INFLUENCE OF EURASIAN SPRING SNOW COVER ON ASIAN SUMMER RAINFALL

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ABSTRACT

The Eurasian snow cover anomaly in spring has been considered as one of the important factors affecting Asian summer monsoon variability. Using the long time series (1922–98) of Eurasian spring (March–April) snow cover (ESSC) reconstructed by Brown (2000. Journal of Climate 13: 2339) and snow cover (1973–98) and depth (1979–87) data from satellite observation, the influences of ESSC on the all-India monsoon (June–September) rainfall (AIMR) and the summer rainfall over all parts of Asia are examined. It is found that the statistical relation between AIMR and ESSC changes over a multi-decadal time scale. The negative correlation between them has increased markedly since the mid 1970s. The region where the summer rainfall has the strongest and most stable negative correlation with the preceding ESSC is located in northern Mongolia, south of Lake Baikal. The correlation between the summer rainfall and ESSC increases after the data are treated with a low-pass filter, showing that the impact of snow cover may be seen more clearly with the removal of the effect of El Niño–southern oscillation. Comparative analyses for contrasting years with excessive and deficient snow cover show that the anomalies of ESSC occur mainly in northwestern Eurasia. In the years of excessive ESSC anomalies, cooling and a cyclonic circulation anomaly in the lower troposphere appear over the northern part of Eurasia, leading to a Rossby-wave-train-like circulation response, then a weakened East Asia summer monsoon and deficient rainfall with an anticyclonic circulation anomaly south of Lake Baikal. Anomalies with opposite signs occur in the years of deficient snow cover. Copyright © 2002 Royal Meteorological Society.

KEY WORDS: Eurasian snow cover; Asian summer rainfall; Asian monsoon; correlation

1. INTRODUCTION

The Eurasian winter/spring snow cover has been considered an important factor influencing the land thermal conditions and then seasonal-to-interannual variabilities of the succeeding Asian summer monsoon and rainfall. Observational studies using satellite-measured snow cover or depth data (Hahn and Shukla, 1976; Kripalani et al., 1996; Sankar-Rao et al., 1996; Bamzai and Shukla, 1999) and historical snow depth data (Kripalani and Kulkarni, 1999) showed that there generally exists an inverse relationship between the year-to-year variations of snow cover or depth over Eurasia and the Indian monsoon rainfall, indicating that excessive (deficient) Eurasian snow cover in winter/spring is followed by weak (strong) Indian monsoon rainfall. Moreover, the winter snow anomaly of western Eurasia seems to have the strongest inverse correlation with the Indian summer rainfall (Bamzai and Shukla, 1999; Kripalani and Kulkarni, 1999). The linkage between Eurasian snow cover and summer rainfall over China was shown to be more complicated (Yang and Xu, 1994). Additionally, the snow–monsoon relation is influenced by El Niño–southern oscillation (ENSO) events (Sankar-Rao et al., 1996; Yang, 1996).

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A number of numerical experiments based on general circulation models (GCMs) (e.g. Barnett et al., 1989; Yasunari et al., 1991; Ose, 1996; Ferranti and Molteni, 1999) have been carried out to simulate the inverse snow–monsoon relation found in the observational analyses. Most of the modelling results confirmed the statistical relationship between the snow and the monsoon. The effect of seasonal snow on the summer monsoon circulation is generally regarded as a result of two dominant feedback mechanisms (Shukla, 1987; Barnett et al., 1989; Yasunari et al., 1991; Meehl, 1994). Persistent Eurasian snow anomalies in winter and spring may induce changes in the surface and atmospheric temperatures with the surface albedo and soil hydrological effects, leading to variations in the large-scale land–sea thermal contrast and the following summer monsoon.

Despite the numerous studies on the relation between the Eurasian winter/spring snow anomaly and Asian summer rainfall, we still find some aspects of the relation worth investigating further. There is relatively little documentation of snow–rainfall statistics on a multi-decadal time scale owing to a lack of comprehensive data. Most of the published observational studies, except Kripalani and Kulkarni (1999), were limited to the records of the recent 20 years or so for which the snow data from satellites are available. Moreover, these studies were mainly concerned with the relation of the winter/spring snow anomaly to the Indian or Chinese monsoon rainfall, without examining its possible relation with the summer rainfall over other parts of the Asian continent. In addition, most of the previous observation studies, except Sankar-Rao et al. (1996), focused on the effect of snow anomaly on the monsoon rainfall, and rarely examined the effect on the large-scale atmospheric circulation.

In this paper we shall use a long time series (1922–98) of the Eurasian spring snow cover reconstructed by Brown (2000) and recent snow cover (1973–98) and depth (1979–87) data from satellite observations to examine further the influences of the snow anomaly on the Asian summer rainfall, atmospheric temperature and circulation over Eurasia. The data used in this study are described in Section 2. The correlation analyses on multi-decadal scales and case studies for years of excessive and deficient snow cover are detailed in Sections 3 and 4 respectively. Finally, in Section 5, we summarize the results and discuss briefly the importance, as well as limitation, of the snow anomaly in modulating the atmospheric circulation and rainfall variabilities.

2. DATA

The 1950–98 monthly land precipitation data on a 2.5° × 2.5° grid (Chen et al., 2002) are used as a basic dataset of precipitation for this paper. This gridded field of monthly precipitation is defined by interpolating gauge observations at over 15 000 stations in the world using the optimum interpolation algorithm. Additionally, the all-India monsoon (June–September) rainfall (AIMR) series for 1922–98 is also used. The gauge-based AIMR was prepared by the Indian Institute of Tropical Meteorology (Parthasarathy et al., 1987; Parthasarathy et al., 1994).

The monthly mean tropospheric temperature and wind data at several levels with 2.5° × 2.5° grid are extracted from the National Centers for Environment Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis dataset (Kalnay et al., 1996). We use these data to show possible impacts of the Eurasian snow cover anomaly on the temperature, circulation and, eventually, rainfall.

Two time series related to the underlying surface forcing are employed: the 1922–98 June–September Nino3 (5°N–5°S, 150°–90°W) sea surface temperature anomaly (SSTA) series is taken from Latif (2000); and a 1922–98 time series of the Eurasian land surface temperature anomaly (ELTA) averaged for March and April is calculated with monthly land surface air temperature on a 5° × 5° grid provided by the Climatic Research Unit, University of East Anglia, UK (Jones, 1994). Here, both SSTA and ELTA are calculated with 1961–90 as the base period.

Three types of snow data are used in this study. The first is the monthly snow cover obtained from the National Oceanic and Atmospheric Administration (NOAA) satellites. The original data were described in Robinson et al. (1993). The satellite-measured snow-cover data used in the present paper are on a 2° × 2° Northern Hemisphere grid from 1973 to 1998. The second is the 1979–87 monthly 1° × 1° snow depth obtained by the National Aeronautics and Space Administration (NASA) with the Scanning Multi-channel
Microwave Radiometer (SMMR) on the Nimbus-7 satellite (Chang et al., 1992). The third is the reconstructed time series of the Eurasian spring (March–April, the same in the following) snow cover (ESSC) for 1922–98 (Brown, 2000). This series of ESSC was derived from a snow-cover area index (1922–71) reconstructed with available in situ snow-depth data over western Eurasia and satellite-observed snow cover (1972–97) over the entire Eurasian continent. The method for reconstructing ESSC series was detailed in Brown (2000). The latest year of the original series was 1997. In this paper the value of 1998 is constructed from the regression of the original series with the satellite snow-cover data. Because of the poor coverage of in situ snow-depth data over eastern Eurasia, Brown (2000) reconstructed the ESSC before 1972 by using only a snow-cover area index for western Eurasia. It was found, however, using the satellite observations, that the variation of the snow-cover extent over western Eurasia is closely correlated with that over the entire continent, especially in March and April. Thus, the time series of reconstructed ESSC is considered sufficiently accurate for climate studies in reflecting fluctuations of the snow cover on a continental scale (Brown, 2000). Here, we can further confirm the reliability of the reconstructed ESSC data. By comparing the reconstructed ESSC anomaly with ELTA, the simultaneous land temperature for the entire Eurasian continent (Figure 1), we find a significant and stable inverse relationship between them, with a correlation coefficient of −0.67 for 77 years. This suggests that the reconstructed ESSC series is reliable even in the early years.

3. CORRELATION ANALYSES

3.1. Interdecadal change of the relation of Indian monsoon rainfall to the Eurasian spring snow cover

Although the relationship between ESSC and AIMR has been investigated in previous studies, it has never been examined on a multi-decadal time scale. Here, we shall use the data from 1922 to 1998 to re-examine their relationship. Figure 2(a) shows year-to-year variations of AIMR and ESSC. For the sake of comparison, variations of AIMR and SSTA are shown in Figure 2(b). Note that the SSTA here means an average for June–September, the same months as the AIMR uses. Thus we consider only the simultaneous relation between AIMR and SSTA here, because the AIMR has been found to be correlated with the simultaneous and time-lagged ENSO indices rather than time-lead ENSO indices (Elliott and Angell, 1987; Prasad and Singh, 1996). As shown in Figure 2, AIMR has a significant negative correlation with the simultaneous SSTA (correlation coefficient −0.51), but it has no clear correlation with the preceding ESSC (−0.08) for
Figure 2. (a) Year-to-year variations of the AIMR and ESSC. The thick solid lines represent the trends fitted with sixth-order polynomials. (b) Same as (a) but for AIMR and Nino3 summer (June–September) SSTA.

The lack of significant correlation between AIMR and ESSC on the longer time scale is in sharp contrast to the snow–monsoon inverse relation found in many previous studies for relatively short periods of recent years (e.g. Sankar-Rao et al., 1996; Bamzai and Shukla, 1999). After excluding possible unreliability of the reconstructed ESSC data in early years (Section 2), the lack of a clear relationship over the 77 years is likely attributed to the non-stationary nature of the correlation between AIMR and ESSC. It is found that the 11 year moving window correlations of AIMR with ESSC were generally weak or even positive before the 1970s, but became strongly negative after 1975 (not shown). This means that the relationship between ESSC and AIMR
has undergone considerable interdecadal changes during the past 77 years, and that the inverse relation between
the snow and the monsoon has intensified over the recent 20 years or so. The interdecadal change between
ESSC and AIMR has also been noted by Kripalani and Kulkarni (1999). On the other hand, 11 year moving
window correlations between AIMR and SSTA (not shown) indicate that the strong negative correlation
between them has been interrupted since the late 1970s, as several investigators (Kumar et al., 1999; Miyakoda
et al., 2000) have pointed out. Because of no long-term trend in AIMR and only a weakly increasing trend
in SSTA, as shown in Figure 2, we speculate that the clear descending trend in ESSC with the recent global
warming since the early 1970s may have made the Asian summer monsoon increasingly sensitive to the change
in ESSC and eventually brought interdecadal changes of AIMR–ESSC and even AIMR–ENSO relationships.

3.2. Correlation between interannual variations of the Eurasian spring snow cover and summer rainfall over
Asia

In this section we shall examine the relation of ESSC with the summer (June–September, the same in
the following) rainfall everywhere over the Asian continent, including the Indian subcontinent. Figure 3(a)
shows the correlation coefficients between ESSC and the summer rainfall at every grid-box for 1950–98.
The correlation was not strong for the past half century as a whole. However, there is a systematic
northeast–southwest band with negative correlation from northeast China to northwest India. The strongest
negative correlation (less than $-0.2$) is located in northern Mongolia, south of Lake Baikal. In contrast, summer
rainfall over Asia, especially in north India and north China, shows much stronger negative correlation with
the simultaneous SSTA (Figure 3(b)).

When we divide the 49 years under discussion into two stages 1950–73 (Figure 4(a)) and 1974–98
(Figure 4(b)), it is found that the ESSC–rainfall correlation pattern has changed and the overall correlation has
intensified in the later stage (Figure 4(b)), especially in northwest India. At the same time, northern Mongolia
is a region with relatively stable negative correlation, and the correlation there has also been strengthened in
the later stage.

Another interesting finding is that the correlation of the Asian summer rainfall with ESSC becomes stronger
when the data of both the rainfall and ESSC are treated with a low-pass filter to retain only those variations
with periods longer than 7 years (thus excluding the components of ENSO related to SSTA). In the patterns
of the correlation between the low-pass filtered components of the rainfall and ESSC (not shown), it is seen

Figure 3. Distributions of the correlations of the grid-box summer rainfall (a) with the ESSC and (b) with the Nino3 SSTA for 1950–98.
The areas of negative correlation are shaded

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Figure 4. Distributions of the correlations between the grid-box summer rainfall and the ESSC (a) for 1950–73, and (b) for 1974–98. The areas of negative correlation are shaded.

Figure 5. Year-to-year variations of the low-pass-filtered ESSC and summer rainfall rate averaged for a box (47.5–52.5°N, 102.5–107.5°E) over northern Mongolia.

that the negative correlations near Lake Baikal, with central values less than −0.6 for both stages, are much stronger than those before the filtering. This means that a closer relation between the summer rainfall and ESSC may exist at a time scale longer than that of ENSO.

As seen in Figure 5, year-to-year variations of the low-pass filtered ESSC and the rainfall averaged for a box (47.5–52.5°N, 102.5–107.5°E) over northern Mongolia show a distinct inverse relationship. During the last 49 years, there are three major out-of-phase stages: the middle 1950s, late 1970s–early 1980s and late 1980s–early 1990s. Excluding the first stage, when satellite data were not available, we will choose 3 years each in the second stage (1979–81) and the third stage (1989–91) to analyse the characteristics and
impacts of ESSC. It should be pointed out that these 6 years are those without large anomalies in SSTA (see Figure 2(b)). This is important because the Asian monsoon variability is so closely linked to ENSO that erroneous conclusions may be drawn in exploring the impact and response of the Asian monsoon without excluding the effect of ENSO (Goswami and Jayavelu, 2001).

4. CASE STUDIES

4.1. Some characteristics of the Eurasian snow cover anomalies for typical years

In order to show the annual cycle and persistence characteristics of the Eurasian snow cover, we examine monthly snow-cover anomalies (relative to 1973–98) from satellite observations since 1973 and find that the 3 years of 1979–81 (1989–91) are typical with extensive (deficient) spring snow cover (Figure 6). For example, the snow-cover anomaly averaged for April of 1979, 1980 and 1981 reaches 3.03 × 10^6 km², but that for March of 1989, 1990 and 1991 is −1.83 × 10^6 km². Moreover, the strong snow-cover anomalies show good persistence, maintaining the same sign at least until June.

The climatological (1979–87) distribution of spring (March and April) snow depth on the Eurasian continent (Figure 7) shows that the seasonal snow mainly covers northern Eurasia north of about 45 °N and the Tibetan

Figure 6. Annual march of monthly snow cover anomalies for (a) three heavy-snow years (1979–81) and (b) three light-snow years (1989–91). The thick grey lines indicate the mean for the 3 years

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Plateau. An area with snow depth more than 40 cm is found in the northeastern part of the continent. This pattern compares with that from historical snow-depth data (Kripalani and Kulkarni, 1999). However, the difference in snow-cover area between the extensive and deficient snow years occurs mainly in the northwestern part of the continent. For example, Europe experienced more snow cover in the first week of March 1979 than in the first week of March 1990 (not shown). Figure 8 shows the difference of spring snow-cover frequency (total number of weeks with snow cover during March and April) averaged for 1979–81 and for 1989–91. The snow cover in the heavy snow years lasted 2 weeks or longer than in the light snow years in eastern Europe, eastern Tibetan Plateau and northeast China. Thus, the snow-cover anomaly in northwestern Eurasia appears
to be a major factor in accounting for the interannual variability of ESSC. This may be a reason why the reconstructed ESSC series, even though only the snow cover data on western Eurasia were used for its reconstruction in the early years (Brown, 2000), can approximate the interannual variation of the whole Eurasian snow cover.

4.2. Impacts of the Eurasian snow cover anomalies on the atmospheric temperature, circulation and rainfall

Previous observational (Sankar-Rao et al., 1996) and modelling (Walland and Simmonds, 1997) studies showed that there is a significant reduction in the lower tropospheric temperature with an increase in snow cover. Figure 9(a)–(c) shows differences between the air temperatures averaged for March–April of 1979–81 and those of 1989–91 at three levels. At 850 hPa, a centre of large negative values lower than −4 °C appears.

![Figure 9](image_url)
over northern Europe and a band of negative values is present over the whole of northern Eurasia extending from Europe to Asia. The magnitude of temperature difference values is reduced with height. For example, the value at the centre over northern Europe is about $-3^\circ$C at 700 hPa and $-2^\circ$C at 500 hPa. This decreasing temperature difference with height suggests that the excessive snow cover occurring mainly in northwestern Eurasia (Figure 8) causes the cooling in the lower troposphere. The fact that the influence of the Eurasian snow cover is mainly limited in the lower troposphere is consistent with the finding of Liu and Yanai (2001); that there is little relationship between the upper tropospheric temperature and land surface temperature over Eurasia on an interannual time scale. In the following May–June (Figure 9(d)–(f)), the centre of negative values over northwestern Eurasia has shifted eastwards, with reduced values of temperature difference. The cooling over northeastern Eurasia still continues. The cooling over the Eurasian continent in spring will delay the development of the Asian summer monsoon.

Figure 10. Same as Figure 9, but for the wind differences shown with streamlines
The ESSC anomaly also exerts marked influences on the atmospheric circulation. For example, in the differences between the March–April wind of 1979–81 and those for 1989–91 (Figures 10(a)–(c)), there is a cyclonic anomaly over northwestern Eurasia at all three levels, suggesting a response of the wind field to the extensive snow-cover anomaly. The centre of the cyclonic anomaly generally corresponds to the cooling centres (see Figure 9(a)–(c)) and is located on the northeastern side of the main ESSC anomaly in northwestern Eurasia (see Figure 8). From late spring to early summer (May–June), a cyclonic circulation anomaly persists over central Russia (Figure 10(d)–(f)). We notice that another cyclonic circulation anomaly appears near the eastern coast of Asia. Between the two cyclonic anomalies, there exists an anticyclonic anomaly on the northeast side of the Tibetan Plateau (south of Lake Baikal). In the wind field anomaly maps at all three levels, one can observe a chain of successive anticyclonic and cyclonic anomalies with a barotropic structure over the northern part of Eurasia. This pattern may be interpreted as a Rossby wave train, which makes atmospheric disturbances induced by the snow-cover anomaly propagate from Europe to Asia.

We also notice that the March–April temperature over the northwestern Tibetan Plateau is higher in the heavy ESSC years than in the light ESSC years (Figure 9(a) and (b)). A similar warming in April–May surface temperature over the Tibetan Plateau has been recognized by Bamzai and Shukla (1999: figure 7). They suggested that the winter and spring anomaly in snow cover over the Tibetan Plateau tends to have a sign opposite to that in northern Eurasia. In the 6 years chosen for the present case studies, however, we observe more frequent snow cover over the eastern Tibetan Plateau, as well as over northwestern Eurasia (see Figure 8). It is difficult to investigate the reasons for the observed warming over the Tibetan Plateau owing to the lack of in situ observations on the western Tibetan Plateau. Warm horizontal temperature advection along the western periphery of the Tibetan Plateau inferred from the southerly circulation anomaly (Figure 10(a) and (b)) responding to the ESSC anomaly may, at least partly, explain the warming over the western Tibetan Plateau in the heavy-snow case.

In summer, the most remarkable feature over Asia in the wind field at 850 hPa is the monsoonal circulations (Figure 11(a)), with prevailing westerly wind over South Asia and southerly wind over East Asia. In the field of the mean wind difference between the high snow-cover years and the low snow-cover years (Figure 11(b)), the anomalous patterns appearing in May–June (Figure 10(f)) still exist in summer. The northerly wind anomaly over East Asia leads to a weaker summer monsoon there in the extensive snow years. The above anomalies in the atmospheric circulation will result in variations in the summer rainfall. In Figure 12, the average difference of summer rainfall rate between the heavy-snow and light-snow years is shown. An area of negative values is found in the region from northern Mongolia to northeastern China. The anticyclonic anomaly south of Lake Baikal (see Figure 11(b)) appears to be responsible for the decreased summer rainfall there for the heavy-snow years. Additionally, the summer rainfall also decreases clearly in Assam and the northernmost part of India. These variations of the summer rainfall shown in Figure 12 are consistent with the results of correlation analyses using multi-year data (see Figures 3(a) and 4) discussed in Section 3.2.

5. SUMMARY AND DISCUSSION

In this work, we have examined the influence of ESSC on the Asian summer rainfall and large-scale temperature and circulation fields in the lower troposphere. Our main findings can be summarized as follows. (i) The relation between AIMR and ESSC is changing over a multi-decadal time scale. The negative correlation between them has markedly intensified since the middle 1970s. (ii) The region where the summer rainfall has the strongest and most stable negative correlation with the preceding ESSC is located in northern Mongolia, south of Lake Baikal. The correlation between the low-pass filtered summer rainfall and ESSC is stronger than that between the original data, showing that the effect of snow cover may be seen more clearly when the effect of ENSO is removed. (iii) Case studies of typical years with extensive and deficient snow cover show that the anomaly of ESSC occurs mainly in northwestern Eurasia and has significant simultaneous and delayed influences on the atmospheric conditions. Heavy ESSC will lead to cooling and a cyclonic circulation.
One interesting finding in this work is the clear interdecadal change of the relationship between ESSC and AIMR in the mid 1970s. Previous studies (Kumar et al., 1999; Krishnamurthy and Goswami, 2000; Miyakoda et al., 2000) reported that the connection of the Indian monsoon to ENSO has considerably weakened in recent decades. Miyakoda et al. (2000) further pointed out that the transition year was 1976. A number of possible mechanisms have been proposed to explain the interdecadal change of the monsoon–ENSO connection. For example, Kumar et al. (1999) thought that a shift in the Walker circulation and increased surface temperature over Eurasia in winter and spring had disrupted the monsoon–ENSO relation. Miyakoda et al. (2000) ascribed the recent weakening of the monsoon–ENSO connection to the climate shift associated with interdecadal variations of global sea temperatures. Kripalani and Kulkarni (1997a,b) found that the impact of ENSO events on AIMR is modulated by the decadal behaviour of AIMR and depends on the prevailing epoch. In addition, a shift of the jet stream over the North Atlantic (Chang et al., 2001) and the change in

anomaly in the lower troposphere over the northern part of Eurasia in spring, forming a Rossby-wave-train-like circulation response, eventually leading to a weak East Asia summer monsoon and deficient rainfall with an anticyclonic circulation anomaly south of Lake Baikal.

Figure 11. (a) The June–September 850 hPa streamlines averaged for the selected six typical years. (b) The difference of June–September 850 hPa wind (the average for 1979–81 minus that for 1989–91). The black areas indicate the Tibetan Plateau above the 850 hPa isobaric surface.
The interactions and feedbacks between the global sea surface temperatures, Eurasian snow cover, and Asian rainfall are worth investigating further in the future.

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REFERENCES


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