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Effects of pan-Arctic snow cover and air temperature changes on soil heat content

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Abstract

Soil heat content (SHC) provides an estimate of the integrated effect of changes in the land surface energy balance. It considers the specific heat capacity, soil temperature, and phase changes of soil moisture as a function of depth. In contrast, soil temperature provides a much more limited view of land surface energy flux changes. This is particularly important at high latitudes, which have and are undergoing surface energy flux changes as a result of changes in seasonal variations of snow cover extent (SCE) and hence surface albedo changes, among other factors. Using the Variable Infiltration Capacity (VIC) land surface model forced with gridded climate observations, we simulate spatial and temporal variations of SCE and SHC over the pan-Arctic land region for the last half-century. On the basis of the SCE trends derived from NOAA satellite observations in 5° latitude bands from April through June for the period 1972-2006, we define a snow covered sensitivity zone (SCSZ), a snow covered non-sensitivity zone (SCNZ), and a non-snow covered zone (NSCZ) for North America and Eurasia. We then explore long-term trends in SHC, SCE, and surface air temperature (SAT) and their corresponding correlations in NSCZ, SCSZ and SCNZ for both North America and Eurasia. We find that snow cover downtrends have a significant impact on SHC changes in SCSZ for North America and Eurasia from April through June. SHC changes in the SCSZ over North America are dominated by downtrends in SCE rather than increasing SAT. Over Eurasia, increasing SAT more strongly affects SHC than in North America. Overall, increasing SAT during late spring and early summer is the dominant factor that has resulted in SHC changes over the pan-Arctic domain, whereas reduced SCE plays a secondary role that is only important in the SCSZ.

Keywords: pan-Arctic, snow cover extent, surface air temperature, soil heat content, Variable Infiltration Capacity model, land surface energy budget

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1. Introduction

Over the pan-Arctic land region, the rise in surface air temperature (SAT) in recent decades has been almost twice as large as the global average (e.g., Serreze et al., 2000; Jones and Moberg, 2003; Overland et al., 2004; Hinzman et al., 2005; White et al., 2007; Solomon et al., 2007; Trenberth et al., 2007; Serreze et al., 2009; Screen and Simmonds, 2010; Cohen et al., 2013; Walsh, 2014). Increases in SAT have been accompanied by increasing soil temperatures with deeper active layer thickness across permafrost regions and decreasing frozen soil depths in the seasonally frozen ground regions (e.g., Hinzman and Kane, 1992; Zhang et al., 2001; Frauenfeld et al., 2004; Osterkamp, 2007; Qian et al., 2011; Smith et al., 2010, 2012; Romanovsky et al., 2007, 2010, 2014; Streletskiy et al., 2015; Peng et al., 2016). Given the potential for releases of soil carbon to the atmosphere at warmer ground temperatures, the land surface warming at high latitudes has attracted considerable scientific attention (e.g., Stieglitz et al., 2003; Zhang, 2005; Osterkamp, 2007; Heimann and Reichstein, 2008; Lawrence and Slater, 2010; Frauenfeld and Zhang, 2011; Park et al., 2014; Kim et al., 2015; Yi et al., 2015).

Because snow is a strong insulator, it can limit the efficient transport of heat between the atmosphere and the ground. Thus, snow plays an important role in determining how air temperature signals propagate into and out of the soil column (Gold, 1963; Goodrich, 1982; Osterkamp and Romanovsky, 1996; Stieglitz et al., 2003; Mann and Schmidt, 2003; Zhang, 2005; Bartlett et al., 2005; Iwata et al., 2008; Lawrence and Slater, 2010; Park et al., 2015). In general, seasonal snow cover results in relatively higher mean annual ground temperatures, especially at high latitudes where stable snow cover lasts from a few weeks to several months

(Zhang, 2005; Frauenfeld and Zhang, 2011). In recent decades, a substantial retreat of snow cover extent (SCE) has been observed during late spring and early summer (Groisman et al., 1994; Déry and Brown, 2007; Brown et al., 2010; Derksen and Brown, 2012; Shi et al., 2011, 2013) in the visible satellite imagery produced by the National Oceanic and Atmospheric Administration (NOAA) (Robinson et al., 1993; Frei and Robinson, 1999). Moreover, these negative SCE trends are well reproduced by simulations using the Variable Infiltration Capacity (VIC) land surface model (Liang et al., 1994; Cherkauer and Lettenmaier, 1999) for both North America and Eurasia (Shi et al., 2011, 2013).

Over the past decades, observed soil temperatures across the pan-Arctic domain have been used as an indicator of climate change (e.g., Osterkamp and Romanovsky, 1999; Zhang et al., 2001; Frauenfeld et al., 2004; Smith et al., 2004; Beltrami et al., 2006; Romanovsky et al., 2002, 2007; Frauenfeld and Zhang, 2011). Recent studies have also attempted to explain the impact of changing seasonal snow cover and air temperature on the ground thermal regime over the pan-Arctic by using soil temperature as an index (e.g., Zhang et al., 1997; Zhang and Stamnes, 1998; Romanovsky et al., 2002; Bartlett et al., 2004, 2005; Lawrence and Slater, 2010; Park et al., 2015). However, *in situ* soil temperatures are problematic because some important physical processes are basically neglected, such as the change in mass due to soil moisture changes and the latent heat effects of freezing and thawing. These effects can be important at high latitudes with seasonally and permanently frozen soils and result in the complicated interpretation for the effects of seasonal snow cover and air temperature on the ground thermal regime.

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To provide a more complete understanding of the effects of high latitude land surface warming, we used an alternative approach based on soil heat content (SHC) as an indicator of changes in the ground thermal regime, which can provide an integrated measure that accounts for changes in temperature, moisture, and latent heat effects. SHC has been used in various studies to document how the land surface responds to atmospheric changes (e.g., Levitus et al., 2001, 2005; Beltrami et al., 2002, 2006; Hu and Feng, 2004; Hansen et al., 2005; Mottaghy and Rath, 2006; Troy, 2010).

We explore here the effects of downtrends in SCE and increases in SAT on SHC over the pan-Arctic land region, with particular emphasis on trends and variability during the late spring and early summer seasons. In Section 2, we describe the observations and model-derived data sets on which our analyses are based. In Section 3, we define three study zones and the computation of SHC. In Section 4, we evaluate the model results, explore trends in SHC, and examine correlations between SCE, SAT, and SHC and the relative roles of snow cover downtrends and increasing SAT on SHC changes. We summarize our findings in Section 5.

2. Data sets

2.1. Observed SCE and SAT data

As described in Shi et al. (2013), we used observed monthly values of SCE, which were extracted from the weekly snow cover extent for the Northern Hemisphere maintained at the National Snow and Ice Data Center (NSIDC). These data span the period October 1966 through June 2014 (Brodzik and Armstrong, 2013), with a spatial resolution of 25 km. We restricted our period of analysis to begin in 1972 because some charts between 1967 and 1971 are missing (Robinson, 2000). The data set has become a widely used tool for deriving trends in climate-related studies (Groisman et al., 1994; Déry and Brown, 2007; Flanner et al., 2009; Derksen et al., 2010; Derksen and Brown, 2011, 2012; Shi et al., 2011, 2013), notwithstanding uncertainties in some parts of the domain for certain times of the year, such as summertime over northern Canada (Wang et al., 2005). Monthly SAT anomaly data were taken from the Climatic Research Unit (CRU, Brohan et al., 2006), and are based on anomalies from the long-term mean temperature for the period 1961-1990 for each month since 1850. The land-based monthly data are on a regular 0.5° by 0.5° global grid. We regridded these data, including the NOAA SCE observations that were aggregated from the 25 km product, to the 100 km EASE grid using an inverse distance interpolation as implemented in Shi et al. (2013).

2.2. Modeled SHC from VIC

The version of VIC used for this study is 4.1.2, which includes some updates to the model's algorithms for cold land processes. For instance, the model includes a snow parameterization that represents snow accumulation and ablation processes using a two-layer energy and mass balance approach (Andreadis et al., 2009), a canopy snow interception algorithm when an overstory is present (Storck et al., 2002), a finite-difference frozen soils algorithm (Cherkauer and Lettenmaier, 1999) with sub-grid frost variability (Cherkauer and Lettenmaier, 2003), and an algorithm for the sublimation and redistribution of blowing snow (Bowling et al., 2004), as well as a lakes and wetlands model (Bowling and Lettenmaier, 2010).

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The snow parameterization in VIC represents snow accumulation and ablation processes using a two-layer energy and mass balance approach (Andreadis et al. 2009) and a canopy snow interception algorithm (Storck et al. 2002) when an overstory is present. In the VIC model setup for this study, each grid cell was partitioned into five elevation (snow) bands, which can include multiple land cover types. The snow model was applied to each grid cell and elevation band separately. When snow water equivalent is greater than a threshold, the model assumes that snow fully covers that elevation band. For each grid cell, the simulated SCE is calculated as the average over the elevation bands. The current version of the frozen soils algorithm uses a finite difference solution in the algorithm that dates to the work of Cherkauer and Lettenmaier (1999). To improve spring peak flow predictions, a parameterization of the spatial distribution of soil frost was developed (Cherkauer and Lettenmaier, 2003). Adam (2007) described some significant modifications to the frozen soils algorithm, including the bottom boundary specification using the observed soil temperature datasets of Zhang et al. (2001), the exponential thermal node distribution, the implicit solver using the Newton-Raphson method, and an excess ground ice and ground subsidence algorithm in VIC 4.1.2.

To model permafrost properly, our implementation used a depth of 15 m with 18 soil thermal nodes (STN) exponentially distributed with depth and a no flux bottom boundary condition (Jennifer Adam, personal communication). When the no flux bottom boundary condition is selected for the soil column, the VIC model solves the ground heat fluxes using the finite difference method. This means that the soil temperature at the bottom boundary can change, but there is no loss or gain of heat energy through the boundary. Compared to the constant heat flux (e.g., Neumann) boundary condition, the no flux bottom boundary condition method adds slightly to the computation time, but is especially useful for very long simulations in climate change studies and permafrost simulations.

We used the same study domain as documented in Shi et al. (2013), which is defined as all land areas draining into the Arctic Ocean, as well as those regions draining into the Hudson Bay, Hudson Strait, and the Bering Strait, but excluding Greenland (because its snow cover is mainly perennial in nature). The model simulations used calibrated parameters, such as soil depths and infiltration characteristics, from Su et al. (2005). The VIC runs are at a three-hour time step in full energy balance mode (meaning that the model closes a full surface energy budget by iterating for the effective surface temperature, as contrasted with water balance mode, in which the surface temperature is assumed to equal the surface air temperature). The model is driven by daily precipitation, maximum and minimum temperatures, and wind speed at a spatial resolution (EASE grid) of 100 km. Also, VIC has an internal algorithm to estimate the incoming shortwave and longwave radiation fluxes that are based on location and meteorological conditions and implicitly used to force the model. The forcing data were constructed from 1948 through 2006 using methods outlined by Adam and Lettenmaier (2008), as described in Shi et al. (2013). To set up the right initial conditions in VIC, especially for the thermal state, we initialized the model with a 100-year climatology created by randomly sampling years from the 1948-1969 meteorological forcings. Also, we calculated the correlation coefficients between SCE, SAT, and SHC using the 35-year time series spatially averaged for the three zones over North America and Eurasia along the 15 m soil profile. Through the above processes, the effects due to the propagation of SAT signals to the deeper soil depth become weak. After validating the VIC simulation, we reconstructed SHC from

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1970 to 2006 for the pan-Arctic land area.

3. Methodology

3.1. Definition of study zones

The NOAA weekly SCE data (hereafter SCE) were analyzed to determine whether or not there are regions with significant changes of SCE in North America and Eurasia. Figure 1(a) shows the spatial distribution of long-term monthly means of SCE from April through June for the 35-year period over North American and Eurasian pan-Arctic domains. Figures 1 (b) and (c) illustrate the latitudinal variations of SCE trends and their area fractions over the North American and Eurasian study domains from April through June. The percentage under each bar chart is the trend significance expressed as a confidence level for each 5° of latitude, while the solid line shows the latitudinal patterns in the snow cover area fractions for each month, which in general are at a minimum for the lowest latitude band, and then increase with latitude poleward.

Based on the latitudinal changes of NOAA SCE as shown in Figure 1, we identified different study zones for North America and Eurasia. From April through June, snow mostly covers latitude bands north of 45°N over the pan-Arctic land area, which are denoted as snow covered zones (SCZs) in Figure 1. The rest of the study domains were denoted as non-snow covered zones (NSCZs) (see Figure 1(a) for North America and Eurasia, respectively). Within the SCZs, we selected only those latitudinal bands within which SCE trends were statistically significant for further analyses. For each month, we denoted these bands as snow cover sensitivity zones (SCSZs). For the remaining bands in the SCZs, there is no significant snow

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cover downtrends, and these bands are defined as snow covered non-sensitivity zones (SCNZs). In Figures 1(b) and 1(c), we use different gray-shaded arrows to highlight the North American and Eurasian SCZs, which include the SCSZs and SCNZs. For example, the SCSZ for May in North America has six latitude bands from 45-50°N to 70-75°N, whereas there is only one band (45-50°N) for the Eurasian SCSZ in April. Given the large-scale snow cover in Eurasia and North America, the effects caused by the differences in snow depth can be neglected.

3.2. Definition of SHC for the soil thermal nodes

SHC, also called soil enthalpy in the literature, is a measure of the heat stored in the soil column. It has been applied in many previous studies (e.g., Hu and Feng, 2004; Hansen et al., 2005; Levitus et al., 2005; Beltrami et al., 2006; Mottaghy and Rath, 2006; Troy, 2010). At its simplest, it is the written as a vertical integral as in Hu and Feng (2004):

$$H = \int_0^z C_s T(z) dz \qquad (1)$$

where H is the soil heat content, C_s is the specific heat capacity, and $T(z)$ is the soil temperature as a function of depth z. This formulation neglects two important physical processes that are important at high latitudes. First, the change in mass due to soil moisture (both liquid and frozen) changes, which changes the heat energy stored in the soil column, is not included. Second, the latent heat effects of freezing and thawing are neglected. These effects can be important at high latitudes with seasonally and permanently frozen soils. Consequently, we calculated the SHC changes as follows for each STN (n):

$$H_n = \int (C_s f_s \rho_s + C_l f_l \rho_l + C_i f_i \rho_i)\, dT - \int L_f\, df_i \qquad (2)$$

where H_n is the SHC value at node n, C is the specific heat capacity of soil, liquid water and frozen water (subscripts s, l, and i, respectively), f is the fraction of soil and liquid and frozen water, ρ is density, dT is the change in temperature, and L_f is the latent heat of fusion. We neglect the heat content of air in the soil pores. We computed the change in SHC was calculated for each model time step, accounting for changes in the fraction of liquid and frozen water with each time step. The fraction of soil remains constant in time. This is essentially the same formulation as Mottaghy and Rath (2006) with the addition of the heat capacity of the soil matrix included. To calculate the total change in SHC for the soil column, the vertical integral was calculated accounting for the spacing of the nodes. Therefore, the total change of SHC is an integrated value over a soil thermal profile from the surface to a specified depth z, rather than an average value for the specific layer. The calculation process for the SHC change is consistent with our motivation here, which is to investigate the impacts of snow cover and air temperature changes on SHC as a function of soil depth.

As described in Section 2.2, we used a depth of 15 m with 18 STNs in the VIC implementation. To maximize computational efficiency, the spacing of soil thermal nodes in the frozen soils framework in VIC should reflect the variability in soil temperature (Adam, 2007). Because the greatest variability in soil temperature occurs near the surface, it is preferable to have tighter node spacing near the surface and wider node spacing near the bottom boundary where temperature variability is reduced. Therefore, these 18 STNs were distributed exponentially with depth as indicated in Table 1. The SHC for each STN in the soil column was calculated for each model time step (three hours) and then aggregated for each month from April through June. Along the soil profile from the top to the bottom, the first STN was named as STN0 with a depth of 0 m indicating it is at the surface, while the deepest one is STN17 with a soil depth of 15 m. The SHC for STN17 represents an integrated thermal value for the soil profile. To simplify the analyses, we calculated SHC changes relative to 1970, using the start of our historical VIC runs as our datum. All monthly SHC values are relative to this datum and as such represent the change in SHC since January 1, 1970. In addition, we calculated monthly SHC anomalies on the basis of monthly means averaged over each NSCZ, SCSZ and SCNZ of North America and Eurasia by removing the 1981-1990 mean. Figure 2 shows the area percentages of NSCZ, SCSZ, and SCNZ in North America and Eurasia from April through June for the period 1972-2006. The experimental design for assessing the effects of pan-Arctic snow cover and air temperature changes on SHC in NSCZ, SCSZ, and SCNZ over North America and Eurasia from April through June for the period 1972-2006 is shown in Figure 3. For example, we can isolate the impact of increasing SAT on SHC changes in NSCZ, which has no presence of snow. Within SCSZ, the effects of both SCE and SAT changes on SHC can be compared. Moreover, we can investigate the effect of snow cover changes by comparing SCSZ and SCNZ, as there are snow cover downtrends in SCSZ whereas none is in SCNZ.

3.3. Mann-Kendall trend test

To analyze long-term changes in the monthly time series of SCE, SAT and SHC over the North America and Eurasia study zones, we used the non-parametric Mann-Kendall (MK) trend test (Mann, 1945; Kendall, 1975), a rank-based method applicable for trend significance. In addition, we used the Sen slope estimator (Sen, 1968) to estimate trend slopes. The MK trend test has been applied in many previous studies for identifying trends in meteorological

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and hydrologic variables (e.g., Lettenmaier et al., 1994; Zhang et al., 2001; Burn et al., 2004; Déry and Brown, 2007; Shi et al., 2011, 2013), and has been found to perform well. We used a 5% significance level (two-sided test).

3.4. Pearson's product-moment correlation coefficient

We used the Pearson's product-moment correlation coefficient was used to assess relationships and computed it separately for each study zone from April through June. Given the 35-year record, correlations are statistically significant at a level of $p < 0.025$ (two-sided test) when the absolute value of the sample correlation is greater than 0.34 based on the Student t-test with 33 degrees of freedom. Through correlation analyses, it is possible to identify the relative roles of downtrends in SCE and increasing SAT on pan-Arctic SHC changes.

4. Results

4.1. Model validation

There are no unified temperature data sets that fully cover our study domain. Here we used a monthly historical soil temperature dataset, which is available across the former Soviet Union, with the earliest observation beginning in 1882 and the last in 1990 (Zhang et al., 2001). Soil temperature was measured at various depths between 0.2 and 3.2 meters, with the largest number of measurements at 0.2, 0.8, 1.6, and 3.2 meters. The temporal coverage varies by station, with some stations only having a few years of data and others with decades of continuous measurements (Troy, 2010). Because there are gaps in the observational data, the VIC results were screened to only include a grid cell when the corresponding station had data.

In this study, we compared soil temperature predicted by VIC against historical soil temperature observations at 146 stations across the former Soviet Union as shown in Figure 4(a). When calculating the annual mean, a year was only included in the analysis if data existed for all twelve months. To compare across the region, we calculated the annual soil temperature anomaly at 0.2, 0.8, 1.6, and 3.2 meters for these 146 stations across the former Soviet Union, which is confined to the common period of 1970-1990 between VIC and observations.

Figure 4(b) compares modeled and observed soil temperature anomalies to evaluate the ability of the model to replicate observed trends. The results reveal that the model slightly underestimates the trend in temperature but captures the interannual variability of the soil temperature dynamics, which are similar to Troy (2010). In addition, Figure 4(b) also shows the correlation coefficients between modeled and observed time series for the period from 1970 to 1990. The significance level is based on a two-tailed Student's t test with 19 degrees of freedom. The VIC and observed soil temperature time series at 0.2, 0.8, 1.6, and 3.2 meters are highly correlated (two-sided $p < 0.01$). Therefore, we conclude that VIC is able to reproduce soil temperature profiles and provides a surrogate for scarce observations for estimation of long-term changes in SHC at high latitudes.

4.2. SCE and SAT trends

The MK trend tests were performed on the monthly time series of SCE and SAT area-averaged over the North America and Eurasia study zones. Tables 2 and 3 summarize the SCE and SAT trends and their significance levels in NSCZ, SCSZ, and SCNZ for both continents from April through June for the entire study period (1972-2006). Table 2 shows that statistically significant ($p < 0.025$) negative trends were detected in SCE for both North American and Eurasian SCSZs, as found in many previous studies. In SCNZ, the decreasing trends in SCE are all non-significant, and the absolute values of trend slopes are much smaller than that in SCSZ. As reported in Table 3, increasing SAT trends were detected for both continents except for North America in May. For June in North America and for May and June in Eurasia, these SAT trends are statistically significant in NSCZ, SCSZ, and SCNZ. In SCNZ, increasing SAT trends are all statistically significant for both continents except for Eurasia in April.

Based on the above long-term trends in SCE and SAT for NSCZ, SCSZ, and SCNZ, it is clear that the impact of increasing SAT on SHC changes can be isolated in NSCZ as there is no presence of snow. In SCSZ, the effects of both SCE and SAT changes on SHC can be compared as indicated in Figure 3. By comparing SCSZ and SCNZ, we can investigate the effect of snow cover changes on SHC changes, as there are snow cover downtrends in SCSZ whereas SCE is not a factor in SCNZ. Figure 2 shows the area percentages for NSCZ, SCSZ and SCNZ in North America and Eurasia from April through June. In Eurasia, the SCNZ dominates in April as there is no significant snow cover change for most portions of the study domain. When snow cover retreats, the SCSZ and NSCZ in Eurasia expands significantly in May and June. Over North America, the SCE retreat occurs earlier than in Eurasia. Especially for May, most regions in North America have snow cover downtrends. In June, Figure 2 clearly illusrates that SCE is already gone for most portions of Eurasia.

4.3. SHC trends

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Figure 5 shows the trends and significance levels of the VIC-derived SHC for each STN in NSCZ, SCSZ and SCNZ over North America and Eurasia from April through June. Figures 5a and 5b show that there are obvious differences for trends and significance levels between North America and Eurasia. For North America (Figure 5a), the SHC in SCSZ increases significantly from the top thermal nodes to the deeper ones, whereas in NSCZ and SCNZ, most thermal nodes have increasing trends, which are not statistically significant. Over Eurasia, this is quite different. In Figure 5b, almost all the thermal nodes in NSCZ, SCSZ, and SCNZ over Eurasia from April through June show statistically significant increasing trends in SHC, indicating that there are different effects of downtrends in SCE and increasing SAT on SHC changes between Eurasia and North America.

4.4. Effects of SCE and SAT changes on SHC

To identify the relative roles of decreasing SCE and increasing SAT on pan-Arctic SHC changes, we examined the correlations among SHC, SCE, and SAT over NSCZ, SCSZ and SCNZ for both North America and Eurasia. Figure 6 shows correlations between observed SCE and VIC-derived SHC in NSCZ, SCSZ, and SCNZ over North America and Eurasia from April through June. In SCSZ, the correlations between SCE and SHC are all statistically significant over both continents from April through June. Over SCNZ, however, the correlations are much smaller and are not statistically significant. These results imply that the static snow cover insulation in SCNZ does not significantly impact SHC changes over the pan-Arctic. Additionally, no correlation exists between SHC and SCE in NSCZ. Furthermore, the implied impact of snow cover extent changes on SHC is similar for North America and Eurasia.

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Figure 7 shows correlations between observed SAT and simulated SHC monthly time series in NSCZ, SCSZ, and SCNZ over North America and Eurasia from April through June. Overall, the results indicate that SAT has a statistically significant impact on SHC changes in NSCZ. Moreover, SAT has greater influence on SHC over Eurasia than in North America as shown in Figure 7. All the correlations over Eurasia are statistically significant except for SCSZ and SCNZ in April, for which the increasing trends in SAT are not statistically significant.

The correlations described in Figures 6 and 7 were calculated on the time series of variables using the Pearson's product-moment method. Both the effects of secular trend and variability are included. We separated these two components and explored the relative roles of the linear trend and the variability (detrended) in the corresponding correlations. Table 4 summarizes correlation coefficients due to the linear trend and the variability between SHC derived from VIC and SCE observations in SCSZ and SCNZ over North America and Eurasia for the period 1972-2006. The significance level (*p*-value) was calculated using a two-tailed Student t-test with 33 degrees of freedom. Basically, SHC and SCE in NSCZ, SCSZ, and SCNZ are highly correlated due to the secular trend, except for May and June in SCNZ over North America, where the SCE trends are zero. In contrast, the variability components are small and not statistically significant. Clearly, the relationships between SHC and SCE time series are mainly dominated by snow cover changes in each study zone over North America and Eurasia. We also applied the same analyses for the VIC-derived SHC and CRU SAT, as reported in Table 5. The linear trends in SAT dominate the correlations between SHC derived from VIC and CRU SAT in NSCZ, SCSZ, and SCNZ over North America and Eurasia. In contrast, the effect of SAT variability is weak and not statistically significant. Therefore, the relationships between the SHC and SAT time series as shown in Figure 7 are mainly due to increasing SAT in each study zone over North America and Eurasia.

As described above, SHC changes are significantly affected by downtrends in SCE and increasing SAT from April through June over North America and Eurasia for the period 1972-2006. But the variablity in SCE and SAT have insignificant effects on SHC. Comparing the correlations in Figures 6 and 7 suggests that: (1) downtrends in SCE have a significant impact on SHC changes in SCSZ, which is similar for both continents; (2) over North America, SHC changes in SCSZ during late spring and early summer are dominated by snow cover downtrends rather than increasing SAT; (3) over Eurasia, increasing SAT more strongly affects SHC than in North America; and (4) overall, increasing SAT has the dominant influence on SHC for North America and Eurasia, and reduced SCE plays a secondary role that is only important in SCSZ.

5. Discussion and Conclusions

We defined three study zones (NSCZ, SCSZ, and SCNZ) within the North American and Eurasian portions of the pan-Arctic land area based on observed SCE trends. Using these definitions of zones, we focused on the effects of pan-Arctic snow cover and air temperature changes on SHC by exploring long-term trends in SHC, SCE, and SAT and their corresponding correlations in NSCZ, SCSZ, and SCNZ for North America and Eurasia. We find that North American and Eurasian late spring and early summer (from April through June) SHC has increasing trends for the period 1972-2006. However, there are obvious differences between North America and Eurasia as to the magnitudes of SHC trend slopes and significance levels. For North America, SHC in SCSZ has mostly increased significantly, whereas in NSCZ and SCNZ, most thermal nodes show non-significant increasing trends. For Eurasia, almost all the thermal nodes in NSCZ, SCSZ, and SCNZ have statistically significant increasing trends, indicating that there are different effects of snow cover downtrends and increasing SAT on SHC changes between North America and Eurasia. By analyzing the corresponding correlations, we conclude that snow cover downtrends have a significant impact on SHC changes in SCSZ for North America and Eurasia from April through June. SHC changes in SCSZ over North America are dominated by snow cover downtrends rather than increasing SAT. Over Eurasia, increasing SAT more strongly affects SHC than in North America. Overall, increasing SAT during late spring and early summer has the dominant influence on SHC changes over the pan-Arctic, and reduced SCE plays a secondary role that is only important in SCSZ.

In this article, we mainly focused on the impacts of snow cover and air temperature changes on SHC for the temporal scale. The value of SHC is that it estimates the heat stored in the soil column by considering the specific heat capacity and soil temperature as a function of depth. In the calculation, the specific heat capacity of soil, liquid water and frozen water were included, as well as the latent heat of fusion, rather than soil temperature alone. Therefore, it provides an estimate of the integrated changes in heat content for the vertical soil column, whereas soil temperature only gives point measurements at specific depths. Given that soil temperature can sit at the freezing point while soil freeze/thaw is going on, this also gives a better integrative estimate of the heat budget. It would be interesting to investigate the spatial

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distribution of SHC trends across the pan-Arctic domain in our future work. In addition, we formulated our estimate (Equation 2) to be based on SCE instead of snow depth due to the following reasons. First, both snow depth and SCE can affect the amount of energy absorbed by the ground, but they are at different levels. Snow depth affects the insulating properties of the snowpack, whereas changes in SCE (whether or not snow is present) have a first order effect on the amount of energy absorbed by the ground and hence the rate of soil warming (e.g., Euskirchen et al., 2007), owing to the large difference in the amount of downward solar radiation absorbed by snow covered land as contrasted with snow free land, particularly in spring (e.g., Déry and Brown, 2007). Second, trends in snow depth over the pan-Arctic domain are highly heterogeneous (Dyer and Mote, 2006; Park et al., 2012, Yi et al., 2015), whereas SCE trends have been much more coherent on a regional basis (Brown and Mote, 2009). For this reason, the decreasing trends in SCE have a dominant impact on the SHC trends, while trends in snow depth are comparatively minor. Therefore, the big SHC changes are associated with transitions from snow cover to snow free in the spring, when downward solar radiation is increasing rapidly.

In the 1970s, global warming first became evident beyond the bounds of natural variability, but increases in global mean surface temperatures have stalled in the 2000s. This pause is commonly called the "hiatus." We know that Earth's climate system is accumulating excess solar energy owing to the build-up of greenhouse gases in the atmosphere. However, global mean surface temperatures fluctuate much more than these can account for. Therefore, the energy imbalance is manifested not just as surface atmospheric and ground warming, but also as melting sea and land ice and heating of the oceans. Especially, more than 90% of the heat goes into the oceans (Trenberth and Fasullo, 2013).

Notwithstanding that the primary focus of this work is on technical issues associated with trends in high latitude moisture and energy fluxes, the work has broader implications that deserve mention. The importance of snow to high latitude energy fluxes, especially as a result of strong contrasts between the albedo of snow covered and snow free surfaces and hence the potential for positive climate impacts as the high latitudes warm, is well known. Furthermore, the low thermal conductivity of snow insulates the ground, and resultant contrasts in the snow surface temperature as contrasted with bare ground affect the transfer of heat to and from the atmosphere (Barry et al., 2007). These factors can influence the climate not only of high latitude regions, but also, via teleconnections, of lower latitudes. More directly, changes in snow cover patterns have major effects on water availability, industry, agriculture, and infrastructure and affect the livelihoods of the inhabitants of high latitude land regions. Furthermore, certain industries depend heavily on snow cover and frozen soils. Oil and gas companies, for example, use ice roads in the Arctic to gain access to resource fields, and are negatively impacted by permafrost changes. Another impact of snow cover changes is increased heat storage in frozen soils, which ultimately results in permafrost thawing. Thawing permafrost already has affected the stability of infrastructure over parts of the pan-Arctic domain, such as buildings, roads, railways, and pipelines (Osterkamp and Romanovsky, 1999). In addition, the northern high latitudes contain about twice as much carbon as the global atmosphere, largely stored in permafrost and seasonally thawed soil active layers (Hugelius et al., 2014). The rising soil heat content may affect future soil carbon releases with potential feedback on climate change (Schuur et al., 2015). Furthermore, permafrost degradation in many high latitude regions is of concern for hydrological processes (Romanovsky et al., 2010; Watts et al., 2012) and for changes in vegetation composition and establishment (Tape et al., 2006; Sturm et al., 2005; Shi et al., 2015). All of these potential impacts are related to the interaction of air, snow and soil freeze-thaw processes, and point to the importance of work like that reported herein in a broader context.

Acknowledgements

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through June for the period 1972-2006. The significance level (expressed as a confidence level) was calculated using a two-sided Mann-Kendall trend test. Trend slope units are MJm^{-2} $year^{-1}$.

Figure 6. Correlations between NOAA SCE and simulated SHC in NSCZ, SCSZ, and SCNZ over North America and Eurasia from April through June for the period 1972-2006. The correlation with asterisks is statistically significant at a level of $p < 0.025$ when its absolute value is greater than 0.34.

Figure 7. Correlations between observed SAT and simulated SHC in NSCZ, SCSZ and SCNZ over North America and Eurasia from April through June for the period 1972-2006. The correlation with asterisks is statistically significant at a level of $p < 0.025$ when its absolute value is greater than 0.34.

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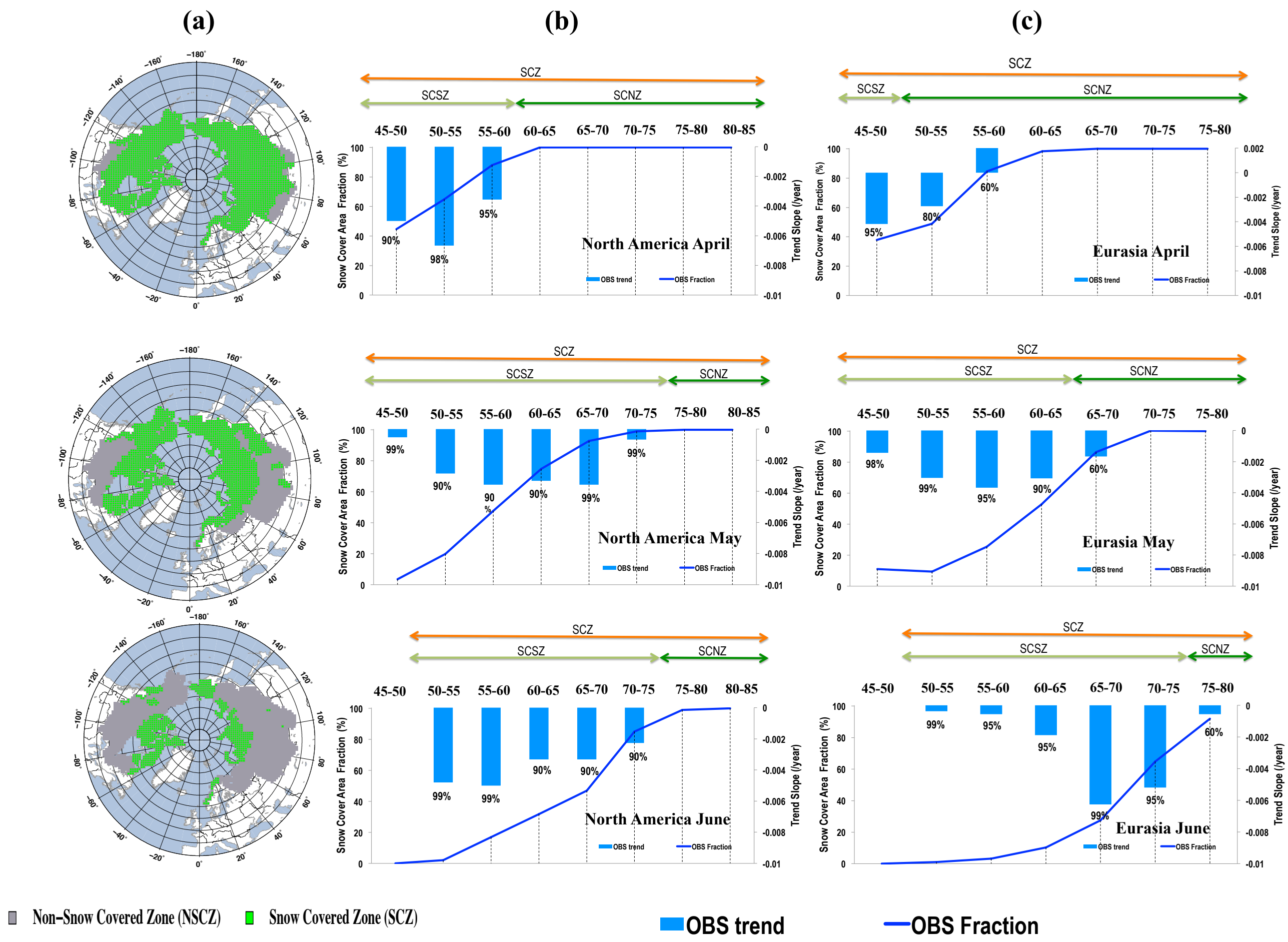

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Figure 1. (a) Spatial distribution of monthly mean snow cover extent (SCE) from NOAA satellite observations (OBS) over North America and Eurasia in the pan-Arctic land region (non-snow covered zone (NSCZ) and snow covered zone (SCZ)) for April (top panel), May (middle panel), and June (bottom panel) for the period 1972-2006. The SCE trends in 5° (N) latitude bands and their area fractions over (b) North American and (c) Eurasian SCZ, including the snow covered sensitivity zone (SCSZ) and snow covered non-sensitivity zones (SCNZ) as indicated by the arrows. The percentage under each bar chart is the trend significance for each 5° (N) of latitude (expressed as a confidence level).

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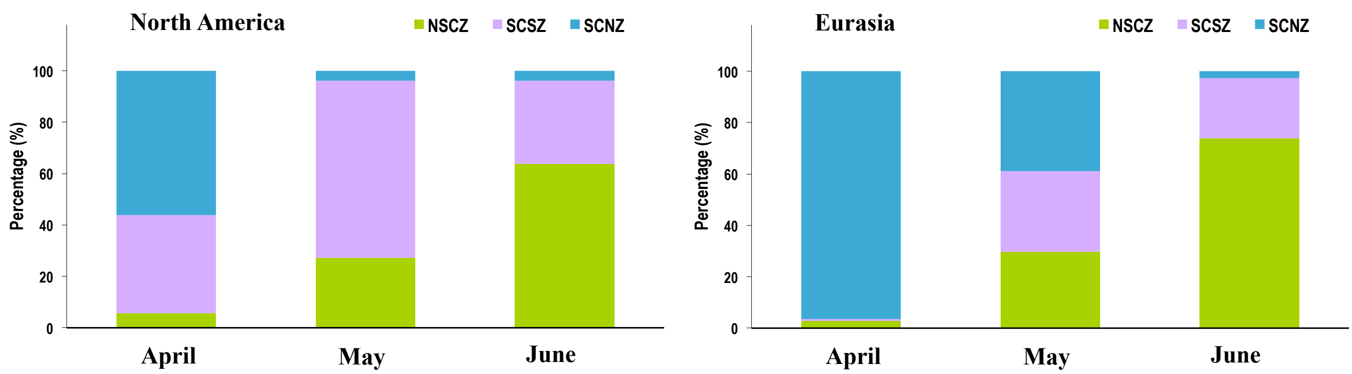


Figure 2. Area percentages of NSCZ, SCSZ, and SCNZ in North America and Eurasia from April through June for the period 1972-2006.

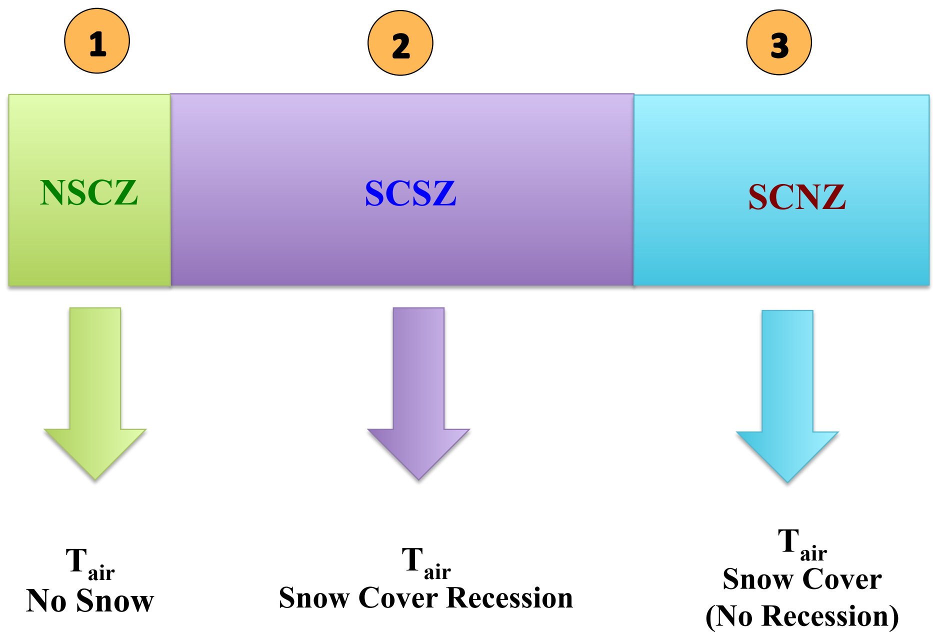
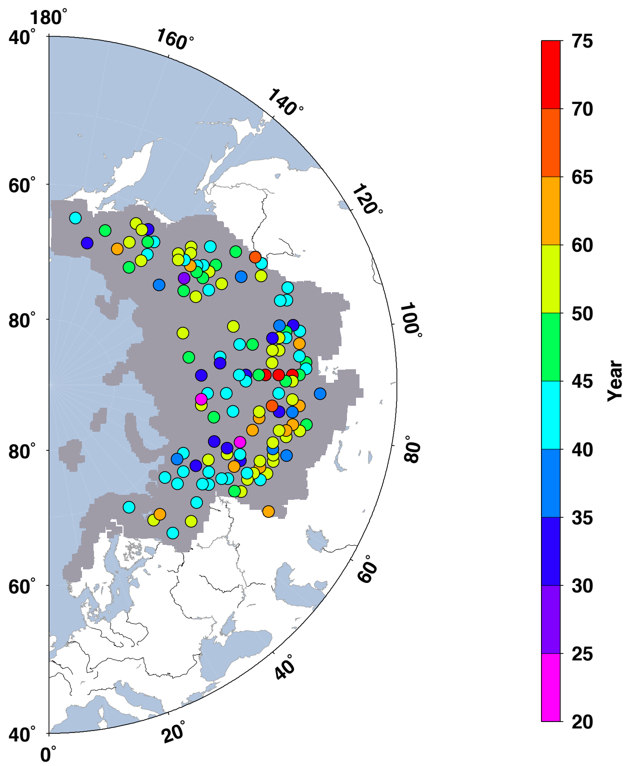


Figure 3. Experimental design for assessing the effects of pan-Arctic snow cover and air temperature changes on frozen soil heat content (SHC) in NSCZ, SCSZ, and SCNZ over North America and Eurasia from April through June for the period 1972-2006.

(a)


180°
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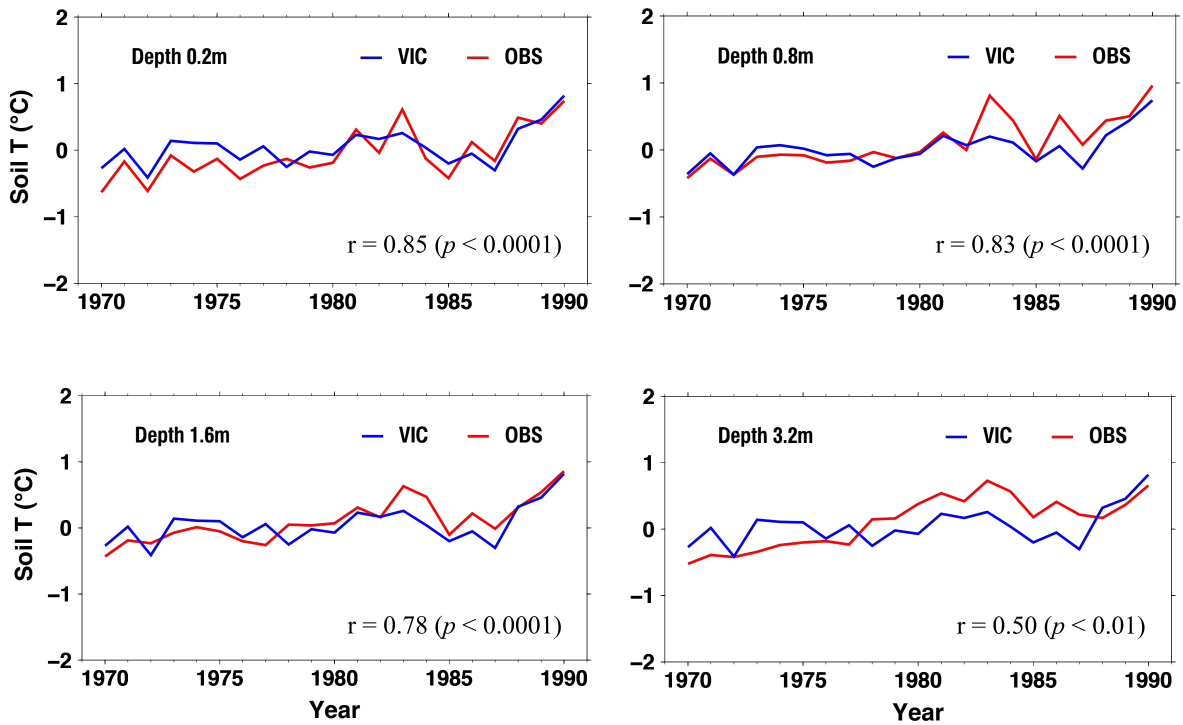
20
Year

(b)



b)

Figure 4. (a) Geographical distribution of 146 observation sites for soil temperature across the former Soviet Union. The color bar at right indicates the number of archived years of data ending in 1990. (b) Comparisons between observed and modeled soil temperature anomalies averaged over 146 observation sites across the former Soviet Union for the period of 1970-1990 at the depths of 0.2 m, 0.8 m, 1.6 m, and 3.2 m, respectively. The correlation is statistically significant at a level of $p < 0.025$.

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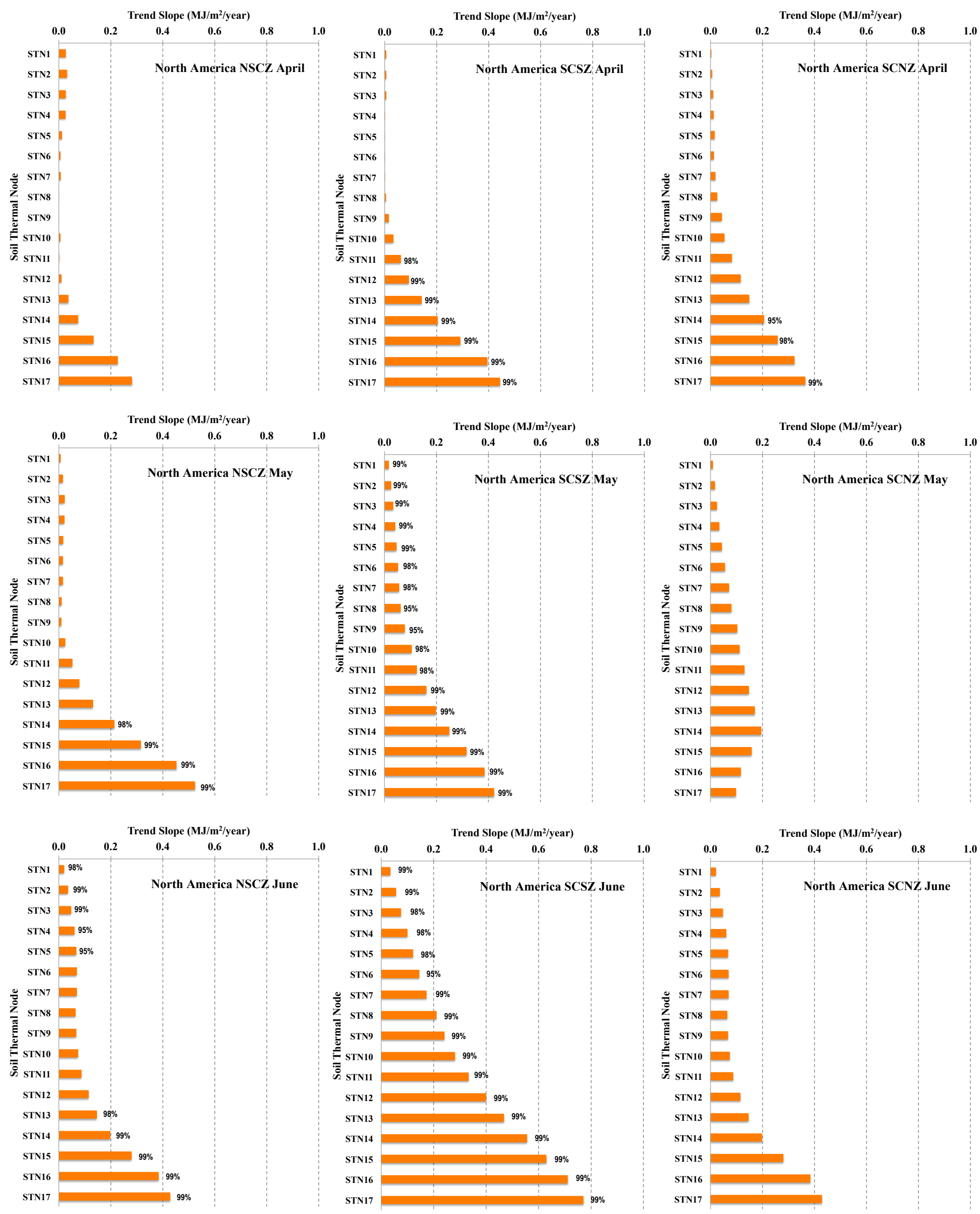
(a)



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(b)

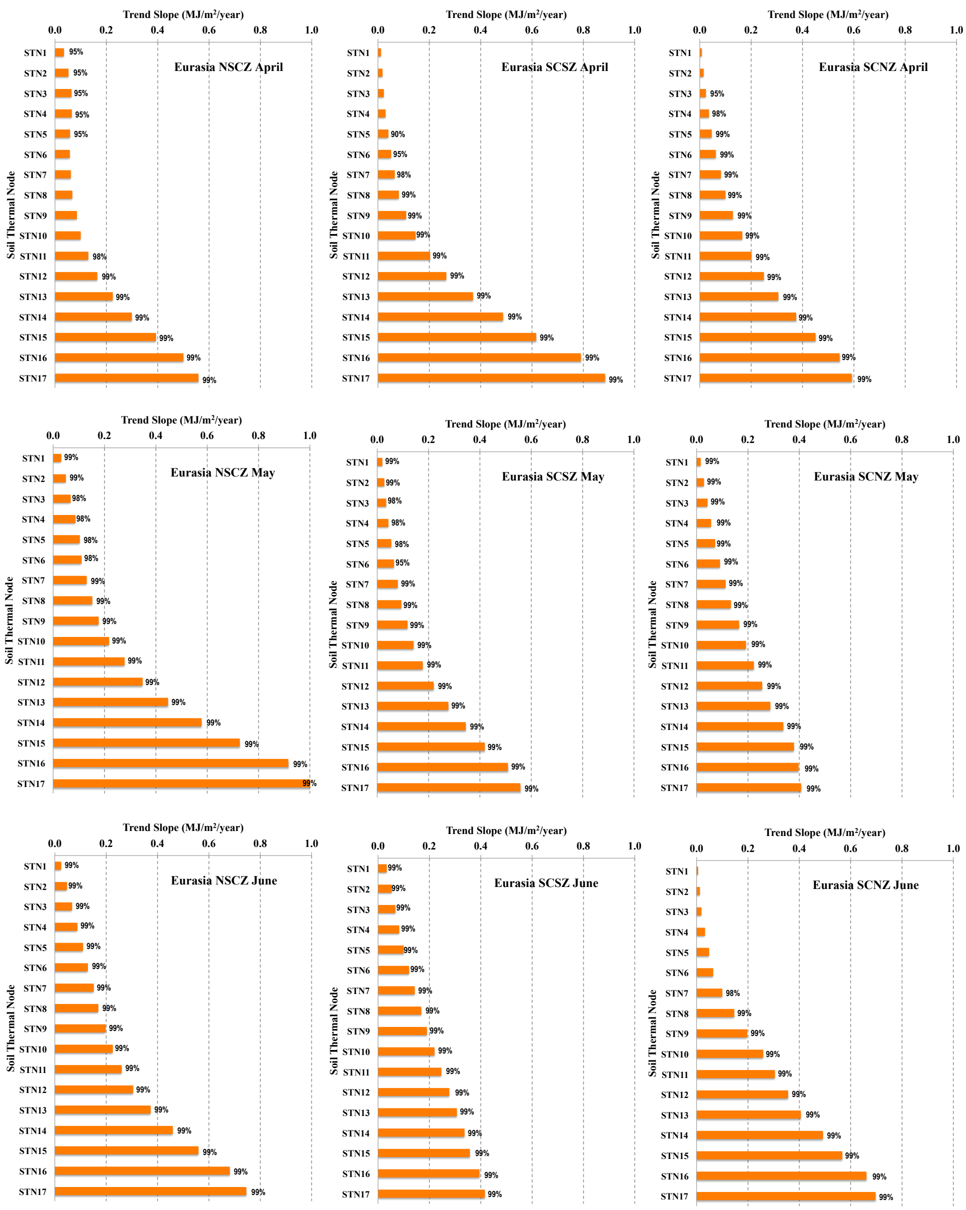

Figure 5. Trend analyses for SHC at the depth of each soil thermal node derived from the VIC model in NSCZ, SCSZ, and SCNZ over (a) North America and (b) Eurasia from April through June for the period 1972-2006. The significance level (expressed as a confidence level) was calculated using a two-sided Mann-Kendall trend test. Trend slope units are MJm^{-2} year^{-1}.

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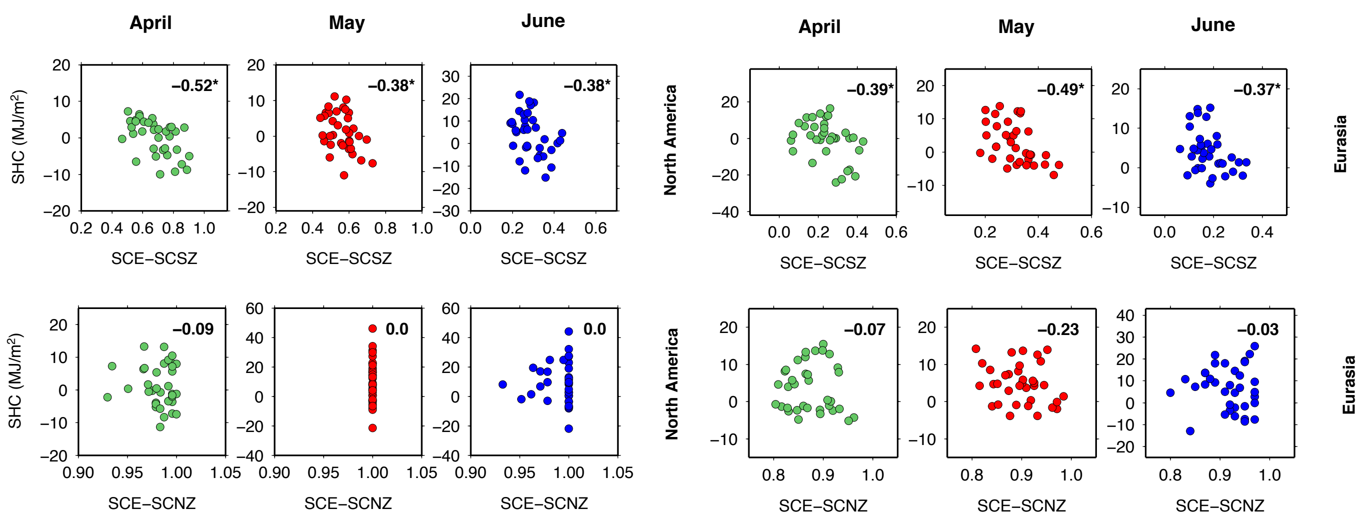

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68 **Figure 6.** Correlations between observed SCE and simulated SHC in SCSZ and SCNZ over North America and Eurasia from April through
69 June for the period 1972-2006. The correlation with asterisks is statistically significant at a level of $p < 0.025$ when its absolute value is greater
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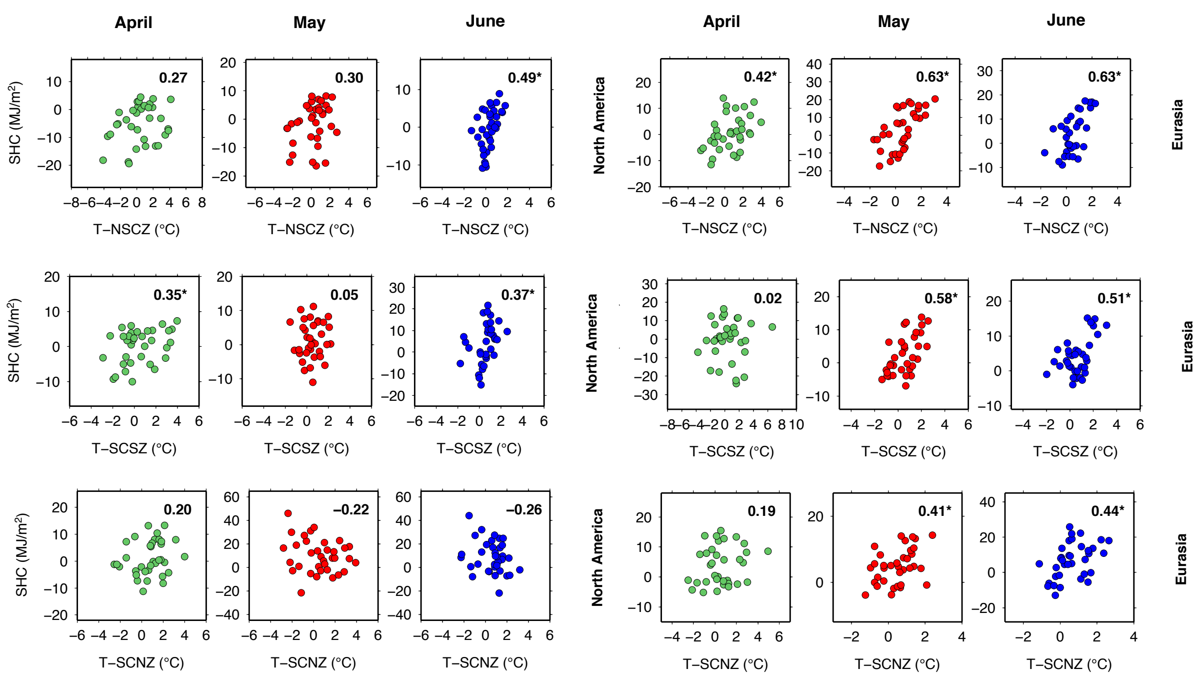

Figure 7. Correlations between observed SAT and simulated SHC in NSCZ, SCSZ and SCNZ over North America and Eurasia from April through June for the period 1972-2006. The correlation with asterisks is statistically significant at a level of $p < 0.025$ when its absolute value is greater than 0.34.

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Table 2. Trend analyses for observed snow cover extent (SCE) in the snow covered sensitivity zone (SCSZ) and the snow covered non-sensitivity zone (SCNZ) over North America and Eurasia from April through June for the period 1972-2006. The significance level (*p*-value) was calculated using a two-sided Mann-Kendall trend test. Trend slope (ts) units are year^{-1}.

Table 3. Trend analyses for CRU monthly surface air temperature (SAT) in the non-snow covered zone (NSCZ), SCSZ, and SCNZ over North America and Eurasia from April through June for the period 1972-2006. The significance level (*p*-value) was calculated using a two-sided Mann-Kendall trend test. Ts units are °Cyear^{-1}.

Table 4. Correlation coefficients due to the linear trend and variability for SHC derived from VIC and NOAA SCE observations in SCSZ and SCNZ over North America and Eurasia from April to June for the period 1972-2006. The significance level (*p*-value) was calculated using a two-tailed Student t-test with 33 degrees of freedom.

Table 5. Correlation coefficients due to the linear trend and variability for SHC derived from VIC and CRU SAT in NSCZ, SCSZ, and SCNZ over North America and Eurasia from April

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Soil Thermal Node	Depth (m)
STN0	0.0
STN1	0.2
STN2	0.4
STN3	0.6
STN4	0.9
STN5	1.3
STN6	1.7
STN7	2.1
STN8	2.7
STN9	3.3
STN10	4.1
STN11	5.1
STN12	6.1
STN13	7.3
STN14	8.8
STN15	10.6
STN16	12.6
STN17	15.0

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	North America						Eurasia					
	April		May		June		April		May		June	
	$p <$	ts	$p <$	Ts	$p <$	ts	$p <$	ts	$p <$	ts	$p <$	ts
SCE-SCSZ	0.025	-0.0052	0.01	-0.0026	0.01	-0.0029	0.025	-0.0042	0.01	-0.0035	0.005	-0.0034
SCE-SCNZ	--	-0.0002	--	-0.0000	--	-0.0000	--	0.0003	--	-0.0010	--	-0.0006

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	North America						Eurasia					
	April		May		June		April		May		June	
	p <	ts	*p* <	Ts	*p* <	ts	*p* <	ts	*p* <	ts	*p* <	ts
SAT-NSCZ	--	0.0345	--	-0.0243	0.005	0.0323	--	0.0531	0.005	0.0663	0.005	0.0412
SAT-SCSZ	--	0.0400	--	0.0243	0.005	0.0415	--	0.0143	0.005	0.0500	0.005	0.0552
SAT-SCNZ	0.005	0.0657	0.005	0.0806	0.01	0.0467	--	0.0044	0.005	0.0435	0.005	0.0471

Table 4. Correlation coefficients due to the linear trend and variability for SHC derived from VIC and NOAA SCE observations in SCSZ and SCNZ over North America and Eurasia from April to June for the period 1972-2006. The significance level (p-value) was calculated using a two-tailed Student t-test with 33 degrees of freedom.

			April		**May**		**June**	
			$p <$	r	$p <$	r	$p <$	r
Correlation (Trend)	**North America**	SCSZ	0.025	-0.9	0.025	-0.7	0.025	-0.8
		SCNZ	0.025	-0.5	--	0.0	--	0.0
	Eurasia	SCSZ	0.025	-0.9	0.025	-1.0	0.025	-0.8
		SCNZ	0.025	0.9	0.025	-0.7	0.025	-0.7
Correlation (Variability)	**North America**	SCSZ	--	-0.2	--	-0.1	--	0.0
		SCNZ	--	0.0	--	0.0	--	0.0
	Eurasia	SCSZ	--	-0.0	--	-0.0	--	0.1
		SCNZ	--	-0.1	--	-0.1	--	0.1

Table 5. Correlation coefficients due to the linear trend and variability for SHC derived from VIC and CRU SAT in NSCZ, SCSZ, and SCN over North America and Eurasia from April to June for the period 1972-2006. The significance level (*p*-value) was calculated using a two-tail Student t-test with 33 degrees of freedom.

			April		May		June	
			$p <$	r	$p <$	r	$p <$	r
Correlation (Trend)	**North America**	NSCZ	0.025	0.3	0.025	-0.7	0.025	0.7
		SCSZ	0.025	0.9	0.025	0.7	0.025	0.8
		SCNZ	0.025	0.5	--	0.0	--	0.0
	Eurasia	NSCZ	0.025	0.9	0.025	1.0	0.025	1.0
		SCSZ	0.025	0.9	0.025	1.0	0.025	0.8
		SCNZ	0.025	1.0	0.025	0.7	0.025	0.7
Correlation (Variability)	**North America**	NSCZ	--	0.2	0.025	0.4	--	0.37
		SCSZ	--	0.1	--	-0.2	--	-0.1
		SCNZ	--	-0.0	--	-0.2	--	-0.2
	Eurasia	NSCZ	--	0.2	--	0.1	--	0.2
		SCSZ	--	0.0	--	0.2	--	0.1
		SCNZ	--	0.1	--	0.1	--	0.1