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SIMULATING THE INFLUENCE OF GREAT PLAINS SNOW COVER ON THE THERMAL CHARACTERISTICS OF A COLD AIR MASS

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ABSTRACT: A better understanding of the influence of snow cover on the thermal properties of the overlying atmosphere may assist in future improvement of climate and weather prediction models. This study utilizes an adapted one-dimensional snow pack model to simulate and analyze the relationship between snow cover and the thermal properties of a cold air mass moving across the central United States in winter. Thermal characteristics of the air mass are derived from model equations governing the heat balance between the underlying surface and the lower atmosphere.

INTRODUCTION

Snow cover and its interaction with boundary layer climate variables has been the subject of much research over the past several decades. In particular, there is increasing interest in the role of snow cover in climate diagnostics and climate change. Much of the attention corresponds with an increase in concerns associated with potential changes in the global environment as a consequence of anthropogenic and natural causes. Consequently, a greater emphasis has been placed on the accuracy of the representation of the interaction between the atmosphere and snow cover in climate and weather prediction models.

At its peak in winter, snow covers approximately 46% of the land surface in the Northern Hemisphere, or about 46 million km^2 (Robinson et al., 1995). Due to the radiative effects of snow, large changes in hemispheric or global snow cover can result in climate modifications through much of earth's troposphere. Research has indicated that snow cover can lower surface air temperatures over time scales of days to months by increasing the surface albedo and through the latent heat of melting (e.g. Dewey, 1977; Heim and Dewey, 1984; Walsh et al., 1985; Namias, 1985; Baker et al., 1992; Leathers and Robinson, 1993; Leathers et al., 1995). Several empirical studies have investigated atmosphere-snow cover relationships on local and regional scales, and large-scale climate models incorporate snow cover as an important variable. Little research has been conducted in the area of modeling the synergistic relationship between snow cover and the lower atmosphere on synoptic spatial and temporal scales.

This study uses a one-dimensional snow pack model to simulate and analyze the relationship between snow cover and the thermal properties of a cold air mass moving across the central United States in winter. The study area encompasses the Great Plains of the United States (Figure 1), a region favorable for snow cover research due to its homogeneous topography and vegetation characteristics. The Great Plains region is characterized by a flat topography, low grass vegetation, and a frequency of winter cold air mass intrusions conducive to sustained snow cover.

SNOW PACK MODEL

A one-dimensional mass and energy balance model for a snow pack (SNTHERM; Jordan, 1991) is adapted for use in this study. In predicting temperature profiles within snow and soil the model addresses the various conditions common in winter, from initial ground freezing in fall to snow ablation in spring, and is capable of adapting to a full range of meteorological conditions (Jordan, 1991). The transport of liquid water and water vapor is included in the model for use in the heat

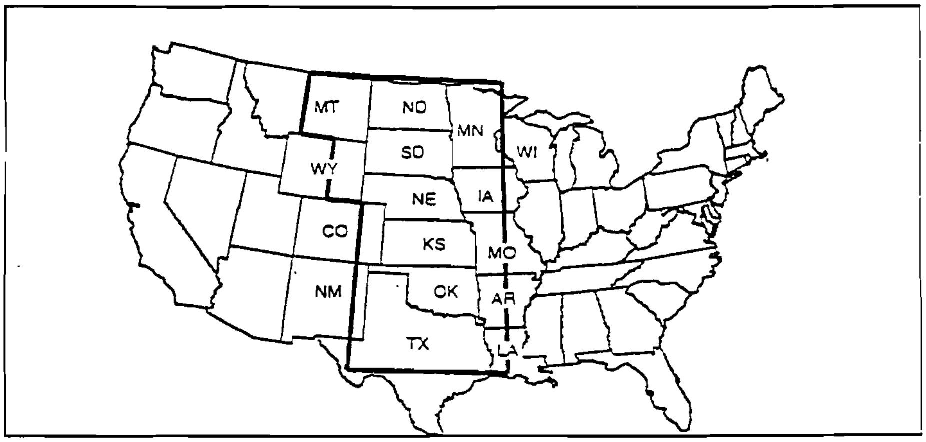

Figure 1. The study region encompassing the Great Plains.

balance equations.

The governing equations for heat and mass balance are applied to horizontally-infinite control volumes in order to obtain a numerical solution of the temperature profile. The boundary conditions for the model are determined by the meteorological conditions at the air/surface interface. Surface energy fluxes are calculated from measured meteorological data including air temperature, relative humidity, wind speed, precipitation, and measured or calculated solar and incoming infrared radiation. The model is capable of calculating radiation values through routines which consider the solar aspect, cloud conditions, albedo, and inclination of the surface (Jordan, 1991). Profiles of temperature and water content for the various control volumes initialize the model, and the physical strata can be supplied by the user or extracted from internal data bases for snow, sand, and clay.

The surface energy balance within the model contains the turbulent fluxes for sensible and latent heat, shortwave and longwave radiation, and convected heat from precipitation (Jordan, 1991). The surface energy flux can be written as:

$$I = -R_{sw}\downarrow(1 - \alpha_s) - R_{lw}\downarrow + R_{lw}\uparrow + H + LE + CV \quad (1)$$

where $R_{sw}\downarrow$ is the downward flux of energy from solar radiation (Wm^{-2}), α_s is the surface albedo, $R_{lw}\downarrow$ and $R_{lw}\uparrow$ are the downward and upward components of longwave radiation (Wm^{-2}), and H, LE, and CV are the turbulent fluxes of sensible, latent, and convective heat (Wm^{-2}; Rowe et al., 1995). The upward flux of longwave radiation can be written as:

$$R_{lw}\uparrow = \epsilon_s \sigma T_0^4 + (1-\epsilon_s) R_{sw}\downarrow \quad (2)$$

where ϵ_s is the emissivity of the surface, σ is the Stefan-Boltzmann constant (5.669 x 10^{-8} $Wm^{-2}K^{-4}$), T_0 is the surface temperature (K), and $R_{sw}\downarrow$ is the downward flux of shortwave radiation (Wm^{-2}). The sensible heat flux can be expressed as:

$$H = \rho_a C_p C_H w (T_a - T_0) \quad (3)$$

where ρ_a is the density of air (kgm^{-3}), C_p is the specific heat of air at constant pressure (1005 $Jkg^{-1}K^{-1}$), C_H is the bulk transfer coefficient for sensible heat, w is the wind speed (ms^{-1}), and T_a and T_0 are the air and surface temperatures (K). The latent heat flux can be expressed as:

$$LE = L_{vi} C_E w (\rho_{v,a} - \rho_{0v,sat}) \quad (4)$$

where L_{vi} is the latent heat of sublimation (2.838x10^6 Jkg^{-1}), C_E is the bulk transfer coefficient for latent heat, $\rho_{v,a}$ is the vapor density in air (kgm^{-3}), and $\rho_{0v,sat}$ is the intrinsic water vapor density at saturation at the surface (kgm^{-3}). The bulk transfer coefficients are calculated using the roughness length and are considered to be equal. For neutral stability:

$$C_H = C_E = k^2/[\ln(z/z_0)]^2 \quad (5)$$

where z is the measurement height above the surface (m), z_0 is the roughness length (m), and k is von Karman's constant (0.40). For non-neutral conditions, the standard stability adjustment is made through the calculation of the bulk Richardson number (Rowe et al., 1995; Jordan, 1991).

METHODOLOGY

The design of this research is based on the premise that a pure cold air mass is a homogeneous, self-contained volume of air possessing a core which is not subject to advection from outside. Thus, the core of a simulated air mass possesses thermal characteristics that change only in response to variations in the energy flux across the air-surface interface. The energy flux varies with changing surface and meteorological conditions, and with changes in the amount of radiation within the system. In order to study snow cover-cold air mass interactions in this manner, it is necessary to alter the methodology of the snow pack model.

The model equation for the surface energy balance contains variables for air temperature within the turbulent heat flux terms. In this study, the core of a cold air mass is assumed to be precipitation free, eliminating the term for the turbulent flux of convective heat. Thus, the model equation for the energy flux across the air-surface interface becomes:

$$I_{air/surf} = R_{nlw} + H + LE \qquad (6)$$

where R_{nlw} is the net longwave radiation, H is the sensible heat flux, and LE is the latent heat flux. Within the model, the sensible heat flux is defined as:

$$H = Q_{sen}(T_a - T_0) \qquad (7)$$

where T_a and T_0 are the air and surface temperatures respectively. The term Q_{sen} is a wind function for sensible heat exchange and is defined as:

$$Q_{sen} = C_{sk} + \rho_a C_p C_H S w \qquad (8)$$

where ρ_a is the air density, C_p is the specific heat of air at constant pressure, C_H is the bulk transfer coefficient for sensible heat, S is a stability function, w is the wind speed, and C_{sk} is the windless exchange coefficient for heat (Jordan, 1991). The latent heat flux can be written as:

$$LE = Q_{lat}(e_a - rh_{fo}e_s) \qquad (9)$$

where e_a is the vapor pressure in the air, rh_{fo} is the fractional humidity of the surface relative to steady state, and e_s is the vapor pressure at the surface. The term Q_{lat} represents a wind function for latent heat exchange and is defined as:

$$Q_{lat} = (C_k + wSC_e(2.5045x10^8))/(R_w T_a) \qquad (10)$$

where C_k is the windless exchange coefficient for water vapor, S is a stability function, C_e is the bulk transfer coefficient for water vapor at neutral stability, w is the wind speed, R_w is the gas constant for water vapor, and T_a is the air temperature.

Upon substituting the expanded terms for H and LE (7-10) into the energy flux equation (6), a quadratic function for air temperature can be derived in the form $ax^2 + bx + c = 0$ where:

$$a = R_w Q_{sen} \qquad (11)$$

$$b = -R_w Q_{sen} T_0 - R_w I_{air/surf} + R_w R_{nlw} \qquad (12)$$

$$c = (C_k + wSC_e(2.5045x10^8))(e_a - rh_{fo}e_s) \qquad (13)$$

and $x = T_a$.

The quadratic formula can be used to solve for air temperature:

$$x = (-b + (b^2 - 4ac)^{1/2})/(2a) \qquad (14)$$

Measurements of surface air temperature, relative humidity, wind speed, atmospheric pressure, fraction of cloud cover, and cloud height, in combination with specified control volume characteristics describing the surface conditions, serve as the initial conditions for the model. In simulating the surface air temperature at subsequent time steps, the simulated surface air temperature and measured meteorological conditions from the previous time step are coupled with the surface conditions at the

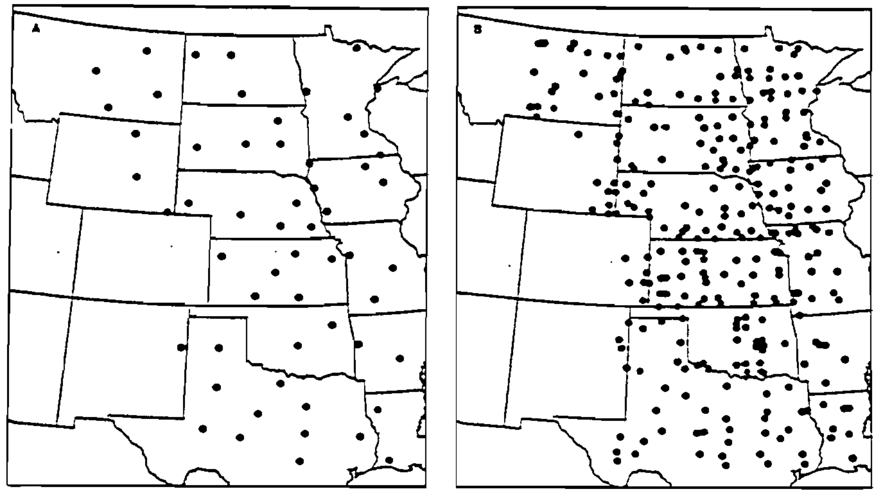

Figure 2. Station distributions for meteorological (a) and snow cover (b) data.

current location in order to calculate a surface energy flux and adjust the surface conditions. Subsequently, the surface energy flux is combined with the current meteorological and surface conditions in the quadratic function in order to calculate the surface air temperature for the current time step. It is assumed that over very short time increments the energy flow between the lowest levels of the air mass and the surface varies only with changes in the surface and atmospheric conditions, and with changes in radiation input. At each time step, measured surface air temperature and relative humidity values are used to specify the current water vapor pressure in the atmosphere.

The cold air mass used for analysis in this research affected the study region during the period 1-4 January, 1991. Characterized by a strong anticyclone, the cold air mass moved from the southwestern plains of Canada, southeastward over the plains region of the United States. Maximum daily temperatures across the study region on 3 January generally ranged from -20°C to 5°C and the central pressure of the anticyclone exceeded 1040 mb. Meteorological data used to characterize the air mass and serve as model input are obtained from the EarthInfo, Inc. National Climatic Data Center (NCDC) Surface Airways database as a subset of the NCDC TD-3280 and TD-3281 databases. Data extracted for this study include surface measurements of air temperature, dew point temperature, sea-level pressure, relative humidity, and wind speed, along with fraction of sky covered by clouds and cloud height. Data are obtained for 57 stations that are relatively homogeneously distributed across the study region (Figure 2a). Missing data for up to three consecutive hours are linearly interpolated while missing cloud height values are determined through a convective cloud base scheme. Stations missing data for greater than three consecutive hours are eliminated from the study.

During the period 1-4 January, 1991 a uniform snow cover of at least 2.5 cm characterized approximately two-thirds of the study region. For this study, daily snow depth data are extracted from the Historical Daily Climate Dataset (HDCD; Robinson, 1988) for 286 stations located across the study area (Figure 2b). During the development of the HDCD, the data were quality controlled using several parameters formulated by Robinson (1988). For the stations used in this study, a missing day of snow depth data is linearly interpolated using data from each shoulder day. Stations with two or more consecutive days of missing data are omitted from the study.

The *Geographic Resource Analysis and Support System* (GRASS) geographic information system (GIS) is used to interpolate meteorological and snow cover data, using an inverse distance weighted averaging method, to a 1° latitude by 1°

longitude grid covering the study region. The hourly position of the center of the anticyclone associated with the cold air mass is monitored using gridded hourly sea-level pressure output from GRASS. The time and position of the entrance and exit of the anticyclone to and from the study region is used to determine a linear, hourly change in latitude and longitude, and is subsequently used to position the air mass during simulation. The core of the cold air mass is defined by a 5° latitude by 5° longitude box around the center of the anticyclone. The result is a cold air mass core represented by 25 1° latitude by 1° longitude grid cells.

In order to determine the surface conditions at each grid cell over which the simulated air mass passes, snow cover and meteorological data are entered into the original SNTHERM snow pack model. For surface grid cells with no previous interaction with the simulated air mass, the initial temperature profile for snow is defined as isothermal and equal to the measured air temperature. For soil, the initial temperature of the top control volume is set equal to the measured air temperature and increased by 0.5°C for each control volume below. The remaining meteorological and climatological conditions for the particular time and location are used in the snow pack model to adjust the snow/soil temperature profile. The one-hour time step is repeated four times using profile output from the previous adjustment. Within the model, temperature convergence for the control volumes is required at time steps much below one hour, producing a temperature profile that reaches equilibrium early in the four-hour iterative procedure. For grid cells previously affected by the simulated air mass, initial temperature profiles for the iterative procedure are determined through the process outlined above, only instituting the simulated air temperature occupying the grid cell during the previous hour.

In simulating surface air temperatures, the 25 grid cells representing the core of the cold air mass are positioned hourly through the linear change in latitude and longitude. Temperature simulations along a grid cell path are initialized with a measured air temperature and temperature simulations end as a grid cell exits the study region. The core of the cold air mass analyzed in this research entered the study region over northeastern Montana from southern Canada at 1600 CST on 1 January 1991. The last grid cell representing a portion of the air mass exited the study region on 1600 CST 3 January 1991 over central Missouri. For the 49-hour period during which the air mass was positioned over the study region, the surface over which it passed possessed a uniform snow cover with depths ranging from 2.5 cm along the southern boundary, to 15 cm along portions of the northern boundary. Simulated air mass temperatures are calculated using measured meteorological and snow cover data and are compared to measured air temperatures. Additionally, air mass temperature simulations are conducted over uniform snow depths of 0 cm, 2.5 cm, 15 cm, and 30 cm. Mean simulated and mean measured air mass temperatures are calculated using values from all of the air mass grid cells within the study region at each hour of the simulation. Comparisons are made by compiling time series and model output statistics. Isotherms are constructed in order to identify spatial differences in the thermal properties of the air mass for the various snow cover conditions.

RESULTS AND DISCUSSION

Time series of the mean measured and mean simulated hourly air mass temperatures over the 49-hour period 1600 CST 1 January through 1600 CST 2 January 1991 show the ability of the model to simulate the air mass temperatures (Figure 3). The diurnal cycle of temperature is reasonably reproduced as are many of the hourly temperature variations. Additionally, the general warming of the air mass with the decrease in latitude is simulated well. Diagnostic statistics (Table 1) further reveal the performance of the model in simulating the mean surface air temperature of the air mass. A high correlation coefficient along with small root mean squared and mean absolute errors indicate an effectiveness of the model in reproducing the variability and magnitude of the measured mean air mass temperatures. Over the full 49-hour period, the mean simulated air mass temperature is 0.44°C cooler than the mean measured temperature of the air mass, and the standard deviation is slightly smaller in the case of

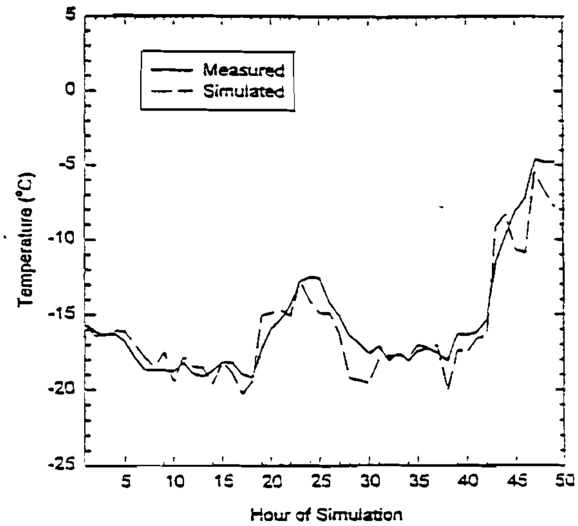

Figure 3. Time series of the mean measured and mean simulated air mass temperatures from 1600 CST 1 January to 1600 CST 2 January.

Table 1 Root mean squared error (RMSE), mean absolute error (MAE), and the correlation coefficient (r) between the simulated and measured air mass temperatures, and the standard deviations (SD) and average temperatures (T) for the simulated and measured air masses.

RMSE	1.37
MAE	1.04
r	0.94
T_{sim} (°C)	-15.95
T_{mes} (°C)	-15.51
SD_{sim} (°C)	3.55
SD_{mes} (°C)	3.87

the simulated air mass (Table 1). A smaller standard deviation is expected based on the assumption within the methodology that the airmass is homogeneous and not subject to advection. Thus, large variance in the thermal characteristics of the simulated air mass are not expected.

Simulated air mass temperatures under uniformly bare ground conditions are substantially warmer than simulated temperatures using measured snow cover (Figure 4a). The temperature difference is most pronounced during daylight hours when surface albedo is most important. This corresponds with the findings of Baker et al. (1992) and Leathers et al. (1995) that differences between days with snow cover and days with no snow cover are greater for maximum daily temperatures than for minimum daily temperatures. Additionally, the differences in mean air mass temperatures are slightly greater during the second diurnal peak when the air mass is positioned six degrees of latitude farther south than 24 hours earlier (Figure 4a). For the full 49-hour period, the mean air mass temperature is 3.52°C warmer under bare ground conditions than under measured snow cover

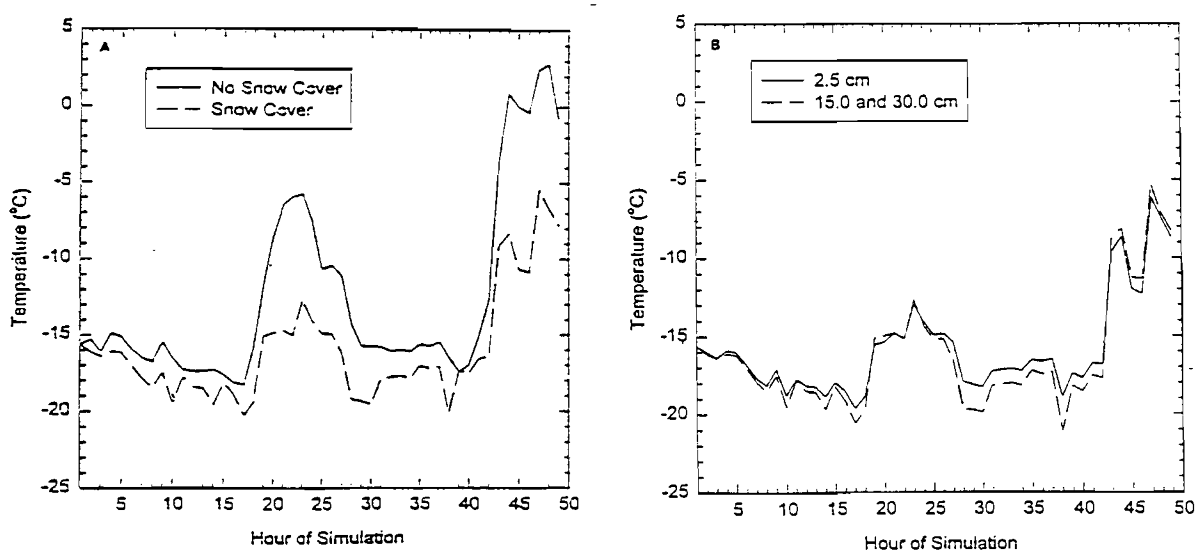

Figure 4. Simulated air mass temperatures under measured snow cover and no snow cover conditions (a), and under 2.5 cm, 15.0 cm, and 30.0 cm snow depth conditions (b), for the period 1600 CST 1 January through 1600 CST 2 January.

Table 2 Mean temperatures and standard deviations for the 49-hour air mass simulations using measured and fixed snow cover conditions.

Snow cover	T_{mean}(°C)	SD(°C)
0.0 cm	-12.43	6.08
2.5 cm	-15.80	3.10
measured	-15.95	3.55
15.0 cm	-16.19	3.63
30.0 cm	-16.20	3.61

conditions (Table 2). The temperature difference is less than those found by Baker et al. (6.4-8.4°C) and Leathers et al. (5-6°C). However, each of these studies examined daily maximum and minimum temperature data for individual stations for a large sampling of winter days with and without snow cover. Thus, several types of synoptic weather patterns much different from the cold air mass studied here (e.g. anomalously warm patterns) were likely included in their studies. The standard deviation of the simulated temperatures is considerably higher without a snow cover (Table 2) as a result of the ground surface warming dramatically during daylight hours, but cooling substantially at night.

Varying the depth of snow cover uniformly across the study region produces minimal differences in mean air mass temperatures (Figure 4b). A small snow depth of 2.5 cm produces slightly warmer temperatures than snow depths of 15 cm and 30 cm, predominantly during nighttime hours. Small daytime differences are likely the result of the nature of the modeled snow/ground surface. Within the model the ground surface is portrayed as flat and without vegetation or any other surface feature which could protrude through a shallow snow cover. Thus, the surface characteristics (e.g. albedo) of a snow cover within the model are consistent regardless of the depth. As a result, variations in snow depth alter only the capacity of the snow pack to transmit heat through to an adjacent body. The magnitude of the soil heat flux into an overlying snow pack becomes more significant during nighttime hours when insolation values are zero. Therefore, warmer air temperatures associated with the 2.5 cm snow depth, particularly at night, are likely the result of a minor influence by the soil heat flux. Over the course of the 49-hour simulation period, the difference in air mass temperatures using a 2.5 cm snow depth versus a 15 cm or 30 cm snow depth increases slightly (Figure 4b). This is likely the result of an increase in the temperature throughout the profile of the thinner snow pack as a result of the first diurnal peak in air temperature, but also the result of an increase in the solar angle resulting from the decrease in latitude. The difference in air temperatures is essentially zero between simulations using 15 cm and 30 cm snow depths and is represented as such here (Figure 4b). Interestingly, Baker et al. (1991) concluded that snow depths of 5 cm or greater effectively mask an underlying bare soil surface. Mean air mass temperatures and standard deviations for the 49-hour simulation period under the varying snow depth conditions are given in Table 2.

Spatial analyses at various times within the 49-hour period (not shown) reveal that temperature differences between simulations using measured snow cover and bare ground increase from southeast to northwest along the axis of the direction of air mass movement. This signifies the cumulative effect of a positive feedback process under bare ground conditions. Warmer air mass temperatures increase soil surface temperatures, which further affect the temperatures of the upstream portion of the air mass as it passes over the warmer surface.

To summarize the findings of this research, simulating the thermal properties of the cold air mass passing over the Great Plains of the United States during 1-3 January 1991 reveals: (1) average air mass temperatures for the 49 hours of simulation are 3.52°C warmer under bare ground conditions than under measured snow cover conditions, with the greatest difference occurring during daylight hours (5-10°C); (2) varying the depth of a uniform snow cover across the region produces only minimal differences in mean air mass temperatures, especially between snow depths larger than 2.5 cm which presumably mask the flux of heat from the underlying ground; and (3) the effects of the variations in snow cover become greater as the cold air mass passes into the lower latitudes of the Great Plains, reflecting the importance of sun angle in surface heating.

This research is currently being expanded through investigations involving variations in snow cover and climatological conditions. Additionally, it is necessary to confirm the findings contained in this paper and the results of further investigations through work with additional cold air mass scenarios.

ACKNOWLEDGEMENTS

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