Circulation Response to Eurasian versus North American Anomalous Snow Scenarios in the Northern Hemisphere with an AGCM Coupled to a Slab Ocean Model

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ABSTRACT

The difference between snow-covered and snow-free conditions is the most climatically significant natural seasonal change the land surface can experience. Most GCM studies investigating snow–atmosphere interactions have focused on impacts of Eurasian snow anomalies caused by the magnitude of snow mass, while North American snow has been shown to have a weaker relationship with downstream climate. Experiment design of recent snow–atmosphere interactions studies has been limited to atmosphere-only models, with sea surface temperature (SST) and sea ice extent represented as boundary conditions. The authors explore the circulation response to anomalous snow scenarios, for both North America and Eurasia, using a slab ocean model. Surface response include significant SST cooling directly downstream of each individual forcing region in addition to upstream centers of remote cooling under maximum snow conditions. Atmospheric response to anomalous snow conditions is consistent through multiple levels in the lower troposphere under maximum snow conditions throughout much of the midlatitudes in both experiments during early winter. Areas of strengthened midtropospheric eddy kinetic energy correlate well with steep geopotential height gradient differences and increased zonal wind at 250 hPa over the western Pacific. Both experiments show similar atmospheric response pathways; however, circulation response to maximum Eurasian snow is focused downstream in early winter, whereas upstream response is particularly evident from the North American experiment. This paper focuses on differences as a result of Eurasian versus North American snow forcing in atmospheric circulation response using an AGCM with a slab ocean model.

1. Introduction

a. Previous work: Snow–atmosphere interactions

Snow cover strongly influences the terrestrial energy balance, most obviously via increased surface albedo but also with increased thermal emissivity, with increased insulation leading to decreased soil heat flux, and as a latent heat sink (Wagner 1973; Robinson and Kukla 1984; Walland and Simmonds 1997; Cohen and Rind 1991). Snow cover is also the frozen storage term in the terrestrial water balance (Derkson and LeDrew 2000). These effects make snow cover a major factor in modulating climate variability and change (Scialdone and Robock 1987).

Regional cooling over snow could potentially create atmospheric effects that propagate to other regions through atmospheric teleconnections (Namias 1985), but the nature of snow–atmosphere interactions cannot be completely elucidated using observed data (Cohen and Entekhabi 2001). Therefore, numerical models have been the primary tool for exploring interactions between snow cover and the atmosphere.

Early GCM studies investigating the influence of anomalous snow cover on the atmosphere, using extreme snow extent scenarios to compensate for resolution limitations, showed an expected surface temperature decrease over the forcing region (Walsh and Ross 1988; Walland and Simmonds 1997). Significant remote anomalies included increased surface air temperature (SAT) and changes in observed storm tracks across the North
Atlantic and in the northeast Pacific. A study using a coupled ocean–atmosphere model linked increased snow cover over Eurasia to a decrease in precipitation over Southeast Asia in spring and early summer and showed that amplification of a snow-induced perturbation could potentially serve as a trigger of an El Niño–Southern Oscillation event (Barnett et al. 1989).

More recently, Allen and Zender (2010) found an initial though with a weaker feedback than found in Eurasia. Cover modestly strengthened the prevailing stationary North American snow forcing has also been modeled. Its vast area produces a significant model response, or freely evolving snow conditions. Added snow cover over western Siberia was negatively correlated with the leading empirical orthogonal function (EOF) of sea level pressure in the North Atlantic, implying that excess snow over Siberia could lead to higher pressure over the North Atlantic and a subsequent negative North Atlantic Oscillation (NAO) phase. Gong et al. (2002) modeled ensembles of autumn/winter climate in which one set responded to a fixed snow climatology and the other set freely modeled its snow conditions. Added snow cover over western Siberia was negatively correlated with the leading empirical orthogonal function (EOF) of sea level pressure in the North Atlantic, implying that excess snow over Siberia could lead to higher pressure over the North Atlantic and a subsequent negative North Atlantic Oscillation (NAO) phase. Gong et al. (2003) further proposed that a positive feedback of vertical propagation of stationary waves over Siberia could weaken the polar vortex in the stratosphere, resulting in a negative Arctic Oscillation (AO) anomaly in the troposphere. The certainty of this mechanism was limited by vertical resolution in the models stratosphere, with only 4 of 19 layers above 100 hPa. The ability of an atmospheric GCM to simulate snow–atmospheric interactions will thus depend on its coupling of the troposphere and stratosphere (Hardiman et al. 2008; Fletcher et al. 2009), as well as deficiencies of snow parameterization. This latter factor was identified in a study by Hardiman et al. (2008) and also in a recent study by Allen and Zender (2011), which found that model underestimation of Eurasian snow extent and variability was likely a major factor and the use of a satellite-based snow cover resulted in an improved AO simulation over the latter part of the twentieth century. Based on such results, it is clear that reproduction of snow–AO relationships in idealized experiments is possible, but proves extremely difficult when using realistic or freely evolving snow conditions.

While snow in Eurasia has been most studied because its vast area produces a significant model response, North American snow forcing has also been modeled. Gong et al. (2003) showed that North American snow cover modestly strengthened the prevailing stationary wave fluxes throughout the Northern Hemisphere, although with a weaker feedback than found in Eurasia. More recently, Allen and Zender (2010) found an initial negative AO response to North American snow forcing, which gave way to a positive AO response in late winter/early spring. Anomalous North American snow cover and atmospheric teleconnections have been linked via enhanced North Atlantic storm-track activity, although the relationship depended on seasonal timing and the persistence of snow (Sobolowski et al. 2010). On a regional scale, Klingaman et al. (2008) found that forced snow cover in the northern Great Plains of the United States modified downstream tropospheric circulation, producing a positive NAO phase. This response was much stronger in January and February than earlier in the season.

b. Snow and the NAM

Numerous investigations connect snow cover variability and low-frequency atmospheric circulation patterns in the Northern Hemisphere. The northern annular mode (NAM) refers to the leading mode of wintertime atmospheric variability in the Northern Hemisphere. Throughout most of the twentieth century, the NAM was exclusively referred to as the NAO, which focused on describing atmospheric variability in the Atlantic sector. Thompson and Wallace (1998) described the leading EOF of wintertime sea level pressure as having a pan-Arctic center of action, which they referred to as the AO. In recent years, the AO/NAO nomenclature has been mostly replaced by the NAM, partly because the term “oscillation” implies regular fluctuations in time, which does not necessarily occur (Baldwin and Thompson 2009). Within this study however, we will refer to the original modal descriptions (AO/NAO), as their regional specificity aids in dynamical description of circulation phenomena.

Numerous GCM studies have sought to capture the recent variability in the AO/NAO through simulations bounded by historical sea surface temperatures (SSTs), with varying degrees of success. The phase and approximately 50% of the amplitude of the long-term variability in wintertime NAO could be accounted for in this way (Rodwell et al. 1999; Mehta et al. 2000; Hoerling et al. 2001). The location of oceanic influence on NAO variability, however, is not clear, as it has been demonstrated that both extratropical and tropical SSTs play a role. However, this can lead to the circular argument as the AO/NAO itself is the dominant driver of upper-ocean thermal anomalies over the extratropical North Atlantic (Deser and Timlin 1997; Visbeck et al. 2003). Greater understanding of causality in midlatitude air–sea interaction is necessary to the understanding of atmospheric predictability as the life cycle of thermal anomalies in the ocean mixed layer is on the order of several months to a year (Frankignoul 1985).
Snow cover, sea ice, and their potential impact on Northern Hemisphere atmospheric variability are understudied, having been overshadowed by oceanic forcings (Hurrell et al. 2006). As discussed above, changes of snow extent on land have been linked to subsequent shifts in AO/NAO (Wantanabe and Nitta 1998; Gong et al. 2002; Klingaman et al. 2008). Changes in SST and sea ice could produce changes of sensible and latent heat fluxes greater than the changes produced by adding or subtracting snow cover over land. Therefore, the oceanic response to land-based snow cover changes contains an enormous potential feedback that has not been previously explored. This points to the need for coupled ocean–atmosphere model experiments to clarify the different roles of the mechanisms affecting the AO/NAO interannual variability.

The effects of SST and snow cover boundary conditions in Atmospheric Model Intercomparison Project, version 2 (AMIP-2), on the evolution of the NAO were explored by Cohen et al. (2005). Although the models used in AMIP-2 reproduced a stable NAO pattern of variability, model-simulated NAO indices were statistically insignificant and inconsistent from model to model. The influence of forced SSTs on the phase of the NAO for an interannual time scale proved insignificant compared to the influence of stochastic variations (Cohen et al. 2005).

c. The present study

Previous modeling studies using GCMs have investigated the potential impacts of anomalous snow cover on the atmosphere and atmospheric teleconnection modes. However, these studies have exclusively been carried out using a data ocean model, in which SSTs are forced by historical average of the seasonal cycle and cannot be modified by atmospheric processes. The data ocean model creates an infinite sink for heat energy and eliminates feedback over ocean surfaces. Another shortcoming of the static ocean surface is that cause and effect relationships between climate variables can only be inferred. Furthermore, studies investigating the influence of SST on atmospheric circulation have shown statistically significant lagged relationships between large-scale SST anomalies and subsequent midtropospheric anomalies in the North Atlantic (Czaja and Frankignoul 1999). Significant lagged relationships between North American snow cover and depth and the NAO were found, where the duration of snow cover was found to be an important factor (D. Kluever and D. Leathers 2008, unpublished manuscript). The locations of significant lagged relationships occurred along the U.S./Canada border and in some cases the northeastern United States and Great Lakes region.

This study investigates the impact of realistic snow perturbations over North America and Eurasia and the subsequent atmospheric circulation pattern response. An interactive limited-layer slab ocean model is used to enable energy exchange over ocean surfaces, using model-predicted SSTs and sea ice fraction rather than prescribed values. Realistic snow extent and depth anomalies are based on historical observations, satellite, and gridded data. A comprehensive study on the differences between the inclusion of the limited-layer slab ocean model versus the traditional experiment design using a data ocean model has been completed (G. R. Henderson et al. 2010, unpublished manuscript), showing that the magnitude of snow cover responses is generally larger and of greater statistical significance on inclusion of the slab model.

2. Experiment design

a. The Slab Ocean Model

These experiments were carried out using three modules—land, atmosphere, and ocean—of the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 3.1 (CCSM3.1) at T42 resolution. In its stand-alone mode, the Community Atmosphere Model (CAM) is integrated together with the Community Land Model (CLM) and a thermodynamic sea ice model [NCAR’s CSM Sea Ice Model (CSIM)], and has the option for either a static data ocean or an interactive slab ocean model. This study uses the Slab Ocean Model (SOM), where ocean mixed-layer temperature is predicted rather than prescribed.

In the SOM, seasonal deep water exchange and horizontal ocean heat transport are simulated by an internal heat source $Q$, which is estimated from a control run. Deep-water energy exchange in the SOM is emulated both spatially and temporally throughout the annual cycle from monthly values of $Q$, which are linearly interpolated in time between prescribed mid-monthly values. Specific mixed-layer depths and seasonally and geographically varying ocean heat fluxes factor into the calculation of SST, resulting in the mixed-layer temperature output variable $T_o$. Sea ice fraction and depth are predicted rather than prescribed from a boundary dataset.

Considerable spin-up time is required when using the SOM. A 60-yr control run was performed to reach model equilibrium. As with fully coupled atmosphere–ocean models (e.g., Meehl et al. 2012), a final equilibrium is not necessarily reached. However, as shown by calculation of net top-of-atmosphere (TOA) radiation of 2 W m$^{-2}$ and SST drift of 0.05 K yr$^{-1}$, our simulations
are stable enough after 20 yr to make fair comparisons between different runs. Snow cover and depth were model predicted in the SOM control run. Upon comparison of the SOM control run snow field with that produced by running CAM with the Data Ocean Model (DOM), no significant differences emerged between the two resulting model snow climatologies.

b. Anomalous snow grid production

To force an anomalous snow scenario within the CLM, snow depth, snow water equivalent (SWE), and snow age are treated as boundary conditions. Snow depth and SWE were calculated based on observed snow data, while snow age values were calculated from a model control run.

Eurasian snow depth data are based on the National Snow and Ice Data Center (NSIDC) Historical Soviet Daily Snow Depth (HSBSD) product. This product spans from 1881 to 1995 and is based on observations made at 284 World Meteorological Organization stations throughout Russia and the former Soviet Union (Ye et al. 1998). The original data were interpolated to a regular 2.5° × 2.5° grid using a distance-weighted algorithm developed by Willmott et al. (1984). The North American data were based on a 1° × 1° interpolated snow depth dataset (Dyer and Mote 2006) from U.S. National Weather Service (NWS) cooperative stations and the Canadian daily surface observations. The interpolation was completed using the same distance-weighted algorithm (Willmott et al. 1984; Shepard 1968) and quality controlled using criteria from Robinson (1989). The period of record was 1900–2000 with daily resolution; however, prior to 1950, the number of reporting stations is greatly diminished.

In an effort for consistency and in recognition of sparse early station records in both series, anomalous observed snow depths were limited to the period 1967–95. This overlaps with the National Oceanic and Atmospheric Administration (NOAA) Northern Hemisphere satellite snow cover extent product period of record, allowing for validation of snow extent fields from a third data source. Depth data at monthly resolution were considered for creation of anomalous snow grid production over both the North American and Eurasian continents.

Both observed snow datasets were regridded to a T42 Gaussian grid using the International Mathematical Subroutine Library (IMSL) SURF implementation of the Akima (1978) bivariate interpolation algorithm for irregularly distributed data points. For each grid box, the five largest and smallest snow depth values for each month during each dataset’s time period were identified. Extreme maximum/minimum snow depth conditions for Eurasia and North America were calculated by taking the means of the five largest and five smallest depth values at each grid box for each month. Extreme depths were then expressed as anomalies from each record’s mean monthly grid value, creating a field of snow depth offsets from the mean depth field. Four different snow scenarios were created using this method: extreme maximum/minimum Eurasian snow depth and the same for North America. Each perturbed snow scenario was simulated in a 40-yr realization that was branched off the coupled control run at approximately year 20. Each of the 40 years was considered independent, where the snow forcing was constant from year to year but varies monthly from September through May and reverts to a model climatology during summer months. This experiment design is of similar structure to previous modeling studies using NCAR’s model (Alexander et al. 2010; Lawrence and Slater 2009).

A global snow depth and SWE monthly climatology was produced from a 200-yr model control run. This model run was performed at NCAR using the CAM coupled to a DOM (NCAR 2011). This monthly climatology was used as the baseline from which to offset snow values, where maximum or minimum depth offset anomalies described previously were either added or subtracted from the monthly model climatology. For each of the four snow scenarios, snow depth was altered over Eurasia or North America only, leaving snow depth over the remaining land areas set at the model climatology. Depths were only perturbed from September through May, leaving summer months set to the model climatology.

SWE for any perturbed snow depth grid box was calculated using a SWE/depth ratio. A typical SWE/snow depth ratio is 1 part liquid equivalent to 10 parts snow depth; however, this ratio varies greatly with snow age, density, temperature, and numerous other variables. The SWE/snow depth ratio applied to any perturbed grid box was based on the calculated ratio of snow depth to SWE values specific to each grid box, estimated from the 200-yr NCAR model climatology. This allows for both spatial and seasonal consideration to factor into the SWE/depth ratio of each grid box. As with snow depth values, SWE values remained set to the model climatology for grid boxes that were not perturbed.

Figure 1 depicts snow depth differences between maximum and minimum scenarios used for both the Eurasian (EUR) and North American (NA) experiments, from November to March. Snow area and depth difference between extreme scenarios over Eurasia are concentrated on the southern boundary of the ephemeral snow line throughout the winter season. In contrast, differences between maximum and minimum prescribed snow depth occur from October through March throughout North America, with some anomalies persisting into
May (not shown) in the northern high-latitude, high-elevation regions. The difference in snow depth ranges from 0 to 0.7 m, with the largest anomalous snow occurring over Alaska and northeast North America.

Regression coefficients of monthly SST against time over the 40-yr model integration were calculated and mapped. On average over the world oceans, SSTs cool slightly over the simulation period. However, the pattern of cooling and warming areas does not coincide with the pattern of SST responses to snow forcing, leading us to conclude that our snow cover forcing results are not a result of model drift. Similar trend analysis was conducted on midtropospheric height fields to ascertain if atmospheric response is influenced by simulation trends or model drift over the 40-yr integration period. Again, little to no coherence in sign or location of trends versus height differences as a result of extreme snow forcing was evident, suggesting that the atmosphere is responding to forced snow scenarios independent of SST trends.

3. Results

All results are expressed as difference fields, maximum snow scenario minus minimum scenario, for both EUR and NA experiments. Values were tested for significance using a Student’s t test and unless otherwise stated only values significant at the 95% level are shown.

a. Surface and low cloud response

The resulting surface energy flux and low cloud differences resulting from the EUR and NA snow scenarios are shown in Fig. 2. Values are averaged temporally over each experiment’s land area only (10°E–180°E for Eurasia and 165°W–55°W for North America). Nonsolar energy flux (Fig. 2, top) represents the sum of the sensible and latent heat fluxes and longwave radiation and is positive upward. Net shortwave energy flux differences (Fig. 2, middle) are shown positive downward. Both nonsolar heat and net shortwave energy fluxes are significantly weakened (<0) under maximum snow scenarios over both continents. This decrease is most intense at the beginning and end of each snow season, most likely a factor of minimum solar zenith angles at this time within the seasonal cycle. Increased surface albedo (not shown) and the insulating effect of snow are most likely factoring into the reduction of absorbed net shortwave radiation and surface heat flux, respectively; however, atmospheric response in the form of cloud coverage is also a factor. Low cloud fraction, from the surface to 700 hPa (Fig. 2, bottom), is greatly enhanced across the majority of the winter season in the EUR experiment, whereas it is more focused in the midwinter for the NA scenario. This increase in low clouds over greater snow extent conditions was noted by Alexander et al. (2010) also, but exact causes of this cloud pattern are unclear.

Response of SAT at 2 m was calculated for both EUR and NA experiment runs (Fig. 3). The anomalous snow during the maximum experiments cools the overlying atmosphere in winter in both experiments. Temperature depressions of up to −6°C are evident along the southern snow boundary across the Eurasian continent (Fig. 3, top), intensifying throughout the winter season. Surface temperature response is more modest for the NA
experiment (Fig. 3, bottom), but temperature depressions are still evident in central and southern North America, peaking in March–May (MAM) with values of up to $-4^\circ$C. Modification of air masses by increased snow to the north is most likely due to a combination of factors such as a potential increase in soil moisture as a result of greater snow extent and proximity from coastal areas. This magnitude increase in maximum temperature...

FIG. 2. (top) Nonsolar surface energy flux, (middle) net shortwave energy flux, and (bottom) fraction of low cloud at the level from the surface to approximately 700 hPa. These are averaged over each experiment’s land area: (left) 10°E–180° for Eurasia and (right) 165°–55°W for North America. Values are expressed as differences, maximum minus minimum snow scenarios, for each of the North American and Eurasian experiments. Only values significant at the 95% level are shown.
depressions as a result of surface snow cover has been identified in southern regions in particular (Leathers et al. 1995). Greatest latitudinal extent in SAT difference is evident from the EUR zonally averaged SAT response in later winter/early spring (Fig. 3, top, far right).

b. SST response

If we consider the SST response to anomalous snow forcing shown for the snow season as 3-month averages (Fig. 4), both experiments display evidence of local surface cooling. SST cooling is present in the western Pacific throughout coastal regions adjacent to the forcing region for the EUR experiment (Fig. 4, top) and in the Gulf of Mexico and along the east coast of North America for the NA experiment (Fig. 4, bottom). In these coastal regions, SST depressions are present in October and intensify through the winter season, ranging from $-1^\circ$ to $-2^\circ$C. This implies colder SST values are associated with maximum snow depth and extent. Both experiments exhibit upstream centers of remote cooling: Pacific cooling in the North American experiment and Atlantic cooling in the Eurasian experiment. This upstream response is preserved throughout the winter season and varies from $-0.4^\circ$ to $-1.4^\circ$C, spanning the North Atlantic from eastern North America to the coast of western Europe in the EUR experiment and centering around the Aleutian low region for the NA experiment. Local and downstream cooling of SST as a result of excess snow mass over each forcing region is intuitive, but upstream centers of remote cooling require more analysis.

To investigate further the seasonal nature of this response, zonally averaged SST differences across each ocean basin are shown in Fig. 5. In each experiment, the basin immediately downstream of the anomalous snow forcing region produces similarly shaped zonal averages through time. In particular, SST anomalies are greatest in early and late winter throughout the ocean basins downstream of the snow forcing: the Pacific in the case of the EUR experiment and the Atlantic in the case of the NA experiment.

c. Atmospheric response

We have clearly established a surface response to snow cover forcing and now will investigate the influence of snow forcing on the lower atmosphere. Presence of snow cools the overlying atmosphere relative to snow-free land surfaces, leading to higher surface pressures over the southern snow margins. Organized patterns of remote pressure response occur from December to March in the EUR experiment, focused upstream of the forcing region but coherent with the SST response described previously. This upstream anomaly persists through January and is joined by an area of significant positive anomalous sea level pressure ranging from 1 to 6 hPa over the high latitudes (not shown). In contrast,
negative pressure anomalies in excess of 5 hPa are present around the Aleutian low region, to the south and west of Alaska, in the NA experiment. This anomaly persists with lesser magnitude through March.

When considering the midtropospheric geopotential height response to excess EUR snow scenarios (Fig. 6, top), lower heights dominate the Northern Hemisphere winter at this level as well as throughout the midtroposphere (other pressure levels not shown). Centers of height anomalies are again focused upstream of the snow cover forcing region, with difference values up to 40 gpm over the North Atlantic during spring. For the NA experiment, however (Fig. 6, bottom), negative height anomalies are consistently present upstream of the snow forcing region in the vicinity of the North Pacific and Aleutian low. In this case, height anomalies are most significant in early (across the North Pacific) and late winter (across the North Atlantic), ranging from 10 to 50 gpm. In both experiments, persistent lower heights in late winter and spring are present over each forcing region, suggesting that surface forcing not only affects overlying SAT but also extends vertically into the lower atmosphere, as indicated by persistent lower geopotential heights under maximum snow conditions.

Remote responses, particularly upstream responses, to anomalous snow extent and depth forcings warranted investigation using alternative metrics. To capture sub-monthly transient response to snow forcing, eddy kinetic energy (EKE) at 500 hPa was analyzed. Transient eddy response describes the redistribution of heat and vorticity in a systematic fashion (Lau and Holopainen 1984), where eddies are relatively small in low and high latitudes but are more important in the midlatitudes, having maximum values near the tropopause during the winter season. In the Northern Hemisphere, large values of EKE coincide with regions of developing and mature mid-latitude depressions to the western edge of the Pacific and Atlantic Ocean basins (James 1995).

Calculation of EKE from daily model output used:

\[
EKE = \frac{1}{2} |\mathbf{\omega}|^2, \tag{1}
\]
where \( \mathbf{u}' = (u', v') \) represents daily perturbations from monthly-mean zonal and meridional wind fields \( (u = \text{zonal wind}) \) and \( \langle \rangle = T^{-1} \int_0^T \langle \rangle \, dt \) is the average over the time interval, taken to be \( T = 1 \) month.

Figure 7 displays differences in EKE at 500 hPa for maximum minus minimum forced snow scenarios for both experiments from December through March. Significant transient eddy differences are present during midwinter months in both experiments; however, contrasting regions of maximum response are evident. Overall, transient eddy activity as a result of EUR snow forcing is focused downstream of the forcing region (Fig. 7, top). In January, strengthened EKE (10–30 m\(^2\) s\(^{-2}\)) is present to the east of Eurasia, extending across the length of the Pacific. On comparison of strengthened EKE locations to those of geopotential height differences (Fig. 6), eddy response is collocated with substantial height gradients, which would likely be accompanied by a comparable temperature gradient. Such a robust response suggests a zone of increased baroclinicity, leading to a strengthening of the East Asia jet under maximum snow conditions. Positive EKE values across the Pacific during midwinter months suggest an excitement of the East Asia jet associated with excess snow mass over Eurasia, contributing to the positive EKE response and subsequent strengthening of the associated storm track throughout the region. A possible southward shift in the storm track of this region is indicated from a north–south dipole arrangement of EKE anomalies in January and February. Isolated areas of EKE response are present over the North Atlantic; however, these patches are sporadic.

In the NA experiment during early winter (Fig. 7, bottom), a dipole of EKE response with negative departures to the north (\(-10 \text{ m}^2\text{s}^{-2}\)) and larger positive values to the south (10–30 \text{ m}^2\text{s}^{-2}\) under maximum snow forcing is present in the western Pacific. This area of increased EKE is collocated with an area of reduced geopotential height at 500 hPa (Fig. 6, bottom), suggesting that an area of increased baroclinicity to the

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**Fig. 5.** Difference in zonal averages of SST over time, averaged over the (left) Pacific and (right) Atlantic basins for the (top) Eurasian and (bottom) North American experiments.
south results from a steeper geopotential gradient and associated temperature gradients. The dipole pattern evident from this early winter EKE response indicates that NA snow forcing shifts storm tracks southward over this region, potentially contributing to decreased warm SST advection from the south and strengthening the negative SST anomaly upstream of the snow forcing previously documented. As the storm track shifts south, the trough off of Eurasia deepens along the baroclinic zone, which in turn could cause a positive feedback to both the sea surface and the atmosphere through greater SST cooling and subsequent lower geopotential heights within the winter season.

The implied link between increased baroclinicity, strengthening of the upper-level jet, and subsequent increase in transient eddy activity resulting from forced snow scenarios was further investigated by considering zonal winds at 250 hPa. The 250-hPa level was chosen as an indicator of jet stream location and strength during the winter season.

When considering the upper-level zonal wind response under EUR snow forcing (Fig. 8, top), increased zonal flow at 250 hPa (1–5 m s\(^{-1}\)) is evident throughout the midlatitudes, in mid-Pacific and mid-Atlantic regions. Downstream increases in zonal wind under maximum EUR snow conditions are evident in late winter and early spring. This downstream intensification of zonal wind correlates well with the increased gradient in geopotential height (Fig. 6, top). A southward shift of zonal winds is suggested under maximum EUR snow conditions from January through March, by a north–south dipole arrangement of wind anomalies over the Pacific, which would consequently enhance lower atmospheric storm tracks. This theory is corroborated by a similar collocated positive–negative dipole pattern of EKE in January and February (Fig. 7, top).

For the NA experiment, upper-level zonal wind anomalies corroborate EKE response under maximum snow conditions in early winter (Fig. 8, bottom). Significant increased zonal winds are present upstream of the forcing region throughout the winter season, ranging from 1 to 7 m s\(^{-1}\), being strongest from October through January. When comparing this increase in zonal wind to that of the mean flow, as estimated from the control run, this represents a 10%–20% increase in most cases and up to a 40% increase over the Pacific in December. Downstream
strengthening and a southward shift of zonal winds associated with maximum snow conditions are evident from March onward. In all cases, regions displaying strengthened upper-level zonal winds are collocated with reduced geopotential heights, implying a greater temperature gradient under maximum snow conditions. Change in downstream height response is present in early versus late winter season, as February through April exhibit greater height depressions throughout the Atlantic sector than in early winter. A steeper height

**FIG. 7.** The EKE difference at 500 hPa during forced snow scenarios, maximum minus minimum, for both the (top) Eurasian and (bottom) North American experiments. Positive contours (red dashed) imply greater EKE under maximum snow depth and extent. Contours are at 10 m² s⁻² intervals with stippled areas significant at the 95% level.

**FIG. 8.** Zonal wind difference at 250 hPa during forced snow scenarios, maximum minus minimum, for both (top) Eurasian and (bottom) North American experiments. Positive contours (red) imply strengthened winds under maximum snow depth and extent conditions. Contours are at 2 m s⁻¹ intervals with stippled areas significant at the 95% level.
gradient would cause an increase in baroclinicity and associated increase in the strength of the midlatitude jet during these months.

4. Summary and conclusions

Previous GCM investigations of the effects of anomalous snow on the overlying atmosphere and associated circulation have used atmosphere-only models. SST and sea ice concentration in these models are boundary conditions constructed from seasonally varying historical values. A common criticism of experiment design for such studies is that specification of boundary conditions is a response to atmospheric conditions rather than the cause (Bretherton and Battisti 2000). Because snow itself is being treated as a forced boundary condition, an imbalanced hydrological cycle is already a factor when interpreting results. When considering how snow can influence the atmosphere let us first consider what drives much of the earth’s large-scale circulation. A seminal paper by Hoskins and Karoly (1981) showed that much of the large-scale atmospheric circulation is modulated by orography. SST is another fundamental modulator of large-scale atmospheric circulation; inclusion of energy exchange over ocean surfaces in such land–atmosphere studies represents an important physical link to better understand previously documented teleconnection pathways and circulation response to anomalous snow.

Surface response to snow cover forcing in our experiments includes significant SST cooling up to \(-1.8^\circ\text{C}\) under maximum snow conditions in both the EUR and NA experiments. Locations of SST cooling include local coastal cooling directly downstream of each individual forcing region in addition to upstream centers of remote cooling: in the Atlantic under anomalously high EUR snow conditions and in the Pacific under anomalously high NA snow conditions. Significant cooling of SAT under maximum snow conditions local to each forcing region was evident in each experiment.

Anomalous maximum snow conditions made the atmosphere cooler and reduced geopotential heights through the midtroposphere and throughout much of the midlatitudes in both experiments. Consistently reduced midtropospheric geopotential heights upstream of the forcing and reduced sea level pressures over the Pacific during early winter were evident from the NA experiment, implying a reduced north–south gradient and a negative AO phase under maximum snow extent and depth.

The impact of extreme snow conditions on surface energy flux variables was generally limited to the vicinity of each forcing region. Areal averaged flux values over continental areas displayed seasonal differences between maximum and minimum snow scenarios. However, no distinct values between EUR and NA simulations were found.

Transient eddy response, EKE, was estimated from daily perturbations from monthly-mean zonal and meridional wind fields. The use of daily model output was necessary to capture transient features, which typically vary on a temporal scale of 2–10 days. In most instances, areas of positive EKE correlate well with steep geopotential height gradient differences between maximum and minimum snow experiments. The barotropic response of atmospheric height over the Pacific in the NA experiment indicated an overall cooling of the lower atmosphere in those locations under maximum snow conditions. A steepened gradient of geopotential height as a result of this cooling would increase the baroclinic zone in the South Pacific and subsequently strengthen the associated upper-level jet. The dipole of EKE in early winter over the Pacific in both experiments, with positive values to the south and negative to the north, indicates a reduced poleward heat flux, which may be contributing to a decrease in warm SST advection northward and the ensuing mid-Pacific SST cooling. This proposed pathway (Fig. 9) is supported by increased zonal wind at 250 hPa collocated with identified regions of sharpened geopotential height gradient, strengthened baroclinicity, and positive EKE.

Results from the EUR experiment show a similar pathway to that proposed for the NA experiment; however, circulation response is found both upstream and downstream of the forcing region in early winter. A southward shift of the prevailing East Asia storm track is indicated from a dipole pattern of EKE in the Pacific. Reduced poleward heat transport associated with a southward shift in the prevailing storm tracks of this region may be influencing the cooling SST trend through reduced warm SST advection to the Aleutian area of the North Pacific. This southward shift is mirrored in the upper-level flow, with significant and strengthened zonal winds in early winter over zones of positive EKE.

![Fig. 9. Schematic of the proposed response pathway to anomalous snow forcing for both Eurasian and North American experiments.](image-url)
This study suggests a similar atmospheric response to continental snow forcing as in previous studies. Results are not directly comparable, however, as the focus of this study was not stratosphere–troposphere coupling or early versus late winter season snow forcing, which were primary features of previous snow–atmosphere interaction studies. Fundamental differences in experiment designs between the current study, an equilibrium climate state design with a seasonally varying continuous snow forcing, and previous studies that consider the transient response to early season snow must also be considered. Transient response studies have been primarily focused on isolating the mechanism by which early season snow anomalies influence midwinter modes of low-frequency atmospheric variability through stratospheric pathways, whereas the current study focuses on the previously unexplored role of ocean surface response to persistent terrestrial snow anomalies. Common emerging themes include significant depression of SAT over excess snow and excitement of chief modes of Northern Hemisphere low-frequency atmospheric variability, such as the AO and/or the NAO. The persistence of such modes through the winter season, however, is variable, which is not surprising as observed atmospheric circulation in the North Atlantic displays a large amount of within-season variance (Hurrell and Deser 2009). Both EUR and NA experiments display response patterns resembling AO and NAO modes of variability. However, the response of these modes is relatively insensitive to the location of the snow forcing. When considering forced EUR snow using a mixed-layer ocean model, atmospheric response is most prevalent over the Pacific and displays stronger and more organized anomalies than evident from the data-forced experiments. In the case of forced NA snow, a persistent upstream center of action is present in the North Pacific on inclusion of a mixed-layer ocean model. The Pacific (Aleutan) sector emerges persistently as a key response region to external forcing in the Northern Hemisphere, suggesting that, once the negative AO or NAO phase is excited in early winter, the response pattern occurs throughout the hemisphere. In this experiment, emphasis on the Pacific sector through SST cooling and associated excitement of the East Asia jet and a southward shift in prevailing storm tracks was evident.

From these results, it is clear that inclusion of an interactive ocean represents a significant addition in such forced snow–atmosphere interaction studies. From the previous discussion of the NAM, AO and NAO, it is also clear that northern Atlantic and Pacific SSTs influence atmospheric circulation on a variety of temporal scales. If this requires the addition of an ocean model to correctly capture land/ocean heating contrasts and the appropriate influence of SST in such forced snow modeling studies, at the time scale needed for these simulations, a mixed-layer “slab” model is sufficient.

In conclusion, circulation responses to external forcing in the Northern Hemisphere typically involve low-frequency circulation patterns of an AO type, which is not a surprise because this is the main mode of atmospheric circulation in the Northern Hemisphere. When perturbing the climate system with anomalous snow extent and depth in these experiments, we see a tendency to enhance or suppress different components of the AO/NAO pattern that are not sensitive to the particular location of the surface forcing. Previous studies investigating snow–atmosphere interactions have relied upon filtering of model output to maximize the signal resulting from snow forcing. In this study, we show impacts of realistic continental snow perturbations from raw model output using an interactive ocean surface. We believe inclusion of an interactive ocean surface for studies of this nature is a valuable contribution to the field of land–atmosphere interactions.

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