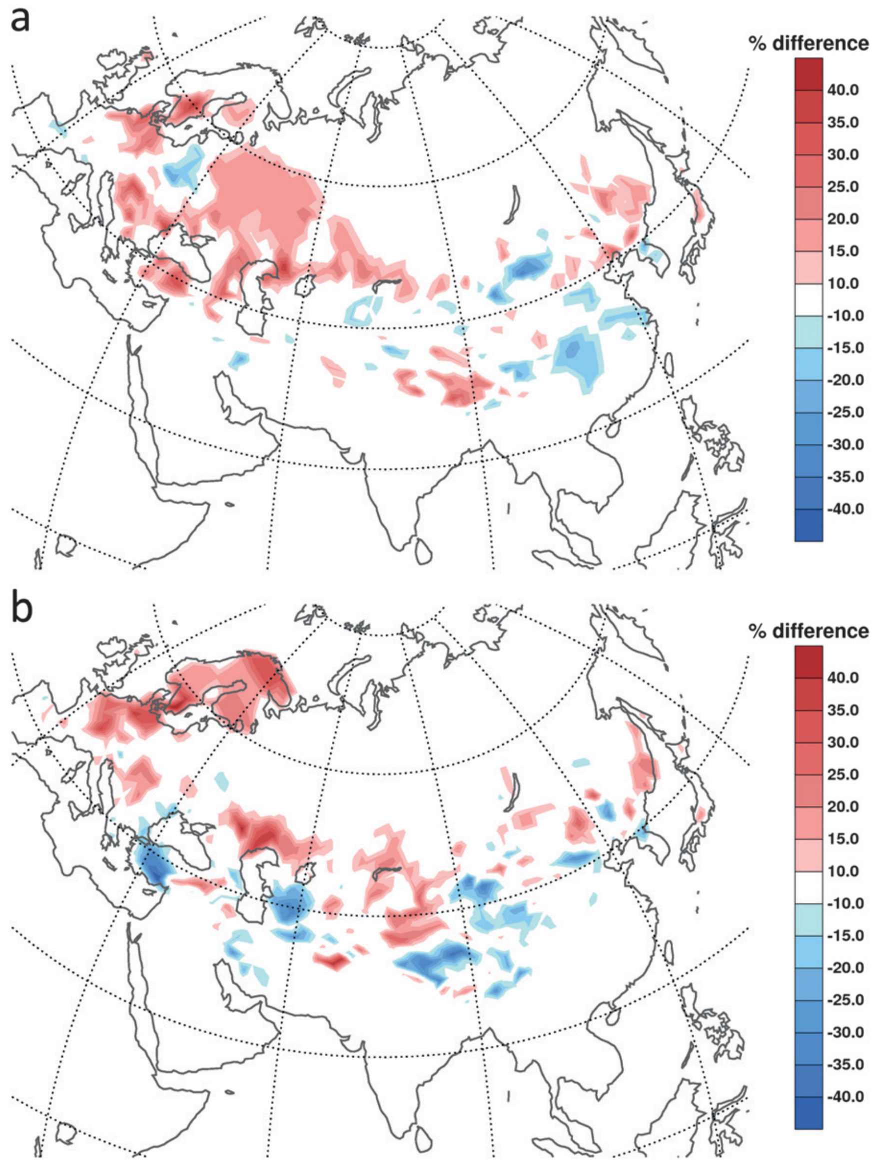

FIG. 7. (a) Change in snow cover extent across Eurasia (a) two to four weeks prior to a vortex displacement and (b) two to four weeks prior to a vortex split.

stratospheric warming and a full season prior to the peak in the surface annular mode event. This signature in the respective fields can be exploited for improved simulations of modeled winter climate and seasonal predictions. Currently, numerical models have been shown to be deficient in predicting atmospheric blocking events (Tracton et al. 1989) and are deficient in simulating troposphere–stratosphere coupling (Hardiman et al. 2008). Identification of precursors may not only advance seasonal forecasting beyond its dependence on ENSO (Cohen and Fletcher 2007), but may also help modelers improve global climate model simulations of present and future winter climate.

Acknowledgments. JC is supported by the National Science Foundation Grants ARC-0909459 and ARC-0909457 and NOAA Grant NA10OAR4310163.

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