Tropospheric Precursors of Anomalous Northern Hemisphere Stratospheric Polar Vortices

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Submitted to Journal of Climate on December 30, 2008

In current form on July 20, 2009

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ABSTRACT

Regional extratropical tropospheric variability affects the wave driving of the Northern Hemisphere wintertime stratospheric polar vortex. Simple reasoning is used to understand the nature of the regional variability that reinforces extratropical planetary waves, and thus vertical EP flux leaving the troposphere. In the European Center for Medium-Range Weather Forecasts (ECMWF) reanalysis record and in WACCM (Whole Atmosphere Community Climate Model), one regional pathway for enhanced planetary wave driving is a deeper low over the North Pacific, and a second pathway is an enhanced high over Eastern Europe. Perturbations in the vortex induced by the two pathways add linearly. These two pathways begin to weaken the upper stratospheric vortex nearly immediately, with a peak influence after a lag of some twenty days. The influence then propagates downwards in time, as expected from wave-mean flow interaction theory. These patterns are influenced by the El-Nino Southern Oscillation (ENSO) and October Eurasian snow cover. These two patterns and the Quasi-Biennial Oscillation (QBO) explain 40% of polar vortex variability during winter.
1. Introduction

a. Background

Much recent work has shown that ENSO has an effect on the polar vortex. Sassi et al. (2004) forced a General Circulation Model (GCM) with observed sea surface temperatures (SSTs) from 1950-2000 and found that the warm phase of ENSO (WENSO) leads to a significantly warmer polar stratosphere. The effect was more pronounced in late winter to early spring. Manzini et al. (2006) and Garcia-Herrera et al. (2006) noted that ENSO’s North Pacific teleconnections propagate to the stratosphere. Taguchi and Hartmann (2006) forced a GCM with perpetual January conditions under both WENSO and CENSO SST conditions in the Pacific, and found more Sudden Stratospheric Warmings, more midlatitude zonal wavenumber-1 (hereafter, wave-1), and a more disturbed vortex under WENSO than CENSO conditions. Limpasuvan et al. (2005) found that in observations, the North Pacific teleconnection typically associated with the cold phase of ENSO (CENSO) leads to vortex intensification. Garfinkel and Hartmann (2007) and Camp and Tung (2007b) demonstrated statistical significance of ENSO’s effect on the polar vortex in the reanalysis record. Garfinkel and Hartmann (2008) (hereafter GH08) showed that the main mechanism through which ENSO modulates the vortex is through its characteristic extratropical teleconnection, which closely resembles the PNA pattern. In particular, GH08 showed that ENSO modifies the wave-1 geopotential height field in the troposphere in such a way that wave-1 height and EP flux are increased in WENSO’s characteristic teleconnection relative to CENSO’s characteristic teleconnection. It is now well established that WENSO weakens the winter stratospheric polar vortex.
A series of papers, beginning with Cohen and Entekhabi (1999), have connected October Eurasian snow cover, the Siberian High in November, and variability in the December and January polar vortex. A summary of earlier work can be found in Gong et al. (2007). Snow directly affects only the lower troposphere overlying the region of the snow anomaly, however, and the pathway through which anomalous snow cover affects the nonlocal circulation has been an open question. Recently, Fletcher et al. (2009) found that the diabatic cooling from the snow causes local isentropic surfaces to dome upwards. In much the same way that a mountain causes an upstream high and downstream low, the domed isentropic surfaces due to the snow induce an upstream high (extending to Europe) and a downstream low (stretching all the way to the dateline). These features, and in particular the Northwestern Pacific Low, propagate upwards into the stratosphere. Hardiman et al. (2008) showed how details of the geographic location of the downstream low over the Northwestern Pacific impact a model’s ability to simulate the effect on the polar vortex of October snow. For the purposes of this article, we assume that a high upstream (extending to Eastern Europe) and a low downstream (extending to the Northwestern Pacific) are associated with Eurasian snow cover anomalies.

The QBO also affects the polar vortex. Holton and Tan (1980) first noted that the zonal mean geopotential height at high latitudes is significantly lower during the westerly phase of the QBO at 50hPa than during the easterly phase. Since then, many modeling-based studies (Hampson and Haynes 2006; Pascoe et al. 2006; Naito and Yoden 2006) and data based studies (Ruzmaikin et al. 2005; Garfinkel and Hartmann 2007) have analyzed the effects the QBO has on the polar vortex, and at the level of detail discussed in this paper, reached similar conclusions. Though the QBO is outside of the main focus of this article, we
will briefly discuss it. Our investigation will center on how anomalies generated by ENSO or October Eurasian snow anomalies, or any other process such as blocking (Martius et al. 2009), can propagate upwards into the stratosphere and weaken the vortex.

2. Data and Diagnostic Tools

The 12 UTC data produced by the ECMWF is used. The ERA-40 dataset is used for the first 45 years (Uppala et al. 2005), and the analysis is extended by using operational ECMWF TOGA analysis. All relevant data from the period September 1957 to August 2007 are included in this analysis, yielding 50 years of data. Randel et al. (2004) found that the ERA-40 data is increasingly inaccurate above 10hPa; here we show all levels.

We also use a 126 year simulation of WACCM version 3.5 to further support results from the ECMWF data. The horizontal resolution is 1.9° by 2.5° degrees (latitude by longitude) with 66 levels in the vertical from the ground up to about 140 km. The physics and chemistry in the middle atmosphere are identical to version 3.0 described in Garcia et al. (2007). The tropospheric convection is upgraded compared to 3.0 to include a new treatment of the dilution of entrainment in convection (Neale et al. 2008) and of the convective momentum transport (Richter and Rasch 2008). The WACCM is run as the atmospheric component of the NCAR Community Climate System Model (Collins et al. 2006). In this configuration, the model interacts with the land, a full depth ocean (which generates an ENSO-like phenomenon), and a sea-ice model. The simulation is a time-slice run with chemical composition corresponding to 1995, spectrally varying solar changes which follows Marsh et al. (2007),
and an imposed QBO as prescribed by the Climate-Chemistry Model Evaluation Activity \(^1\) for the World Meteorological Organization Ozone Assessment. The simulation used in GH08 did not have a QBO and had prescribed sea surface temperatures (and thus a prescribed ENSO). Except where indicated, the hybrid sigma/pressure vertical coordinate is converted into a pressure coordinate before any analysis is performed.

The anomalous polar cap geopotential height area averaged from 70N and poleward\(^2\) and 3hPa to 30hPa (24.5km to 40.7km in our log-p scaling) is used as the index for polar vortex strength (hereafter VSI, vortex strength index). Anomalously low heights indicate a stronger vortex. This index is computed both for daily anomalies (the anomalies are computed as deviations from that calendar date’s climatology, after a ninth order 30 day cutoff low pass Butterworth filter has been applied to smooth the climatology) and monthly anomalies (the anomalies are computed as deviations from that calendar month’s climatology). As in GH08, the QBO is the ECMWF area averaged zonal wind from 10S to 10N at 50hPa, which closely resembles the QBO phase that most strongly affects the early winter vortex (Anstey and Shepherd 2008). The Eurasian snow cover data is from Brown (2000) from 1957-1997 \(^3\) and from Rutgers University from 1967-2007 \(^4\). In the overlap period of 1967-1997 we use an equally weighted sum of the two. The correlation between our index and the index in Cohen et al. (2007) from 1966 to 2004 is 0.85. Indices of geopotential height in the troposphere will be defined in Section 4. The Nino3.4 index from the CPC/NCEP\(^5\) is used in Section 8.

\(^1\)http://www.pa.op.dlr.de/CCMVal/Forcings/CCMVal_Forcings_WMO2010.html
\(^2\)A few figures were created using 65N (Baldwin and Thompson submitted) as the southern latitude; differences were minute
\(^3\)see ftp://sidads.colorado.edu/DATASETS/NOAA/G02131/time_series_sce_swe.txt
\(^4\)see http://climate.rutgers.edu/snowcover/table_area.php?ui_set=1&ui_sort=0
\(^5\)see http://www.cpc.noaa.gov/data/indices/sstoi.indices
as our ENSO index. All indices are defined such that all correlations between the various indices are positive.

Throughout this article, we compute correlations between time series. The time series exhibit autocorrelation, so the number of degrees of freedom (DOFs) used in tests for significance is less than the number of days of data available. To account for this, we compute the DOFs for each index involved in a given correlation (following Bretherton et al. (1999)), and then assign the smallest DOFs of the constituent indices making up that correlation as the DOFs for that correlation.

Three diagnostics are used in Section 4: EP flux diagrams, wave-1 and wave-2 height variance diagrams, and height on a pressure surface. A full description of how these are calculated is in GH08. For the EP flux, the anomaly from the zonal mean for $u$, $T$, $\omega$ and $v$ of the daily 12Z ECMWF data is taken. The zonal Fourier cross spectrum is used to compute the wave-1 and wave-2 components of $\overline{\text{v}'\theta'}$, $\overline{\text{u}'\omega'}$, and $\overline{\text{v}'u'}$. These covariances are then used to compute the EP flux vectors using the equations from Vallis (2006, pp. 537) and Andrews et al. (1987, pp. 128). For Section 4a, daily anomalies are computed as deviations from that calendar date’s climatology, after a ninth order 30 day cutoff low pass Butterworth filter has been applied to smooth the climatology. For Section 4c, we create monthly anomalies as follows. A monthly average of the daily EP flux is computed for each month. The annual cycle is computed by averaging over each calendar month, and the annual cycle is then subtracted from the raw EP fluxes to produce the anomalous EP fluxes. A Monte Carlo test is used to establish significance between different composites of the anomalous EP flux. If either component of the EP flux is significantly different between the two phases, the point is shaded gray. The EP flux vectors are scaled such that plotted
vectors should appear divergent when they are.

Diagrams of wave-1 and wave-2 variance of geopotential height are also shown. The power spectrum is the decomposition by wavenumber of the total variance; thus, a plot of the wave-1 or wave-2 component of the power spectrum shows the variance for each wavenumber. The wave variance, rather than the actual amplitude, is plotted because of the relationship between the total variance of the streamfunction and the EP flux for Rossby waves on a $\beta$ plane with constant static stability and uniform zonal flow (see Vallis (2006, pg. 300) and Andrews et al. (1987, pg. 188)). Like the EP flux, the wave variance is computed from the daily ECMWF data, averaged into monthly means, and then has the climatology removed. The wave variance is multiplied by the density before plotting. A Monte Carlo test is used to test for significance between different appropriately chosen composites, and significant regions are shaded. For both EP flux diagrams and wave variance diagrams, the difference between the 20 most extreme months of each phase of a given index is plotted.

A last diagnostic used is the wave-1, wave-2, and all wave pattern of anomalous geopotential height as compared to climatology. These are produced by regressing relevant indices against a zonal Fourier decomposition of geopotential height. This allows one to visually connect the full pattern of variability associated with a given index to the wave-1 and wave-2 pattern it sets up, and to then compare the anomalies in wave-1 and wave-2 to the climatological wave-1 and wave-2. If the anomalies in tropospheric wave-1 and wave-2 on a pressure level are in (out of) phase with the climatology, we expect an increase (decrease) in EP flux and height variance.
3. Tropospheric Precursors of Vortex Weakening

We begin by objectively searching for tropospheric anomalies well correlated with vortex weakening. To do this, we compute the difference in VSI between each day and ten days later\(^6\); positive (negative) values of this index mark vortex weakening (intensification). We then take the correlation of this vortex weakening index (VWI) with the time series of daily NDJF anomalous (i.e. deviation from the 30 day smoothed climatology for that day and location) geopotential height at every grid point in the mid-troposphere\(^7\). This method pinpoints those locations in the troposphere in NDJF that were associated with a weakening or strengthening polar vortex in the observational record. The tropospheric geopotential height is low pass smoothed with a 6-day cutoff ninth order Butterworth filter to remove synoptic variability; including synoptic variability does not qualitatively affect our results.

Figure 1A shows the correlation of the 500hPa heights with the vortex weakening index in the reanalysis. Figure 1B is created using the WACCM data following this procedure as well (except for the use of the hybrid sigma-pressure level 0.510, instead of the 500hPa level, which is justified because the centers we focus on are away from extreme topography). Three centers of a Central Pacific Rossby wave-train \(^8\), and a high over Eastern Europe, weaken

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\(^6\)Sensitivity to the choice of ten days, as opposed to fewer or more days was explored, without qualitatively affecting the results.

\(^7\)Looking at tropospheric geopotential height halfway through the ten day interval, or a couple of days before the ten day interval starts, indicates that these anomalies propagate slowly westward with time, like large scale Rossby waves; overall, the centers found below still dominate. These centers also dominate composites of the 100 most intense vortex weakening and strengthening days.

\(^8\)The variance of the VWI explained by the subtropical Pacific and Canadian Highs in this wavetrain overlaps strongly with the variance of the VWI explained by the North Pacific Low; in the rest of the paper,
the vortex in both the model and in the reanalysis. The tropospheric influence (especially the high over Eastern Europe) is weaker than in the reanalysis, but overall, the two agree. The rest of this article seeks to understand why Northern Pacific and Eastern European tropospheric variability so strongly modulate the vortex.

The following specific questions are addressed in this article:

i. Can simple reasoning explain why tropospheric anomalies in these two regions would weaken the vortex? Are these two regional anomalies important pathways through which ENSO and Eurasian snow cover affect the vortex? Do these regional anomalies affect the vortex more strongly in early winter or in late winter? See Section 4, 5, and 6 respectively.

ii. Do perturbations of the vortex associated with Eastern European variability and North Pacific variability add linearly? Specifically, is the vortex at its weakest when height over the North Pacific and Eastern Europe are both in a phase that independently would act to weaken the vortex? Camp and Tung (2007a) found that the vortex was little different between EQBO and WQBO under solar maximum conditions, and also between solar minimum and solar maximum under EQBO conditions; does a similar nonlinearity exist when these two tropospheric pathways are examined? See Section 7 and 8.

iii. Matsuno (1970) explained how planetary scale waves in the troposphere can propagate upwards and modulate the polar vortex. But the polar vortex has internally generated variability that is present even with constant tropospheric forcing (Holton and Mass

we focus on the North Pacific Low exclusively.
1976; Scott and Haynes 2000; Scott and Polvani 2006; Gray et al. 2003). How much of the observed variability in the polar vortex is coherent with variability outside of the extratropical stratosphere? See Section 8.

4. Connection Between Regional and Planetary Scale Tropospheric Variability

a. Our Explanation

We now seek to explain why variability in the North Pacific and Eastern Europe are well correlated with vortex weakening. We know that the vortex is weakened by breaking planetary waves (Matsuno 1970). Consider the following problem: where, in the NH extratropical troposphere, can one place regional high or low height anomalies such that stratospheric EP flux is substantially enhanced? The first step in answering this question is to objectively search for those regions in the extratropical troposphere that are well correlated with wave-1 and wave-2 EP flux. We then explain why these regions should have their strong effect on wave-1 and wave-2 EP flux. Finally, we search for regions that enhance both wave-1 and wave-2 EP flux simultaneously.

Our first step is to understand how tropospheric variability affects EP flux in the lower stratosphere. To do this, we correlate the wave-1 anomalous vertical EP flux at 70hPa area averaged from 35N and poleward for each day in NDJF in the ECMWF data (i.e. a daily time series of lower stratospheric wave-1 vertical EP flux anomalies), with the geopotential height anomalies at 500hPa at every gridpoint three days earlier (results are similar if a lag
of two or four days is used). We follow the same procedure for wave-1 EP flux at 500hPa and for wave-2 EP flux as well. In this way, we objectively search for the tropospheric anomalies that precede wave-1(wave-2) EP flux anomalies. Figure 2 shows that wave-1 (wave-2) EP flux is highly correlated with a wave-1(wave-2) pattern of extratropical height anomalies.

Figure 3 shows the climatological stationary eddy (i.e. deviation from the zonal average) height field for NDJF and its wave-1 and wave-2 components. A comparison of Figure 2 and Figure 3A-B shows that the tropospheric anomalies that lead to enhanced wave-1(wave-2) EP flux are collocated with the climatological wave-1 (wave-2) zonal asymmetries. Such a result is expected by the arguments in Dunkerton et al. (1981), who link height anomalies and EP flux. Tropospheric height anomalies that reinforce the climatological stationary waves have a highly significant affect on wave-1 and wave-2 vertical EP flux in the lower stratosphere.

We now return to our original question: why do North Pacific and Eastern European tropospheric anomalies lead to a weakening of the vortex? Our solution is, in essence, that anomalies that reinforce the climatological stationary wave pattern will weaken the vortex. A close examination of Figure 3A-B shows that climatological wave-1 and wave-2 are both low over the Northwestern Pacific and high over Eastern Europe; thus, anomalies that reinforce these climatological asymmetries will weaken the vortex. ⁹ A slightly more quantitative approach is to plot a linear combination of climatological wave-1, wave-2, and

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⁹One caveat should be mentioned. As section 8 will show, much vortex variability has nothing to do with North Pacific and Eastern European variability. In particular, variability external to these two can mask the affect of the two, and the vortex may not actually weaken even with a low over the North Pacific and a high over Eastern Europe. In the rest of this paper, however, we use such strong language.
wave-3, and see where highs and lows appear. Figure 3C shows the extratropical pattern generated by plotting $\frac{6}{11}$*wave1+$\frac{4}{11}$*wave2 +$\frac{1}{11}$*wave3. The 6/11, 4/11, and 1/11 factors are the approximate climatological partition, by wavenumber, of the geopotential variance at the 70hPa level from 30N to 80N in NDJF; the result is insensitive to the exact details of this partitioning. Climatological low wavenumber tropospheric asymmetries are strongest over the North Pacific and over Eastern Europe; thus, anomalies in these locations will affect the vortex.

Figure 4 shows the analog to Figure 3 but for the WACCM data. As in Figure 3C, an enhanced low over the North Pacific will enhance tropospheric wave-1 and wave-2 heights. In Figure 4C, the high that was more confined to Eastern Europe now weakens and spreads into the Atlantic Ocean and into Siberia. Thus, we might not expect the Eastern European High to have quite as strong an effect in WACCM as in the reanalysis. Overall, though, agreement between the two data sources is excellent; because WACCM has realistic stationary waves (i.e. realistic orography and land-sea contrast), the regional tropospheric anomalies that weaken the vortex are nearly identical in WACCM and reanalysis.

Climatological eddies are enhanced by a low over the Northern Pacific, where the climatological wave-1 and wave-2 are both low, and by a high over Eastern Europe, where climatological wave-1 and wave-2 are both high. When the climatological eddies are enhanced, the wave driving of the vortex from the troposphere is expected to increase. Thus, Figure 1 showed that strong anomalies in these two locations weakens the vortex 10.

10We do not mean to imply that these are the only possible tropospheric precursors of vortex variability. As seen in Figure 2, high height anomalies over the Northeastern Pacific (like in January 2009) will lead to anomalous wave-2 EP flux propagating up to the stratosphere, and if the wave-2 EP flux anomaly is strong
b. The AlI and EEI

Before we provide further causal evidence linking regional height variability and vortex variability, we create indices at the two locations where Figures 1, 3C, and 4C have strong zonal asymmetries. A monthly and daily index of the anomalous 500hPa geopotential height at 55°N, 175°E, hereafter the AlI (Aleutian Low Index)\textsuperscript{11}, and at 60°N, 40°E, hereafter the EEI (Eastern European Index), are created in order to track temporal variability at each location. Anomalies for the monthly indices are computed as deviations from that calendar month’s climatology, and anomalies for the daily indices are computed as deviations from climatology after a 30 day smoother has been applied to the daily climatology. The correlation of the monthly(daily) EEI with the AlI in NDJF is -0.05 (-0.01), so the indices are independent of each other. Both indices are defined such that the positive phase results in a weaker vortex. This phase is denoted the ‘W’ phase; the phase that cools and strengthens the vortex is denoted the ‘C’ phase.

The AlI and EEI indices have very low autocorrelation from month to month. For example, the lag 1 month autocorrelation of the Nino3.4 index over the months of NDJFM is .97, whereas the EEI’s and AlI’s lag 1 month autocorrelation is 0.11. These small enough, it can outweigh the lack of wave-1 EP flux. Similarly, a high height anomaly in the Atlantic (like that produced by the blocks in Martius et al. (2009)) can enhance wave-1 EP flux sufficiently that the loss of wave-2 is overwhelmed. Such instances are rare though- Figures 1 and 3 demonstrate that the most common way of weakening the vortex, and the most effective way of increasing the magnitude of the planetary wave pattern (i.e. the way to do it with the smallest anomaly), is to collocate an anomaly with the climatological planetary wave pattern.

\textsuperscript{11}Subsection c will comment on the precise longitudinal position for the North Pacific Low that most strongly affects the vortex.
correlations mean that every month’s EEI and AII is statistically independent. Also, the
correlations of the monthly indices examined here with the QBO are below 0.12; neither the
EEI nor the AII are biased towards one QBO phase.

In GH08, the affect of the PNA on the vortex was examined. Here we focus exclusively
on the North Pacific Low component of the PNA, the Aleutian Low, because the pathway
through which the PNA directly affects the wave driving of the vortex is the Aleutian Low.
In particular, the correlation of the monthly AII in NDJF with the VSI lagged one month
later is 0.26, which is slightly higher than the correlation of the vortex with the PNA index
as defined by the CPC/NCEP used in GH08 (0.20). The correlation between the PNA and
the AII is 0.62.

c. Mechanism for North Pacific and Eastern European Influence

In this subsection, the mechanism through which the AII and EEI affect the vortex is
established. Monthly mean data are used. The first diagnostic used to study how the EEI
affects the vortex is the geopotential height at 500hPa along with its wave-1 and wave-2
components (Figure 5A-C), as compared to climatological wave-1 and wave-2 (Figure 3A-
B). The first diagnostic is generated by regressing the EEI against the 500hPa heights. The
other diagnostics are wave-1 and wave-2 EP flux and height variance (Figures 6A-B and
7A-B), which are obtained by calculating the difference between the 20 biggest positive and
negative extremes in the EEI. The wave-1 anomaly associated with the EEI is mostly in
quadrature with climatology, but wave-2 is in phase in WEEI relative to CEEI. Wave-2 EP
flux and height variance (Figure 7A-B) are significantly increased in WEEI relative to CEEI,
weakening the vortex. Wave-1 (Figures 6A-B) does very little to change the vortex in WEEI relative to CEEI. The net effect of an enhanced high over Eastern Europe is enhanced EP flux convergence at the vortex\textsuperscript{12}.

The second regional pathway is an Aleutian Low. See Figure 5D-F for the wave-1, wave-2, and all wave eddy heights regressed against the AII; these are to be compared to climatological wave-1 and wave-2 (Figure 3A-B). The wave-1 anomaly in WAlI(CAlI) is in phase (out of phase) with climatology. In contrast, the wave-2 anomaly lies along a node in the climatological wave-2 field. Wave-1 EP flux and height variance (Figure 6C-D) are significantly increased in WAlI relative to CAlI. Wave-2 EP flux and height variance (Figure 7C-D) is significantly reduced and anomalous divergence of wave-2 EP flux occurs at the vortex, but the wave-1 convergence overwhelms the wave-2 divergence so that total EP flux convergence at the vortex is increased by WAlI relative to CAlI.

The longitudinal position of the North Pacific low that is expected to most strongly affect the vortex in Figures 3C and 4C differs slightly from the longitude of maximum correlation in Figure 1. In particular, a low near the dateline, not near Russia, is best correlated with vortex weakening. To study why Northwestern Pacific variability has a slightly weaker affect on the vortex than Aleutian low variability, we index Northwestern Pacific variability with the anomalous geopotential height at 55\textdegree N, 130\textdegree E, create the three diagnostics in Figures 5-7(not shown), and compare them to the same three diagnostics for the AII. As might be

\textsuperscript{12}We investigated how lower stratospheric winds affect the propagation of the Eastern European signal to the mid and upper stratosphere in the daily data, but found no significant difference in the modulation of the vortex by Eastern European variability for any of the many configurations of the lower stratospheric zonal winds examined.
expected from Figures 3A-B, the wave-1(wave-2) anomaly associated with a Northwestern Pacific Low is weaker(stronger) than for the All. Though wave-2 EP flux convergence at the vortex is slightly greater for an enhanced Northwestern Pacific low than for the All, wave-1 EP flux, which substantially weakens the vortex during a deep Aleutian low, has only a moderate effect on EP flux convergence at the vortex for a Northwestern Pacific Low. The vortex is sensitive to the total (i.e.- all wave) EP flux convergence at the polar vortex, so a small divergence from wave-2 can be outweighed by the very large convergence from wave-1; thus, a low over the central Pacific is more effective in weakening in the vortex. Nonetheless, an enhanced low anywhere over the central and western North Pacific will result in EP flux convergence at the vortex and weaken the vortex.

d. **Downwards Propagation of Vortex Anomalies**

In this section, we show that the EEI and All affect the vortex in line with expectations from wave-mean flow interaction; namely, vortex anomalies propagate downwards with time. We compute the correlation between the QBO, All, and EEI in NDJF with geopotential height from 70°N poleward at all heights. To investigate the phase lag between variability in these indices and in the polar vortex, the polar cap height is lagged from 0 to 3 months behind the EEI and All indices. Because the EP flux integrated over at least a few prior weeks determines the state of the vortex at a given time (Polvani and Waugh 2004), we correlate smoothed All and EEI with the vortex to better isolate the wave-mean flow interaction.

The procedure is as follows. We first compute the daily EEI, All, QBO, and vortex strength at every level, and then smooth them with a ninth order 30 day cutoff low pass
Butterworth filter. We then generate a daily climatology of the smoothed data, and compute daily anomalies from the day’s climatology. Finally, we compute the lagged correlations between the anomalous EEI, AII, and QBO in NDJF with the anomalous vortex strength at all levels from 0 to 90 days later. In this way, we explore how anomalies in the troposphere and QBO manifest themselves as downwards propagating events in the stratosphere. Because we smooth the indices, we only assign one and a half degrees of freedom per year (one and a half degrees of freedom per winter follows from Bretherton et al. (1999)). The effect of varying the cutoff on the Butterworth filter was examined, without too much qualitative effect on the results below.

Figures 8A,C show the lagged correlation of the vortex strength at every level with the AII and EEI in the ECMWF data, and Figure 8B shows the same for the AII in the WACCM. Significant correlations at 95% using a 1-tailed Student’s t test are in gray. For both the AII and EEI, the upper stratospheric polar vortex starts to weaken nearly immediately. The AII influence on the vortex in ECMWF peaks 20 days after the Aleutian Low peaks and propagates downwards on the timescale of a few weeks. The EEI index is well correlated with the vortex some 20 days later, and shows a similar downward propagation. The EEI appears to influence the lower stratospheric vortex for much longer than the AII, but we do not understand why. The QBO (see Figure 8D) is significantly correlated with the vortex as well, but shows no time lag as the QBO has a much longer time scale. The downwards propagation for the EEI and AII resembles that found in Baldwin and Dunkerton (1999), Kuroda and Kodera (1999), and Reichler et al. (2005).
5. **ENSO and October Eurasian Snow**

We now connect our results from Section 4 with ENSO and Eurasian snow cover. Monthly mean data is used throughout.

**a. ENSO**

A deeper Aleutian Low is part of the characteristic extratropical pattern associated with anomalous convection during WENSO (Horel and Wallace 1981; Hoskins and Karoly 1981). We now examine, in the ECMWF data, whether the Aleutian Low is the main mechanism through which ENSO affects the vortex. To do this, we use regression to remove the influence of the All from the January and February VSI ($VSI_{resid} = VSI - R_{All,VSI,All}$), and then correlate ENSO with the residual VSI. The variance of the vortex explained by ENSO drops by half when the Aleutian low influence is removed. Much of the influence of ENSO on the January and February vortex (the season when Sassi et al. (2004) and Manzini et al. (2006) found a maximum in ENSO's influence on the mid-stratosphere) is due to ENSO’s teleconnection, and once the linear affect of this teleconnection on the VSI has been removed, ENSO’s influence on the vortex is reduced.

The WACCM model run also shows that the dominant mechanism through which ENSO modulates the vortex is the All. We define an ENSO index as the average temperature in the lowest sigma level over the Nino3.4 region, and the All is identical to that used for the ECMWF analysis (this Nino3.4 is correlated with WACCM’s All at the 0.22 level). We again use regression to remove the influence of the All from the January and February VSI ($VSI_{resid} = VSI - R_{All,VSI,All}$), and correlate ENSO with the residual VSI. The correlation
drops from 0.13 to 0.04. Most of the influence of ENSO on the January and February vortex in WACCM is due to ENSO’s teleconnections. We conclude from both ECMWF and WACCM data that the Aleutian Low strength is a good predictor of VSI, and that ENSO contains little independent information about the vortex.

b. October Eurasian Snow

We now examine how the October Eurasian snow effect on the vortex in December and January could be manifested through the EEI or the North Pacific. October Eurasian snow is not well correlated with the AII in early winter. But October Eurasian snow is well correlated with geopotential height over the Northwestern Pacific (not shown, but see Fletcher et al. (2009)), and the Northwestern Pacific could be a conduit through which snow cover anomalies affect the vortex.

The correlation of the Eastern European index in November and December with our Eurasian October snow index is 0.40, which is significant at the 95% confidence level. Such a correlation is consistent with Cohen et al. (2001), and indicates that part of the mechanism through which October Eurasian snow might affect the polar vortex is the upstream high with which it is related.

Finally, we test how much of the effect of October Eurasian snow on the vortex is due to its co-appearance with the Eastern European high, the QBO, and the Northwestern Pacific low. We use regression to remove the influence of the December QBO, EEI, and Northwestern Pacific Low from the January VSI \( (VSI_{resid} = VSI - R_{EEI,VSI}EEI - R_{NW Pac,VSI}NW Pac - R_{QBO,VSI}QBO) \), and correlate October Eurasian snow with the residual VSI. The correlation
drops from 0.34 to 0.18 (which is no longer significant). Most of the influence of October Eurasian snow on the January vortex is due to the presence of the QBO, EEI, and Northwest Pacific Low in December.

6. Intra-Winter Variability

We now investigate whether the vortex responds to the troposphere differently in early and late winter. Figure 3C is very similar for each calendar month, and thus we do not show it for individual months. Because Figure 3C changes so slightly between the different months, we expect that the tropospheric anomalies most strongly correlated with the VWI are the same all winter long. We test this by recreating Figure 1 for the combination of November and December (early winter; Figure 9A-B), and January and February (late winter; Figure 9C-D). Though subtle differences do exist between Figures 1 and 9, the same overall pattern, and the AII and EEI, appear.

An important caveat to our finding that early and late winter are similar needs to be mentioned. We repeated this analysis using monthly mean ECMWF data and found that the troposphere, and especially the EEI, affects the early to mid-winter vortex more strongly than it does the late winter vortex. We are more convinced by our results from correlating daily tropospheric data with the VWI, however, because it maximizes the degrees of freedom available from the short observational record. Thus, we find that the vortex reacts similarly

\[ \text{In the WACCM monthly mean data, however, the effect of the troposphere on the early and late winter vortex is similar. The only significant difference between adjacent calendar months for either index is the difference between the effect of the AII on the March vortex relative to the February vortex.} \]
to early winter and late winter tropospheric variability; as more observational data becomes available, however, this question will merit revisiting.

7. Linearity in Combining the All and EEI

Here we examine whether perturbations in vortex weakening due to the All and EEI add linearly. All days in the record are grouped into the four possible composites based on the value of the All and EEI (the composites are WAll/WEEI, WAll/CEEI, CAll/WEEI, and CAll/CEEI; neutral days are discarded, but the composites with neutral days are included on Table 1 in small type as they also exemplify linearity). The mean VWI is then computed for each of these four composites; see Table 1. Each of these four composites is significantly different from the other three at the 95% level, except for the diagonal comparison of WEEI/CAII to CEEI/WAll. The effect of the EEI is just as significant in any All phase, and the effect of the All is just as significant in any EEI phase. Camp and Tung (2007a) found that the QBO and solar cycle perturbations of the vortex add nonlinearly, but for the All and EEI we find no such nonlinearity.

8. Seasonal Nowcasting

Finally, we wish to analyze how much of the variability of the polar vortex on seasonal timescales is associated with the All, EEI, and QBO. We do this by computing NDJF seasonal averages of the All, EEI, and QBO indices for each year in the ECMWF data, and then correlating each index with a DJFM seasonal average of the VSI. For the longer
WACCM run, we not only examine the externally forced variability of the vortex over the extended winter season, but also examine whether early winter modulation of the vortex by external forcings differs from late winter modulation of the vortex by external forcings. For early winter, a ND average is taken of monthly All, EEI, QBO, and Nino3.4 indices and a DJ average of the monthly VSI, and for late winter a JF average is taken of the monthly All, EEI, QBO, and Nino3.4 indices and a FM average is taken of the monthly VSI.

In addition to comparing the All, EEI, and QBO to the VSI separately, we also compute multiple regression of the three with the VSI to examine how much of the total variability in the VSI is linearly associated with variability outside of the extratropical stratosphere. We correlate the Nino3.4 index and October Eurasian snow with the vortex as well.

Table 2 shows the results for the single regression analysis. Table 2A is for the ECMWF data, and Table 2B is for the WACCM data. The All, QBO, and EEI all have a large effect on the vortex. The correlation of the VSI with the All and EEI is larger than the correlation of the VSI with Nino3.4 and Eurasian snow cover respectively. In WACCM, the difference between early winter and late winter variability for all three indices is not significant at the 95% level. A smaller fraction of vortex variability is well correlated with these three indices in WACCM than in ECWMF, consistent with Figure 1.

We now perform multiple regression of the All, QBO, and EEI with the vortex. Four different multiple regressions are performed (All/QBO, EEI/QBO, EEI/All, and EEI/QBO/All), and the multiple regression coefficient with the VSI for these combinations is in Table 3. Table 3 is similar to Table 2, except the first number in Table 3 gives the correlation between the VSI and the optimally weighted (in a least squared error sense) combination of the All, EEI, and QBO. The EEI does not enhance the multiple regression coefficient of the All and
QBO with the vortex much; the EEI does not covary with the vortex as much as the other two. Forty percent ($0.64^2$) of the total variance of the vortex can be explained by these three predictors.

By comparing the multiple regression coefficients in Table 3 to the individual correlations in Table 2A, we can measure how independent each predictor is from the other two in explaining vortex variability. The double regression coefficient for the All and QBO is 0.63, and the individual correlations are 0.46 and 0.42. If the All and QBO contain entirely independent information on the VSI, then the double regression coefficient for the two would be $\sqrt{0.46^2 + 0.42^2} = 0.6275$. The actual double regression coefficient is 0.6274 (which is rounded to 0.63 on Table 3); this implies that the All and QBO are uncorrelated and explain unique components of the VSI's total variance. A similar calculation can be done for the All and EEI. The maximum possible double regression coefficient is $\sqrt{0.46^2 + 0.27^2} = 0.539$, and the actual double regression coefficient is 0.525; the All and EEI contain mostly independent information about vortex state. The seasonal QBO and EEI, however, are correlated, so that the multiple regression coefficient of the two is noticeably less than the maximum possible correlation. Overall, however, each of the three predictors we consider is fairly independent from the other two.

9. Conclusions

The regional tropospheric anomalies that are expected to most strongly modulate the vortex are found to be those that amplify the climatological extratropical planetary waves. This tropospheric state leads to enhanced vertical EP flux and to a weakened vortex. In the
reanalysis record, the two regional anomalies that most effectively modulate the vortex are the North Pacific Low and the Eastern European High. In WACCM, the result is similar, except that the tropospheric influence is weaker. No significant differences between early winter and late winter are apparent.

A low over the North Pacific results in a dramatic increase in wave-1 leaving the troposphere and converging at the vortex, and a high over Eastern Europe results in a dramatic increase in wave-2 leaving the troposphere and converging at the vortex. A low over the Northwestern Pacific (possibly associated with October Eurasian Snow) has a similar, but weaker, effect than that of a low over the central Pacific. The dominant pathway through which ENSO modulates the vortex is its North Pacific teleconnection. Most of the variance of the polar vortex explainable by October Eurasian snow can be traced back to the Northwestern Pacific, QBO, and Eastern Europe.

These patterns weaken the vortex in a manner consistent with the expectations from wave-mean flow interaction. In particular, the Eastern European high and the North Pacific low affect the upper stratosphere shortly after the anomaly at the surface, and the influence propagates downward over the next month.

The Aleutian low and the high over Eastern Europe are useful forecasters of the polar vortex strength. Combining the two with the QBO explains 40% of the total variance of the polar vortex during winter; 40% of the variance of polar vortex is associated with these three sources of external variability. The remaining variance may be explained by the initial condition of the vortex, the lower stratospheric state, internal stratospheric variability, or other factors.
Acknowledgments.

This work was supported by the Climate Dynamics Program of the National Science Foundation under grant ATM 0409075.
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List of Tables

1 Mean vortex weakening for nine different composites of the EEI and AII. Units are standard deviations of the vortex weakening index. The top left cell gives the mean normalized vortex weakening for the WAI/WEEI composite, the top right cell gives the vortex weakening for WAI/CEEI, etc. A smaller font is used for the composites with either neutral AII or EEI. 33

2 Correlation coefficients of wintertime predictors with the VSI. The first number is the correlation between the VSI and the relevant index, the middle number the probability at which this correlation is significantly different from zero as given by a Student-t test, and the last number the degrees of freedom. Part A is for the ECMWF data, and Part B is for the WACCM data. 34

3 Multiple regression coefficients of seasonal predictors with the VSI. The first number is the multiple correlation coefficient (minimize the squared error), the second number is the probability that the correlation is different from zero, and the last number the degrees of freedom. 35
### Linearity of Vortex Weakening

<table>
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**Table 1.** Mean vortex weakening for nine different composites of the EEI and AlI. Units are standard deviations of the vortex weakening index. The top left cell gives the mean normalized vortex weakening for the WAll/WEEI composite, the top right cell gives the vortex weakening for WAll/CEEI, etc. A smaller font is used for the composites with either neutral AlI or EEI.
Correlation of Individual Patterns with VSI

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<tr>
<th>mnths</th>
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<th>EEI</th>
<th>Nino3.4</th>
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<td></td>
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<td>0.27, 0.971, 47.2</td>
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<td>0.16, 0.965, 119.0</td>
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</table>

Like A, but for WACCM

Table 2. Correlation coefficients of wintertime predictors with the VSI. The first number is the correlation between the VSI and the relevant index, the middle number the probability at which this correlation is significantly different from zero as given by a Student-t test, and the last number the degrees of freedom. Part A is for the ECMWF data, and Part B is for the WACCM data.
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<td>0.53, 0.999, 47.2</td>
<td>0.64, 0.998, 16.0</td>
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</table>

Table 3. Multiple regression coefficients of seasonal predictors with the VSI. The first number is the multiple correlation coefficient (minimize the squared error), the second number is the probability that the correlation is different from zero, and the last number the degrees of freedom.
List of Figures

1  Tropospheric precursors of vortex weakening. Panel A shows the correlation of 500hPa daily anomalous heights with the vortex weakening index (the difference in VSI from day n to day n+10) in NDJF in the ECMWF data. Panel B is identical, but for the WACCM using height at the model’s sigma-pressure level 0.510. Significant regions using a Student-t 2-tailed test with a 95% significance level are shaded.

2  Tropospheric precursors of anomalous wave-1 and wave-2 EP flux. Panel A(C) shows the correlation of 500hPa daily anomalous heights with the wave-1 EP flux at 500hPa (70hPa), and Panel B(D) is identical, but for the wave-2 EP flux. Significant regions using a Student-t 2-tailed test at the 99.9% significance level are shaded. No smoothing is applied to either index. A different lower latitude from elsewhere in this paper is used to ease viewing.

3  Climatological wave-1 (A) and wave-2 (B), \( \frac{6}{\pi} \text{wave1} + \frac{4}{\pi} \text{wave2} + \frac{1}{\pi} \text{wave3} \) (C), and eddy height field (D), in the ECMWF data at 500hPa. Large anomalies are shaded to ease viewing. Contour intervals are 32 meters for A and B, 16 meters for C, and 64 meters for D.

4  Like Figure 3 but for WACCM.
Tropospheric state obtained by regressing 500hPa heights against the Aleutian (55°N,175°E- bottom row) and Eastern European (60°N,40°E- top row) indices. Amplitudes represent a one standard deviation anomaly in height. First column is total (i.e. all wave) height, second column is wave-2 height, third column is wave-1 height. Large anomalies are shaded to ease viewing. The last closed contour in A is at 80m, and the last closed contour in D is at -100m; the contour interval is 20m in A and D, and 10m in B, C, E, and F.

Wave-1 EP flux and height variance associated with the 20 biggest anomalies in the Eastern European feature (A,B) and Aleutian feature(C,D). Regions with significant EP flux are shaded, the divergence of the EP flux is in units of ms$^{-1}$day$^{-1}$, and EP flux arrow lengths are multiplied by a factor of 5 above 100hPa to be visible in the stratosphere. A reference arrow for the stratosphere is located in the top left hand corner of the plot; its vertical component is $1.0879 \times 10^5 kgs^{-2}$, and its horizontal component is $1.25 \times 10^7 kgs^{-2}$. 20 months are in each composite. The 1 hPa and 1000 hPa levels are excluded. The height variance is in units of $m^2$ and has been multiplied by the basic state density. Significantly positive(negative) regions are shaded light gray(dark gray).

Like Figure 6 but for wave-2 EP flux and height variance.

Correlation of North Pacific Low (All), QBO, and Eastern European(EEI) index in NDJF with polar cap heights lagged from 0 to 90 days later, as a function of height. Significant regions are shaded. The WACCM data is plotted into the mesosphere.
Like Figure 1 but for different halves of winter season. Panel A shows ND for the ECMWF, panel B shows ND for WACCM, panel C shows JF for the ECMWF, and panel D shows JF for WACCM.
Fig. 1. Tropospheric precursors of vortex weakening. Panel A shows the correlation of 500hPa daily anomalous heights with the vortex weakening index (the difference in VSI from day n to day n+10) in NDJF in the ECMWF data. Panel B is identical, but for the WACCM using height at the model’s sigma-pressure level 0.510. Significant regions using a Student-t 2-tailed test with a 95% significance level are shaded.
Fig. 2. Tropospheric precursors of anomalous wave-1 and wave-2 EP flux. Panel A(C) shows the correlation of 500hPa daily anomalous heights with the wave-1 EP flux at 500hPa (70hPa), and Panel B(D) is identical, but for the wave-2 EP flux. Significant regions using a Student-t 2-tailed test at the 99.9% significance level are shaded. No smoothing is applied to either index. A different lower latitude from elsewhere in this paper is used to ease viewing.
Fig. 3. Climatological wave-1 (A) and wave-2 (B), $\frac{6}{11} \times \text{wave1} + \frac{4}{11} \times \text{wave2} + \frac{1}{11} \times \text{wave3}$ (C), and eddy height field (D), in the ECMWF data at 500 hPa. Large anomalies are shaded to ease viewing. Contour intervals are 32 meters for A and B, 16 meters for C, and 64 meters for D.
Climatological NDJF, wave 1

Climatological NDJF, wave 2

Avg. of wave1, wave2, and wave3

Full eddy height field

**Fig. 4.** Like Figure 3 but for WACCM.
Fig. 5. Tropospheric state obtained by regressing 500hPa heights against the Aleutian (55° N, 175° E - bottom row) and Eastern European (60° N, 40° E - top row) indices. Amplitudes represent a one standard deviation anomaly in height. First column is total (i.e. all wave) height, second column is wave-2 height, third column is wave-1 height. Large anomalies are shaded to ease viewing. The last closed contour in A is at 80m, and the last closed contour in D is at -100m; the contour interval is 20m in A and D, and 10m in B, C, E, and F.
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Fig. 7. Like Figure 6 but for wave-2 EP flux and height variance.
Fig. 8. Correlation of North Pacific Low (All), QBO, and Eastern European (EEI) index in NDJF with polar cap heights lagged from 0 to 90 days later, as a function of height. Significant regions are shaded. The WACCM data is plotted into the mesosphere.
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