Sensitivity of atmospheric response to modeled snow anomaly characteristics

Gavin Gong,[1] Dara Entekhabi,[2] Judah Cohen,[3] and David Robinson[4]

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[1] The presence of snow over broad land surface regions has been shown to not only suppress local surface temperatures, but also influence various remote climate phenomena. However, the specific mechanisms and snow anomaly characteristics which produce this response are still not well understood. In this study, large-ensemble general circulation model (GCM) experiments are performed to investigate the sensitivity of the atmospheric response to snow cover vs. snow depth anomalies, and the relevant surface thermodynamic processes involved. Realistic, observation-based, autumn-winter snow forcings over Siberia are developed and applied as model boundary conditions, to evaluate the climate response to (1) comprehensive snow forcings including snow cover and snow depth components, (2) snow cover only forcings, and (3) snow forcings in the absence of a surface albedo response. Results indicate that snow cover extent anomalies are not the only significant contributor to the local temperature response; snow depth anomalies are shown to have a comparable effect. Furthermore, the albedo effect is not the predominant thermodynamic mechanism; processes related to the insulative properties of the snowpack (e.g., thermal conductivity and latent heat flux) are also involved. Lastly, we find that realistic snow cover and snow depth anomalies acting in conjunction are required to produce a local temperature response which is strong enough to distinctly modulate the winter Arctic Oscillation (AO) mode as shown in previous studies. Such a detailed understanding of the atmospheric sensitivity to snow anomaly characteristics is beneficial for effectively utilizing any potential climate predictability contained in snow anomaly signals. *INDEX TERMS:* 1863 Hydrology: Snow and ice (1827); 1833 Hydrology: Hydroclimatology; 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; *KEYWORDS:* snow, climate

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1. Introduction

[2] The influence of snow on climate is well documented in the literature [see *Cohen*, 1994, for review]. The principal relationship is a decrease in local surface temperature over snow covered land, relative to snow-free land. For broad snow covered regions, this local thermodynamic response affects remote climate as well. In the Northern Hemisphere, snow anomalies have been linked to seasonally lagged local climate features, and also to various regional and hemispheric-scale phenomena. Therefore snow anomalies have been suggested as a potential predictor of climate on seasonal timescales.

[1]Department of Earth and Environmental Engineering, Columbia University, New York, New York, USA.

[2]Department of Civil and Environmental Engineering, Massachusetts Institute of Technology, Cambridge, Massachusetts, USA.

[3]Atmospheric and Environmental Research, Inc., Lexington, Massachusetts, USA.

[4]Department of Geography, Rutgers University, Piscataway, New Jersey, USA.

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[3] The surface thermodynamic processes which can act to reduce surface temperature in the presence of snow cover include less absorbed shortwave radiation due to the high albedo of snow, more outgoing thermal radiation due to the high emissivity of snow covered land, more outgoing latent heat flux due to snowmelt, evaporation and/or sublimation, and less incoming heat flux from the underlying soil due to the low thermal conductivity of the snowpack. To varying degrees, these processes occur in response to both snow cover and snow depth anomalies. For example, snow covered land has a notably higher albedo than snow free land, but a deep, fresh, or complete snow cover also has a higher albedo than a shallow, old, or partial snow cover.

[4] Although snow cover and snow depth anomalies are inherently related, in nature they do not necessarily coincide in magnitude or even phase. For example, an anomalously large snow cover extent does not necessitate anomalously large snow depths within the snow covered region. Therefore it is of interest to distinguish the sensitivity of the atmospheric response to snow cover vs. snow depth anomalies, and identify the thermodynamic processes associated with each snow anomaly character-

istic. A numerical study is presented here which performs such an analysis, focusing on autumn-winter snow anomalies over Siberia.

2. Literature Review

[5] Early studies used empirical observations to investigate the statistical relationship between snow cover extent and various climate phenomena [*Hahn and Shukla*, 1976; *Dewey*, 1977; *Walsh et al.*, 1982; *Foster et al.*, 1983, *Namias*, 1985]. More recent studies, consisting of both empirical analyses and modeling experiments, have by and large confirmed the early studies. More specifically, snow cover anomalies have been associated with surface temperature [*Leathers and Robinson*, 1993; *Leathers et al.*, 1995], tropospheric circulation [*Walland and Simmonds*, 1997; *Clark and Serreze*, 2000], the Indian summer monsoon [*Douville and Royer*, 1996; *Bamzai and Shukla*, 1999], and several modes of atmospheric variability [*Gutzler and Rosen*, 1992; *Watanabe and Nitta*, 1998; *Cohen and Entekhabi*, 1999; *Frei and Robinson*, 1999; *Bojariu and Gimeno*, 2003; *Saunders et al.*, 2003], including the Arctic Oscillation [*Gong et al.*, 2002, 2003a; *Saito and Cohen*, 2003]. Hereafter, *Gong et al.* [2003a] will be referred to as GEC03.

[6] The historical emphasis on snow cover extent as a climate modulator is reasonable, given that a continuous record of visible satellite-based snow cover extent data over the Northern Hemisphere has been available since about 1966 [*Robinson et al.*, 1993; *Robinson and Frei*, 2000]. However, empirical studies are limited by the fact that snow cover extent does not fully characterize snow anomalies. Regions which exhibit minimal snow cover variability may in fact exhibit extensive snow depth variability, the climatic effects of which would not be captured in the empirical correlations to snow cover extent [*Bamzai and Shukla*, 1999]. Empirical investigations that account for snow depth have been limited in number and robustness, due in part to a lack of reliable snow depth data coverage. The development of passive microwave based satellite sensors offers the potential to provide spatially complete snow depth observations, but their accuracy is not yet sufficient to produce a consistent, high-quality data set [*Armstrong and Brodzik*, 2001; *Frei et al.*, 2003].

[7] Modeling studies provide a suitable platform for explicitly distinguishing the atmospheric sensitivity to snow cover and snow depth anomalies. However, perhaps due to the precedent established by empirical studies, modeling studies have similarly focused on the climate response to snow cover extent. Some studies prescribe a snow cover only forcing, in which snow depths are held constant between different experiments [*Cohen and Rind*, 1991; *Walland and Simmonds*, 1997]. Others do prescribe a combined snow cover and snow depth forcing [*Yeh et al.*, 1983; *Douville and Royer*, 1996; *Cohen and Entekhabi*, 2001; *Marshall et al.*, 2003; GEC03]. However, these studies by and large limit their analysis to the climatic response to this aggregate forcing, and make no effort to differentiate the response to the snow cover and snow depth components.

[8] A limited number of studies have attempted to distinguish snow cover and snow depth anomalies, and they have generally indicated that snow depth makes a distinct contribution to the overlying atmosphere thermodynamics. *Baker et al.* [1992] used point station data to demonstrate that a deep snow cover produces a substantial local temperature depression, while a shallow snow cover results in a notably smaller depression. *Fallot et al.* [1997] analyzed a long-term data set of cold season snow depth, temperature and precipitation at 110 point stations within the Former Soviet Union, and found modest correlations between snow depth and the climate variables. Much earlier, *Wagner* [1973] found a regression relationship between snow depth and temperature using a much smaller station data set. *Watanabe and Nitta* [1998] performed a model study with snow depth only forcing over Siberia, and reported broad dipole anomalies reminiscent of the dominant mode of extratropical Northern Hemisphere climate variability. Model studies by *Barnett et al.* [1989] and *Yasunari et al.* [1991] found snow depth to be more influential than snow cover extent with respect to the Indian summer monsoon.

[9] Clearly, snow depth anomalies have a non-negligible effect on both local and remote climate conditions. This study presents a comprehensive investigation of the relative climatic contributions of snow cover and snow depth anomalies, via a suite of large-ensemble general circulation model (GCM) experiments during the autumn-winter season. First, the climatic impact of a realistic, observation-based snow (extent and depth) anomaly over Siberia is simulated (referred to as the SIB experiments). Next, the Siberia boundary conditions are modified to represent snow cover only forcing (referred to as the COV experiments), and the resulting local surface response for this partial snow forcing is compared to that for the total snow anomaly forcing. Third, the sensitivity of surface albedo to snow is explicitly suppressed within the model (referred to as the INS experiments), to gain insight to the partial influence of non-albedo processes associated with the snowpack. Finally, remote climate impacts are briefly investigated, in the context of modulations to the dominant mode of wintertime extratropical Northern Hemisphere climate variability, specifically the Arctic Oscillation (AO; *Thompson and Wallace* [1998]).

3. General Circulation Model Experiments

3.1. Model Description and Experiment Design

[10] The Max-Planck Institute for Meteorology ECHAM3 atmospheric GCM [*Roeckner et al.*, 1992] is used for this study. The model climate compares favorably with other GCMs and observed data sets, within the guidelines of the Atmospheric Model Intercomparison Project (AMIP; *Gates et al.* [1998]). Climate features of particular relevance to this study have been extensively tested and reported in the literature. Snow cover and snow mass climatology is reproduced reasonably well, on both a hemispheric scale [*Foster et al.*, 1996; *Frei and Robinson*, 1998] and specifically within the Siberia region (GEC03). Basic features of the wintertime AO mode of variability, defined as the leading empirical orthogonal function of the winter December–February (DJF) season Northern Hemisphere sea level pressure field, are successfully captured by a twenty-year control integration of the base model with climatological sea surface temperatures [*Gong et al.*, 2002].

Thus the ECHAM3 GCM serves as a reasonably suitable platform for evaluating the atmospheric sensitivity to Siberian snow anomalies.

[11] ECHAM3's land surface parameterization is derived from the Simple Biosphere model [*Sellers et al.*, 1986], and snow is parameterized in a straightforward manner (DKRZ 1994). Snow depth is maintained at each gridcell as an internal state variable *S*, in the form of snow water equivalent (SWE). A distinct snowpack layer resides above the surface soil layers, and evolves according to a budget equation:

$$\frac{\partial}{\partial t}S = \frac{J_S + P_S - M_S}{\rho_w} \tag{1}$$

where J_S = evaporation rate over the snowpack, P_S = snowfall rate over the snowpack, M_S = snowmelt rate over the snowpack, and ρ_w = density of water. Melted snow M_S is intercepted by vegetation and infiltrated into soil via separate budget equations for each surface reservoir. Evaporation J_S over snow occurs at the potential rate. For snow depths greater than 0.025 m SWE, snowpack temperature T_S evolves via a separate heat conduction equation for the snowpack:

$$\frac{\partial}{\partial t}T_S = \frac{F}{\rho_S C_S S} \tag{2}$$

where F = sum of radiative and turbulent fluxes at the surface, ρ_S = snow density, and C_S = heat capacity of snow. For snow depths less than 0.025 m SWE, T_S is not updated, and the basic surface heat conduction equation is solved irrespective of the snowpack. Surface albedo α_{surf} is related to the snow depth via:

$$\alpha_{surf} = \alpha_b + (\alpha_S - \alpha_b)\frac{S}{S + S^*} \tag{3}$$

where α_b = surface background albedo, α_S = snow albedo, and S^* = 0.01 m SWE. A minimum (maximum) value for snow albedo over land is assigned according to fractional forest area and ranges from 0.4 (0.8) for unforested land to 0.3 (0.4) for fully forested land. The actual α_S value is interpolated from these limits based on surface temperature. In this parameterization, thermodynamic processes are unaffected by snow age and metamorphosis, and land surface thermal emissivity is unaffected by the presence of snow cover. Finally, the snow covered gridcell fraction C_S is parameterized as:

$$C_S = \min\left(1, \frac{S}{S_c}\right) \tag{4}$$

where S_c = 0.015 m SWE.

[12] All model experiments are run at T42 spectral truncation (roughly 2.8° gridcell resolution), and integrated from September through February. This resolution is generally adequate to evaluate continental and hemispheric scale atmospheric response, and more importantly allows for computationally efficient large-ensemble simulations. Each experiment consists of twenty realizations of this six-month integration period, with independent September 1 initial conditions obtained from the twenty-year control simulation. These ensemble experiments are performed in pairs, in which one experiment prescribes extensive snow conditions over Siberia, while the other prescribes limited snow conditions over Siberia. Snow forcings are prescribed at the beginning of each timestep, during which the prescribed snow is subject to melting and evaporation as dictated by the model, and the model atmosphere is allowed to respond accordingly. Ensemble mean diagnostics are computed for each experiment, and significant differences (extensive snow–limited snow) are evaluated using the statistical t-test, to investigate the model response to a positive Siberian snow anomaly.

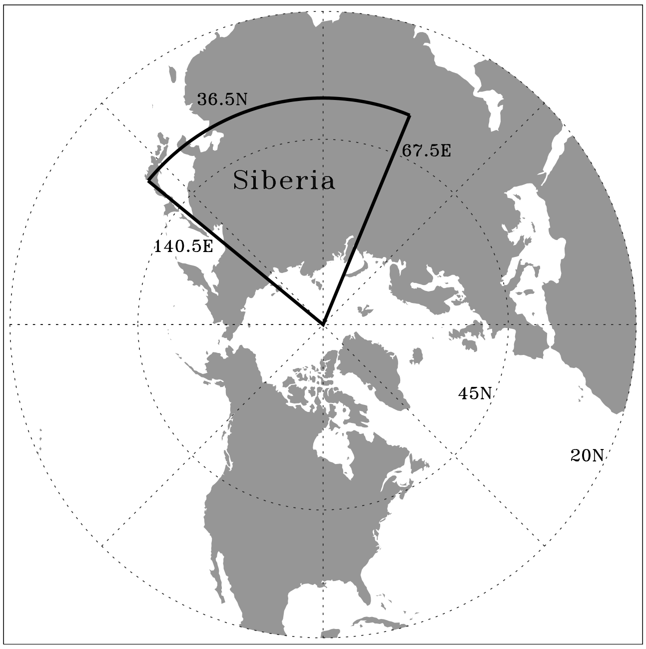

Figure 1. Prescribed snow forcing region over Siberia applied for all experiments. Snow outside of this region is maintained by the model as an internal state variable.

3.2. Realistic Snow Forcing: SIB

[13] The first pair of experiments, denoted as SIB, prescribes realistic, observation–based snow conditions over the Siberia forcing region indicated in Figure 1. Note that this set of experiments is extensively documented in GEC03. The selected forcing region represents one of the world's largest contiguous land surface areas, consisting primarily of taiga forest and treeless tundra. It is subject to considerable and consistent snow cover during winter, but interannually varying onset of snow cover during autumn. It is further characterized by a sizable mountain chain extending northeast from the Tibetan Plateau (elevations consistently over 1000 meters and a peak of roughly 4500 meters at Mt. Belukha), which have been shown to influence stationary wave activity and atmospheric circulation throughout the Northern Hemisphere [*Plumb*, 1985; *Gong et al.*, 2004]. Therefore this region holds considerable potential for snow-forced climate modulation over broad spatial scales. Snow cover forcings are taken from the National Oceanographic and Atmospheric Adimnistration (NOAA) visible satellite weekly data set [*Robinson et al.*,

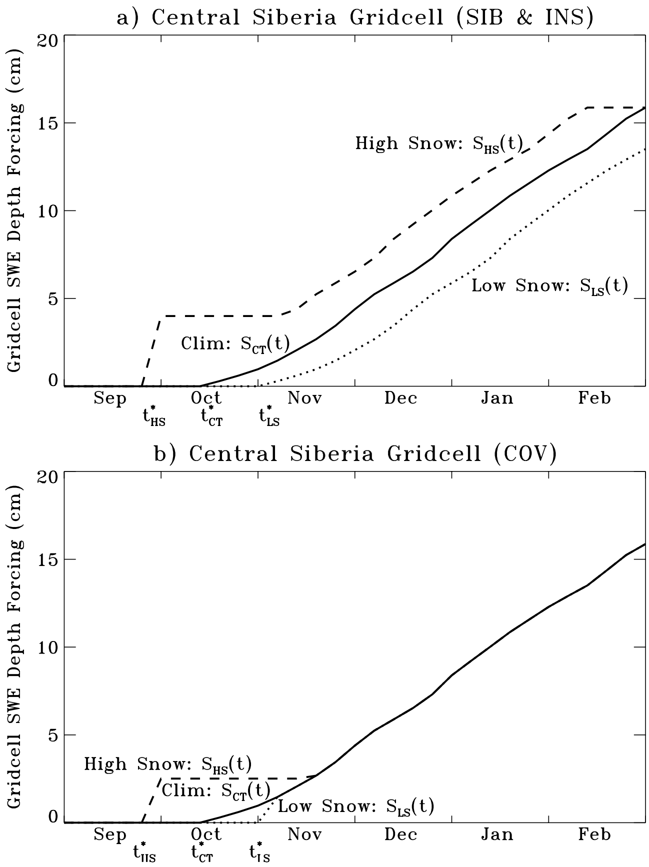

Figure 2. Weekly SWE depth forcing time series applied at a gridcell in central Siberia, for (a) SIB and INS, and (b) COV. Dashed (dotted) line represents high (low) snow experiments. Also shown is the SWE depth time series from the model climatology (solid line), and observed snow arrival dates for each experiment.

1993]. Observations from September 1976–February 1977 (September 1988–February 1989) are used for the extensive (limited) snow experiment, as this period exhibited the highest (lowest) autumn season snow cover extent over both Eurasia and Siberia. Snow cover extent variations between these two experiments begin in late September and continue steadily into December, after which the snow cover forcing is minimal since all of the Siberia perturbation region is essentially covered with snow for both experiments. The largest differences occur in mid-October, when the snow cover extent can vary by roughly a factor of five (see GEC03 Figure 2).

[14] The NOAA data set indicates the presence or absence of snow, but does not provide any information regarding snow depth. Prescribing realistic snow depths presents a challenge, since reliable and comprehensive historical snow depth data are not readily available. Therefore an approximate method is applied here, in which the model's weekly gridcell snow water equivalent (SWE) climatology time series (obtained from the twenty-year control simulation) is translated backward (forward) in time to represent SWE for the extensive (limited) snow experiment. The magnitude of the temporal translation is determined by the observation-based gridcell snow arrival date for each experiment, relative to the climatology. This procedure is performed individually at each gridcell within the Siberia forcing region. It is illustrated graphically in Figure 2a, and can be expressed mathematically as:

$$S_{HS}(t) = S_{CT}(t + [t^*_{CT} - t^*_{HS}])$$

$$S_{LS}(t) = S_{CT}(t - [t^*_{LS} - t^*_{CT}])$$

where S_{HS} = SWE depth for the high snow experiment, S_{LS} = SWE depth for the low snow experiment, S_{CT} = SWE depth for control simulation, t = time during model integration period, t^*_{CT} = observed snow arrival date for the high snow experiment, t^*_{LS} = observed snow arrival date for the low snow experiment, and t^*_{CT} = snow arrival date for control experiment. A minimum SWE value of 4.0 cm is also applied to the high snow forcing experiment, in order to maximize the surface thermodynamic contrast between the two experiments.

[15] The resulting snow forcings are checked and adjusted as needed to ensure that the high snow (low snow) forcing always exhibits more (less) extensive snow cover, earlier (later) initial snow occurrence, and larger (smaller) SWE values over Siberia, relative to the climatology, thereby representing a consistent positive (negative) snow perturbation. Also note that the magnitude of the snow depth forcing is directly related to the magnitude of the snow cover forcing, as represented by the observed snow arrival dates. Thus the aim of this approximate procedure is not to precisely represent the actual snow depths that occurred during the periods from which snow cover observations are applied, but rather to prescribe reasonable and consistent snow depth forcings, derived in part from the observed data. Nevertheless, GEC03 demonstrated that the resulting SWE depth forcings averaged over Siberia successfully capture the range of observed values, as derived from a 25-year data set (1966–1990) of thrice-weekly snow depth measurements at roughly 1300 stations in the Former Soviet Union [*Krenke*, 1998].

[16] Overall, these two experiments prescribe extreme but realistic and observation-based snow conditions over Siberia. The ensemble mean response to a positive Siberian snow forcing produced by this pair of experiments will be denoted as SIB. It represents the model's atmospheric sensitivity to a comprehensive snow anomaly, including both snow cover and snow depth characteristics, and all relevant thermodynamic processes.

3.3. Snow Cover Only Forcing: COV

[17] A second pair of snow-forced ensemble experiments is performed, analogous to the SIB experiments. The principal difference is that the snow cover forcing is applied, but the snow depth forcing is not. The atmospheric response for this pair of snow cover only experiments is denoted as COV. Figure 2b shows the corresponding weekly SWE depth time series for a gridcell in central Siberia, for the COV experiments. Wherever snow cover exists, the SWE depth is prescribed at the climatological values. In other words, the temporal shift applied in SIB is omitted in COV. In cases where the high snow experiment calls for snow cover, but the climatology indicates either no snow or

shallow snow (i.e., less than 2.51 cm SWE), a depth of 2.51 cm SWE is prescribed. This value represents a minimum snow depth for the high snow experiment, and was selected to ensure a thermodynamically active snow layer based on the ECHAM3 snow parameterization scheme. In cases where the low snow experiment calls for no snow, but the climatology indicates snow cover, a depth of zero is assigned.

[18] In this way, a snow forcing essentially occurs only when the high snow experiment is snow covered while the low snow experiment is not. Thus the partial snow forcing in COV only occurs during the early stages of the snow season, as exemplified in Figure 2b. Later in the season, when both the high snow and low snow experiments are snow covered, snow depths are identical so the snow forcing ceases. This is in contrast to SIB, where earlier (later) snow is consistently associated with deeper (shallower) snow, so that a snow forcing occurs throughout the model integration, as depicted in Figure 2a. Comparison of the atmospheric response between SIB and COV will indicate the degree to which snow cover only anomalies contribute to the total snow-forced response.

[19] The most obvious thermodynamic mechanism associated with snow anomalies is surface albedo, which can be as high as 0.8 for snow, whereas snow-free land surfaces rarely exceed 0.35. Many studies attribute the atmospheric sensitivity to snow primarily to this albedo effect [*Baker et al.*, 1992; *Marshall et al.*, 2003; *Kumar and Yang*, 2003], which has prompted numerous efforts to study and improve albedo representation in models [*Roesch et al.*, 1999; *Lefebre et al.*, 2003]. Removing snow depth anomalies as in the COV experiments retains the albedo differences due to snow cover extent. Therefore it is tempting to consider the COV experiments as isolating the climatic response to snow-related albedo anomalies, but caution must be exercised when making such a generalization. As mentioned earlier, snowpack characteristics (uniformity, depth, age) also affect albedo, albeit to a more modest degree. Thus a fraction of the total albedo effect is lost when snow depth anomalies are removed. Furthermore, other surface thermodynamic processes are also affected to some degree by the presence or absence of even a thin snow cover, so that the model response for COV is not due entirely to the albedo effect. The COV experiments focus on the atmospheric sensitivity to snow cover only forcings, but do not necessarily isolate specific thermodynamic mechanisms.

3.4. Snow Insulation Only Forcing: INS

[20] To better assess the specific influence of the albedo effect, a third pair of snow-forced ensemble experiments is performed, also analogous to the SIB experiments. The principal difference this time is that the surface albedo is specified at the background (i.e., snow-free) value, for each gridcell in the Siberia forcing region and at every timestep. The snow cover and snow depth forcings themselves are unchanged from SIB (see Figure 2a). In this way, the large albedo differences associated with snow are explicitly suppressed, so that this set of experiments isolates processes related to surface insulation properties of the snowpack. In this paper, insulation refers not just to thermal conductivity changes due to the presence of snow, but also all other non-albedo related properties, such as thermal emissivity and energy sinks due to phase changes. The atmospheric response for this set of insulation only experiments is denoted as INS. Since snow depth anomalies are included for INS, snow forcings are not restricted to the early-season as for COV, but occur throughout the simulation period. Comparison of the atmospheric response between SIB and INS will indicate the degree to which insulation processes, exclusive of the albedo effect, contribute to the total snow-forced response.

[21] A number of modeling studies suggest that in fact albedo may not be the dominant mechanism involved. These studies generally evaluate the full surface energy balance response to some sort of snow forcing, and report a complex set of flux anomalies, of which the albedo effect on shortwave radiation is but one component. For example, *Yeh et al.* [1983] and *Cohen and Rind* [1991] find that albedo and snowmelt processes contribute to an observed local temperature decrease in response to snow, but that longwave radiation, latent heat and sensible heat fluxes produce a slight temperature increase, which dampens the aggregate response. With respect to the Indian summer monsoon, *Barnett et al.* [1989] and *Vernekar et al.* [1995] find that spring snowmelt produces the temperature and subsequent monsoon response, while *Yasunari et al.* [1991] attribute the response to the albedo effect, and *Douville and Royer* [1996] find that both processes contribute significantly. The current investigation of the precise contribution of surface albedo vs. insulation mechanisms, in conjunction with an evaluation of snow cover vs. snow depth, will provide additional insight regarding the atmospheric response to snow anomaly characteristics and associated thermodynamic processes.

4. Local Surface Atmospheric Response to Snow Forcings

4.1. Seasonal Mean Fields

[22] We begin by evaluating the local surface response over the Siberia forcing region during the autumn September–November (SON) season, which contains the largest forcings in terms of both snow cover and snow depth. To present a broad overview, Figure 3 shows the SON surface albedo and temperature response fields over the extratropical Northern Hemisphere, for all three experiment pairs. For SIB (Figures 3a and 3b, repeated from GEC03 Figures 4a and 4b), a strong and statistically significant albedo increase and temperature decrease occur in response to the comprehensive snow forcing, confined to the Siberia forcing region. This indicates that local surface thermodynamic processes, particularly the reduction in shortwave radiation absorbed due to the high albedo of snow, are directly responsible for the temperature response.

[23] This result is consistent with many previous studies, some of which concluded that the snow cover extent anomalies and the associated albedo effect are principally responsible for the temperature decrease [*Watanabe and Nitta*, 1998; *Kumar and Yang*, 2003]. However, Figure 3c and 3d indicates that for COV, removing snow depth forcings results in a clear but notably damped response

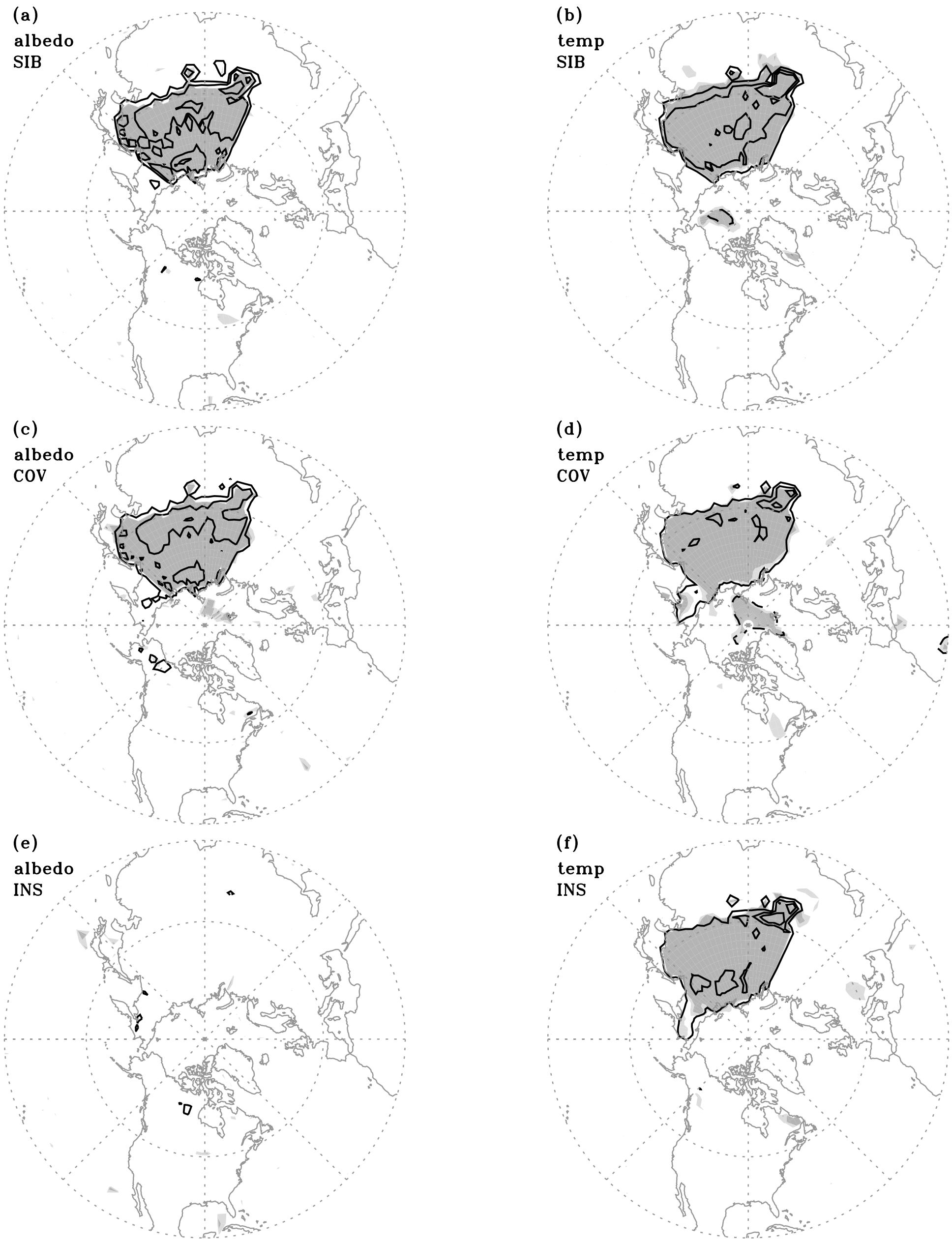

Figure 3. Autumn season (SON) surface response to positive Siberian snow forcing, over the extratropical Northern Hemisphere, for SIB (a, b) repeated from GEC03 Figures 4a and 4b, COV (c,d), and INS (e, f). (a, c, e) Surface albedo: contours drawn at ±0.02, 0.1, 0.25, where solid (dashed) line denotes positive (negative) value. (b, d, f) Surface temperature: contours drawn at ±1, 3, 5°C, where solid (dashed) line denotes negative (positive) contour value. Light (dark) shading indicates 90% (95%) statistical significance.

compared to SIB, for both albedo and temperature. The Siberia-average albedo response decreases by about 33%, which confirms that the albedo effect is largely associated with snow cover, but also with snow depth to a fair degree. The temperature response similarly decreases by roughly a third, which indicates that the snow depth anomalies excluded from COV contribute measurably to the overall temperature response for SIB.

[24] Removing the albedo effect in INS (hence the null albedo response in Figure 3e) nonetheless produces a significant temperature response (Figure 3f). The Siberia average temperature response for INS is smaller in magnitude than for SIB and very close to that for COV. For INS, only the surface insulative thermodynamic processes are active, yet these mechanisms are able to produce a temperature response roughly two-thirds as large as when the albedo effect is also included. Overall, Figure 3 indicates that the albedo effect (mainly associated with snow cover extent) is clearly an important contributor to the local temperature response to snow anomalies. However, it is also apparent that this is not the exclusive mechanism; snow depth anomalies and insulative properties associated with

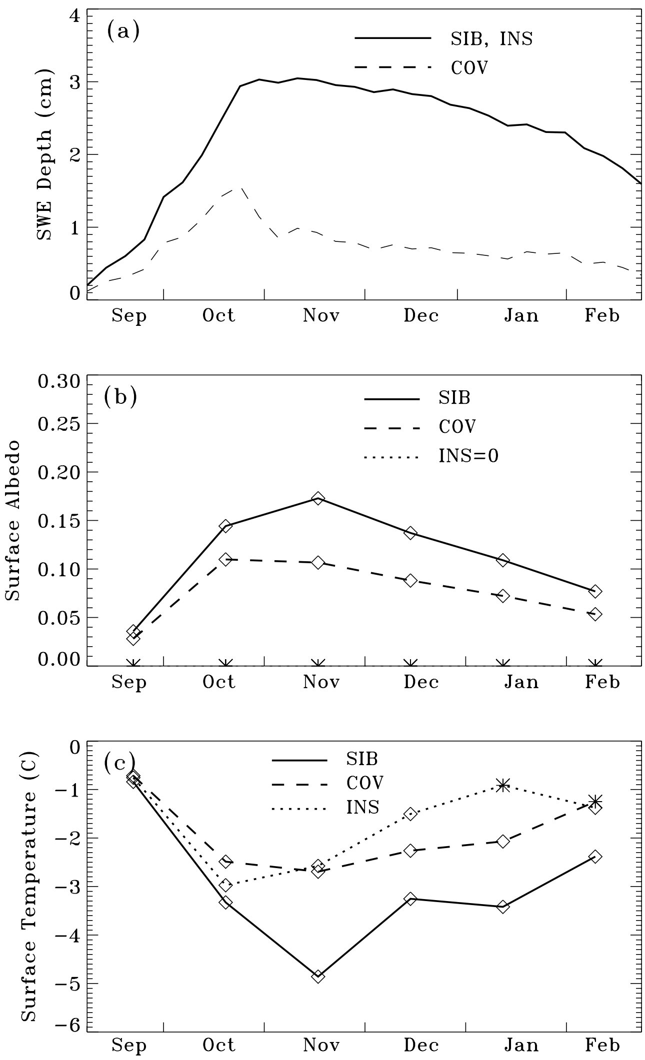

Figure 4. Monthly time series of Siberia average snow-forced response (high snow-low snow), for SIB, COV and INS. (a) Model input weekly SWE depth. (b) Surface albedo. (c) Surface temperature. For Figures 4b and 4c, diamonds (asterisks) denote values which are (are not) statistically significant at the 95% level.

the snowpack are also important contributors to the local surface response.

4.2. Siberia Average Surface Energy Balance

[25] We now present monthly time series of the average response to snow over the Siberia forcing region, for a number of surface diagnostics, to help ascertain the precise surface thermodynamic response for the three sets of experiments. Figures 4 shows the albedo and temperature response, and also the weekly snow forcing. Note that for SIB and INS, the snow forcing builds rapidly during autumn due to snow cover differences, and is maintained during winter months due to snow depth differences. Only snow cover is prescribed for COV, therefore the snow forcing in Figure 4a is notably reduced, and occurs primarily during autumn. In terms of average SWE depth over Siberia, the snow depth forcing excluded in COV constitutes a considerable portion of the overall snow forcing in SIB and INS, throughout the simulation period. Surface albedo (Figure 4b) increases in response to snow for SIB and COV as expected, peaking in mid-late autumn just after the peak snow forcings. Consistent with Figure 3, the damped response for COV indicates that albedo responds to snow depth as well as snow cover. Again, note that the albedo change for INS is by construction nonexistent. The temperature response (Figure 4c) similarly peaks around mid-late autumn, and a clear but damped response is apparent for both COV and INS, beginning in November. Snow cover only forcings (COV) and insulation only processes (INS) each contribute considerably to the total temperature response seen for SIB. Clearly, snow cover anomalies and the associated albedo effect is an important but nonexclusive mechanism for driving the local surface temperature response to snow anomalies.

[26] Figures 5 and 6 show various components of the surface energy balance over Siberia. Figure 5a shows the

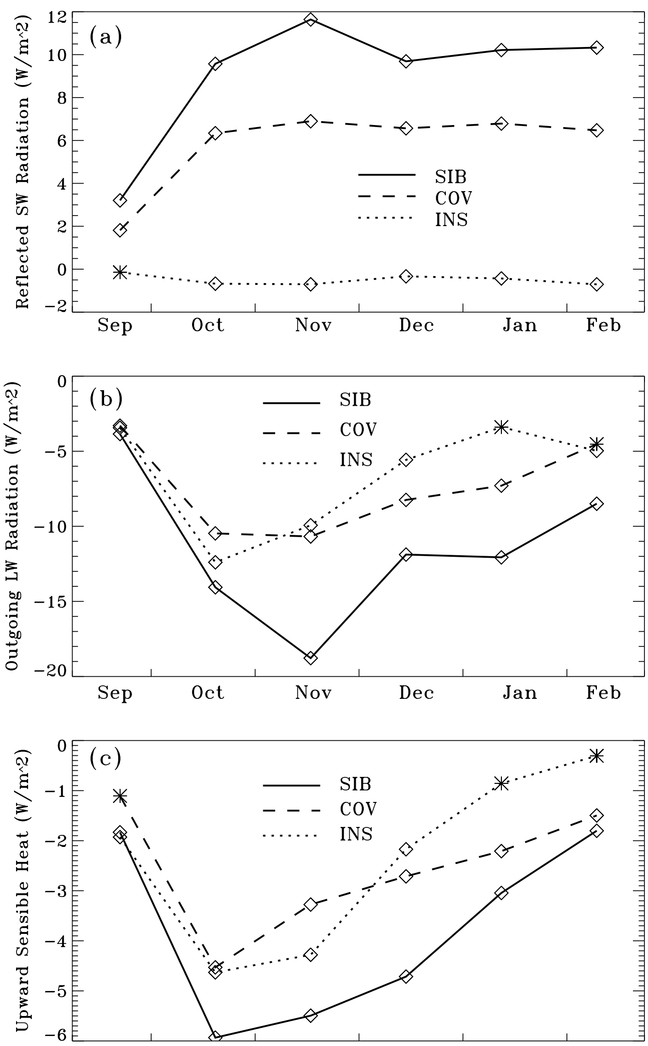

Figure 5. Monthly time series of Siberia average snow-forced response (high snow − low snow), for SIB, COV and INS. Diamonds (asterisks) denote values which are (are not) statistically significant at the 95% level. (a) Reflected (upward) SW radiation away from surface. (b) Upward LW radiation away from surface. (c) Upward sensible heat flux away from surface.

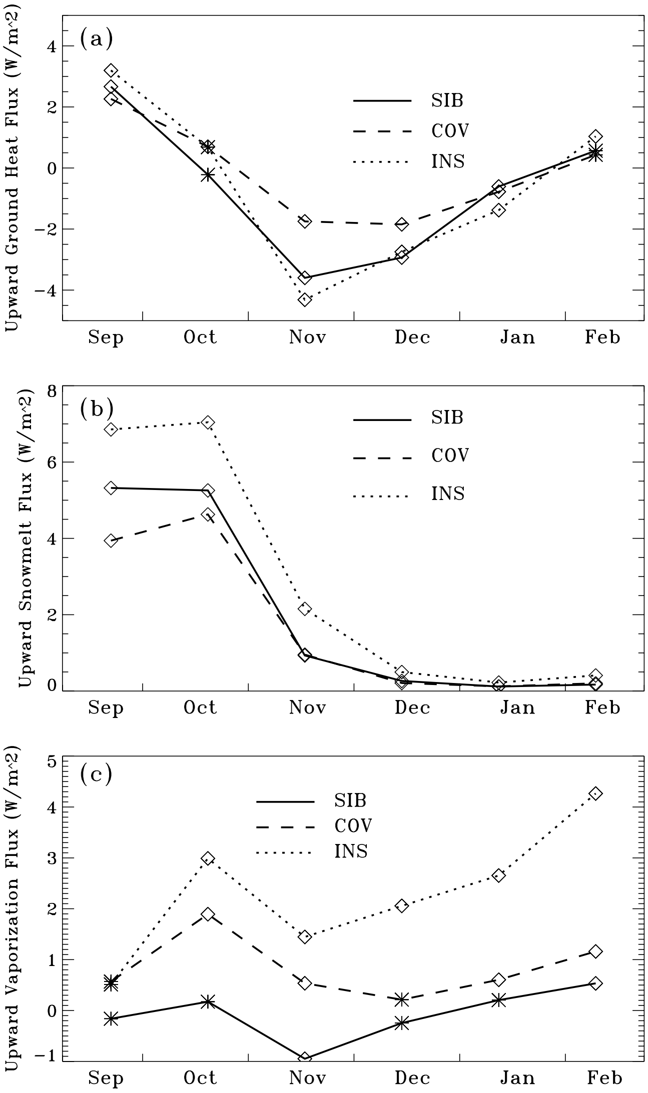

Figure 6. Monthly time series of Siberia average snow-forced response (high snow–low snow), for SIB, COV and INS. Diamonds (asterisks) denote values which are (are not) statistically significant at the 95% level. (a) Upward ground conduction heat flux from soil to surface. (b) Upward latent heat flux due to snowmelt. (c) Upward latent heat flux due to evaporation/sublimation.

monthly time series response for upward or reflected shortwave (SW) radiation away from the surface. As expected, the higher albedo of snow results in more reflected SW radiation, and the response for SIB, COV and INS are all analogous to Figure 4b (albedo). The only exception occurs during the winter months; whereas positive albedo anomalies lessen as Siberia approaches near-complete snow coverage for both high and low snow experiments, reflected SW radiation anomalies are maintained since the snow forcing is concentrated in southern Siberia, where incoming and hence reflected SW radiation are both greater. Overall, it is clear that SW radiation flux anomalies are a direct expression of the albedo effect.

[27] Outgoing longwave (LW) radiation and upward sensible heat flux responses are shown in Figure 5b and 5c. These parameters exhibit a consistently negative (i.e., less outgoing or upward flux) response to snow, which suggests additional energy at the surface, and a surface temperature increase rather than the model simulated decrease in Figure 4c. Note however that the time series patterns for these two surface fluxes resemble the patterns for surface temperature. This is reasonable since both outgoing LW radiation and sensible heat fluxes are inversely proportional to surface temperature. The time series patterns in Figure 5b and 5c are a more or less direct response to the temperature anomalies in Figure 4c. In effect, these fluxes represent a negative feedback which mitigates the ultimate temperature decrease in response to snow, which is consistent with the results of *Yeh et al.* [1983] and *Cohen and Rind* [1991]. Figure 5b also indicates that the higher thermal emissivity of a snowpack relative to snow-free land is not an influential process. The increase in outgoing LW radiation expected due to snow emissivity is overwhelmed by the decrease due to lower surface temperatures.

[28] Figure 6a shows the response to snow of upward ground heat flux due to conduction from the soil to the surface. In general, the low thermal conductivity of a snowpack insulates the soil. After a snow anomaly has formed by mid-autumn for SIB, this inhibits the conduction of heat from the relatively warm soil to the cold surface, and produces the negative snow-forced ground heat flux anomalies from November onward. For COV, the snow depth forcing is not included, so there is less snow to insulate the soil, and consequently the negative ground heat flux anomalies are mitigated during November and December when snow depths are shallow. Note that the ground heat flux time series are similar for SIB and INS, since the suppression of snow albedos at the surface has little bearing on the heat flux from the underlying soil. Ground heat flux exhibits a clear response to the prescribed snow forcings, due to the low thermal conductivity associated with snow.

[29] The upward latent heat flux response due to snowmelt is shown in Figure 6b. For SIB, considerable melting occurs in response to the snow forcing in the early season then gradually tapers off, as surface temperatures fall below 0°C and the prescribed snow is less susceptible to melting. For COV, this snowmelt flux response is consistently smaller, since in the absence of a snow depth forcing at the beginning of each timestep there is less snow available for the model to melt during the timestep. For INS, suppression of the albedo effect means a decrease in reflected SW radiation, which provides additional energy to melt the prescribed snow, so that the early season snowmelt flux is consistently larger. The prescribed snow forcing characteristics have a demonstrable effect on early-season snowmelt fluxes.

[30] Figure 6c shows the upward latent heat flux response due to evaporation/sublimation, which is sensitive to both surface snow conditions and atmospheric conditions. Snow forcing provides an additional moisture source, but any increase in evaporation/sublimation that results is countered to varying degrees by decreases due to increased stability of the lower atmosphere [*Cohen and Rind*, 1991], which result from the lower surface temperatures and decreased sensible heat flux. Other negative feedback mechanisms include decrease evaporative capacity in the atmosphere due to lower temperatures and increased moisture content. For

SIB, what results is an essentially negligible change in this latent heat flux term. For COV, the reduction in surface moisture availability is offset by greater atmospheric instability and other feedbacks, and a modest net increase in evaporation/sublimation occurs. For INS, the substantial decrease in reflected SW radiation that occurs in the absence of the albedo effect disrupts this balance, and provides additional energy for evaporation/sublimation. In the early season this latent heat flux generally occurs as evaporation of melted snow. During the winter months snowmelt is minimal, so the added energy is used for direct sublimation from the snowpack. The latent heat flux response due to evaporation/sublimation is not directly attributable to snow forcings for SIB and COV due to atmospheric interactions, but for INS the large influx of net SW radiation results in a clear positive response.

4.3. Thermodynamic Contributors to the Snow-Forced Temperature Decrease

[31] Note in Figures 5 and 6 that no single component of the surface energy balance dominates the surface thermodynamic response, for any of the three experiment pairs. The ultimate surface temperature response results from the contributions of, and complex interaction between, the various thermodynamic mechanisms. For SIB, the temperature decrease due to a comprehensive snow forcing arises from three processes: decreased net SW radiation due to the albedo effect, decreased upward ground heat flux due to the low thermal conductivity of the snowpack, and increased latent heat flux due to early-season snowmelt. Other processes such as outgoing LW radiation and sensible heat flux decrease in response to the temperature decrease, which serve as negative feedback mechanisms on the overall temperature response.

[32] For COV, only a snow cover forcing is prescribed, and the resulting temperature decrease is notably damped. Responsible processes for this mitigation are: a more moderate decrease in net SW radiation since albedo is somewhat influenced by snow depth, a more moderate decrease in the upward ground heat flux since the snowpack is effectively thinned, and a more moderate increase in latent heat flux due to early-season snowmelt since there is less snow to melt. Negative feedbacks associated with LW radiation and sensible heat flux are also moderated due to the mitigated temperature response. With the removal of snow depth forcings, all of the snow-forced thermodynamic responses identified for SIB still occur, but the magnitudes are smaller. Consequently the surface temperature response with a snow cover only forcing is notably less than that for a comprehensive snow forcing. This result underscores the contribution of snow depth anomalies to local climate fluctuations, and confirms that the response to snow cover anomalies is not an albedo effect only.

[33] For INS, a comprehensive snow forcing is prescribed but the albedo effect is explicitly suppressed. The most obvious consequence is a nullified snow-forced decrease in net SW radiation, since the prescribed snow is prevented from reflecting away any SW radiation. If albedo had been the sole forcing mechanism on the climate associated with snow anomalies, then its suppression would have resulted in little or no snow-forced change in the surface energy balance, and consequently a minimal change in surface temperature. However, the results for SIB confirm that other mechanisms such as ground heat flux due to snow insulation and latent heat flux due to snowmelt are also affected by snow, and these processes produce a noticable decrease in surface temperature for INS despite the lack of an albedo effect. Furthermore, the increase in latent heat flux due to early-season snowmelt is amplified, since SW radiation that would normally be reflected away instead provides more energy for snowmelt. Similarly, the added SW surface energy input also generates latent heat flux due to early-season evaporation of melted snow and late-season sublimation. These enhanced latent heat fluxes further contribute to the modeled surface temperature decrease for INS. Thus it is clear that albedo is not the exclusive thermodynamic process involved in snow-forced local climate fluctuations.

5. Hemispheric Climate Mode Response to Snow Forcings

[34] In this section we present a summary evaluation of the snow-forced response to remote climate features over the extratropical Northern Hemisphere, for all three experiment pairs. In GEC03, it was shown that the local surface temperature decrease over Siberia during autumn produced by SIB, in response to a comprehensive and realistic snow forcing, also initiates an atmospheric teleconnection pathway. This dynamical pathway involves stationary wave flux and mean flow anomaly interactions throughout the troposphere and stratosphere, enabled by the unique co-location of a major center of stationary wave activity (partially forced by mountainous conditions in eastern Siberia) within the snow anomaly [*Gong et al.*, 2004]. A downward propagating, hemispheric-scale signal results during the winter season, comprised of a southward shift in the zonal wind field, and positive (negative) geopotential height and sea level pressure anomalies at high (mid) latitudes. These characteristics depict the negative phase of the Arctic Oscillation (AO), one of the dominant modes of atmospheric variability in the extratropical Northern Hemisphere [*Thompson and Wallace*, 1998]. It is of interest to see whether the mitigated local surface temperature response for COV and INS, associated with partial snow/thermodynamic forcings, is sufficient to trigger the teleconnection pathway and ultimate negative AO mode modulation.

[35] Figure 7a, 7c, and 7e shows the vertical wave activity flux (WAF; *Plumb* [1985]) response to snow at 850 hPa elevation during autumn, for SIB (repeated from GEC03 Figure 5), COV and INS. The WAF diagnostic describes the three-dimensional transmission of stationary wave energy throughout the atmospheric system, and is produced in part by large-scale diabatic heating anomalies. A clear snow-forced enhancement of upward WAF over Siberia occurs for SIB, propelled by the strong local temperature decrease. This upward wave activity propagation represents the first leg of the teleconnection pathway. For both COV and INS, the upward WAF response still occurs but is notably damped, which is not surprising given the mitigated temperature response. Hence with only partial snow/thermodynamic forcings the teleconnection pathway is compromised at its incipient stages.

[36] Figure 7b, 7d, and 7f shows the corresponding sea level pressure (SLP) response to snow during winter, which

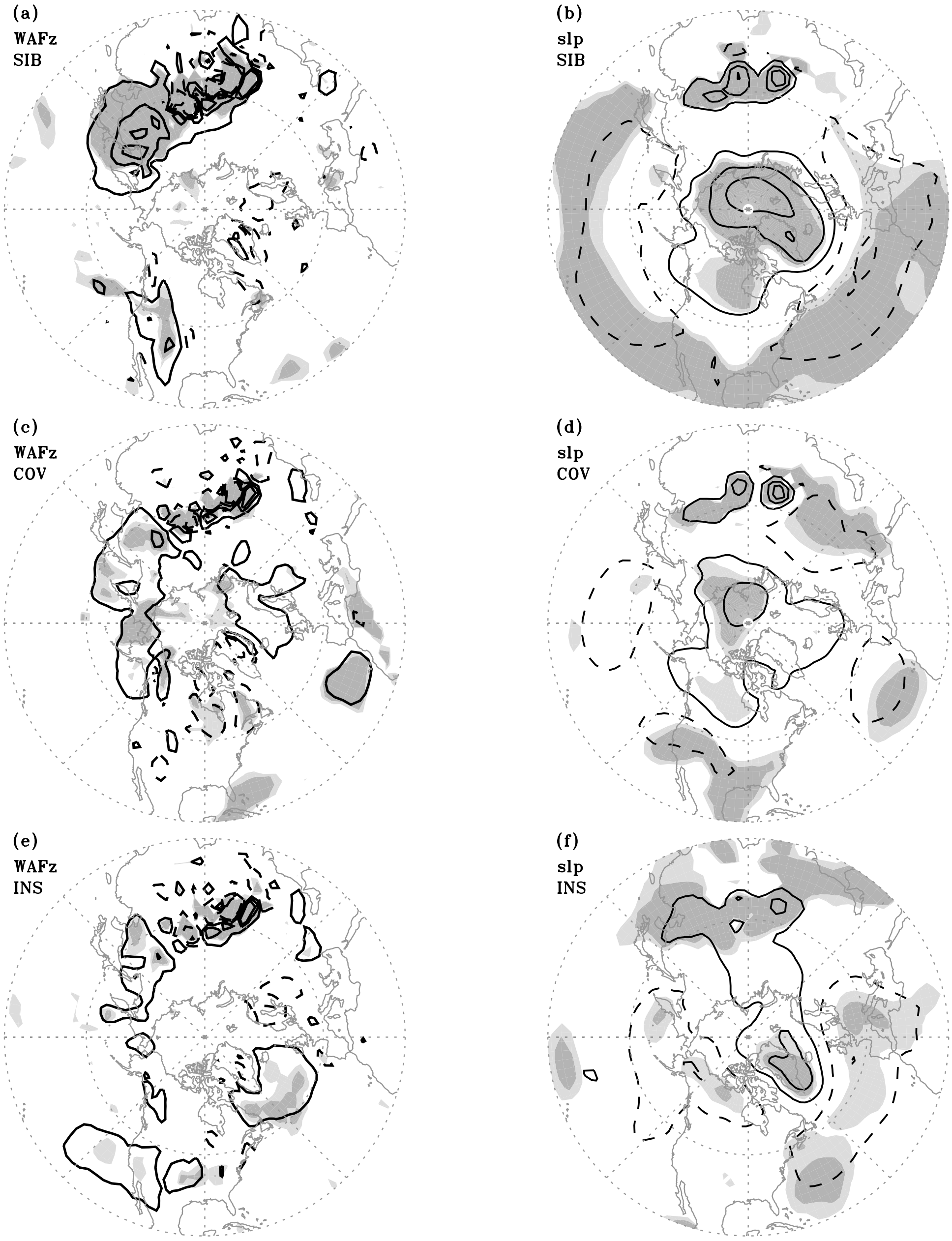

Figure 7. Seasonal average response to positive Siberian snow forcing, over the extratropical Northern Hemisphere, for SIB (a, b) repeated from GEC03 Figures 5 and 4f, COV (c, d), and INS (e, f). (a, c, e) Autumn (SON) vertical wave activity flux contours drawn at ±0.01, 0.04, 0.08 m^2s^{-2}. (b, d, f) Winter (DJF) sea level pressure contours drawn at ±1, 3, 5 hPa. Dashed line denotes negative contour value. Light (dark) shading indicates 90% (95%) statistical significance.

represents the ultimate winter AO mode modulation produced by the teleconnection pathway. For SIB (repeated from GEC03 Figure 4f), the comprehensive and realistic Siberian snow forcing results in a dipole SLP anomaly field that is clearly indicative of the negative AO pattern. Compared to the AO mode response between the two snow-forcing years derived from reanalysis data, the spatial pattern is very similar (+0.87 grid point correlation coefficient), but the modeled aggregate snow forcing only accounts for roughly 30% of the observed total AO mode response magnitude (see GEC03 for details). For COV and INS, the SLP anomaly fields somewhat resemble the negative AO mode, but the centers of action are noticeably weaker, and the midlatitude anomalies are also less coherent. It is unclear whether Figures 7d and 7f represent a snow-forced AO mode modulation at all. Thus when only partial snow/thermodynamic forcings are applied over Siberia, the compromised teleconnection pathway results in at best a considerably diminished winter climate mode response. Clearly the full suite of snow cover and snow depth anomalies over Siberia, and all relevant thermodynamic processes, are required to distinctly modulate hemispheric-scale climate as represented by the winter AO mode.

[37] Although the seasonally-averaged WAF and SLP anomaly fields shown in Figures 7c–7f are similar for COV and INS, these partial snow/thermodynamic forcings may not necessarily have an equivalent effect on the precise nature of the teleconnection pathway and AO mode response. Snow cover anomalies are transient in nature, and produce abrupt but short-lived temperature responses as the snow line migrates southward during the autumn season. Hence for COV, Siberia as a whole does not experience a sustained snow forcing and temperature response. For INS, insulation mechanisms are in effect throughout the simulation period, since snow depth forcing is included. Thus Siberia as a whole experiences a less abrupt but temporally sustained temperature response. These subtle differences may lead to noticeably different remote responses, but are likely not captured in the seasonal, monthly, and Siberian averages presented here. A more complete evaluation of the dynamical atmospheric response to partial snow/thermodynamic forcings and its temporal evolution is not addressed here but is the subject of ongoing research.

6. Conclusions

[38] Recent studies demonstrate that land surface snow anomalies can influence both local and remote climate features. To fully utilize any potential climate predictability contained in snow anomaly signals, it is important to understand the atmospheric sensitivity to snow anomaly characteristics (cover vs. depth) and the relevant thermodynamic processes. Such an investigation is conducted in this study, via three sets of large-ensemble numerical GCM experiments. The SIB experiments evaluate the atmospheric response to a realistic, observation-based positive snow forcing, which include both snow cover and snow depth anomalies. For COV, only snow cover forcings are incorporated. Finally for INS, the albedo response to snow is explicitly suppressed, to evaluate the atmospheric response to a comprehensive snow forcing, but via snow insulation processes only.

[39] One conclusion that can be drawn from this study is that both snow cover and snow depth anomalies are important contributors to the local surface temperature response. When snow depth forcings are removed in COV, the resulting temperature decrease is notably damped compared to SIB. As summarized in Section 2, numerous studies which detect a statistical or modeled relationship between snow and climate attribute the climate response to snow cover extent anomalies. However, these studies by and large neglect to consider the relative contribution of snow depth. The few studies which do so have generally concluded that snow depth is an important contributor, to both local and remote climate. The results of this study corroborate the importance of snow depth anomalies on climate variability.

[40] A second conclusion is that surface albedo is not the exclusive thermodynamic mechanism for producing snow-forced local temperature anomalies. For SIB, the high albedo of snow, the low thermal conductivity of snow, and the occurrence of snowmelt during the early season each impact a different component of the surface energy balance, and are all found to contribute measurably to the resulting temperature decrease. Thus when the albedo effect is suppressed in INS, the other mechanisms still occur, resulting in a diminished but still significant temperature decrease. Many studies cite the predominance of the albedo effect in producing snow-forced climate atmospheric anomalies (see section 3.3). However, studies which include a surface energy balance assessment generally recognize other mechanisms as well (see section 3.4). The surface energy balance analysis in this study provides further evidence that albedo is not the sole responsible mechanism.

[41] Finally, our experiments indicate that realistic snow cover and snow depth anomalies acting in conjunction are required to produce a local temperature response which is strong enough to distinctly modulate the winter AO mode. This result supports the theory that recent efforts to correlate observed snow cover fluctuations to various climate phenomena may be neglecting a critical component of the forcing mechanism, namely snow depth [*Bamzai and Shukla*, 1999]. The focus on snow cover has been unavoidable to a degree due to the historical lack of reliable spatially comprehensive snow depth data. The ongoing development of passive microwave, remotely-sensed snow depth products over broad spatial scales may be of tremendous potential benefit in this regard, both by providing independent data with which to verify the model results to partial snow forcings presented here, and by facilitating even more insightful observational and modeling studies.

[42] Of course, these conclusions are predicated on the accurate parameterization of snow albedo and other surface thermodynamic processes in the ECHAM3 GCM. Although the relevant snow and climate features are all reproduced reasonably well, the model's land surface parameterization, derived from the Simple Biosphere model (SiB; *Sellers et al.* [1986]), handles snow thermodynamics in a fairly simplistic manner. *Foster et al.* [1996] and *Frei and Robinson* [1998] report that the interannual variability of snow cover and snow mass is generally underestimated by ECHAM3, as well as other GCMs. *Roesch et al.* [1999] asserts that the ECHAM4 GCM underestimates snow albedo due to the simplifications inherent in the land surface scheme, and *Roesch et al.* [2001] presents an improved snow cover fraction parameterization. *Stieglitz et al.* [2001] describes a detailed snow physics scheme which leads to improvements in snowmelt and insulation processes. *Frei et al.* [2003] reports that GCM snow cover variability is improved considerably in AMIP2 vs. AMIP simulations.

[43] These studies suggest that the snow thermodynamic processes, and hence the local and remote climate response to snow, may be underrepresented in our experiments. One possibility is that the incorporation of more sophisticated land surface and snow schemes will enhance the atmospheric sensitivity to the prescribed snow forcings, and produce a stronger climate response. However, due to the number and complexity of the thermodynamic processes involved, it is difficult to surmise the effect of improved physics. This study demonstrates that our hypothesized snow cover/depth - climate relationships can be discerned by a representative and respected current-generation GCM and its default snow scheme. An explicit assessment and comparison between various GCMs and snow schemes is the subject of ongoing research, and will be addressed in future papers by the authors.

[44] Another complicating factor involves the snow forcing specification procedure. Since snow is maintained by the model as an internal state variable, prescribing surface snow depths at every timestep as a model boundary condition necessarily encroaches on the model's surface water and energy balances. Other studies follow a somewhat different approach by adjusting the amount of snowfall produced by the model atmosphere, but in either case snow mass is artificially created or destroyed. All snow-forced modeling studies are faced with this difficulty, however these practices become more questionable given the importance of latent heat flux responses in this study. For example, the large and sudden introduction of prescribed snow forcings during the warm autumn months (Figure 2) is likely to induce more snowmelt than if such a snow anomaly had occurred naturally within the model. Thus the latent heat flux response to snow may be somewhat exaggerated in our externally snow-forced study. More sophisticated GCMs with improved snow parameterizations may be able to better capture observed snow characteristics and variability, and facilitate a similar analysis of snow-forced climate response using internally-generated snow anomalies.

[45] Finally, it should be noted that our modeled winter climate response to snow cover and snow depth forcings over Siberia may not be applicable to other regions and times. The modeled perturbation region and season was selected precisely because of its previously demonstrated atmospheric response to snow anomalies and potential for distinct snow cover and snow depth forcings. Other snow forcing scenarios such as winter snow anomalies in polar regions and ephemeral snow anomalies during the warm season may not exhibit the same combination of thermodynamic responses produced here. Moreover, two related papers by the authors [*Gong et al.*, 2003b, 2004] identify geographic and orographic features unique to Siberia which enable the hemispheric-scale climate response described in Section 5. Taken together, our results specify autumn in Siberia as a particularly influential season and region for local and remote atmospheric sensitivity to land surface snow anomalies, which therefore holds considerable potential for snow-forced climate predictability.

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J. Cohen, Atmospheric and Environmental Research, Inc., Lexington, MA 02421, USA.

D. Entekhabi, Department of Civil and Environmental Engineering, Massachusetts Institute of Technology, Cambridge, MA 02139, USA.

G. Gong, Department of Earth and Environmental Engineering, Columbia University, New York, NY 10027, USA. (gg2138@columbia.edu)

D. Robinson, Department of Geography, Rutgers University, Piscataway, NJ 08854, USA.