Relationship between anomalies of Eurasian snow and southern China rainfall in winter

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Abstract
Characteristics of the snow water equivalent (SWE) over high-latitude Eurasia and its relation with precipitation in China during January, February and March (JFM) are investigated. The JFM Eurasian SWE exhibited a decadal downward shift in the late 1990s, marked by a frequently positive phase in 1979–98 and a negative phase afterward. The decadal shift corresponds to anomalous northeasterly flow over southeastern China. Consequently, warm and moist airflow from tropical oceans is weakened, accompanied by reduced rainfall over southeastern China. The US National Centers for the Environmental Prediction Climate Forecast System (CFS) capture both the interannual variation and the decreasing trend of JFM Eurasian SWE reasonably well for several months in advance. The relationship between Eurasian SWE and southeastern China rainfall is also captured by the CFS in the prediction.

Keywords: Eurasian snow, China precipitation, NCEP CFS

1. Introduction
The characteristics of Eurasian snow and their variations have been analyzed in various studies (e.g. Ye and Bao 2001). Gutzler and Rosen (1992) showed a decreasing trend of February Eurasian snow cover for 1972–90. Brown (2000) reported that Eurasian SWE was characterized by a significant reduction in April coverage during 1922–97. Bulygina et al (2009) showed a decrease in the duration of snow cover in the northern regions of European Russia and in the mountainous regions of southern Siberia from 1966 to 2007. Wu et al (2009) and Zuo et al (2011) also reported a significant decadal shift of springtime Eurasian snow water equivalent (SWE) in the late 1980s based on an analysis for 1979–2004.

Snow can influence the climate system in several ways: (1) increasing the surface albedo, (2) increasing the amount of infrared radiation lost via high emissivity, (3) serving as a thermal insulator due to its low thermal conductivity, and (4) melting snow as a sink for latent heat (Sellers 1965, Wagner 1973, Yasunari et al 1991, Jones 2001, Yasunari 2007, IPCC 2007, Li et al 2009). Furthermore, these snow effects have strong spatial and seasonal dependence. For example, Cohen and Rind (1991) showed that snow cover via the reduction in absorbed shortwave radiation and the increased latent heat sink of melting snow only caused a short-term local decrease in surface temperature because of the negative feedback between snow and the emitted longwave radiation, and between snow and sensible and latent heat flux. Groisman et al (1994) reported that the feedback of snow cover on the planetary radiative balance may be positive during autumn and early winter, although the annual-mean response of snow cover feedback on radiative balance is negative.

More than a century ago, Blanford (1884) documented an inverse relationship between winter snow over the Himalayas and subsequent all-India monsoon rainfall (AIMR). There has been increasing evidence that snow may generate anomalous atmospheric forcing via changing the process of energy and water transfer between the land surface and the atmosphere.
The existence of a snow–monsoon relationship was supported by subsequent studies with updated snow data and numerical models (Hahn and Shukla 1976, Dey and Bhanu Kumar 1983, Dickson 1984, Barnett et al 1989, Yang and Lau 1998, Bamzai and Shukla 1999, Faluollo 2004, Chen and Wu 2000, Wu and Qian 2003, Zhang et al 2004). These studies focused on the relationship between the snow over Eurasia and the Tibetan Plateau and the Asian summer monsoon rainfall. Some studies emphasized the complex nature of the relationship between Eurasian snow and the broader-scale Asian monsoon which is also strongly influenced by El Niño–Southern Oscillation and Arctic Oscillation (AO) (Yang 1996, Sankar-Rao et al 1996, Liu and Yanai 2002, Yang et al 2003, Kumar and Yang 2003, Xin et al 2010). Cohen and Saïto (2003) demonstrated that during the observational period of continental snow cover, a constructed leading snow cover index (based on summer and autumn snow cover) was more skillful in predicting land surface temperature in eastern North America than the AO index.

Another aspect of the influence of land surface condition is the impact of snow over mid–high–latitudes on the rainfall over China. During winter, 60% of the Eurasian continent is covered with snow (Moriagha et al 2003), which affects both local and large-scale atmospheric circulation patterns and hydrological processes. Yang and Xu (1994) found that wintertime Eurasian snow was positively correlated with the summer rainfall in southern and northern China but negatively correlated with the summer rainfall in central and northeastern China. Wu and Kirtman (2007) showed that the enhanced spring snow cover in western Siberia corresponded to above-normal spring rainfall in southern China. Wu et al (2009) showed that negative Eurasian SWE anomalies in spring were associated with enhanced rainfall in southern China and reduced rainfall in the upper reach of the Yellow River in summer.

Both snow cover and snow depth are important quantities for analyzing snow variability. Snow cover directly affects surface albedo and snow depth provides information for the insulation properties of snow. Because SWE is the product of snow density times snow depth divided by water density, it is a useful variable to analyze snow variability, especially for hydrological purposes (Armstrong and Brodzik 2005, Déry et al 2005). However, available in situ SWE observations are sparse and may not be suitable for systematic analysis of large-scale variability. The National Center for Environmental Prediction (NCEP) has recently completed a new coupled global reanalysis, the Climate Forecast System Reanalysis (CFSR), for 1979 to the present (Saha et al 2010). The SWE in the CFSR includes the assimilation of observational and analysis data from the Air Force Weather Agency’s Snow Depth model (SNODEP) (Kopp and Kiess 1996) and the National Environmental Satellite Data and Information Service (NESDIS) Interactive Multisensor Snow and Ice Mapping System (IMS) (Helfrich et al 2007). Accordingly, the CFSR provides an optimal SWE synthesis of observation and the first guess from CFSR model integration. In addition to the CFSR, NCEP has also produced retrospective seasonal forecasts (or hindcasts) with a new version of the NCEP coupled Climate Forecast System (CFSv2) initialized from CFSR for 1982–2009, allowing an assessment of the seasonal predictability of various climate variabilities.

In this study, we investigate the characteristics of Eurasian SWE and its relationship with precipitation over China in January, February and March (JFM) using the CFSR SWE data and the Climate Prediction Center (CPC) unified precipitation data. We also assess the predictability of snow and the snow–precipitation relationship using the CFSv2 hindcasts.

2. Data and methods

The CFSR has a global horizontal resolution of ~38 km and vertical resolution of 64 levels with top pressure at ~0.266 hPa. The land analysis data are from the Global Land Data Assimilation System (GLDAS). The SNODEP mode and the NESDIS IMS are used in updating the CFSR SWE. SNODEP has been operational since 1975 and is available for the entire reanalysis period. The IMS data are manually generated Northern Hemisphere snow cover analysis (determined by surface data, geostationary and polar-orbiting imagery, and microwave-based detection algorithms) produced once per day. They are available at 23 km resolution starting in February 1997 and at 4 km resolution starting in February 2004 (Saha et al 2010). We analyze the monthly CFSR SWE and winds at 700 hPa. For precipitation, we use the CPC unified global monthly gauge, constructed on a 0.5° latitude × 0.5° longitude grid over the global land through the interpolation of quality-controlled rain gauge reports from ~30 000 stations collected from the Global Telecommunication System and many other national and international analyses (Xie and Arkin 1997, Xie et al 2007). Our analysis is based on the datasets from 1979 to 2010.

The NCEP Climate Forecast System version 1 (CFSv1) has been used for operational forecast since 2004 (Saha et al 2006). A new system (CFSv2) based on the same coupled atmosphere–ocean–land model for CFSR was implemented in March 2011. The atmospheric component of CFSv2 is the NCEP Global Forecast System used for operational weather forecasting (Mooithi et al 2001) with T126 horizontal and 64-level vertical resolutions. The oceanic component is the NOAA Geophysical Fluid Dynamics Laboratory Modular Ocean Model version 4 and the land component is the Noah Land Surface Model (Saha et al 2010). The CFSv2 hindcasts for nine target months were produced with four runs every fifth day initialized from the CFSR. For each initial month, an ensemble of 16 members is used. For example, for 0-month lead (LM0) CFSv2 JFM data, the initial conditions of days 7, 12, 17, and 22 December are run for January, days 11, 16, 21, and 26 January are run for February, and days 5, 10, 15, and 20 February are run for March. We analyze the CFSv2 hindcasts from 1983 to 2009. In addition, correlation and composite analyses are used to reveal the relationship between SWE and rainfall in China.

3. Results and discussion

We defined the normalized JFM SWE over high-latitude Eurasia (HLEA, 55–70°N, 0–180°E) as a SWE index to describe the characteristics of Eurasian snow in winter. Clearly, the SWE index shows a decreasing tendency. The decadal shift
occurred in the late 1990s, with the dominance of negative (positive) phase during 1999–2010 (1979–98) (figure 1(a)). Based on the decadal variation in JFM SWE over HLEA, the period 1979–98 is selected as a high snow index episode (hereafter referred to as HSWI) and 1999–2010 as a low snow index episode (hereafter referred to as LSWI). The values of JFM SWE (also precipitation and horizontal winds at 700 hPa) were averaged in the two episodes, 1999–2010 and 1979–98 (for CFSv2 dataset, the two episodes are 1999–2009 and 1983–98) respectively, and the discrepancies between them represent the differences between LSWI and HSWI. Figure 1(b) shows the difference in JFM SWE between LSWI and HSWI over the Eurasian continent. SWE reduces over most of Eurasia, with a reduction of 40 kg m$^{-2}$ in East Europe and Siberia. Increased SWE is found only over a few small areas such as the regions around Turkey, Lake Balkhash, and Baikal. Previous studies reported a decrease in Eurasian spring snow with SWE data from the National Ice and Snow Data Center (Wu et al. 2009). Both this current study and Wu et al. (2009) showed an evident decreasing trend in Eurasian snow in winter and spring.

Corresponding to the reduced SWE over HLEA, JFM precipitation decreased over southern China, with a reduction of about 1.0 mm/day in southeastern China (figure 2). Winds at 700 hPa are used to examine the change in atmospheric circulation associated with the changes in snow over HLEA and precipitation in southeastern China. The difference in 700 hPa winds between LSWI and HSWI shows strong anomalous northeasterly flow over southeastern China and the surrounding oceans (figure 3). Consequently, warm and moist flow from the tropical oceans is weakened, accompanied by reduced rainfall over southeastern China. The in-phase relationship between the snow over HLEA and the rainfall over southeastern China was also found by Wu and Kirtman (2007) using observed snow cover and precipitation data for 1979–2000 in spring, indicating the robustness of the relationship between the high-latitude snow and the tropical–subtropical rainfall.

The CFSv2 JFM SWE for LM0, one-month lead (LM1), and two-month lead (LM2, hereafter a one-month lead is defined as LM1, two-month lead as LM2, and so on) captures the variations (both the interannual variability and
Figure 3. Difference in horizontal winds at 700 hPa between LSWI and HSWI (1999–2010 minus 1979–98; unit: m s$^{-1}$).

the decreasing trend) of observed HLEA SWE quite well, although the CFSv2 overestimates the value of SWE slightly (figure 4(a)). As the lead time increases, the interannual variability of the ensemble mean becomes weaker. The time scale beyond which the model loses the initial observed anomalies is quantified in figure 4(b) which shows the correlation between the CFSv2 forecast and CFSR SWE. It is seen that the model is capable of capturing the observed anomalies reasonably well at LM0 to LM3 with a correlation coefficient of 0.97, 0.89, 0.60 and 0.42 (figure 4(b)), respectively, indicating that the CFSv2 can capture the Eurasian SWE three months ahead. No useful skill is found in predicting at a lead time of four months (LM4, using the ICs of September (for January), October (for February), and November (for March)) or longer. Variations of CFSv2 SWE from LM4 to LM8 are very small, with a climatological-mean value of SWE around 90 kg m$^{-2}$. Also, the correlation coefficients are insignificant between the CFSR SWE and the CFSv2 SWE from LM4 to LM8. Another feature in the CFSv2 is that model SWE increases with lead time, indicating larger snowfall rate in the model than that in the observation.

The CFSv2 captures the decreasing trend of JFM SWE over HLEA quite well (figure 5(a)), with maximum negative anomalies of $-40$ kg m$^{-2}$ in East Europe and Siberia. However, the CFSv2 simulates slightly positive SWE anomalies between LSWI and HSWI over part of West Europe, which is not seen in the CFSR SWE. In addition, the positive SWE anomalies around Turkey, Lake Balkhash, and Baikal are larger than those in the CFSR. The reduced precipitation over southern China and surrounding oceans is also captured by the CFSv2 successfully, with maximum anomalies $-0.6$ mm/day over southeastern China (figure 5(b)), which is consistent with the feature in the CPC unified precipitation (figure 2), in spite of a slightly smaller magnitude. Furthermore, the CFSv2 captures the anomalous northeasterly wind around southeastern China and the easterly anomalies along 20°N, but are slightly smaller than that in CFSR. The characteristics aforementioned indicate that the CFSv2 is skilful for predicting the relationships of the SWE over HLEA with the rainfall in southeastern China and the horizontal winds over East Asia.

4. Conclusions

Using the SWE data from the recently completed coupled global NCEP Climate Forecast System Reanalysis (CFSR) at high spatial resolution (T382L64), we found that the JFM SWE over high-latitude Eurasia continent has exhibited a decreasing trend in the past few decades. Large positive SWE anomalies frequently appeared in 1979–98, whereas negative anomalies mostly occurred in 1999–2010. Corresponding to the reduced JFM SWE over HLEA, lower-tropospheric northeasterly anomalies occurred over southeastern China and the surrounding oceans. As a result, warm and moist airflow from tropical oceans was blocked over the South China Sea, Philippines, and the Philippine Sea, leading to negative rainfall anomalies over southeastern China.

The NCEP CFS provides monthly and seasonal climate predictions over the globe, and the CFS products are now becoming an important source of information for regional climate predictions. Retrospective forecasts from the new version of CFS, CFSv2, are analyzed and it is found to be
Figure 5. Differences in CFSv2 LM0 values: (a) SWE (unit: kg m\(^{-2}\)), (b) precipitation (unit: 0.1 mm/day), and (c) horizontal wind at 700 hPa (m s\(^{-1}\)), between LSWI and HSWI (1999–2009 minus 1983–98).

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