Study on Snow Depth Anomaly over Eurasia, Indian Rainfall and Circulations

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Abstract

Based on Historical Soviet Daily Snow Depth (HSDSD-II) data for the period 1941–1995, it is found that winter/spring snow depth anomalies over east Eurasia (70 E–160 E, 35 N–65 N) is positively correlated, while west Eurasia (25 E–70 E, 35 N–65 N) is negatively correlated with subsequent Indian summer monsoon rainfall (ISMR). The correlations become stronger and negative with winter snow depth anomaly over west Eurasia during the recent period of study (1975–1995). Four years of high snow (1966, 1968, 1979 and 1986), and two years of low snow (1961 and 1975), are selected on the basis of standardized winter snow depth anomaly over west Eurasia. The Pentad (Fivedays average) analysis of snow depth during the above contrasting snow years show that high/low snow is associated with late/early snow disappearance. NCEP/NCAR reanalyzed data are used for the study of some important atmospheric characteristics of seasonal monsoon circulation during high/low snow years, followed by deficient/excess ISMR.

The important findings of this study are: (1) the anomalous persistence of winter/spring snow delays the spring time continental heating, results in weak thermal low, and weak monsoon westerlies over India, (2) the difference fields of Velocity Potential shows complete phase reversal in the dipole structure from high to low snow years, and (3) the temperature and related circulation fields show that low level jet over east Asia is also considerably influenced by Eurasian snow anomaly.

1. Introduction

The large variation in the seasonal rainfall affects the agricultural production, drinking water, energy generation and the over all economy of many tropical countries. Therefore detail study on the understanding, and accurate prediction of monsoon variation, have been a topic of intense scientific research for more than a century. There are numerous studies on snow-monsoon relationship which clearly show that snow plays an important role in the climatic fluctuations. More than a century ago, Blanford (1884) found inverse relationship between Himalayan winter/spring snow accumulation and following Indian Summer Monsoon rainfall (ISMR). Later, inverse relationship was confirmed by Walker (1910) by using more number of years. The Asian summer monsoon, of which ISMR is a significant part, is a major phenomenon affecting the lives of large number of the population who live in the tropics. Hahn and Shukla (1976) revived the interest in snow-monsoon linkage by using satellite derived snow cover data, to show that correlation coefficient (CC) between winter Eurasian snow cover south of 52°N, and the following ISMR, is statistically significant and negative. Later this result was supported by Dey and Bhanukumar (1982), Dickson (1984), Vernekar et al. (1995), Dash et al. (2003), Kripalani et al. (2003a) and Dash et al. (2004).
Kripalani et al. (1996) using snow depth data from Nimbus-7 satellite, showed that two regions over Eurasia had negative CCs with ISMR; one region to the north-east of Moscow, and the other between Mongolia and Siberia. Further studies of Kripalani and Kulkarni (1999), based on Historical Soviet Daily Snow Depth (HSDSD-I) data for the period 1881–1984, inferred that wintertime snow depth over western Eurasia surrounding Moscow shows significant negative relationship with subsequent ISMR, whereas that over the east Eurasia in central Siberia has significant positive relationship with ISMR. They conjectured the existence of a midlatitude long wave pattern, with an anomalous ridge (trough) over Asia during winter prior to a strong (weak) ISMR. For better understanding of snow monsoon relationship, and for easy comparison with all the previous studies, Bamzai and Shukla (1999) correlated December, January, February and March (DJFM) mean snow cover anomalies for four regions with the subsequent ISMR. Their four regions of study were: (i) west Eurasia (40°N–60°N, 10°W–30°E), (ii) the whole of Eurasia (20°N–90°N, 0°–190°E), (iii) southern Eurasia (20°N–50°N, 0°–190°E) and (iv) the Himalayas (30°N–45°N, 60°E–105°E). They used satellite-derived snow cover data for 22 years (1973 to 1994). Bamzai and Shukla (1999) found that ISMR had the highest CC of 0.63 with the western Eurasia snow cover, whereas it was −0.34 with the snow over whole Eurasia. The winter and spring snow cover of southern Eurasia and the Himalayas show high interannual variability, but are poorly correlated with the subsequent ISMR.

Using several simulations of the European Centre for Medium range Weather Forecasts (ECMWF) model, Ferranti and Molteni (1999) have concluded that interannual variability of Eurasian snow depth in early spring is influenced by the boundary forcing arising from sea surface temperature (SST) anomalies over the tropical eastern Pacific, during the previous winter. Their results also indicate that Eurasian snow depth influences the seasonal mean monsoon, independently of El-Niño Southern Oscillation (ENSO). Recently Corti et al. (2000) performed 12 ensemble runs of ECMWF atmospheric General Circulation Model (GCM), and concluded that the snow-monsoon relationship is just an artifact of the influence of El-Niño anomalies, on both winter and summer circulations. However, Corti et al. (2000) did not rule out an active role of the snow anomaly in causing the persistence of the Eurasian wind anomaly. They further cited 1994 monsoon case (Soman and Slingo 1997), indicating that the snow related wind anomaly over Eurasia persists from winter to summer, independently from the Pacific SST anomaly. From this study the relationship between SST anomaly, and snow depth anomaly over Eurasia is not clear.

Almost all the observational studies of snow-monsoon relationships are based on the relationship between snow and monsoon rainfall only. Despite several studies in the past, the mechanism of the winter and spring time signals are not clearly understood at present. How the antecedent snow is linked to the large-scale monsoon circulation is not clear. To understand the physical mechanism responsible for the snow-monsoon connection, it is necessary to understand the related changes in monsoon circulation patterns, which may be linked to the mid-latitude circulation. Sankar-Rao et al. (1996) using National Oceanic and Atmospheric Administration (NOAA), National Environmental Satellite Data and Information Service (NESDIS) data for the period 1967 to 1992, concluded that following the winter of high snow, stationary perturbations with higher pressures over central Asia north of India are produced in the lower atmosphere, and the following Asian summer monsoon is weaker. Simultaneously, in the upper atmosphere, lower anomalous pressure occurs during summer, which weakens the upper level monsoon high. The anomalous upper tropospheric low pressure system cover a large area, extending from Asian middle latitude to India.

In the present study, based on Soviet snow depth (HSDSD-II), and ISMR of India Meteorological Department (IMD) during the period 1941–1995, two years have been considered, with low Eurasian snow followed by excess ISMR, and four years have been considered, with high snow followed by deficient ISMR. The difference fields of temperature, wind, stream function and velocity potential between the above two extreme cases have been examined in detail, with a view to study the relationship
between the Eurasian snow, midlatitude circulation and the Indian summer monsoon circulation features. Section 2 gives details of data used in this study. Section 3 describes the relationship between decadal variation of seasonal snow depth anomaly and ISMR. The identification of high and low snow years have been made in Section 4. Section 5 examines the seasonal differences in circulation characteristics in extreme years, and Section 6 describes the important conclusions of this study.

2. Data

NCEP/NCAR reanalysed data have been used for the period 1948–1998, which include temperature, wind, stream function and velocity potential fields at upper and lower atmospheric levels. A detailed analysis of the reanalysed data has been given by Kalnay et al. (1996). We have not used NCEP/NCAR reanalysed snow since this data set does not contain snow depth, but the water equivalent of snow. Kripalani et al. (1998) analysed the NCEP/NCAR reanalysed snow data set and did not find significant relationship with ISMR. ISMR data from June to September for the period 1881–1994, from Parthasarthy et al. (1995), and the ISMR for 1995 from the Journal Mau- sam, have been used in classifying the excess, normal and deficient rain years.

The HSDSD-II data updates and replaces the original HSDSD-I data that was previously available from National Snow and Ice Data Center (NSIDC) on CD, and provides a long-term climatological data for the periods 1881–1995. The HSDSD data were extracted from the Soviet Meteorological archive, which contains daily data from Russian\USSR World Meteorological Organisation (WMO) stations. HSDSD includes daily snow depth and daily state of snow cover (percentage of surrounding area that is covered by snow). Products derived at NSIDC, and included on the CD, are the daily, monthly, seasonal and climatological summaries. In this study daily data summaries are used to compute the monthly mean snow depth for all the 284 WMO defined stations. These stations are located in the midlatitude (Fig. 1), mostly inhabited, area of Eurasia. The geographical distribution of the stations lie between 35°N and 72°N latitude lines, and between 20°E and 180°E meridians, while the elevation of the stations varies from 15 meters to 2100 meters.

3. Decadal variation in snow depth anomaly and ISMR

Based on the spatial correlation coefficients (CCs) between winter/spring snow depth and ISMR, entire Eurasia is divided into two parts, namely the west (25°E–70°E, 35°N–65°N), and the east (70°E–160°E, 35°N–65°N) Eurasia where consistence data are available. All the stations north of 65°N are not used in the present study, because the number of stations are scare, and also the occurrence of missing data is high.

Figures 2(a), (b), (c) and (d) show spatial distribution of CCs between winter (DJF) and spring (MAM) snow depth anomalies and ISMR for the period 1948–1974 and 1975–1995, respectively. Two distinct periods for CCs analysis have been selected on the basis of Climate Change (IPCC; 2001). For the period 1948–1974, CCs between preceding winter snow depth anomaly and following ISMR, is weaker over the entire domain (Fig. 2a). While during the recent period of study i.e. 1975–1995, stronger CCs are seen over large area of west Eurasia (Fig. 2c). Figure 2(a) also shows that the winter (DJF) snow depth anomaly over west Eurasia lying between 25°E–70°E shows negative CCs, while those over east Eurasia lying between 70°E–160°E depicts positive CCs with subsequent ISMR. This is consistent with the results of Bamzai and Shukla (1999), who were surprised to find that the only significant and inverse CCs, between 1973–1994 winter snow depth anomaly and following ISMR, occurs over west Europe. The present study also confirms the relationship of two coherent regions based on the HSDSD-II. Snow cover is not a quite representative parameter to assess the amount of snow. The snow cover area in any two cases may be the same, but snow depths may be quite different, and the affect on monsoon would be quite different in the two cases. Snow depth is physically more meaningful than snow cover for better understanding of the inverse snow-monsoon relationship. Since the former is a better measure of the total precipitation that finally results in the soil memory.

Although the relationship between Eurasian snow depth and ISMR have been investigated
in many studies, but it has never been examined on multidecadal time scale. This study has used the data from 1941–1995 for such a study. Figure 3(a) presents the year to year variation between DJF snow depth anomaly over west Eurasia and subsequent ISMR. Figure 3(a) clearly shows inverse relationship between winter snow depth and following ISMR. Figure 3(b) presents 10 years sliding window of CCs, between 11 years moving average of DJF snow depth over the west and the east Eurasia, and subsequent ISMR. Figure 3(b) depicts that winter snow depth anomaly over west Eurasia is strongly associated, than over east Eurasia. It is also noticed that snow depth anomaly over east Eurasia is positively correlated, while west Eurasia is negatively correlated with subsequent ISMR. The interdecadal changes between Eurasian snow depth and ISMR has been also noted by Kripalani and Kulkarni (1999).

4. Selection of contrasting years

Using the DJF snow depth data for the year 1941–1995, mean values for each year over west Eurasia are computed, and mean of the series and standard deviations are also calculated. The snow depth of each year for the period 1941–1995 is expressed as a standardised snow depth anomaly, by dividing the departure of each year from normal, by standard deviation. The standardised snow depth anomaly, considered for west Eurasia, for the period 1941–1995. The years having snow depth anomaly between \( \pm 1 \) standard deviation, are considered as normal snow years. Similarly the
years having snow depth anomaly equal to or above $+1$ standard deviation are taken as high snow years, and those having equal to or less than $-1$ standard deviation snow depth anomaly are identified as low snow years. Based on this criterion, it is found that the years 1945, 1949, 1950, 1954, 1955, 1961, 1972, 1973 and 1975 are low snow years and 1942, 1957, 1966, 1968, 1979, 1986 and 1989 are high snow years. The rest of the years in the period 1941 to 1995 had normal snow depth anomalies. These years are consistent with the years shown in Kripalani et al. (2003b). Similarly, the ISMR rainfall anomaly for each year has been computed. The years having ISMR anomaly more than or equal to $+1$ standard deviation are termed as excess monsoon years, and those less than or equal to $-1$ standard deviation are considered as deficient monsoon years. The years having ISMR anomaly between $-1$ and $+1$ standard deviation are termed as normal monsoon years. Based on this criterion the years 1942, 1947, 1956, 1959, 1961, 1970, 1975, 1983 and 1988 are excess monsoon years and 1941, 1951, 1965, 1966, 1968, 1972, 1974, 1979, 1982, 1985, 1986 and 1987 are deficient monsoon years.

It may be noted that the major boundary forces affecting the Indian monsoon (Krishnan et al. 1999) are the Pacific, southern Indian ocean SST, and Eurasian snow. With a view to examine whether the above cases of high (low) Eurasian snow followed by deficient (excess) ISMR are free from the influence of SST. Com-

Fig. 2. Correlation coefficients (CCs) between Eurasian snow depth anomalies at locations mentioned in Fig. 1 with the following ISMR during the period 1948–1974 in (a) DJF, (b) MAM seasons and in (c) DJF and (d) MAM seasons during the period 1975–1995 respectively.
posite SST difference fields, between the years of high and low snow years are shown in Fig. 4. It is seen that between the years of high and low Eurasian snow, SST difference over the Pacific Ocean is less than 1°C, and over the south Indian Ocean SST difference is about 0.5°C. Normally ENSO years are identified with SST difference > 1°C. In order to choose the cases not influenced by SST anomalies in the Pacific and Indian Oceans, we have considered cases in (i) 1961 and 1975 where low winter snow is followed by excess ISMR, and (ii) 1966, 1968, 1979 and 1986 where high snow is followed by deficient ISMR.

5. Atmospheric circulation characteristics

The composite difference in the circulation characteristics in high snow (1966, 1968, 1979 and 1986) and low snow (1961 and 1975) years are studied in detail by analysing the composite difference in temperature, wind, stream function and velocity potential from NCEP/NCAR monthly mean reanalysis at 850 hPa and

![Graphs showing correlation and snow depth anomalies](image-url)
200 hPa. Bamzai and Shukla (1999) emphasised that the inverse snow-monsoon relationship holds especially in those years when snow is anomalously high or low for both the winter as well as the consecutive spring season. If there is heavy snow cover in winter, it is likely that it will affect the snow cover in spring. Heavy snow cover in midwinter usually does not easily melt, because of low level solar insolation. If there is already snow on the ground, the precipitation in February, March and April is likely to be in the form of snow (because of the surface temperature close to 0°C), rather than rain. Such a process could explain the results of Bamzai and Shukla (1999). Hence, we have computed the circulation characteristics and plotted the difference of mean fields in December, January and February (DJF), March, April and May (MAM) and June, July, August and September (JJAS) as the seasonal means, with a view to examine the evolution of the mean monsoon fields in advance, in response to the snow depth over west Eurasia.

The composite seasonal differences of the above four years of high snow, and two years of low snow, show that snow anomalies persist from DJF to MAM (not shown) seasons. However, the persistence of snow is more prominent from winter to spring especially over west Eurasia. Section 3 also shows that the inverse relationship of snow depth anomaly over west Eurasia with ISMR is strong from winter to spring season. A detailed analysis of how winter snow depth anomaly is maintained in spring season. We have shown the daily variations of the composites of snow depth, and temperatures in high and low snow cases. The important mechanism of snow accumulation, maintaining and melting are explained in great detail in Shinoda (2001), Shinoda et al. (2001) and Ueda et al. (2003). In the present analysis, all the data for each station over west Eurasia in the years of high and low snow, are averaged over the periods of five days (pentad) separately. The early (late) snow-disappearance years are generally characterised by high (low) temperature. Figure 5(a and b) presents the composite time series of snow depth (cm) and temperature (°C) in the years of high and low snow over west Eurasia. Major changes in snow depth in high and low snow cases are noticed from mid February to early spring (Fig. 5a). Figure 5(b) reveals that west Eurasia is cooler in high snow years compared to low snow years. Shinoda et al. (2001) have also mentioned more cooling in high snow (late snow disappearance) case as compared to low snow (early snow disappearance). Interestingly, the significant temperature difference (Fig. 5b) was noticed during almost the entire period from January to May (except in the end of May), between high and low snow cases. This indicates that the albedo effect of snow act for both cases
(high and low years), but more prominent in high snow years. This seems that the low(late snow disappearance years)/high(early snow disappearance years) temperature are associated with northerly cold/southerly warm air conditions, led to large/small snow (Shinoda et al. 2001).

Figure 6(a) shows that the entire Eurasia above 45°N is cooler in high snow winter compared to low snow winter. In high snow winter, the west region was cooled by 4°C at 850 hPa, whereas east Eurasia was cooled by 6°C. The cooling is visible up to the 500 hPa level, with 3°C in the west and 1°C in the east (Fig. 7a). In spring high snow, east Eurasia remains cooler by 2°C (Fig. 6b), and the temperature difference is consistent at 500 hPa. The cooling over the west, and east Eurasia in winter (Fig. 6a), undergoes a southward shift in spring and monsoon seasons as shown in Figs. 6(b) and (c). Similar shifting in temperature has been also noted in Sankar-Rao (1996) and Douville and Royer (1996). Southward shifts of temperature

Fig. 5. Seasonal changes in 5 days (a) snow depth (cm) and (b) temperature (°C) over the west Eurasia (♦ represents low snow years and ▲ represents high snow years).

Fig. 6. Composite temperature difference (°C) between high and low snow years at 850 hPa level (a) mean of December, January and February (DJF), (b) mean of March, April and May (MAM) and (c) mean of June, July, August and September (JJAS).
anomalies may be directly related to land-surface processes including snow. This impact of land-surface processes may be propagated at the upper troposphere. Figure 6(c) and Fig. 7(c) also reveal that the temperature anomalies persist till summer, even over East Asia, in particular over the Korean region. Figure 7(c) also suggests that snow distribution could affect monsoon variability over East Asia, in particular the Korean region. This is consistent with the results of Kripalani et al. (2002).

The composite seasonal wind difference fields at 850 hPa are shown in Figs. 8(a), (b) and (c) during winter, spring and summer monsoon seasons respectively. The circulation patterns in these figures are consistent with the corresponding seasonal snow depth temperature difference fields. Figure 8(a) indicates that the western part of west Eurasia is dominated by an anomalous cyclonic circulation. The westerly anomalies to the west of the Caspian Sea are much stronger during high snow years than low snow years. In spring (Fig. 8(b)), there is an anomalous anticyclonic circulation over west Eurasia, and anomalous cyclonic circulation over east Eurasia. Such anomalous circulations are well organised over midlatitude belt of 40°N to 70°N. Figure 8(c) shows an anomalous anticyclonic circulation over the north Arabian Sea and west India, centered at about 20°N and 75°E which contributes to the anomalous easterlies over the Arabian Sea during JJAS. This circulation feature can be traced back to similar wind anomalies, but less organised in spring as shown in Fig. 8(b). A large part of India comes under the influence of anomalous anticyclonic circulation during monsoon season at the lower level (Fig. 8c). Results weak monsoon circulation in high snow, compared to low snow years. Figure 8(c) also shows that strong north-easterlies west of Korea/east China, with suppressed southwesterly cross-equatorial flow. This indicates dry winds from north, and less moisture from the west Pacific towards East
Asia, in particular Korea. This is consistent with the results of Kriapalani et al. (2002). The composite seasonal wind difference fields at 200 hPa are shown in Figs. 9(a), (b) and (c) during winter, spring and summer monsoon seasons, respectively. Figure 9(a) shows anomalous cyclonic circulation at the upper level over east and west Eurasia during winter. Figure 9(c) shows the anomalous westerlies over the Arabian Sea indicating weaker upper level easterlies in high snow years, compared to low snow. These anomalies have their genesis in the spring as seen in Fig. 9(b). Figure 9(b) also shows well developed anticyclonic circulation over Southeast Asia in spring, which persists over central Asia in JJAS as seen in Fig. 9(c). These anomalous winds at upper level correspond to weaker easterly in the deficient monsoon year in high snow, compared to the excess rain year in low snow. Thus during the year of high snow, easterly started weakening in spring to give rise to weak monsoon, upper level easterly, during JJAS (Fig. 9c). The anomalous westerlies over the South China Sea in DJF (Fig. 9a) persist up to MAM (Fig. 9b). Such westerlies during high snow years arrest the northwestnorth movement of the Tibetan Anticyclone from its winter position over the equator, to its summer position over Tibet. Krishnan and Majumdar (1999) have shown the prominent southward incursion of mid-latitude westerlies over northwest India during May at 200 hPa. Such anomalous features suppress the development of upper tropospheric anticyclone, and hence the monsoonal circulation over India. Earlier Joseph (1978) and Joseph et al. (1981) had also noted similar anomalous features, which lead to weak monsoon over India.

Another important field considered in this study, to examine the influence of high snow on monsoon circulation, is the evolution of a negative stream function anomaly at 200 hPa. The formation of negative anomaly at upper level
can be seen in spring season (Fig. 10b). The gradual growth of negative anomaly during monsoon season occupied a large area north of the Indian subcontinent (Fig. 10c), and negative stream function anomaly at the upper level moved somewhat eastward; however the westerly wind anomaly remained over a large area of the summer monsoon. It was, in fact, formed by the occurrences of strong westerly winds aloft that entered northwestern India from the desert areas of Africa and Arabia to the west. These winds are very dry, and may act to inhibit deep convection. The weakening of the upper level divergent circulation on the planetary scale, and the establishment of a weak cyclonic circulation anomaly at the lower levels (not shown), and a weak easterly anomaly aloft, as found from the analysis of the stream function anomaly, seems to have set the stage for weak monsoon over India during high snow years. Figures 10(a) and (b) also show strong negative stream function anomalies over East Asia, in particular over the Korean region in high snow years.

The upper level seasonal tropospheric velocity potential difference fields in winter, spring and JJAS are shown in Figs. 11(a), (b), and (c) respectively. It is well known that there is a strong upper level divergent centre (Krishnamurti et al. 1972), associated with the Asian monsoon. Normally the upstream side of tropical easterly is associated with the largest divergent centre and the downstream side, west of India is associated with convergence (Chen and Van Loon 1987). The positive velocity potential difference field over the Indian subcontinent, and the negative difference field to the east in Fig. 11(c), indicate that the upper level divergence centre was weaker over India in high snow compared to low snow. This divergent circulation changed in such a way that the intensity of tropical easterly over the monsoon regions of southern Asia and Africa was weak (Fig. 11c) in high snow year. As shown by
Chen and van Loon (1987), the anomalous divergent circulation during weak tropical easterly year reduces the generation of kinetic energy on the upstream side of the jet, and the destruction of kinetic energy on the downstream side of the jet.

6. Conclusions

Based on the present study the following important conclusions can be made.

(1) Lag CCs indicate that winter snow depth anomaly over west Eurasia (25°E–70°E, 35°N–65°N) is negatively correlated, and east (70°E–160°E, 35°N–65°N) Eurasia is positively correlated with the subsequent ISMR.

(2) Results also show that composite low level atmospheric temperature difference field, between the years of high and low Eurasian snow in winter, can be as large as 6°C and the cooling persist up to the spring season, with a gradual southward shift during the year of more snow. The easterlies start weakening in spring to give rise to weak monsoon upper level easterlies during JJAS. Low level monsoon westerlies over the Arabian Sea also become weaker in high snow year, compared with low snow years.

(3) Composite wind fields at lower level shows that low snow over west Eurasia, and high snow over East Eurasia, may intensify the North Pacific subtropical high and the low-level Jet, whereas heavy snow over west Eurasia and light snow over eastern Eurasia may weaken these features.

(4) The formation of large negative stream function anomaly at upper level north of the Indian subcontinent, and associated wind dryness of the upper level westerly winds during high snow years, induce weak monsoon over India. Difference fields of velocity potential in high and low snow years in-

Fig. 10. Same as in Fig. 6 except for the stream function difference ($10^7$ m$^2$/s) at 200 hPa.
Indicate that the anomalous convergence center in winter over India gradually becomes weaker as spring comes, and gives way to anomalous divergence which persists in JJAS. The anomalous convergence/divergence center shifts from the Northern Hemisphere in winter, to the Southern Hemisphere over Australia in spring, and is almost maintained there in JJAS.

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References

Fig. 11. Same as in Fig. 6 except for the velocity potential difference ($10^6$ m$^2$/s) at 200 hPa.


