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# Stratospheric control of the extratropical circulation response to surface forcing

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[1] The transient atmospheric circulation response to fall season Siberia region snow forcing in an atmospheric general circulation model is investigated. The forcing generates a stratosphere-troposphere interaction response whose coupling to the surface on multiple week timescales depends on the state of the stratosphere prior to the initiation of the forcing. An initially weak stratospheric polar vortex increases the likelihood of a negative Northern Annular Mode response at the surface. These results may provide useful insight for seasonal prediction of circulation anomalies arising from surface forcing in the extratropics.
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## 1. Introduction

[2] Anomalous fall season Siberian snow cover has been proposed as a potentially skillful predictor of winter Northern Hemisphere (NH) climate anomalies [*Cohen and Entekhabi*, 1999; *Gong et al.*, 2002; *Cohen et al.*, 2007]. The underlying idea is that Siberian snow cover anomalies can generate planetary wave activity flux pulses that propagate into the stratosphere and induce stratospheric zonal mean circulation anomalies [*Saito et al.*, 2001]. Through eddy mean-flow interactions these circulation anomalies then descend into the troposphere [*Baldwin and Dunkerton*, 1999, 2001]. The enhanced predictability comes about from the multiple week separation between the initial snow forcing and the return signal at the surface.

[3] However, the observed predictability derived from the snow-stratosphere-surface link exhibits large uncertainty and several authors have questioned these results [*Limpasuvan et al.*, 2005; *Kushnir et al.*, 2006]. One problem is that stratospheric circulation anomalies can begin without a clear lower tropospheric precursor. And even though previous work suggests that stratospheric anomalies can be used to predict the wintertime surface circulation via a downward coupling [*Charlton et al.*, 2003; *Christiansen*, 2005], the same snow forcing can produce markedly different stratospheric responses [*Limpasuvan et al.*, 2005] and, consequently, very different predictions for the surface response.

[4] In this letter, we analyze the NH extratropical circulation response to Siberia region snow forcing in an atmospheric general circulation model (AGCM, section 2). Our novel result is that the long-term (multiple week) tropospheric response to snow forcing depends significantly on the prior state of the polar stratosphere (section 3). The AGCM is run using a coarsely resolved stratosphere, as is typical in climate assessment studies. Nevertheless, the mechanism we describe here is robust and, we argue, does not depend qualitatively upon the details of the AGCM stratospheric representation (section 4).

## 2. Methods

[5] We investigate the transient atmospheric circulation response to a switch-on snow perturbation in the Geophysical Fluid Dynamics Laboratory (GFDL) atmospheric/land GCM AM2/LM2 [*Anderson et al.*, 2004]. This is the atmospheric component of the coupled ocean-atmosphere model used for climate and seasonal prediction studies at GFDL. The model is run with 24 vertical levels, a lid height of 1 hPa, and with 4 layers above 100 hPa and 7 layers above 300 hPa. The horizontal resolution is 2.5° longitude × 2° latitude.

[6] The time of the transition to snow-covered conditions in the Siberian region varies greatly from year to year [*Robinson et al.*, 1993]. Our snow perturbation experiment represents in a simple way an early start to the snow season in this region. To obtain robust statistics we use a 100-member ensemble of realizations started from independent initial conditions. Each ensemble member is branched from a 100-year simulation of AM2/LM2 forced with observed climatological sea surface temperatures (SST) and sea ice (SI) and fixed atmospheric composition.

[7] Every October 1 of this 100-year control run, we branch two additional simulations that run from October 1 to December 31. In the first branch simulations, we hold the snow mass field fixed throughout the run at its October 1 value at each land-surface point. In the second, we do the same thing, except we add to the October 1 snow mass field a 100 kg m$^{-2}$ perturbation over the Siberia region 60°E–140°E, 40°N–80°N (see Figure 1). The perturbation is equivalent to a 0.4 m deep snow layer [*Frei et al.*, 2003], a characteristic wintertime value. We have verified that our results are insensitive to varying this depth between 0.04 m and 4 m.

[8] Once switched on, the snow perturbation is constant throughout each branch run and the forcing is local to the Siberia region alone. We define the response to the snow perturbation as the difference between the two branch simulations.

## 3. Results

[9] The snow perturbation cools the surface, driving changes in the regional and large-scale atmospheric circu-

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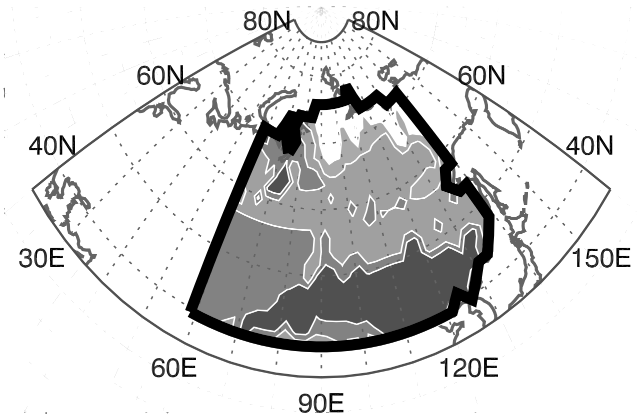


**Figure 1.** Ensemble mean response in surface albedo (grey shading, contour interval = 0.05, minimum shading = 0.35). The snow mass perturbation region is shown by a thick black line.

lation. Figure 1 shows that the snow forcing increases ensemble mean surface albedo by $\sim$0.4–0.5 and local net (upward) shortwave radiation by 35 W $m^{-2}$ during the days 1–30. This causes strong surface cooling ($\sim$0.33 K $d^{-1}$), which induces a circulation response over the forcing region. A surface-trapped high is overlain by a deep low extending up to 100 hPa (not shown), which is the classical large-scale response to midlatitude surface diabatic forcing [*Hoskins and Karoly*, 1981].

[10] To investigate the dynamical response to snow forcing, we examine the response of the geopotential height averaged (with appropriate area weighting) over the polar-cap region poleward of 60°N ($Z_{PCAP}$). Following *Cohen et al.* [2002], we also use $Z_{PCAP}$ as a proxy for the Northern Annular Mode (NAM) [*Thompson and Wallace*, 2000]. Figure 2a shows the evolution of the ensemble mean $Z_{PCAP}$ response ($\Delta Z_{PCAP}$). The ensemble mean time-evolving $\Delta Z_{PCAP}$ is robustly reproduced in two randomly selected 50 member subsets (not shown).

[11] Initially, the most significant response is a tropospheric low centered at 400 hPa that persists until day 50. However, from day 30 onwards the strongest signal is a positive $\Delta Z_{PCAP}$ response in the stratosphere above 100 hPa, corresponding to a warming of the polar stratosphere. As seen in many other studies [*Baldwin and Dunkerton*, 2001; *Polvani and Waugh*, 2004; Cohen et al., submitted manuscript, 2007] this circulation response progresses downward into the troposphere and a weak like-signed anomaly reaches the surface around day 65. We find (not shown) that the stratospheric $\Delta Z_{PCAP}$ is preceded by increased Eliassen-Palm wave activity flux (WAF) from the troposphere that converges in the stratosphere. This WAF response drives a dipolar negative-NAM response in the zonal mean geopotential (Figure 2b), in agreement with the cited observational studies. Thus, despite its coarsely resolved stratosphere, this model is capable of capturing the main features of a stratosphere-troposphere interaction event.

[12] Our principal novel result follows from the variability in the ensemble mean response seen in Figure 2. This variability is large: the peak amplitude of $\Delta Z_{PCAP}$ ($\sim$70 m at 30 hPa at day 65) is only half of the model climatological standard deviation and the ensemble standard deviation of $\Delta Z_{PCAP}$ is 112 m at 100 hPa for days 15–92. But we can significantly reduce the uncertainty in the expected response by using information on the state of the atmosphere prior to the perturbation.

[13] First, following *Reichler et al.* [2005], we ask whether, given the sign of the tropospheric response, robust precursors exist in the initial atmospheric state. We identify cases for which the average lower tropospheric response $\Delta Z_{PCAP}$ (p = 925 hPa, t = Oct. 15–Dec. 31) is positive and those for which it is negative; there are 50 in each set. We then average the zonal-mean geopotential for each set over the five days prior to the perturbation (Sept. 26–30, '$t_{INIT}$' hereafter); we call this quantity $Z_{INIT}$. The composite difference in $Z_{INIT}$ between the positive set and the negative set is shown in Figure 3. The region of the strongest and most statistically significant difference is in the polar stratosphere, poleward of 75°N and above 300 hPa. Despite the relatively poor stratospheric resolution in our model, we claim it is sufficient to resolve this vertically deep structure. Figure 3 implies that realizations in which the polar vortex is initially relatively weak (positive $Z_{PCAP}$) produce a

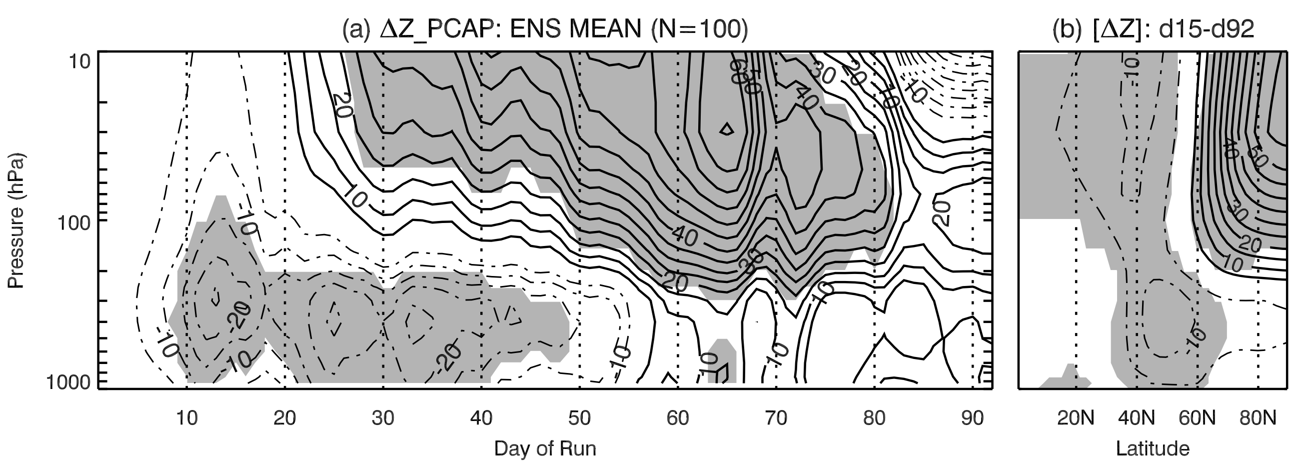


**Figure 2.** (a) Time-pressure cross-section of the ensemble mean $Z_{PCAP}$ response (ΔZ_PCAP). Contour interval = 5 m and negative contours are dashed. Shading indicates significant responses according to a Student's *t*-test ($p(t) < 0.05$). (b) The zonal mean geopotential height response ([ΔZ]) time-averaged between days 15–92. Contours and shading as in Figure 2a.

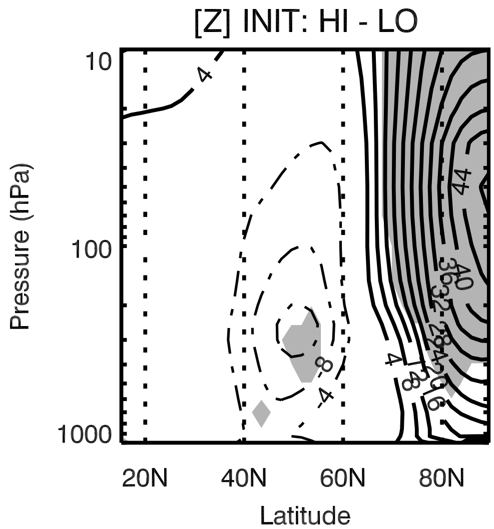


**Figure 3.** The difference in $t_{INIT}$ zonal mean geopotential height ([Z]) between the positive-response and the negative response cases. Contour interval is 4 m and negative contours are dashed. Differences significantly different from zero ($p(t) < 0.05$) are shaded.

relatively strong positive tropospheric response to snow forcing.

[14] Second, we ask whether, for a given initial atmospheric state, we can predict the tropospheric response to the snow perturbation. Figure 3 implies that the polar lower stratosphere is a good location to look for robust predictors of the tropospheric response. Accordingly, we carry out a complementary calculation in which we use $Z_{INIT}$ at 50 hPa as an a priori predictor of how individual realizations will respond to the snow forcing. This predictor describes the initial state of the stratospheric polar vortex, with positive height anomalies indicating a warmer polar stratosphere with a weaker polar vortex and vice versa.

[15] Henceforth, we refer to ensemble members with $Z_{INIT} > 0$ m as ‘‘initially warm’’ cases and those with $Z_{INIT} < 0$ m as ‘‘initially cold’’ cases. Unsurprisingly given Figure 3, there is some overlap with the cases based on tropospheric response: about 2/3 of the initially warm cases are also positive tropospheric-response cases, and similarly for the initially cold and negative tropospheric-response cases. However, we emphasize that the calculations are distinct: the first analysis is based on a posteriori information (the response is given) and the second is based on a priori information (the initial state is given).

[16] Figures 4a and 4b show the composite average $\Delta\ Z_{PCAP}$ time evolution for the initially warm and the initially cold cases. For the warm cases (Figure 4a) the pattern is similar to the ensemble mean (Figure 2a) except that the warming response and the coupling to the surface are much stronger and more statistically significant. We see a more sharply defined warming event that begins in the stratosphere and propagates downwards to the surface on a timescale of several weeks.

[17] The initially cold cases (Figure 4b) exhibit a different time evolution to the initially warm cases. The response pattern shows a more prolonged cooling in the troposphere and the warming in the stratosphere is weaker, shorter in duration, and less vertically coherent. The coupling to the surface occurs later and is weaker relative to the initially warm cases, and the response is not significant at all levels. The difference between the two subsets is highly significant in the troposphere and stratosphere from days 35–70 (Figure 4c).

[18] Stratosphere-troposphere interactions are associated with WAF pulses whose timing is difficult to predict. Thus, even though the snow perturbation is switched on at a fixed date, we cannot determine precisely when the WAF response it induces will occur. But using our database of realizations, we can ask whether, given the time of the WAF response from each realization, $Z_{INIT}$ will influence the downward propagating zonal-mean circulation response to each WAF event.

[19] To answer this question, we define $t_{\mathrm{MAX}}$ as the day when the 50 hPa vertical WAF response ($\Delta$WAF) reaches its maximum amplitude during days 1–50 of each realization. This is the period when the ensemble mean stratospheric response develops (see Figure 2a). Our measure of WAF is the eddy meridional heat flux poleward of 60°N. We examine only the strongest WAF events where upward $\Delta$WAF exceeds two standard deviations of the daily ensemble mean. This selection yields two new subsets, each containing 18 (20) of the initially warm (cold) realizations. Next, we examine the coupling of these WAF events to the surface using the lagged relationship between WAF and sea-level pressure (SLP) [e.g., *Polvani and Waugh*, 2004].

[20] Figures 4d and 4e show composites of the SLP response ($\Delta$SLP) for the new subsets during days $t_{\mathrm{MAX+11}}$ to $t_{\mathrm{MAX+40}}$. In other words, 10 days after each WAF event we compute the 30-day mean $\Delta$SLP; the results are insensitive to moderate changes in the lag and averaging times. The composite for the initially warm cases (Figure 4d) closely resembles a strong negative NAM pattern. In the composite of the initially cold cases (Figure 4d), $\Delta$SLP is weaker and the pattern is less spatially coherent. The significant high SLP response located over Siberia in both composites is related to the snow forcing and not to the stratosphere. Therefore, compositing relative to $t_{\mathrm{MAX}}$ emphasizes more clearly than Figures 4a and 4b that the character of individual stratosphere-troposphere coupling events is strongly influenced by the prior stratospheric state.

[21] We have carried out several analyses to better understand Figures 4d and 4e and their implications. First, notice, as stated in the title of Figures 4d and 4e, that the frequency of strong WAF events is almost the same for the initially warm and initially cold cases. This shows that $Z_{INIT}$ influences the absorption but not the size of the WAF response. Second, we have found that the amplitude of the SLP response to strong WAF events is not significantly different in the two cases, but the SLP pattern in the initially warm cases more consistently resembles a negative NAM. Thus, Figures 4d and 4e do not indicate that the surface responses are weaker in the initially cold cases, but instead that they are more variable and unpredictable.

## 4. Discussion and Conclusions

[22] The significant difference between the initially cold and initially warm cases suggests that, in response to our snow perturbation, there is dynamical information in the

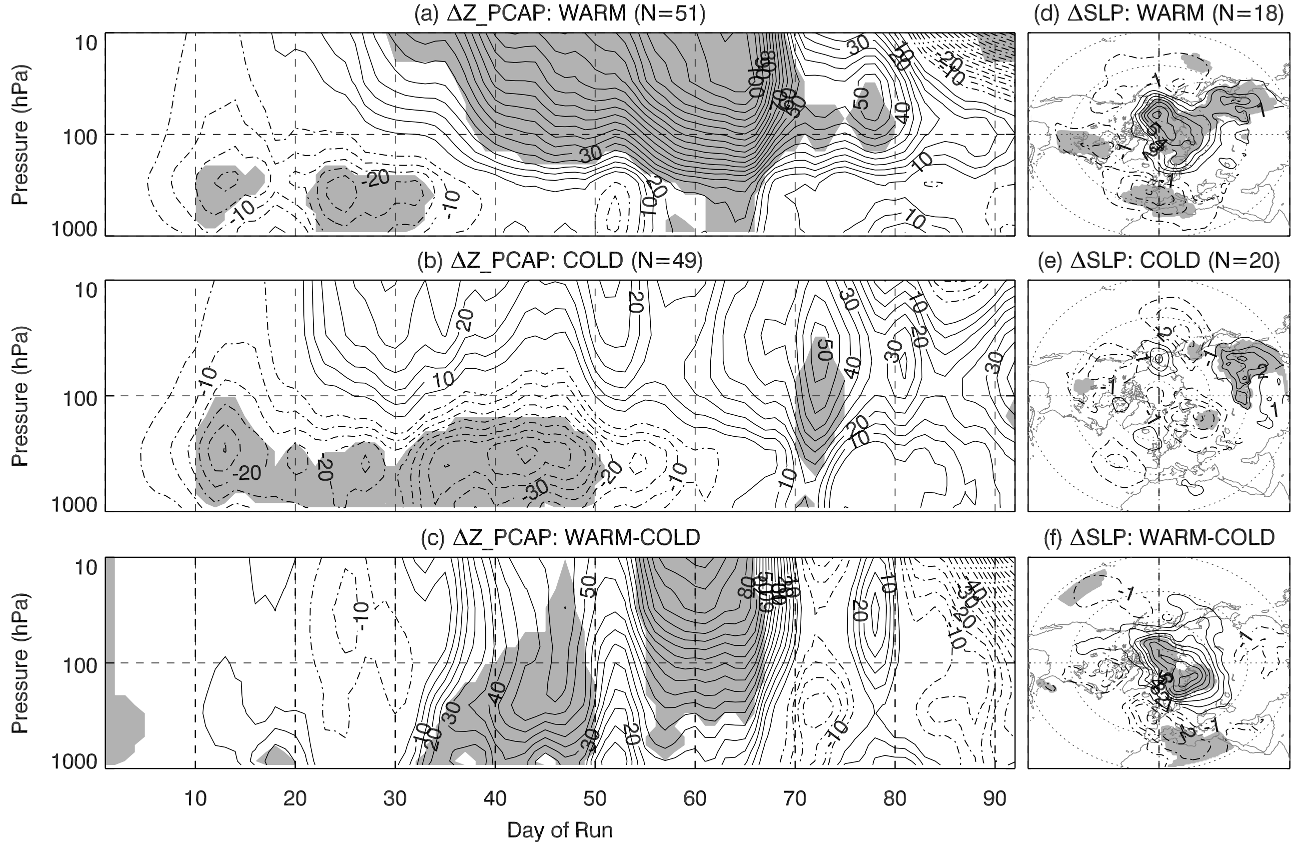


**Figure 4.** Left hand panels are as Figure 2a but for (a) initially warm cases, (b) initially cold cases and (c) is Figure 4a minus 4b. Righthand panels are composite Northern Hemisphere SLP response (ΔSLP) during days $t_{\mathrm{MAX+11}}$ to $t_{\mathrm{MAX+40}}$ in (d) initially warm cases, (e) initially cold cases and (f) is Figure 4d minus 4e (see text for more details). Number of ensemble members in each composite is indicated in figure title. Contour interval is 1 hPa and negative contours are dashed. Significant responses ($p(t) < 0.05$) are shaded.

initial state of the stratosphere that yields predictability extending well beyond the classical 14-day limit for the troposphere. Our tentative hypothesis is that if the stratospheric polar vortex is weak and of limited latitudinal extent, stratosphere-troposphere coupling is more predictable. Under these conditions the vortex is able to absorb relatively more of the WAF pulse induced by the forcing. In the case that the vortex is stronger and larger, the WAF pulse is sometimes deflected equatorward and sometimes ducted along the edge of the vortex, and overall the degree of WAF absorption and subsequent stratosphere-troposphere coupling is less predictable.

[23] We have shown that when the polar stratosphere is initially warm, the polar stratospheric response to snow forcing exhibits further warming. However, this appears to be more than persistence of the stratospheric state. In our 100-year control simulation (section 2), during initially warm cases the stratosphere is equally likely to become warm or cold. Clearly, predictive information about coupled stratosphere-troposphere responses is derived from the combination of a weak vortex and the presence of the snow perturbation.

[24] We are exploring the practical implications of these results for seasonal prediction. For example, we have found that strong planetary wave-driving events produce a significant negative NAM response in 20% (5%) of cases with an initially weak (strong) vortex. An important research question is to determine from the observations whether the combinations of surface and stratospheric predictors lead to a significant improvement in our ability to forecast circulation anomalies on longer timescales.

[25] The stratospheric influence demonstrated here is observed using GFDL AM2 in its standard configuration without a well resolved stratosphere. At this point, we cannot say whether the resulting influence is weaker or stronger than in a model with a better represented stratosphere. However, we claim that the basic mechanism linking the surface response to the strength of the polar vortex should be captured in some form in GFDL AM2. Furthermore, the structures in Figures 2–4 are deep and should be at least nominally resolved by this model. In any case, it is imperative to evaluate these results in a GCM with an improved stratosphere and this is the subject of ongoing work.

[26] Finally, we believe these findings have implications for other problems involving midlatitude surface forcing, for example the response to midlatitude SST or sea ice anomalies [*Peng et al.*, 2003; *Deser et al.*, 2007]. In principle, wherever a surface forcing generates a planetary wave response, the initial polar vortex state could yield predictive information about the transient tropospheric response. Understanding these time evolution problems

could also help better describe the atmospheric circulation response to climate change, in which the surface and stratospheric environments are simultaneously affected.

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