



## Investigating the ability of general circulation models to capture the effects of Eurasian snow cover on winter climate

Steven C. Hardiman,<sup>1</sup> Paul J. Kushner,<sup>1</sup> and Judah Cohen<sup>2</sup>

Received 18 June 2008; revised 9 September 2008; accepted 22 September 2008; published 15 November 2008.

[1] The ability of general circulation models (GCMs) to reproduce the observed strong correlations of Eurasian snow extent in the fall to wave activity and Northern Annular Mode anomalies in the following winter is studied. The observed correlations have been hypothesized to involve two parts: a Rossby wave pulse generated in the troposphere in response to snow-forced surface cooling and a coupled zonal-mean stratosphere-troposphere response to this Rossby wave pulse involving eddy mean flow interactions. It is found that all coupled ocean atmosphere GCMs used within the Coupled Model Intercomparison Project 3 (CMIP3) fail to capture the observed correlations. Using the CMIP3 GCMs and two versions of a particular GCM forced by prescribed sea surface temperatures, possible reasons for this are considered. The snow forcing, as represented in the spatial extent and interannual variability of snow cover area, is found to be reasonable although somewhat weak in the GCMs, as is the relationship between snow cover and the zonal-mean circulation. However, the anomaly of eddy geopotential height associated with Eurasian snow cover anomalies is found to be too localized longitudinally in the GCMs. It is proposed that the reduced longitudinal scale of the snow-forced Rossby wave pulse prevents it from propagating into the stratosphere, thus inhibiting the observed wave-driven stratosphere-troposphere response to the pulse.

**Citation:** Hardiman, S. C., P. J. Kushner, and J. Cohen (2008), Investigating the ability of general circulation models to capture the effects of Eurasian snow cover on winter climate, *J. Geophys. Res.*, *113*, D21123, doi:10.1029/2008JD010623.

### 1. Introduction

[2] The ability of snow cover in Eurasia during the fall to influence the Northern Hemisphere atmosphere the following winter has been the topic of several observational and modeling studies [Cohen and Entekhabi, 1999, 2001; Cohen et al., 2001, 2007; Saito et al., 2001; Gong et al., 2003, 2004; Cohen and Fletcher, 2007; Fletcher et al., 2008]. In observations, it has been found that Eurasian snow extent in October correlates well with the Northern Annular Mode (NAM) Pattern in the following January [Cohen and Entekhabi, 1999] and that the October snow-January NAM connection involves anomalous vertical propagation of Rossby wave activity [Cohen et al., 2007]. The snow-circulation connection involves two parts: first, the set up of a snow-forced circulation anomaly in the troposphere and its propagation as a Rossby wave pulse into the stratosphere, and, second, the stratospheric zonal-mean NAM response to wave driving from the pulse and its subsequent downward progression from the stratosphere to the troposphere.

[3] Some aspects of the snow-circulation connection have been successfully simulated with general circulation models (GCMs) [Gong et al., 2002, 2003; Fletcher et al., 2007, 2008]. For example, Gong et al. have simulated the circulation response to prescribed Eurasian snow anomalies and have found a coupled stratosphere-troposphere NAM response similar to the observed. Using a similar setup, Fletcher et al. have shown that the strength and timing of the NAM response is sensitive to stratospheric conditions and to GCM representation of the stratosphere.

[4] The cited Gong et al. and Fletcher et al. studies prescribe snow anomalies to set up the initial Rossby wave pulse. Once the pulse is set up, it appears that the stratosphere-troposphere NAM response follows reliably. However, in this study we ask whether or not current generation GCMs are capable of simulating observed snow-circulation events spontaneously, from the appearance of anomalous snow cover in October to the accompanying NAM anomaly in the following winter. The answer appears to be “no”: the observed connection between October Eurasian snow cover and wintertime circulation is not captured in the set of GCMs we investigate. This is perhaps surprising, because current generation GCMs are capable of simulating seasonal and interannual variations in snow [Slater et al., 1998; Frei et al., 2003], because interannual variations in snow cover in GCMs do exhibit NAM-related signatures [Gong et al. 2002], and because GCMs have been shown to be capable

<sup>1</sup>Department of Physics, University of Toronto, Toronto, Ontario, Canada.

<sup>2</sup>Seasonal Forecasting, AER, Inc., Lexington, Massachusetts, USA.

of simulating stratosphere-troposphere connections in other contexts [e.g., Charlton *et al.*, 2004; Gillett and Thompson, 2003].

[5] In this paper, we document and begin to explain the inability of GCMs to simulate the snow-circulation connection, which is potentially important for seasonal and long time scale climate prediction [Cohen and Fletcher, 2007; Cohen and Barlow, 2005]. A description of the GCMs and observational data used in this study is given in section 2, the results are presented in section 3, and discussion and conclusions are given in section 4.

## 2. Models and Observational Data

[6] As stated in the Introduction, this study presents a primarily negative result: that current generation GCMs do not simulate the observed connection between October snow anomalies and winter NAM anomalies. We have analyzed a variety of GCMs in different configurations, including coupled ocean atmosphere GCMs and GCMs integrated with prescribed SSTs. We will not present a comprehensive analysis but instead highlight the following representative model configurations:

[7] 1. We provide a general characterization of the control run simulations, with preindustrial greenhouse gases and fixed radiative forcings, from the Coupled Model Intercomparison Project 3's (CMIP3) archive <http://www-pcmdi.llnl.gov/projects/cmip/index.php>. The lengths of these model runs vary from 100 years to 150 years. All the CMIP3 models are fully coupled atmosphere-ocean GCMs.

[8] 2. We present more detailed results from two versions of the Geophysical Fluid Dynamics Laboratory (GFDL) model AM2.1. The first version, which we call "AM2\_STANDARD", is the atmospheric component of the coupled ocean-atmosphere GCM CM2.1 in the CMIP3 archive. It is run with observed SSTs over the period 1983–1998, with radiative forcing held fixed at 1990 levels. AM2\_STANDARD is a "low-top" GCM with 24 vertical levels (4 levels above 100 hPa; model lid at 3 hPa) and is documented in Anderson *et al.* [2004] and Delworth *et al.* [2006].

[9] 3. The second version of AM2.1, which we call "AM2\_HI", is one with improved stratospheric representation, with 48 vertical levels (21 above 100 hPa) and a model lid at 0.003 hPa. While in AM2\_STANDARD a Rayleigh drag scheme acts to damp winds in the upper layers of the model, in AM2\_HI a more physically realistic non-orographic gravity wave drag scheme [Alexander and Dunkerton, 1999] is used in the stratosphere. Other characteristics of AM2\_HI are described by Fletcher *et al.* [2008]. The particular integration we use is the control run from the Fletcher *et al.* study. It is forced with climatological SSTs that repeat each year. The simulation is run for 50 years.

[10] In this study we use three sources of observational data:

[11] 1. NCEP reanalysis data from 1948–2004 (57 years) available from <http://www.cdc.noaa.gov/cdc/reanalysis/>.

[12] 2. A measure of October-mean snow cover extent over Eurasia that combines satellite-based observations [Robinson *et al.*, 1993] for 1967–2004 and Brown's [2000] historical reconstruction for 1948–1966. Cohen *et*

*al.* [2007, Table 1], provides the numerical values for the index.

[13] 3. The snow depth data used in (Figure 3a) is ERA40 data from 1957–2001 (45 years) obtained from the ECMWF data server [http://data.ecmwf.int/data/d/era40\\_daily/](http://data.ecmwf.int/data/d/era40_daily/).

## 3. Results

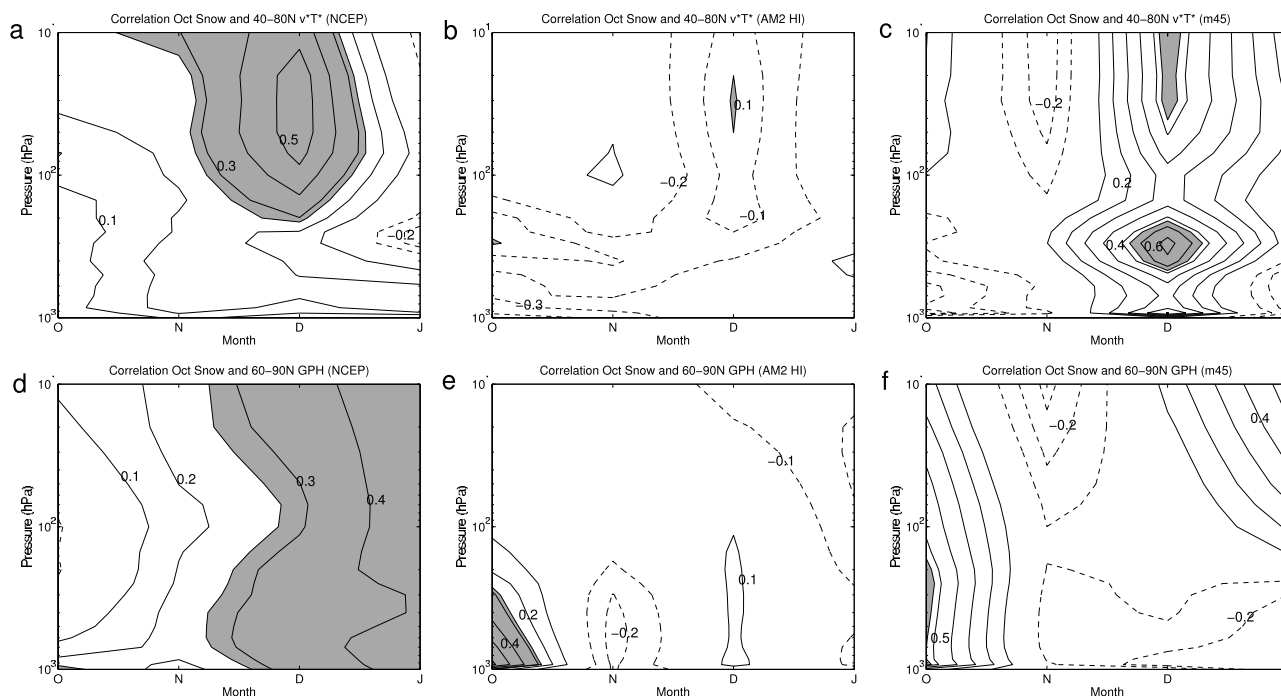
[14] The two-part mechanism of Cohen *et al.* that explains the connection between Eurasian snow extent in the fall and NAM anomalies in the following winter is as follows. Anomalously high snow cover over Eurasia locally cools the land surface and increases the land-sea temperature gradient. This increases the zonal asymmetry in surface and lower tropospheric temperature and hence in the potential vorticity distribution. This results in a Rossby wave pulse that amplifies the climatological stationary wave field. The amplified planetary waves gradually act on the zonal-mean wind until they penetrate into the stratosphere in December. The planetary waves then break on the edge of the stratospheric polar vortex, which exerts a westward torque in the stratosphere. This weakens the winds in the vortex and warms the polar stratosphere. The anomalous zonal-mean circulation projects onto the negative phase of the NAM, and the anomaly in the NAM progresses downward into the troposphere (as described by Baldwin and Dunkerton [2001]) in December and January.

[15] Figures 1a and 1d show this mechanism at work in the observations. Figure 1a shows the correlation coefficient of Eurasian snow extent index with a simple measure of the vertical flux of planetary wave activity from the troposphere to the stratosphere. This measure of wave activity flux is the 40–80N averaged 100 hPa meridional component of the zonal-mean eddy heat flux,  $v^*T^*$ , calculated from the year-to-year departures from the climatological mean of the eddy meridional velocity and temperature fields. The snow index is a measure of the October mean snow covered area in Eurasia, as described in section 2. The meaning of the term "snow index" will remain identical throughout the text. Shaded regions indicate statistical significance at the 95% level as calculated using the Students *t*-Test. The *t* value for this test is calculated from the correlation coefficient, *r*, using the formula

$$t = r\sqrt{\frac{n-2}{1-r^2}} \quad (1)$$

where *n* is the number of model years. Employing the *t*-Test for the field in Figure 1, significant correlations are those above 0.27 for the observations (57 years), 0.28 for AM2\_HI (50 years), and 0.54 for AM2\_STANDARD (16 years).

[16] Figure 1a shows that, in the observations, there is a large correlation, about 0.5, with December  $v^*T^*$  in the upper troposphere and stratosphere, demonstrating that in years where October snow extent in Eurasia is anomalously large, there is an anomalously large upward planetary wave flux in December. This represents the first part of the two-part snow-circulation connection. The second part arises from mean-flow forcing of the zonal-mean circulation by the wave activity flux anomaly. Figure 1d demonstrates the combination of these, showing the statistically significant



**Figure 1.** (a–c) The correlation of the snow index with October mean  $v^*T^*$  area averaged from  $40^{\circ}\text{N}$  to  $80^{\circ}\text{N}$  in NCEP, AM2\_HI, and AM2\_STANDARD, respectively. (d–f) The correlation of the snow index with October mean GPH area averaged from  $60^{\circ}\text{N}$  to  $90^{\circ}\text{N}$  in NCEP, AM2\_HI, and AM2\_STANDARD, respectively. The contour interval is 0.1 with solid contours representing positive values, dashed contours representing negative values, and shaded regions representing statistical significance at the 95% level.

correlation between the snow index and December and January polar cap averaged geopotential height (GPH) throughout the atmosphere.

[17] Figures 1a and 1d are similar to Figure 4 of *Cohen et al.* [2007], but Cohen et al. use daily dynamical fields instead of the monthly fields used here. We find that using daily fields to calculate  $v^*T^*$  and using the monthly mean of the result to calculate correlations with October snow produces an almost identical figure (not shown). Thus interannual variations in the quasi-stationary wave field, rather than in the transient eddy field, are important here. When computing equivalent figures for the GCMs we use monthly mean data, which is much easier to obtain and analyze than daily data.

[18] Figures 1b, 1c, 1e, and 1f demonstrate that in the GFDL GCMs the observed effect of snow on the stratosphere-troposphere circulation is not simulated. In AM2\_HI the correlation of the snow index to  $v^*T^*$  has the wrong sign and is not significant, and the correlation to GPH is too strong in October and negligible thereafter. Thus having a model with a relatively well resolved stratosphere does not guarantee that the snow-circulation link will be captured. In AM2\_STANDARD the correlation of the snow index to December  $v^*T^*$  is slightly more realistic, but has the wrong vertical structure. The correlation to October and November  $v^*T^*$  is of the wrong sign, and there is negligible correlation to GPH.

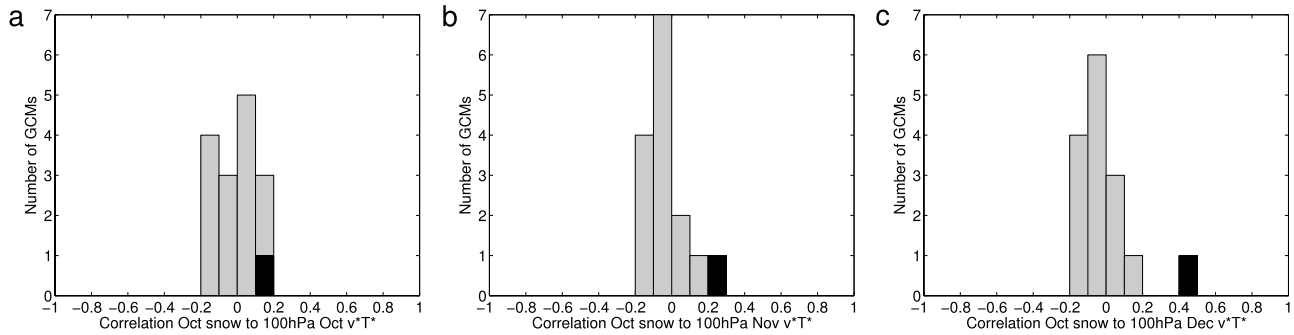
[19] We have tested for these correlations in other versions of the GFDL atmospheric GCMs, including versions under development for coupled climate modeling at GFDL, and find a similar disagreement with observations (not

shown). Furthermore, this disagreement with observations extends to all the CMIP3 GCMs. Figure 2 shows histograms of the correlation of the snow index to October, November and December 100 hPa  $v^*T^*$ . The CMIP3 GCMs are shown by gray bars, and the observations are shown by black bars. About half the models show a positive correlation in October, but by November and December the magnitude of the correlation in all models falls well below that in the observations, demonstrating that none of these models capture the effect of snow on the dynamics.

[20] The snow-circulation connection is a highly temporally and spatially structured phenomenon that provides a stringent test of a GCM. There are several reasons why GCMs might fail to capture this phenomenon. For example, the model might produce a poor simulation of snow variability in October, a poor simulation of the wave field associated with snow forcing, a poor simulation of the zonal-mean flow through which the anomalous waves might propagate, or a poor simulation of the eddy mean flow interactions involved in the downward propagating NAM response. We now examine several of these factors for our set of GCMs.

### 3.1. Snow-Climate Relationship

[21] Many GCMs, such as the GFDL AM2, have a comparatively simple land surface component and snow scheme. Given the sensitivity of models to the details of snow cover [e.g., *Loth and Graf*, 1998; *Hall and Qu*, 2006] it is reasonable to ask whether this adversely affects the representation of snow variability in the model. Generally speaking, however, snow appears to be reasonably well



**Figure 2.** Frequency distribution of the October snow index to (a) October, (b) November, and (c) December  $v^*T^*$  at 100 hPa for the 14 CMIP3 GCMs. The CMIP3 GCMs are shown by gray bars, and the observations are shown by black bars.

simulated in GCMs. For example, Figure 3 shows the spatial pattern of the snow depth regressed on the normalized snow index (normalized so as to have unit standard deviation) for (a) the observations, (b) AM2\_HI, and (c) AM2\_STANDARD. The snow index is calculated for each model using the method of *Brown* [2000]. The models show slightly more extensive snow cover but agree well with the observations.

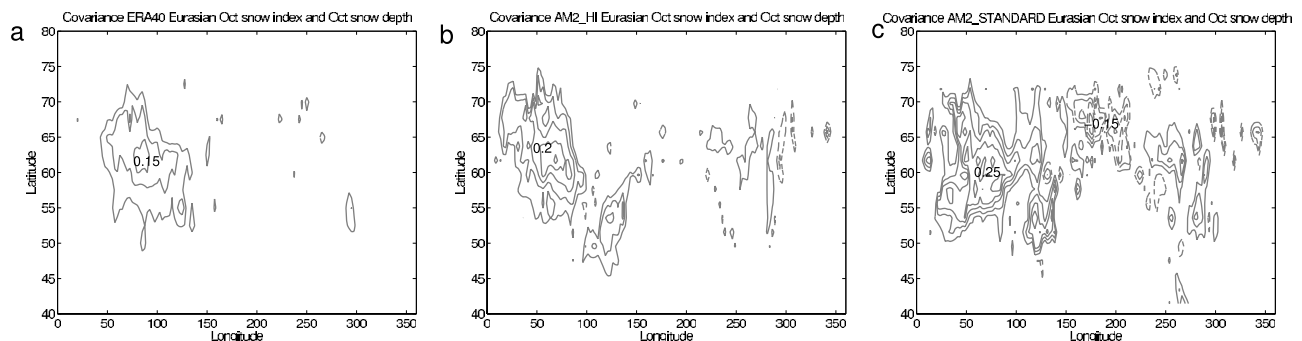
[22] The snow depth variability shown in Figure 3 is a useful measure of snow representation in GCMs. However, once the snow depth is more than a few centimeters further increase in snow depth will not alter the surface albedo or, therefore, the radiative influence on the dynamics. A better measure of how snow cover will influence the dynamics in GCMs is snow extent variability, given by the variability in the snow index. The value of this index is unaffected by increase in snow depth past a threshold value (here 2.5 cm, which although low for observations is appropriate for GCMs). The interannual variability in Eurasian snow extent, studied also in *Frei et al.* [2003], is low in most GCMs (see Figure 4). It is more realistic in AM2\_STANDARD, which uses observed SSTs, than in AM2\_HI, which uses climatological SSTs, and in AM2\_LO, a version of the standard AM2.1 forced with climatological SSTs. The interannual variability in the Eurasian region captured by the CMIP3 models is on average about half the observed. This may partly explain the poor correlation between snow and  $v^*T^*$  in Figure 1b. Other things remaining equal, a reduction by half in the amplitude of the snow forcing

would reduce by a factor of four the amplitude of the eddy forcing of the zonal-mean flow as represented in terms such as  $v^*T^*$ . However, in our model survey we have found little correspondence between stronger interannual variability in October Eurasian snow cover and the ability of a model to more realistically simulate the snow-circulation effect. Nevertheless, this remains one potential factor explaining what prevents the GCMs from capturing this effect.

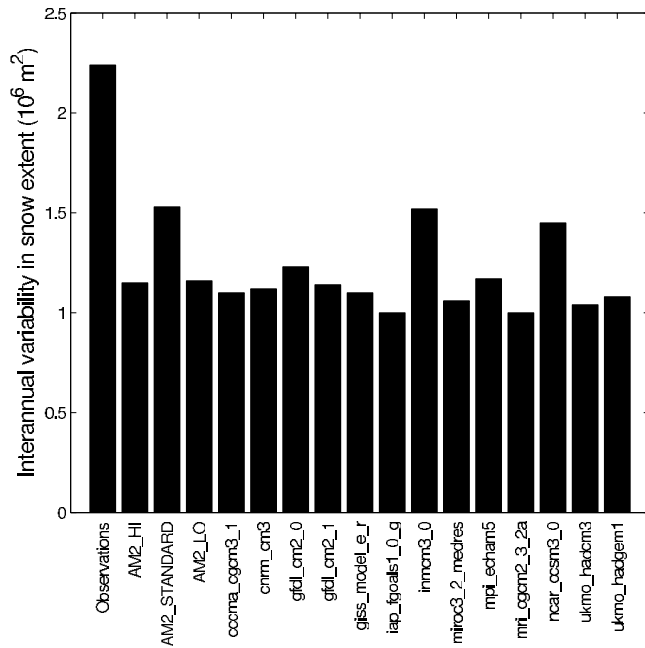
### 3.2. Atmospheric Wave Structure

[23] Another factor that might explain the discrepancy between observed and simulated snow-circulation connections is the character of the wave pattern associated with snow anomalies. Figure 5 shows regressions of October and December mean eddy GPH (m) at 60°N on the normalized snow index for: (a) and (d) the observations, (b) and (e) AM2\_HI, and (c) and (f) AM2\_STANDARD.

[24] The simulated October wave patterns show systematic similarities and differences to the observations. In October the regression in observations is significant over a wide range of longitudes, with maxima centered at 70E and 230E in the troposphere (Figure 5a). The regression in the models shows a similar general pattern but one that is more zonally localized, with maxima at 70E and 150E in AM2\_HI and 70E and 170E in AM2\_STANDARD (Figures 5a and 5c). AM2\_STANDARD is slightly closer to the observations, and also agrees better with the observations in the stratosphere. Both models capture the low GPH centered over the snow forcing region. Whether this



**Figure 3.** October mean snow depth (m) regressed on the normalized snow index for the (a) observations, (b) AM2\_HI, and (c) AM2\_STANDARD. The contour interval is 0.05 m with solid contours representing positive values and dashed contours representing negative values.

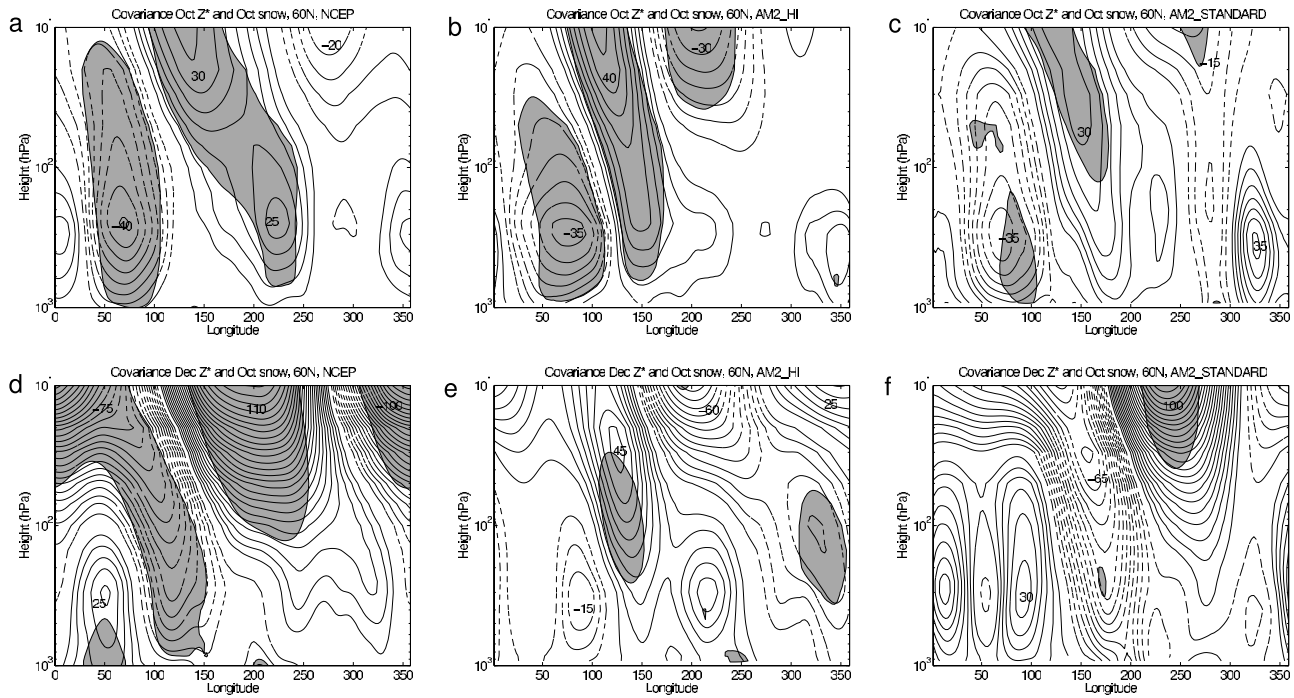


**Figure 4.** Standard deviation of interannual variability in the Eurasian October snow extent ( $10^6 \text{ m}^2$ ) for observations, AM2 models, and CMIP3 models.

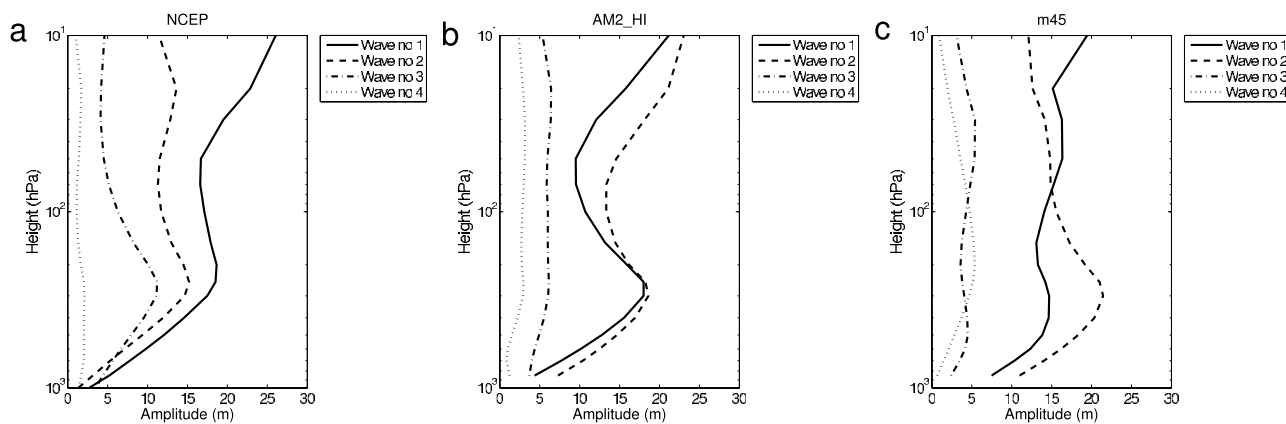
particular feature is a cause of the snow anomaly or a response to the snow anomaly is a potentially interesting dynamical question that is beyond the scope of our current study.

[25] For December, the regression in AM2\_STANDARD shows a high over the snow region which extends too far both horizontally and vertically and is not significant (Figures 5d and 5f). However, the regression is closer to observations than that in AM2\_HI which has the wrong sign (Figure 5e). Neither model captures very well the pattern seen in the observations – a high, confined to the troposphere, upstream of the region of snow cover, and a significant low extending down from the stratosphere, downstream of the region of snow cover. Our current dynamical understanding of the tropospheric response to snow cover is that snow-induced cooling locally decreases the thickness of isentropic layers in the troposphere, and that the upstream-high downstream-low anomaly pattern reflects potential vorticity conservation by eastward travelling fluid columns passing through this locally altered stratification [Fletcher et al., 2008]. The models appear to be capturing neither the correct tropospheric dynamical response nor the correct vertical propagation of planetary waves forced by Eurasian snow cover.

[26] The fact that the wave pattern in October is more longitudinally localized in the models than in the observations might be an important indicator of the subsequent circulation response. By the Charney-Drazin theory [e.g., Andrews et al., 1987] the horizontal length scale of planetary waves controls the extent to which they can propagate into the stratosphere. To further quantify the zonal structure of the wave response, Figure 6 shows the individual zonal wave number components of the regression of the snow index on October mean eddy GPH (corresponding to Figures 5a–5c). This was calculated by performing a Fourier decomposition on the regression pattern in Figure 5



**Figure 5.** Regressions of October and December mean eddy GPH (m) at 60°N on the normalized snow index for the (a and d) observations, (b and e) AM2\_HI, and (c and f) AM2\_STANDARD. The contour interval is 5 m with solid contours representing positive values, dashed contours representing negative values, and shaded regions representing statistical significance at the 95% level.



**Figure 6.** Fourier decomposition into individual wave number components of the regression of the snow index onto the October mean eddy GPH for the (a) observations, (b) AM2\_HI, and (c) AM2\_STANDARD. Figures correspond to Figures 4a–4c. The wave number 1 component is shown with a solid line, wave number 2 with a dashed line, wave number 3 with a dash-dotted line, and wave number 4 with a dotted line. There are only small contributions from wave number 5 and above so these are not shown.

at each vertical level. In both the observations and the models zonal wave numbers 1 and 2 dominate. However, there are some striking differences. In the observations wave number 1 dominates at all altitudes. In AM2\_STANDARD, wave number 1 dominates in the stratosphere and wave number 2 in the troposphere. In AM2\_HI wave number 2 dominates at all altitudes. Focusing on the region from 50 to 100 hPa (the lower stratosphere) the wave number 1 contribution is considerably smaller in AM2\_HI than in either observations or AM2\_STANDARD. Indeed, looking at the individual wave number components of the regression of snow index on November mean eddy GPH shows that in this month, from 50 to 100 hPa, wave numbers 1 and 2 dominate in the observations and AM2\_STANDARD, whereas wave number 3 dominates in AM2\_HI (not shown).

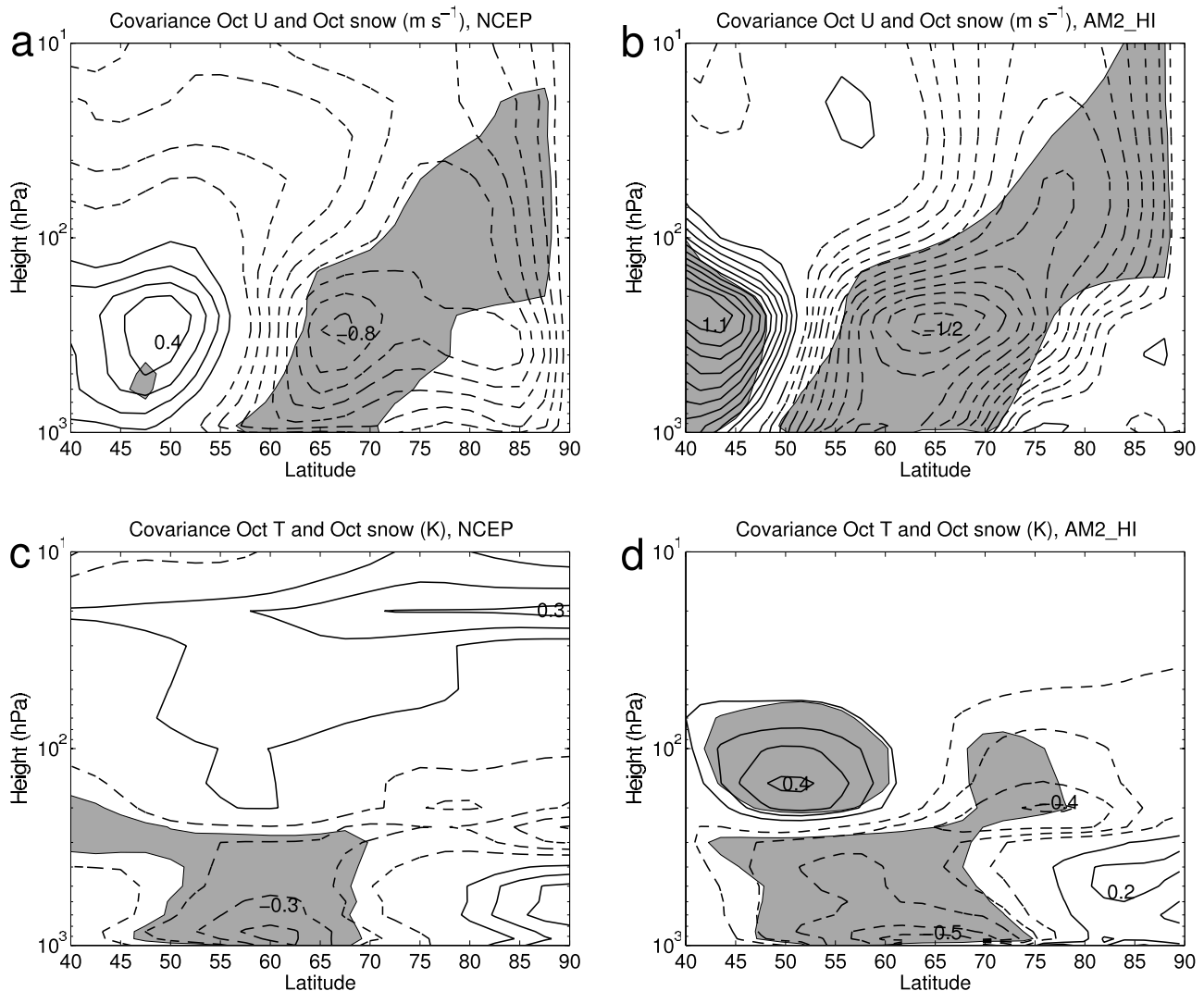
[27] The reduced wave number 1 amplitude of the snow-related eddy field reflects a general problem in the model climatological eddy field. For example, the wave number 1 GPH amplitude is smaller in AM2\_HI than in observations and in AM2\_STANDARD from 50 hPa to 100 hPa throughout October to December (not shown). It is not easy to diagnose why this might be the case. For example, no systematic wind biases exist that would explain the reduction in scale of the stationary eddies in the models. However, it is possible that changes to models that would improve the stationary eddy field would lead to an improved representation of the snow-circulation coupling.

[28] Another possible factor explaining the absence of the snow-circulation connection in models is that the tropospheric zonal-mean circulation variability associated with snow cover might be poorly simulated. This may affect the initial propagation of the wave pulse. Figure 7, which shows regressions of October mean zonal-mean zonal wind ( $U$ ) and temperature ( $T$ ) onto the snow index for observations and AM2\_HI, represents some of the analysis we have carried out along these lines. Regressions for AM2\_STANDARD are very similar to those for AM2\_HI and hence are not shown. The equatorward shift of the tropospheric jet associated with high snow cover is larger in AM2\_HI (Figure 7b) than in

observations (Figure 7a). Consistent with this, regressing October mean EP flux [Andrews *et al.*, 1987, p. 128] onto the snow index shows a greater poleward EP flux in the troposphere of the models than is seen in observations (not shown). Our interpretation is that the models initially respond too strongly to snow cover anomalies. There is also, in October, a large EP flux anomaly into the ground in the extratropics in AM2\_HI and AM2\_STANDARD and no flux into the stratosphere, whereas the observations show upward EP flux in the extratropics through the tropopause and into the upper stratosphere (not shown). This is consistent with the idea of tropospheric trapping of the Rossby pulse in the models and is evident from the sign of the correlations in Figure 1. It could be an important clue to the lack of influence of snow cover later in the year in the models. The temperature increase in the stratosphere in high snow years shown in the observations (Figure 7c), and not shown in AM2\_HI (Figure 7d), is evidence of this. This temperature increase in observations is not statistically significant in October, but becomes more pronounced and widespread throughout November and December (not shown). These differences aside, the structure of the regressions in the models (Figures 7b and 7d) look very similar to those in observations (Figures 7a and 7c) in regions of significance.

### 3.3. Stratosphere-Troposphere Coupling

[29] As mentioned above, the connection between October snow and winter circulation consists of two parts, one part involving the Rossby pulse response to snow and the second part involving eddy mean flow interactions in the stratosphere-troposphere system. Figure 8 demonstrates the second part in observations. It is produced similarly to Figures 1a and 1d except that monthly averaged  $v^*T^*$  and GPH are regressed against the principal component time series of the tropospheric NAM (the leading EOF of January sea level pressure) rather than against the snow index. Figures 8a and 8b show, consistently with Baldwin and Dunkerton [2001] and Polvani and Waugh



**Figure 7.** (a, b) Regressions of the October mean zonal-mean zonal wind ( $U$ ) onto the snow index for the observations and AM2\_HI, respectively. (c, d) The equivalent regressions of the October mean temperature ( $T$ ) onto the snow index. The contour interval in Figures 7a and 7b is  $0.1 \text{ m s}^{-1}$  and in Figures 7c and 7d is  $0.1 \text{ K}$ . Solid contours represent positive values, dashed contours represent negative values, and shaded regions represent statistical significance at the 95% level. Regressions for AM2\_STANDARD are very similar to those for AM2\_HI and hence are not shown.

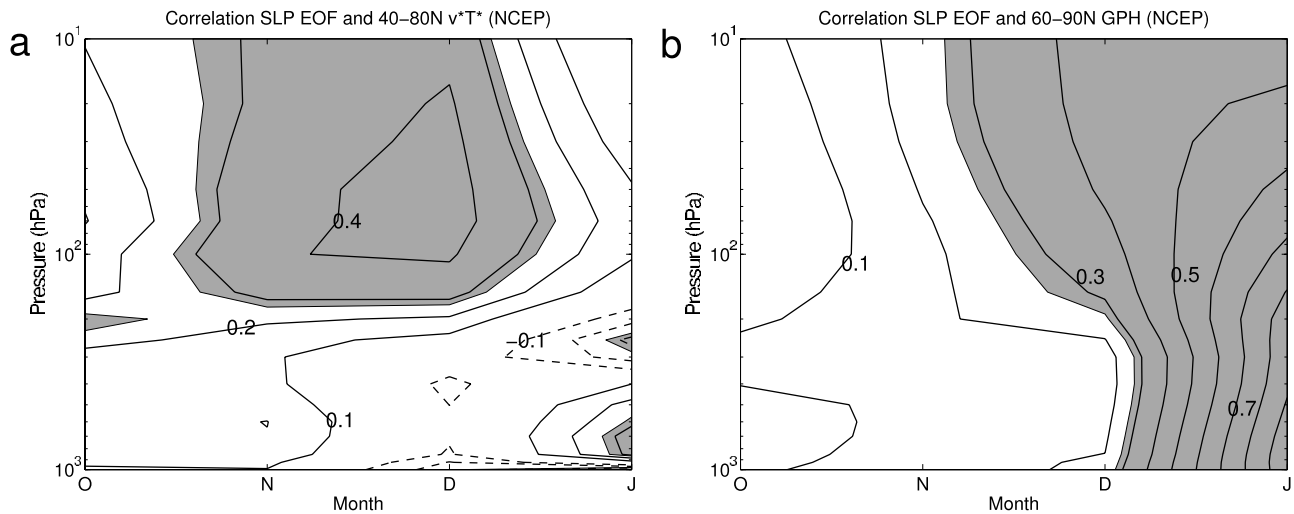
[2004], the statistically significant correlations between January sea level pressure and both (1) upward wave activity flux in the stratosphere in November and December and (2) GPH in the stratosphere in December and in the troposphere and stratosphere in January.

[30] The ability of the CMIP3 GCMs to capture this part of the mechanism is mixed. Figure 9 shows histograms of the correlation of the principal component time series of the tropospheric NAM to October, November and December  $v^*T^*$  at 100 hPa. It can be seen that at least one model matches the observations fairly closely, and that most models display a correlation of the correct sign throughout November–December. This demonstrates that the GCMs, although far from perfect, capture the second part of the mechanism proposed by Cohen *et al.* [2007] far better than the first. AM2\_STANDARD with a correlation of 0.5 in November but negative correlation in December, and

AM2\_HI with no correlation in November and a correlation of 0.1 in December, perform neither better nor worse than the CMIP3 models in this respect.

#### 4. Discussion and Conclusions

[31] In this study we have investigated the mechanism put forward by Cohen *et al.* [2007] by which Eurasian snow cover in October may influence winter climate in the Northern Hemisphere and, particularly, whether this mechanism is captured in state of the art GCMs. Studies in which snow forcing is prescribed in GCMs [Gong *et al.*, 2002, 2003; Fletcher *et al.*, 2007, 2008] have successfully simulated some features of the snow climate relationship. However, this study has shown that when snow is allowed to evolve freely in the GCMs, the simulated response to snow is not as observed. Although the correlation between strato-



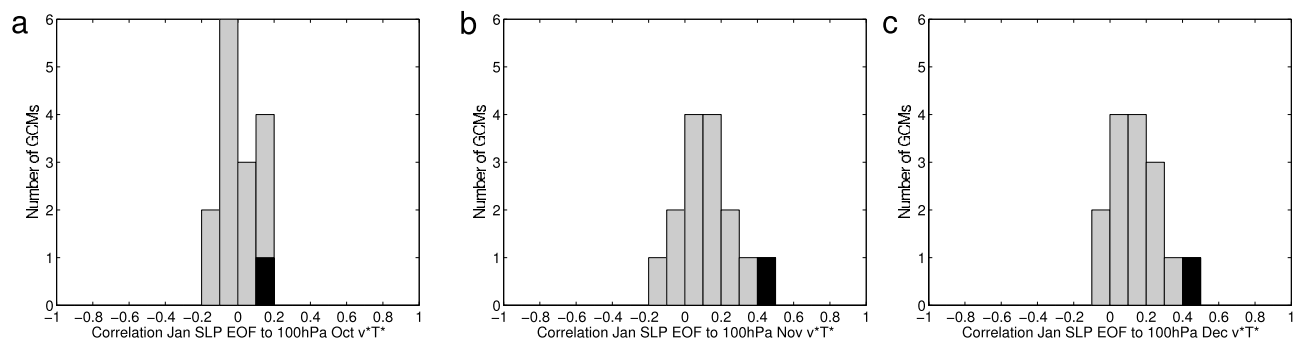
**Figure 8.** Figures 8a and 8b are as Figures 1a and 1d, respectively, except that the regressions are against the tropospheric NAM index as described in the text.

spheric planetary wave flux in December and tropospheric dynamical fields in January is represented in some GCMs, they all fail to capture the effects of October snow cover on these fields.

[32] The reasons why models fail to capture this mechanism turn out to be subtle. For example, we have found that the spatial pattern of the snow cover over Eurasia and regressions of the snow index onto zonal wind,  $U$ , and temperature,  $T$ , are reasonably well represented in the GFDL GCMs. However, the GCMs capture only about half of the observed interannual variability in snow cover which is likely to lead to a weaker response to snow in the dynamical fields. Further, regressing the October mean eddy geopotential height (GPH) at  $60^\circ\text{N}$  on the snow index in the GCMs reveals that the response to snow cover in the models is too zonally localized, perhaps because of a westward bias in the extratropical zonal-mean zonal wind climatology in October. Thus planetary waves forced by anomalously high snow cover may not propagate into the stratosphere as efficiently as they do in the observations. We therefore suggest the possibility that changes to models leading to an improved stationary eddy field would lead to an improved representation of the snow-circulation coupling.

[33] Figures 1a and 1d show the clear effect of Eurasian snow cover on winter climate. That GCMs do not capture this effect means that they are missing a potentially important aspect of winter climate variability. Thus we conclude that, although the failure of GCMs to capture the effects of snow cover is a difficult issue to solve, it is nonetheless an issue which it is important to be aware of.

[34] The work in this paper gives some indication of where the problems may lie. It is suggested that an improved interannual variability in Eurasian snow extent leading to increased variability in planetary waves over Eurasia which, as mentioned above, must have the correct zonal structure, would help to solve this issue. Improvements in the models' precipitation parameterizations and in the snow component of land surface schemes might address some of the problem of the simulation of Eurasian snow extent. Improvements in stratospheric representation might also improve the surface-to-stratosphere coupling. However, it is difficult to develop an effective strategy to improve the simulation of the planetary wave response to the snow variations; instead, such improvements would reflect overall improvements in climate model simulations of the general circulation.



**Figure 9.** Frequency distribution of the correlation of the tropospheric NAM index with (a) October, (b) November, and (c) December  $v^*T^*$  at 100 hPa for the 14 CMIP3 GCMs. The CMIP3 GCMs are shown by gray bars, and the observations are shown by black bars.



[35] We conclude by briefly pointing out some broader implications of this study. Capturing each step of the snow-circulation relationship, from the initial snow anomaly, to the lower tropospheric circulation response, to the vertically coupled stratosphere-troposphere response, represents a strong test of a model's ability to simulate a wealth of physical processes. The effect of Eurasian snow cover on winter climate is probably not the only subtle yet important relationship between the cryosphere and the atmospheric general circulation not captured in current generation climate models. Careful consideration of why such observed relationships do not show up in GCMs will be essential to improve forecasts on seasonal and climate time scales.

[36] **Acknowledgments.** The authors wish to thank Chris Fletcher for his many insightful comments on this work. S.C.H. was funded by the Natural Sciences and Engineering Research Council of Canada. J.C. is supported by the National Science Foundation grant ATM-0443512. Thanks are also due to NOAA/GFDL for the model code and the AM2/STANDARD model output, and to Sabine Stanley and the Canadian Institute for Theoretical Astrophysics for providing computational resources. P.J.K. acknowledges the support of the Kavli Institute for Theoretical Physics. This research was supported in part by the National Science Foundation under grant PHY05-51164.

## References

- Alexander, M. J., and T. J. Dunkerton (1999), A spectral parameterization of mean-flow forcing due to breaking gravity waves, *J. Atmos. Sci.*, *56*(24), 4167–4182.
- Anderson, J. L., et al. (2004), The new GFDL global atmosphere and land model AM2-LM2: Evaluation with prescribed SST simulations, *J. Climate*, *17*(24), 4641–4673.
- Andrews, D. G., J. R. Holton, and C. B. Leovy (1987), *Middle Atmosphere Dynamics*, 489 pp., Elsevier, New York.
- Baldwin, M. P., and T. J. Dunkerton (2001), Stratospheric harbingers of anomalous weather regimes, *Science*, *294*, 581–584.
- Brown, R. D. (2000), Northern Hemisphere snow cover variability and change, 1915–97, *J. Climate*, *13*(13), 2339–2355.
- Charlton, A. J., A. Oneill, W. A. Lahoz, and A. C. Massacand (2004), Sensitivity of tropospheric forecasts to stratospheric initial conditions, *Q. J. Royal. Meteorol. Soc.*, *130*(6), 1771–1792.
- Cohen, J., and M. Barlow (2005), The NAO, the AO, and global warming: How closely related?, *J. Climate*, *18*(21), 4498–4513.
- Cohen, J., and D. Entekhabi (1999), Eurasian snow cover variability and Northern Hemisphere climate predictability, *Geophys. Res. Lett.*, *26*(3), 345–348.
- Cohen, J., and D. Entekhabi (2001), The influence of snow cover on Northern Hemisphere climate variability, *Atmos.-Ocean*, *39*, 35–53.
- Cohen, J., and C. Fletcher (2007), Improved skill of Northern Hemisphere winter surface temperature predictions based on land-atmosphere fall anomalies, *J. Climate*, *20*(16), 4118–4132.
- Cohen, J., K. Saito, and D. Entekhabi (2001), The role of the Siberian High in Northern Hemisphere climate variability, *Geophys. Res. Lett.*, *28*(2), 299–302.
- Cohen, J., M. Barlow, P. J. Kushner, and K. Saito (2007), Stratosphere-troposphere coupling and links with Eurasian land surface variability, *J. Climate*, *20*(21), 5335–5343.
- Delworth, T. L., et al. (2006), GFDL's CM2 global coupled climate models. part I: Formulation and simulation characteristics, *J. Climate*, *19*(5), 643–674.
- Fletcher, C. G., P. J. Kushner, and J. Cohen (2007), Stratospheric control of the extratropical circulation response to surface forcing, *Geophys. Res. Lett.*, *34*, L21802, doi:10.1029/2007GL031626.
- Fletcher, C. G., S. C. Hardiman, P. J. Kushner, and J. Cohen (2008), The dynamical response to snow cover perturbations in a large ensemble of atmospheric GCM integrations, *J. Climate*, in press.
- Frei, A., J. Miller, and D. A. Robinson (2003), Improved simulations of snow extent in the second phase of the Atmospheric Model Intercomparison Project (AMIP-2), *J. Geophys. Res.*, *108*(D12), 4369, doi:10.1029/2002JD003030.
- Gillett, N. P., and D. W. J. Thompson (2003), Simulation of recent Southern Hemisphere climate change, *Science*, *302*(5643), 273–275, doi:10.1126/science.1087440.
- Gong, G., D. Entekhabi, and J. Cohen (2002), A large-ensemble model study of the wintertime AO-NAO and the role of interannual snow perturbations, *J. Climate*, *15*, 3488–3499.
- Gong, G., D. Entekhabi, and J. Cohen (2003), Modeled Northern Hemisphere winter climate response to realistic Siberian snow anomalies, *J. Climate*, *16*(23), 3917–3931.
- Gong, G., D. Entekhabi, J. Cohen, and D. Robinson (2004), Sensitivity of atmospheric response to modeled snow anomaly characteristics, *J. Geophys. Res.*, *109*, D06107, doi:10.1029/2003JD004160.
- Hall, A., and X. Qu (2006), Using the current seasonal cycle to constrain snow albedo feedback in future climate change, *Geophys. Res. Lett.*, *33*, L03502, doi:10.1029/2005GL025127.
- Loth, B., and H.-F. Graf (1998), Modeling the snow cover in climate studies: 2. The sensitivity to internal snow parameters and interface processes, *J. Geophys. Res.*, *103*(D10), 11,329–11,340.
- Polvani, L. M., and D. W. Waugh (2004), Upward wave activity flux as a precursor to extreme stratospheric events and subsequent anomalous surface weather regimes, *J. Climate*, *17*(18), 3548–3554.
- Robinson, D. A., K. F. Dewey, and R. R. Heim (1993), Global snow cover monitoring: An update, *Bull. Amer. Meteorol. Soc.*, *74*(9), 1689–1696.
- Saito, K., J. Cohen, and D. Entekhabi (2001), Evolution of atmospheric response to early-season Eurasian snow cover anomalies, *Mon. Weather Rev.*, *129*(11), 2746–2760.
- Slater, A. G., A. J. Pitman, and C. E. Desborough (1998), The validation of a snow parameterization designed for use in general circulation models, *Int. J. Climatol.*, *18*, 595–617.

J. Cohen, Seasonal Forecasting, AER, Inc., 131 Hartwell Avenue, Lexington, MA 02421, USA.

S. C. Hardiman, Met Office, FitzRoy Road, Exeter, Devon EX1 3PB, UK. (steven.hardiman@metoffice.gov.uk)

P. J. Kushner, Department of Physics, University of Toronto, 60 St. George St., Toronto, ON M5S 1A7, Canada.