CLIMATOLOGY

The weakening relationship between Eurasian spring snow cover and Indian summer monsoon rainfall

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Substantial progress has been made in understanding how Eurasian snow cover variabilities affect the Indian summer monsoon, but the snow-monsoon relationship in a warming atmosphere remains controversial. Using long-term observational snow and rainfall data (1967–2015), we identified that the widely recognized inverse relationship of central Eurasian spring snow cover with the Indian summer monsoon rainfall has disappeared since 1990. The apparent loss of this negative correlation is mainly due to the central Eurasian spring snow cover no longer regulating the summer mid-tropospheric temperature over the Iranian Plateau and surroundings, and hence the land-ocean thermal contrast after 1990. A reduced lagged snow-hydrological effect, resulting from a warming-induced decline in spring snow cover, constitutes the possible mechanism for the breakdown of the snow-air temperature connection after 1990. Our results suggest that, in a changing climate, Eurasian spring snow cover may not be a faithful predictor of the Indian summer monsoon rainfall.

INTRODUCTION

Seasonal snow cover plays a notable role in the Earth's climate system because it directly influences the proportion of solar radiation absorbed (*1*, *2*) and creates soil moisture anomalies that may persist for periods of a few weeks to months (*3*–*5*). It has been recognized as an important contributor to seasonal climate prediction in several regions, particularly in the Indian subcontinent, where the extent of spring snow cover was first proposed as a useful tool for forecasting the following Indian summer monsoon rainfall more than a century ago (*6*, *7*).

The Indian summer monsoon rainfall during June to September contributes more than 80% of the total annual rainfall over India, and its interannual variation of around 10% can have substantial impacts on agriculture and the economy in general (*8*–*10*). Skillful prediction of this variation months in advance would allow improved agricultural planning, drought mitigation, and flood warning (*11*, *12*). Numerous observational (*13*–*18*) and modeling studies (*19*–*24*) have shown that positive Eurasian snow cover anomalies during winter and spring tend to be followed by an anomalous deficit of rainfall over the Indian subcontinent in the subsequent summer monsoon season, while negative snow cover anomalies tend to be followed by abundant rainfall. However, there are studies noting that this negative snow-monsoon teleconnection only holds true when the snow cover in particular regions, such as western Eurasia, is considered (*16*, *25*, *26*) or when the effect of the El Niño Southern Oscillation (ENSO) is excluded (*25*, *27*, *28*). After years of extensive research, exactly which part of the Eurasian region is primarily responsible for the inverse correlation between snow cover and the following Indian summer monsoon rainfall and how robust the relationship is in a changing climate remain unclear. This uncertainty is probably partly due to the use of different period lengths and different snow proxies in different studies (*14*, *16*, *17*).

The extent of spring Eurasian snow cover has been shrinking in recent years, as a consequence of rapid climate warming at mid to high latitudes (*29*). It would seem reasonable to suppose that this large-scale spring snow retreat should lead to higher land temperature and an increased land-sea thermal contrast that would be expected to strengthen the Indian summer monsoon. However, the Indian summer monsoon rainfall has experienced a clear downward trend in the last several decades (*30*), suggesting that our comprehension of the snow-monsoon relationship in a changing climate is incomplete. Previous studies have documented that the snow-monsoon relationship shows secular changes (*28*, *31*, *32*), but the mechanisms through which the altered snow cover affects the snow-monsoon relationship have not been identified. Here, we aim to unravel the causes of an evolving relationship between the Eurasian spring snow cover and the subsequent Indian summer monsoon rainfall at the interannual time scale by synthesizing long-term satellite snow cover data, instrumental rainfall records, and reanalysis datasets.

RESULTS

Changes in the snow-monsoon relationship

We found that changes in spring snow cover fraction (SCF; March to May) over central Eurasia (40 to 80°E, 35 to 65°N; the red box in Fig. 1A), including eastern Europe, western Siberia, central Asia, and the western Himalaya, had the strongest negative impact on the subsequent all-Indian summer monsoon rainfall (AISMR; June to September) for the entire study period. This result is consistent with previous findings, although the time periods considered are different (*6*, *16*, *28*, *33*). Another region with obvious negative correlation is northeast Eurasia (100° to 140°E, 65° to 73°N; the black box in Fig. 1A). However, the spring SCF over this region exhibits smaller interannual variability in comparison with that in central Eurasia (fig. S1), meaning that its importance in the seasonal forecast of the AISMR changes should be limited.

Next, we focus on the central Eurasian region only and examine the stability of the interannual relationship between central Eurasian spring SCF and the subsequent AISMR for the period 1967–2015. For this analysis, the area-weighted average spring SCF over central Eurasia (hereafter CESCF) was calculated as an index of spring snow cover

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variabilities. Both CESCF and AISMR have a prominent interannual variability (Fig. 1D). Correlation analysis between CESCF and AISMR, using a 15-year sliding window, indicated that the relationship between the two variables underwent a dramatic shift starting around 1990 (see inset in Fig. 1D). Consequently, we split the study period into two subperiods (1967–1990 and 1991–2015) for further inspection. We found that $R_{\text{CESCF-AISMR}}$ (both time series are detrended; see Materials and Methods) is significantly negative during the period 1967–1990 ($R = -0.61$, $P < 0.01$) but becomes nonsignificant and positive for the period 1991–2015 ($R = 0.18$, $P = 0.39$) (Fig. 1D). This phenomenon was also reflected in the correlation patterns, where spring SCF over central Eurasia displays a coherent strong negative correlation with AISMR before 1990 but shows no significant correlation thereafter (Fig. 1, B and C).

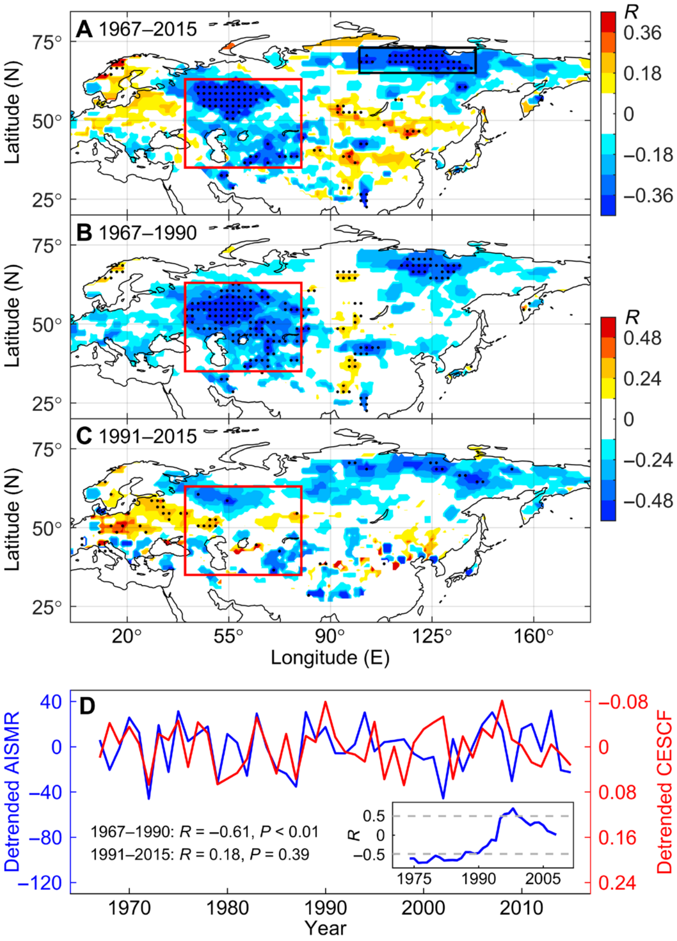

Fig. 1. Shift in the snow-monsoon relationship. (**A**) Spatial distribution of correlation coefficients (*R*) of spring SCF over Eurasia with AISMR for the period 1967–2015. (**B** and **C**) Same as (A) but for the periods 1967–1990 and 1991–2015, respectively. (**D**) Time series of detrended AISMR and CESCF. AISMR is all-Indian summer monsoon rainfall (June to September) from the Indian Institute of Tropical Meteorology (IITM), and CESCF denotes the area-weighted average spring (March to May) SCF over central Eurasia. Note that the order of the right-side axis is reversed to enable easier comparison of the time series. The line chart embedded in (D) denotes a 15-year sliding correlation between AISMR and CESCF. Significant *R* values are identified as gray dashed lines ($P < 0.05$). The black dots in (A) to (C) indicate that the *R* is statistically significant ($P < 0.05$). The red boxes represent the region of central Eurasia (40° to 80°E, 35° to 65°N), and the black box denotes the northeast Eurasia (100° to 140°E, 65° to 73°N).

Robustness test of the shift in the snow-monsoon relationship

To check the robustness of the observed shift in $R_{\text{CESCF-AISMR}}$ between the periods 1967–1990 and 1991–2015, we performed the following tests. First, we randomly selected 20 years for each period (1967–1990 and 1991–2015) and found that $R_{\text{CESCF-AISMR}}$ shifted from -0.61 ± 0.07 (with 98.6% of values being statistically significant and negative during 1967–1990) to 0.18 ± 0.11 (with 99.4% of values being statistically nonsignificant during 1991–2015) (Fig. 2A). The results suggest that the observed shift was not due to certain singular values.

Second, we tested whether the observed shift in the snow-monsoon relationship is due to the weakened ENSO-monsoon association. This is an important test, given that previous studies have indicated that ENSO could affect both the Indian summer monsoon and Eurasian snow cover through atmospheric teleconnections (*34*–*36*), and the fact that the ENSO-AISMR relationship has also collapsed since the 1990s (*37*). To make the test, we calculated a partial correlation between CESCF and AISMR after statistically controlling for the ENSO effect for the two periods. $R_{\text{CESCF-AISMR}}$ calculated by this method also shows a shift from −0.55 ($P < 0.01$) during the earlier period to 0.21 ($P = 0.34$) during the later period (Fig. 2B). Moreover, the correlation

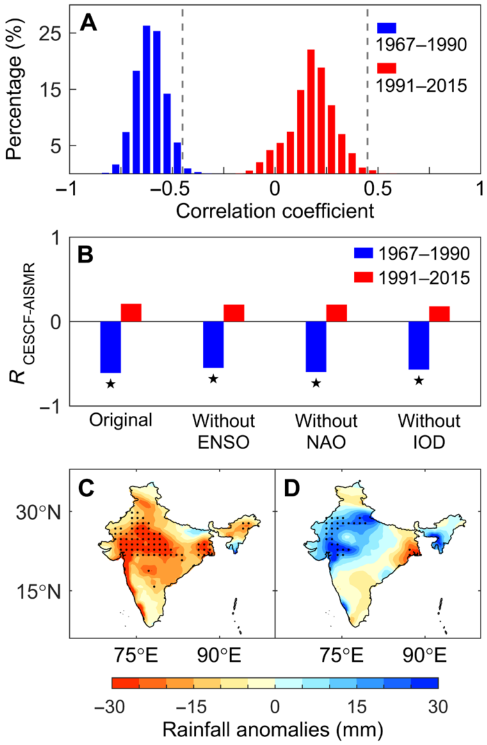

Fig. 2. Robustness tests of the shift in the snow-monsoon relationship. (**A**) Frequency distributions of the correlation coefficients of CESCF with AISMR for 1967–1990 and 1991–2015. We calculated the correlation coefficients by randomly selecting 20 years in each corresponding period. Significant correlation coefficients are indicated by the gray dashed lines ($P < 0.05$). (**B**) Partial correlation coefficients of CESCF and AISMR ($R_{\text{CESCF-AISMR}}$) for 1967–1990 and 1991–2015, calculated by statistically controlling for the effect of ENSO, NAO, and IOD in turn. The original $R_{\text{CESCF-AISMR}}$ denotes the correlation analysis without excluding any other effects. The asterisks indicate that the correlations are statistically significant ($P < 0.05$). (**C** and **D**) Linear regression of CRU-derived Indian summer monsoon rainfall (June to September) with respect to CESCF for 1967–1990 and 1991–2015, respectively. The black dots represent significant rainfall anomalies at the 95% confidence level based on a Student's *t* test.

patterns of spring SCF with AISMR after removing the ENSO signal during the two periods (fig. S2, A and B) are consistent with those that include the ENSO effect (Fig. 1, B and C). Similar results are also obtained when the same analyses are carried out to exclude the effects of NAO (North Atlantic Oscillation) and IOD (Indian Ocean Dipole) (Fig. 2B and fig. S2, C to F). These analyses illustrate that the shift in the snow-monsoon relationship is not an artifact but a real phenomenon.

Third, the change in the snow-monsoon relationship was also found when the Climate Research Unit (CRU) precipitation dataset was adopted to represent the Indian summer monsoon rainfall intensity (Fig. 2, C and D). To obtain further insight into the shift in the snow-monsoon relationship, we performed linear regression analyses to determine how the spatial pattern of Indian summer monsoon rainfall anomalies covaries with CESCF for the two periods. These analyses show that positive anomalies of spring snow cover over central Eurasia precede a significant decline in summer rainfall especially over northwestern and central India (also known as the core Indian monsoon region) during the period 1967–1990 (Fig. 2C), while it precedes an increase in summer rainfall after 1990 (Fig. 2D). Although the shift in $R_{\text{CESCF-AISMR}}$ was originally found when using the time series of area-averaged precipitation over India, the shift in summer monsoon rainfall anomalies in response to CESCF between the two study periods is also observed throughout most of the Indian subcontinent, suggesting that the shift could be related to large-scale changes in monsoon circulation.

Performance of CMIP5 models

Do current climate models capture such a shift in the snow-monsoon connection? To answer this question, we examine the simulation outputs of 30 models participating in the fifth phase of the Coupled Model Intercomparison Project (CMIP5) (table S1). We show that the key region with the strongest impact of spring snow cover on the Indian summer monsoon rainfall varies among models (fig. S3). In terms of the multimodel mean, the strongest impact lies in central Russia rather than central Eurasia (fig. S3). These differences across models and the differences between the model outputs and observations could arise from a combination of several factors and should be interpreted with caution. The CMIP5 climate models have systematic biases in their simulation of snow states (*38*, *39*). Such biases could propagate through the models and lead to errors in the models' atmospheric responses to the underlying snow patterns (*4*), thus contributing to the model-data mismatch. Although the key region with the strongest impact of snow on the monsoon was not captured, around 8 of 30 models and the multimodel mean could reproduce the shift in the snow-monsoon relationship (fig. S4), if their own region where spring SCF strongly affected the Indian summer monsoon was used.

DISCUSSION

The decoupled snow-air temperature connection induces the shift in snow-monsoon relationship

It is widely believed that the Indian summer monsoon originates from the meridional thermal contrast between the Asian continent and the Indian Ocean (*40*, *41*). Changes in spring snow cover could induce anomalous warming or cooling of the atmospheric temperature over the Asian continent through the snow-albedo and snow-hydrological effects (*4*, *5*, *19*, *24*), which would affect the land-sea temperature gradient and thereby the Indian summer monsoon intensity (*23*, *24*, *42*). Snow-induced atmospheric temperature anomalies can thus act as a "bridge" linking snow to monsoon. Moreover, both Eurasian snow cover and air temperature underwent an interdecadal shift around 1990 (*43*). We hypothesize that the weakened inverse snow-monsoon relationship may result from the loss of the coupling between spring snow cover and summer atmospheric thermal conditions after 1990.

To test this hypothesis, we examined the relationships of summer (June to September) air temperature with CESCF and AISMR for both study periods. Since the mid-upper level tropospheric temperature plays a more important role in driving the Indian summer monsoon (*44*, *45*), we used the air temperature at 500 hPa (AT500) to indicate the atmospheric thermal conditions. The regression analyses showed that AT500 over southwestern and northeastern Eurasia has a significant negative association with CESCF during the period 1967–1990 (Fig. 3A), but almost no significant connection occurred after 1990 (Fig. 3B). In contrast, AISMR was positively linked with AT500 over large parts of the Eurasian continent during both periods (Fig. 3, C and D). We further calculated the correlation coefficients of CESCF and AISMR with the area-averaged AT500 over the Iranian Plateau and surroundings (IPS; 55° to 80°E, 25° to 40°N), where the air temperature exerts the strongest influence on the Indian summer monsoon (fig. S5). Our results again showed that CESCF is significantly negatively correlated with AT500 over the IPS during the earlier period ($R = -0.51$, $P < 0.01$), but the correlation becomes insignificant in the later period ($R = 0.14$, $P = 0.50$) (Fig. 3E). The observed shift in the snow-air temperature relationship coincides with that in the snow-monsoon relationship (Fig. 3F). However, the AISMR is significantly correlated with AT500 during both earlier ($R = -0.70$, $P < 0.01$) and later periods ($R = -0.84$, $P < 0.01$) (Fig. 3G). This result was also robust to the use of temperature data from the Japanese 55-year Reanalysis (JRA-55) (fig. S6). The analyses presented above support our hypothesis that the failure of the impact of CESCF on AISMR could be attributed to a decoupling between snow cover and atmospheric thermal conditions over the Asian landmass after 1990.

Potential mechanisms for the decoupled snow-air temperature connection

Changes in Eurasian spring snow cover and the resultant alterations of land surface conditions could exert strong impacts on the hemispheric atmospheric circulation and hence the climate during the following seasons (*5*, *22*, *32*). Therefore, we explored whether the observed shift in the snow-air temperature relationship is explained by changes in the snow-related summer atmospheric circulation. We performed regression analyses of summer wind field at 500 hPa with respect to the CESCF. Our analyses illustrated that there is a prominent anomalous cyclonic circulation centered over the IPS following the excessive anomaly of CESCF during the earlier period (Fig. 3A). Such an anomalous cyclonic circulation, on the one hand, acts to transport cold air from midlatitudes into the IPS via the anomalous northeasterly winds. On the other hand, the cyclonic flow could probably further lower the mid-tropospheric air temperature by favoring upward air motion that leads to an increase in cloud cover and a reduction of downward solar radiation (*46*, *47*). The cooling in the mid-upper troposphere over the IPS would contribute to a decrease of the meridional thermal gradient and thereby a depressed Indian summer monsoon. In the later period, however,

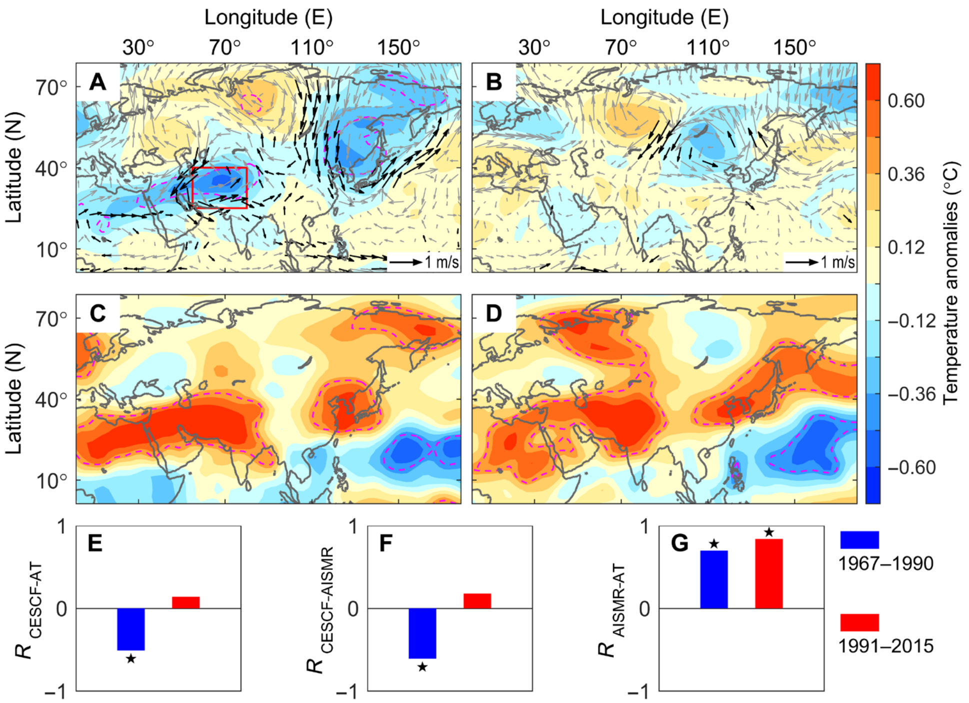

Fig. 3. The relationships between air temperature, atmospheric circulation, AISMR, and CESCF for the periods 1967–1990 and 1991–2015. (**A** and **B**) Linear regression of summer (June to September) air temperature (AT) (°C) and wind fields (m/s) at 500 hPa with respect to CESCF for 1967–1990 and 1991–2015, respectively. (**C** and **D**) Linear regression of air temperature with respect to AISMR for 1967–1990 and 1991–2015, respectively. The pink dashed lines represent significant air temperature anomalies at the 95% confidence level, and black arrows denote that the wind anomalies are statistically significant ($P < 0.05$) based on a Student's *t* test. (**E** to **G**) Correlation coefficients of CESCF with air temperature averaged over the IPS, that of CESCF with AISMR, and that of AISMR with air temperature over the IPS, respectively. The IPS is denoted by the red box in (A). The asterisks indicate that the correlation coefficients are statistically significant ($P < 0.05$). All the temperature and wind data are derived from the NCEP-NCAR reanalysis.

there are almost no significant anomalies of circulations in response to the variability of CESCF (Fig. 3B). Particularly, the cyclonic circulation over the IPS formed in the earlier period has vanished but was replaced by an anomalous weak southerly wind, which is conducive to the advection of warm air from lower latitudes to the IPS and may thus destroy the inverse snow-air temperature correlation (Fig. 3B). These results were found to be similar when using the JRA-55 dataset (fig. S6, A and B).

The potential mechanism responsible for the shift in snow-related circulations and the resultant snow-air temperature decoupling could be related to the change in the lagged snow-hydrological effect. Numerous observational and modeling studies indicated that the snow-hydrological effect has as a crucial role in determining the land-atmosphere coupling through its effects on the soil moisture (*4*, *5*, *48*). The soil moisture anomalies formed from spring snowmelt may take weeks or months to dissipate, known as the "memory," providing a lagged impact on the summer climate both locally and remotely through changing the surface energy budget (*5*, *23*). However, as for global warming, changes in snow mass and snow phenology (*29*) could alter the soil moisture memory, which may disrupt the delayed effect of spring snow on summer climate.

As illustrated in Fig. 4, in the period 1991–2015, mean spring SCF over central Eurasia decreased by 16% (Fig. 4A), and the snowmelt date advanced by at least 2 weeks (Fig. 4B) in comparison to those in the period 1967–1990. Such a decline in spring snow cover and an advance in snowmelt date could lead to less infiltration of meltwater into the soils, resulting in the soil moisture anomalies becoming more susceptible to rainfall and evaporation during summer and hence losing the memory. To test this mechanism, we calculated the memory of spring (March to May) soil moisture anomalies over central Eurasia as the number of months until the correlation coefficient of soil moisture between spring and the following months dropped below the insignificant level ($P > 0.05$). Our analysis showed that the memory length reduced from 4 months (June to September) in the earlier period to 2 months (June and July) in the later period (Fig. 4C). As a result of the decline in soil moisture memory since 1990, the anomaly of CESCF can only efficiently affect the atmospheric circulations and air temperature over Asian landmass during June and July rather than during the whole monsoon season (June to September) (fig. S7). Therefore, the reduced snow-hydrological effect can be regarded as a plausible mechanism for the observed decoupled relationship between snow and air temperature in the later period.

In addition, we note that the atmospheric circulation pattern associated with the anomaly of CESCF during 1991–2015 (Fig. 3B) closely resembled the pattern triggered by the variability of the west-east dipole mode of Eurasian spring snow cover anomalies (fig. S8C) (*48*, *49*). This dipole structure is found to dominate the variations of Eurasian spring snow cover anomalies since 1990 (fig. S8, A and B) (*32*, *49*). It therefore tentatively suggests that the emergence of the west-east dipole pattern may lead to changes in the CESCF-related atmospheric circulations, which, in turn, contribute to the decoupling between CESCF and air temperature over the IPS in the later period. However, uncertainties remain in understanding the

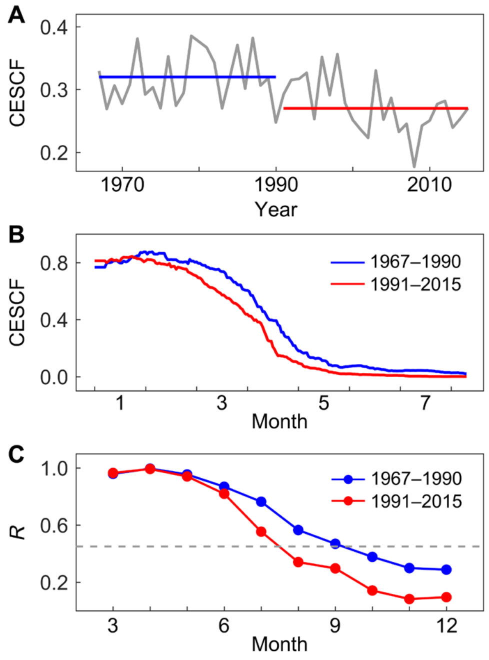

Fig. 4. Changes in spring snow cover and soil moisture memory over central Eurasia for 1967–1990 and 1991–2015. (**A**) Time series of the CESCF. The blue and red lines represent mean CESCF for 1967–1990 and 1991–2015, respectively. (**B**) Seasonal evolution of CESCF during two periods. (**C**) Correlation coefficients (*R*) of monthly mean soil moisture with spring (March, April, and May) soil moisture averaged over central Eurasia for the periods 1967–1990 and 1991–2015. The gray dashed line denotes the *R* significant at 95% confidence level.

linkage between snow and circulation anomalies based on statistical analysis. We could not rule out the possible interference of internal atmospheric dynamics and other boundary conditions, such as sea surface temperature over the North Atlantic, with the influence of snow (*32*). Further model simulation works are needed to quantify the actual role of the west-east dipole pattern of snow cover in shaping the circulation.

In summary, we have used the long-term satellite-based snow cover data and instrumental rainfall records (1967–2015) to show that the classical inverse relationship between Eurasian spring snow cover and subsequent Indian summer monsoon rainfall has disappeared after 1990. This shift in the snow-monsoon relationship is found to be related to the reduced lagged snow-hydrological effect, which could be further attributed to the reduction of spring snow cover and an advance in snowmelt date due to the climate warming. However, our understanding of the mechanisms behind the loss of the snow-monsoon correlation is still limited. Further studies using the state-of-the-art earth system models, which include a physical representation of the interactions between snow and atmospheric dynamics, are required to develop our comprehension of these mechanisms. In addition, similar behavior to the observed weakening of the snow-monsoon correlation is found in the relationship between the Indian summer monsoon and ENSO, suggesting a changing paradigm for Indian summer monsoon precipitation in a warming atmosphere. The CMIP5 models fail to capture the connection between snow cover and the monsoon and its shift at the decadal time scale, emphasizing the necessity to improve the models' ability to represent snow status and atmospheric responses to snow anomalies in a warming world.

MATERIALS AND METHODS

Observational and reanalysis datasets

We used AISMR data from the Indian Institute of Tropical Meteorology (IITM) to represent the variability of the Indian summer monsoon rainfall (June to September). This dataset was derived from the area-weighted average of 306 rain-gauge stations that are almost uniformly distributed throughout India (*50*). To confirm the reliability of the IITM data, we also used rainfall datasets from the Climatic Research Unit Time Series (0.5° × 0.5°;CRU TS v4.00) (*51*) and the Global Precipitation Climatology Centre (0.5° × 0.5°; GPCC v7) (*52*).

The snow cover data were from the Northern Hemisphere EASE-Grid 2.0 weekly snow cover and sea ice extent version 4, available at the National Snow and Ice Data Center (NSIDC). The data were stored as binary files in arrays of 720 columns by 720 rows and had a spatial resolution of 25 km. To make the comparison with other data easier, we converted the EASE-Grid 2.0 data to a 1° × 1° longitude-latitude grid. The snow cover data were in binary form (snow or no snow), and so we used the SCF, rather than snow cover extent, to represent the variability of snow cover. The monthly SCF was calculated by summing the number of weeks in that month for which snow was present for a pixel and by expressing this number as a fraction of the total number of weeks in that month (*16*). In this study, we used only the spring (March, April, and May) snow cover to investigate the snow-monsoon relationship. There are two reasons for this restriction. First, the snow cover has little interannual variability during the winter. Second, the snow-albedo effect was found to be high in the spring because of the rapid retreat of snow cover. In addition, the snow-hydrological effect, controlled by anomalies in snow water equivalent, emerges in spring, meaning that the degree to which the atmosphere responds to changes in snow should be strong at this time of year (*4*, *5*).

The data of 500 hPa air temperature and wind fields with a resolution of 2.5° × 2.5° for the summer months (June to September) were obtained from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (*53*). The air temperature and wind data from the JRA-55 were also used (*54*). The soil moisture data were provided by the National Oceanic and Atmospheric Administration (NOAA) Climate Prediction Center (CPC) and had a resolution of 0.5° × 0.5°. In addition, we used the 3-month mean (June, July, and August) Oceanic Niño Index in the Niño 3.4 region to represent the ENSO, and the average monthly NAO index for January, February, and March to represent the cold season NAO index. Both the ENSO index and the NAO index are also available at the NOAA CPC. We used summer (June to September) Dipole Mode Index (DMI) from the NOAA Earth System Research Laboratory to represent the IOD. The IOD was considered since it exerts an important impact on both the Indian summer monsoon rainfall (*55*) and Eurasian snow cover (*56*).

The snow cover and precipitation data from 30 earth system models participating in the CMIP5 were also used in our study (table S1). Because of the uncertainties in the climate forcing data for CMIP5 after 2005 (*57*), we only used the historical simulations data (1967–2005). The CMIP5 models had various spatial resolutions, and so, for convenience, we converted all the CMIP5 model output to a 1° × 1° longitude-latitude grid.

Statistical methods

To diagnose the impact of spring snow cover on the Indian summer monsoon rainfall, we performed a correlation analysis between AISMR

(June to September) and satellite-based Eurasian spring (March to May) SCF for each grid square. We also computed the partial correlation between spring SCF and AISMR while controlling for the effects of ENSO, NAO, and IOD. In addition, we performed a linear regression analysis to reveal the response of the air temperature and atmospheric circulation to the snow anomalies. All variables were detrended before the statistical analyses. The significance of all the analysis was determined by the standard two-tailed Student's *t* test method.

SUPPLEMENTARY MATERIALS

Supplementary material for this article is available at http://advances.sciencemag.org/cgi/content/full/5/3/eaau8932/DC1

Fig. S1. The 11-year sliding SD of spring SCF averaged over central (40° to 80°E, 35° to 65°N; the red box in Fig. 1A) and northeast (100° to 140°E, 65° to 73°N; the black box in Fig. 1A) Eurasia for the period 1967–2015.

Fig. S2. Spatial distribution of the partial correlation coefficients of Eurasian spring SCF with AISMR during 1967–1990 and 1991–2015.

Fig. S3. Spatial distribution of the correlation coefficients of Eurasian spring SCF with Indian summer (June to September) monsoon rainfall for the period of 1967–2005 in the CMIP5 models.

Fig. S4. Scatter plot of the correlations between Indian summer monsoon rainfall and area-averaged spring snow cover for 1967–1990 versus that for 1991–2005.

Fig. S5. Spatial distribution of the correlation coefficients of 500 hPa summer (June to September) air temperature over Eurasia with AISMR for 1967–2015.

Fig. S6. Same as Fig. 3 but using air temperature and wind fields from the JRA-55 reanalysis.

Fig. S7. Anomalies of air temperature and circulations related to the variabilities of central Eurasian spring snow cover.

Fig. S8. Shift in the dominant pattern of Eurasian spring snow cover anomalies and its associated atmospheric circulation anomalies after 1990.

Table S1. Summary of CMIP5 models used in this study.

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