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Brief communication "Modeled rain on snow in CLM3 warms soil under thick snow cover and cools it under thin"

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Abstract

Rain-on-snow has decimated ungulate herds in the North and warmed permafrost significantly in Spitsbergen. As the permafrost temperatures are used as an integrated signal of the climate change, there is an urgent need to characterize the relationship between rain-on-snow and permafrost temperatures. By incorporating reanalysis based (ERA40) climate forcing into the land model (CLM3) and introducing an artificial rain on snow event on all model pixels the areas with thick snow cover (>0.5 m) experienced season average permafrost warming, sites with intermediate snow depths (0.15–0.5 m) experienced cooling, while sites with thin snow cover were more sensitive to other factors.

1 Introduction

A large body of modeling and observational data shows that the high latitudes are leading in the warming of the globe (e.g. Serreze et al., 2000; IPCC, 2007). Since these vast areas are subject to rapidly changing climatic forcing it is important to better understand how they function to prepare us for the future and allow accurate detection of changes as manifested for example in the soil temperatures and annual soil thaw depths (Brown et al., 2008).

The soil and it's temperatures in the north are the subject of much interest because they are: (1) the recorder and archive of past near surface temperatures to aid in the characterization of climate change (e.g. Lachenbruch and Marshall, 1986), (2) the substrate of much of vulnerable infrastructure due to potential thaw of soil ice and resulting ground subsidence (e.g. Nelson et al., 2001), and (3) the reservoir of greenhouse gases that may be released into atmosphere due to soil warming (e.g. Oechel et al., 1993).

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The relations between climate, snow, active layer and permafrost temperatures have received increasing attention since the second world war and engineering challenges posed by the construction of the Alcan highway and the Prudhoe bay oil pipeline (for a recent compilation of the past research see: Hallet et al., 2004). However, rain on snow has just recently been recognized as an important thermal forcing for the seasonally snow covered soil temperatures at least where it has been studied (Putkonen, 1998).

Rain on snow (hereafter ROS) is well known to inhabitants of the cold climates, but is poorly documented due to the inherent difficulty in detecting liquid precipitation in a generally frozen environment. The routine detection of ROS is poor and consequently the true spatial distribution and frequency is not well known. Global Climate models predict significant increase in the area affected by ROS within the next 50–100 years as the Arctic is leading in the general warming of the globe (Putkonen and Roe, 2003). It has also been shown that the sparse station network in the north as well as the climate models have entirely missed severe ROS events in the past (Rennert et al., 2008) whose magnitude has been determined by anecdotal evidence (Nagy and Gunn, 2004) and remotely sensed passive microwave signature (Grenfell and Putkonen, 2008).

Using surface observations from a well-established field site in Arctic Spitsbergen in conjunction with nearby meteorological station data, Putkonen (1998) and Putkonen and Roe (2003) demonstrated that a large wintertime rain penetrates through the snow pack, and is followed by a slow freezing of the moisture within and below the snowpack. They showed that the rainfall exerts a much stronger influence on the thermal structure of the snowpack and soil than just an increase in the air temperature without the rain. Although infrequent, these events are capable of exerting considerable control over the mean wintertime soil temperature under the snow pack in their study site in Spitsbergen.

To extend the climatological work of Putkonen and Roe (2003) on snow and soil temperatures we use a well documented Community Land Model 3 (CLM3) (Oleson et al., 2004) and the coupled Community Climate System Model 3 (CCSM3). This allows us to study the effect of ROS on soil temperatures in the entire Northern Hemisphere

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seasonally snow covered land area. We use the CLM3 driven by current reanalysis data and compare the resulting soil temperatures to a separate CLM3 run where all snow covered land area is affected by a one time, artificially induced large ROS event. Our expectation is that ROS will generally warm the season averaged soil temperatures.

Although the current spatial pattern and frequency of ROS are not well known, it is much more likely that ROS occurs on coastal areas downwind of open ocean than in the generally drier and colder interiors of continents that are far from open ocean (Putkonen and Roe, 2003). However, large events have been documented for example in the Banks Island, Arctic Canada that in the winter is surrounded by sea ice and large continent.

In our modeling of soil thermal effects of ROS we are not concerned of the probability of a ROS in a given area or its likely magnitude, but the characteristic soil thermal response that can be expected if and when a large ROS happens. For this reason we model a large ROS event of constant magnitude that occurs at all Northern Hemisphere snow covered model pixels at the same time. This allows for a clear comparison between model pixels.

2 Model and analysis

The Community Land Model (CLM3.0) (Oleson et al., 2004) was used to examine the effect of significant ROS events on soil temperatures of Northern Hemisphere seasonally snow covered land area. CLM3 is developed as part of the NCAR Community Climate System Model (CCSM3), and can be used in offline mode using atmospheric forcing data, such as that from the ECMWF ERA40 reanalysis project. CLM3 divides the world into grid cells. Each cell in turn is composed of land types (glacier, wetlands, lakes, and vegetated). Vegetation in each cell is modeled by four plant functional types (PFTs), such as Arctic grasses, and boreal trees and brush, as well as bare ground. The PFT portion of the model handles hydrologic and thermal processes including

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through-fall of liquid and solid precipitation, evaporation and transpiration. Heat transfers from the atmosphere to the vegetation and ground take into account incoming solar radiation, incoming and outgoing long wave radiation, as well as atmospheric stability, and sensible and latent heat fluxes.

The soil is represented in the model by ten layers which increase in thickness with depth. The lower thermal boundary of the soil model, at 3.43 meters, has zero heat flux. Snow, if present, is modeled by up to five layers, with thinnest layers at the soil/snow boundary. Water flow in both the snow and soil is based on Darcy's law of porous media, with effective porosity derived from soil properties, as well as ice content and degree of liquid water saturation. Runoff and subsurface drainage are also calculated. At each time step, phase change is computed based on available ice and water as well as temperatures and heat fluxes.

In our analysis those components of the CLM3 that provide feedback to the climate model, but do not affect temperatures and water content in the snow and soil have been turned off. These include the River Transport Model (RTM), dynamic global vegetation model (CLM-DGVM) and the volatile organic compound (VOC) model.

Atmospheric data was downloaded from ECMWF for the decade of 1975–1984, at six hour intervals. These data include air temperature at 6 meters above surface, dew point temperature, wind velocities, incoming solar radiation, total precipitation and total snow fall. Preprocessing of these data require distinguishing between instantaneous values such as temperature, and cumulative ones such as precipitation, as well as packaging the data into the monthly files that CLM3 expects. By default the offline CLM3 routines use the air temperature around freezing to partition total precipitation into snow and rain. We modified this code to use the snow fall variable available in the ERA40 database. This has also allowed us to directly modify the relative snow and rain fall amounts, variously removing and adding ROS events to/from the model runs.

The CLM3 allows water to pond in the top soil layer. However we found that when large amounts of ROS were introduced, all of the liquid that passed through the snow was immediately routed to runoff, which is unrealistic in the Arctic where the soil is

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typically frozen under the seasonal snow cover. In an attempt to replicate the ponding of water and related increase in soil temperature that was observed in Spitsbergen (Putkonen and Roe, 2003; Putkonen, 1997), we modified the model to allow only spring melt to be routed to runoff which produced results that qualitatively correspond to published observations.

In order to compare the thermal insulation provided by the snow pack between the model pixels it is useful to define the snow pack thermal resistance (Bejan, 1993):

$$R = L/k \quad (1)$$

where: R is thermal resistance per unit area ($\mathrm{K\,W^{-1}}$), L is the thickness of the snow pack (m), and k is thermal conductivity ($\mathrm{W\,m^{-1}\,K^{-1}}$). Thermal conductivity is calculated based on snow density after equation 2a by Sturm et al. (2002):

$$k(\rho) = 0.138 - 1.01\,\rho + 3.233\,\rho^2 \quad (2)$$

where ρ is snow density ($\mathrm{g\,cm^{-3}}$).

3 Results

Earlier field work and modeling (Putkonen, 1997;Putkonen and Roe, 2003; Rennert et al., 2008; Grenfell and Putkonen, 2008) showed that ponding of liquid water from a ROS event at the soil surface under a thick insulating layer of snow, could raise the temperature of the soil to freezing for a number of days. The ROS water effectively introduces a source of latent heat within and below the snow, while the snow pack limits the flux of this heat to the atmosphere.

We have experimented with the CLM in order to reproduce the effect of ROS that was observed in Spitsbergen (Putkonen, 1997; Putkonen and Roe, 2003). By artificially introducing a rain on snow event on the order of 50 mm in a one day period, we were able to produce a similar soil warming effect at a number of locations in the modeled test region. However, when we quantified the effect of such a ROS event on soil

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temperatures by taking the January to April average of the temperature for the top soil layer, it was found that some locations exhibited a net cooling, rather than the expected net warming of the soil. This is illustrated in Fig. 1, where the lighter shades of gray indicate a net warming due to the ROS event, and darker ones a net cooling.

Detailed examination of the results showed that in the areas of thick season average snow depth the artificial rain-on-snow event brings the soil surface temperature to freezing point and decreases the snow thermal resistance temporarily as the snow is compacted and its density increases. But due to further snow fall, the mean snow thermal resistance returns to values comparable to those before the ROS event, and the mean January–April soil temperatures remain above the control case.

Figure 2 shows some of the model variables for a location in central Siberia (67.5° N, 135° E). The ROS event is shown by air temperature (TBOT), which was raised to freezing for four days surrounding the rain fall. As shown by the SOILLIQ variable, a portion of this rain fall reaches the soil, where it is allowed to pond. While this water remains liquid, the soil temperature (TSOI) is restricted at freezing, even though the average snow temperature (not shown) is below freezing. At this location, the thermal resistance of the snow pack remains lower than the control case for nearly a month following the ROS event due to a limited snow fall. This allows the liquid water that reaches the soil to freeze rapidly, and allows increased heat flux from the soil to the cold atmosphere because of the denser snow pack. The net effect is that the soil temperatures are colder than the control case for the rest of the season, leading to the average cooling for this area.

4 Discussion

It was unexpected to find out that the same amount of ROS at various parts of the seasonally snow covered Northern Hemisphere resulted in either warming or cooling of the season averaged soil temperature in the CLM3. The main reason for the cooling is the increased thermal conductivity of the snow pack which through the following months overwhelms the initial warming that was created by the phase change of the rain and the related release of latent heat at soil surface. To gain more insights in the two opposite thermal cases (the warming and the cooling) those map pixels that experienced the largest warming and cooling were studied separately.

The total range of artificially induced season average cooling and warming is from –2.0 °C to 1.0 °C. The extreme thermal cases were defined as those that warmed or cooled for more than half of the maximum. The extreme warming pixels are those with season average temperature difference >0.5 °C. The extreme cooling pixels are those with season average temperature difference > –1.0 °C. The season average temperature difference is defined as (T_{ROS}–T_{ref}) where T_{ROS} is season average soil temperature when artificial ROS is included in the model, and T_{ref} is the unaltered, natural season average reference temperature.

Figure 3 shows all the data separated in three classes: extreme warming, extreme cooling, and minor thermal response. The minimum season average snow depth where extreme cooling does not occur and extreme warming does is 0.65 m, and the depth where extreme cooling occurs but extreme warming does not is 0.35 m–0.15 m. At season average snow depths <0.15 m the snow soil system is sensitive to small changes in the timing of subsequent snow precipitation events and climatic fluctuations and both extreme warming and cooling are seen. Seventeen and half percent of all the extreme cases are found within the transitional zone: 0.35 m < snow depth < 0.65 m.

We suggest that the season average snow depth 0.5 m (center of transition zone) can be used as a cutoff between extreme warming and cooling, excluding all cases where season average snow depth is less than 0.15 m.

It is important to note that the modeled artificial ROS event of 50 mm is relatively large and should not be expected to occur across the Arctic in any given year. However, large events have been encountered far from open ocean in the Arctic. Therefore, we hypothesize that events of the same order of magnitude may occur anywhere in the Arctic.

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5 Conclusions

Rain on snow (ROS) is known to occur in the northern seasonally snow covered land area and has been shown to significantly warm the season average soil temperatures in Spitsbergen. The seasonally frozen and snow covered ground in the North and especially the permafrost are sensitive archives of past temperatures, thermally sensitive reservoirs of greenhouse gases, and substrate for vulnerable infrastructure. Although the current spatial pattern and frequency of ROS is not well known the climate models predict significantly larger area in the future becoming affected by large ROS events as the climate changes. Therefore an ample motivation exists to characterize the effects that ROS has on Northern Hemisphere soil temperatures.

We set out to induce an artificial, constant ROS event throughout the Northern Hemisphere seasonally snow covered land area in the CLM 3 to characterize the related response of soil temperatures. It was found that counter intuitively in most pixels the season average soil temperatures cooled due to the ROS. The analyses of the spatial patterns and individual model pixels revealed that the extreme season average warming is related to thick snow cover approximately >0.5 m where the water under snow pack is effectively insulated from the atmospheric cooling and may remain liquid for weeks or months. These areas are typically found near coasts. On the other hand the extreme cooling was typically related to thinner snow pack (<0.5 m). In these areas the water froze rapidly because of the enhanced heat flux through the wetted snow pack. The long term net effects were the enhanced heat flux out of the soil and resulting cooler season average soil temperatures.

The relation of snow depth to extreme cooling/warming broke down in model pixels with the thinnest season average snow packs (<0.15 m) that are typically found at the fringes of the seasonally snow covered land area. Pixels with extreme warming and cooling were found in this group due to the ephemeral nature of the snow cover and resulting sensitivity of soil temperatures to timing and duration of the snow cover.

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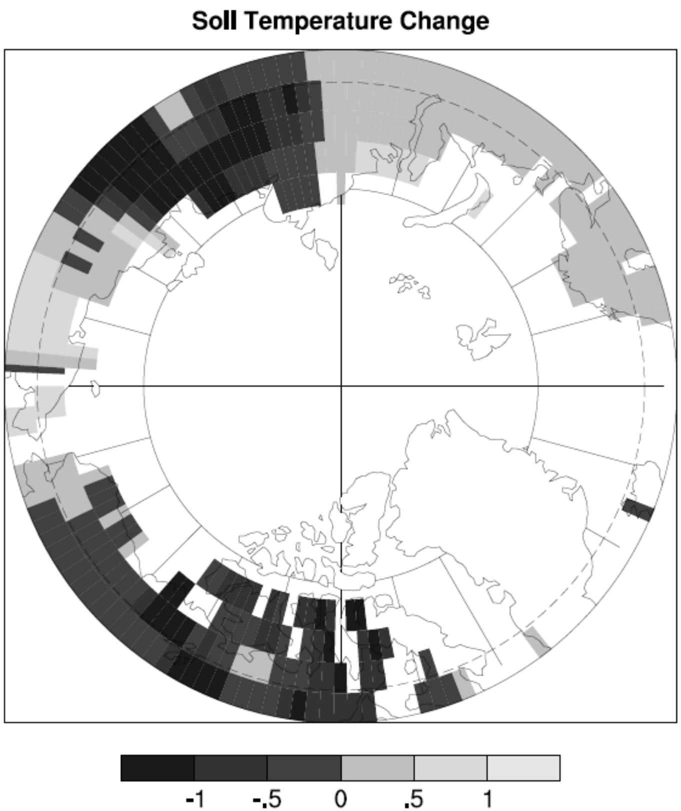

Fig. 1. Soil temperature change due to a modeled large January 1982 ROS event; positive is a net warming for the season. The same amount of ROS was introduced to every cell in order to determine the complex thermal response in the Arctic. The scale is given in °C difference in mean surface soil temperature January–April between ROS and control case. Surprisingly some parts of the Arctic respond to ROS by cooling rather than warming.

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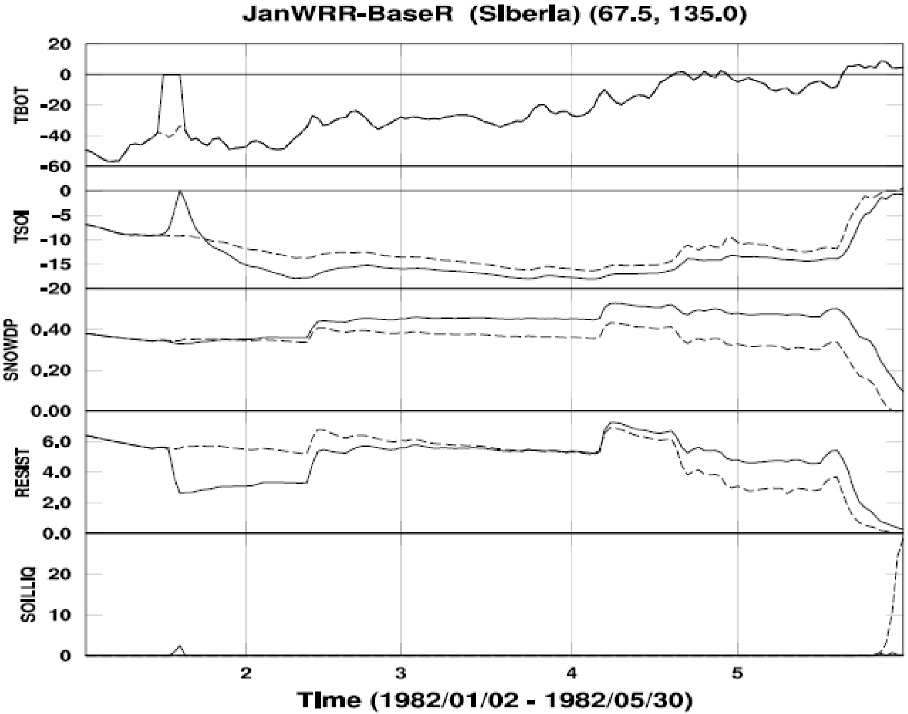

Fig. 2. A central Siberia location (67.5° N, 135° E). Solid lines are the values with the large modeled ROS event; dashed lines are the base case. TBOT is the reference air temperature °C (note the four day introduced warm period in January 1982). TSOI is temperature in the top soil layer °C. SNOWDP is snow depth in meters. RESIST is thermal resistance of the snow. SOILLIQ is liquid in the top soil layer (note the influx with the ROS event). In this case there was no significant snow fall for a month after the modeled ROS event. The reduced thermal resistance of the snow allowed greater cooling of the soil, resulting in a net cooling for the season, compared to the base case.

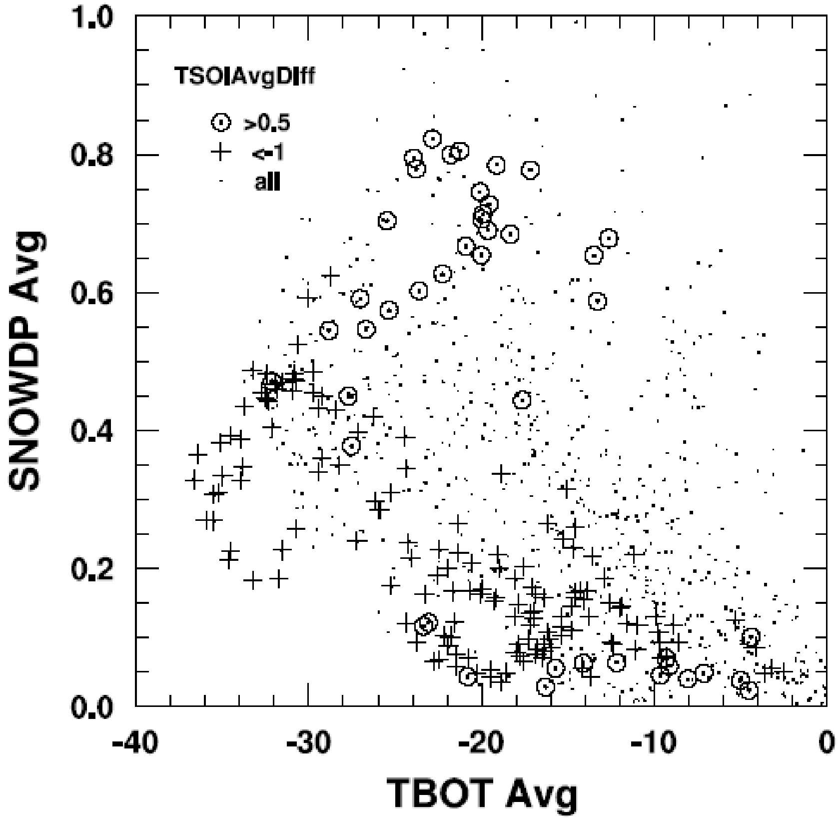

Fig. 3. All model pixels separated into three categories: (1) circle, mean season average soil temperature with ROS >0.5 °C warmer than reference case, (2) cross, mean season average soil temperature with ROS < −1 °C cooler than reference case, (3) dot, all remaining cases that show only intermediate cooling or warming. TBOT is the reference air temperature °C. SNOWDP is snow depth in meters.