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Atmospheric circulation processes contributing to a multidecadal variation in reconstructed and modeled Indian monsoon precipitation

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Abstract An analysis of the recently reconstructed gridded May–September total precipitation in the Indian monsoon region for the past half millennium discloses significant variations at multidecadal timescales. Meanwhile, paleo-climate modeling outputs from the National Center for Atmospheric Research Community Climate System Model 4.0 show similar multidecadal variations in the monsoon precipitation. One of those variations at the frequency of 40–50 years per cycle is examined in this study. Major results show that this variation is a product of the processes in that the meridional gradient of the atmospheric enthalpy is strengthened by radiation loss in the high-latitude and polar region. Driven by this gradient and associated baroclinicity in the atmosphere, more heat/energy is generated in the tropical and subtropical (monsoon) region and transported poleward. This transport relaxes the meridional enthalpy gradient and, subsequently, the need for heat production in the monsoon region. The multidecadal timescale of these processes results from atmospheric circulation-radiation interactions and the inefficiency in generation of kinetic energy from the potential energy in the atmosphere to drive the eddies that transport heat poleward. This inefficiency creates a time delay between the meridional gradient of the enthalpy and the poleward transport. The monsoon precipitation variation lags that in the meridional gradient of enthalpy but leads that of the poleward heat transport. This phase relationship, and underlining chasing process by the transport of heat to the need for it driven by the meridional enthalpy gradient, sustains this multidecadal variation. This mechanism suggests that atmospheric circulation processes can contribute to multidecadal timescale variations. Interactions of these processes with other forcing, such as sea surface temperature or solar irradiance anomalies, can result in resonant or suppressed variations in the Indian monsoon precipitation.

1. Introduction

The Indian summer monsoon is the world’s largest and most powerful monsoon system. While the monsoon develops in southern Asia and its adjoining Indian Ocean and the Arabian Sea, its influence on atmospheric circulation and global water and energy cycles is profound [Webster, 2006]. During the summer monsoon season from early June through most of September, large amount of moisture is transported to the Indian subcontinent by the southwesterly monsoon flow off the east coasts of Somali and contributes to about 80% of the annual precipitation in India [Ashfaq et al., 2009; Berkelhammer et al., 2010]. A 10% deficit or surplus in the monsoon precipitation would result in severe drought or flood in India [Sikka, 2003; Gadgil et al., 2005]. These statistical facts show the dominant roles of the monsoon circulation and precipitation for the weather and climate in India and also indicate the vulnerability of the densely populated monsoon region to variations in the Indian summer monsoon [e.g., Sinha et al., 2011].

In a global scale, the Indian summer monsoon is considered as a part of the warm season northward migration of the Intertropical Convergence Zone (ITCZ) in the eastern Northern Hemisphere [e.g., Chao and Chen, 2001; Gadgil, 2003; Fleitmann et al., 2007] and is a key circulation component that transports heat/energy from the tropical and subtropical regions to the high-latitude and polar region [e.g., Trenberth and Stepaniak, 2003, 2004; Hazeleger et al., 2005]. As a part of the rising branch of the mean Hadley cell, the rising motion and associated precipitation in the Indian summer monsoon increase the total potential energy in the tropical and subtropical atmosphere. The increased energy is transported northward after a portion of it is converted to the kinetic energy that maintains the stationary eddies which sustain the northward energy transport [Hazeleger et al., 2005] and also the transient eddies associated with the Ferrel cell in the midlatitude. This role of
the Indian summer monsoon in energy generation and poleward energy transport reiterates the importance in understanding the monsoon variation for prediction of the climate in the midlatitude and high-latitude regions in the Northern Hemisphere, as well as for the social wellbeing of the population in the monsoon region.

Great efforts have been devoted to gaining such understanding in the past decades. While progress has been made in modeling and understanding seasonal and subseasonal variations in the Indian monsoon [e.g., Shukla et al., 2009], it remains debatable on how the observed decadal-multidecadal variations in the monsoon circulation and precipitation may have developed.

Among the studies that examined multidecadal variations in Indian monsoon, some have investigated and suggested the multidecadal variation in the solar irradiance as a possible cause [e.g., Mehta and Lau, 1997; Neff et al., 2001; Agnihotri et al., 2002; Burns et al., 2002; Fleitmann et al., 2003; Kodera, 2004; Bhattacharyya and Narasimha, 2005; Berkelhammer et al., 2010]. Many of these studies show that the solar irradiance and monsoon intensity variations have been comparable in phase at decadal-centennial timescales. Positive anomaly in solar irradiance corresponds to heavy monsoon rainfall. However, as also indicated in Neff et al. [2001], Burns et al. [2002], and Kodera [2004], a direct effect of changing solar irradiance on the monsoon is unlikely, and the irradiance anomaly has to go through the lower boundary of the atmosphere, e.g., via the surface, including sea surface temperature anomalies, to influence the monsoon.

The effects of persistent sea surface temperature (SST) anomalies in the North Atlantic Ocean, e.g., the Atlantic Multidecadal Oscillation, on (downstream) eastern Northern Hemisphere summer circulation and the Indian monsoon were examined in Dugan et al. [1997], Wang et al. [2005], Goswami et al. [2006], and Feng and Hu [2008], among others. Wang et al. [2005] suggest that the multidecadal fluctuations in the Indian monsoon occurred during the Holocene from 8400 to 8100 years and again from 4500 to 4000 years before present were likely triggered by extra freshwater recharge from the Greenland ice melt to the North Atlantic Ocean and subsequent changes in global oceanic circulation. Feng and Hu [2008] show that the North Atlantic SST and the all-India monsoon rainfall varied in phase at the frequency of 19 years per cycle during A.D. 1880–2000. Meanwhile, the multidecadal variations in the North Pacific SST, such as the Pacific Decadal Oscillation (PDO), also have been examined for their effects on the decadal timescale variations in Indian monsoon rainfall [e.g., Sen Roy, 2011; Krishnamurthy and Krishnamurthy, 2014a]. Krishnamurthy and Krishnamurthy [2014a] show that in the past century PDO modulated both the Walker and the Hadley circulations in the subtropics of East Asia and, via them, indirectly affected the decadal-scale variations in Indian monsoon circulation and precipitation.

Another indirect cause alluded to the North Atlantic SST variation is the Eurasian continent snow cover anomalies [e.g., Hahn and Shukla, 1976; Barnett et al., 1989; Meehl, 1994; Anderson et al., 2002]. Hahn and Shukla [1976] examined the snow cover-monsoon intensity relationship, using satellite data, and found a negative relationship. Both the modeling studies of Barnett et al. [1989] and Meehl [1994] show that more extended Eurasian snow cover in boreal spring leads to weaker Indian monsoon in the following summer.

While these studies have disclosed factors/forces influencing the Indian summer monsoon at multidecadal timescales, they face a common issue; i.e., the relationships between the forcing and the response in Indian monsoon precipitation vary in time. For instance, the multidecadal solar irradiance-monsoon relationship appeared in about 10,000 years before present [Fleitmann et al., 2003] and reemerged in the Medieval warm period from A.D. 800 to 1290 and also in the last century [Berkelhammer et al., 2010]. Similar on-and-off/unsteady relationships also are shown between the North Atlantic SST/the PDO/the Eurasian snow cover and the Indian monsoon variation. While these varying relationships are showing that those forcing factors may have influenced multidecadal variation in the Indian monsoon in certain periods and under specific conditions in the past millennia, they indicate that those external forcings are not the only causes for the multidecadal variation in the Indian summer monsoon. Consequently, they raise the questions if and how the internal physical processes in the atmospheric circulation could have supported and sustained multidecadal variations in Indian monsoon in some periods and under certain conditions. These are the questions that we try to address in this study.

As an effort to address these questions, we examine multidecadal variations in the Indian summer monsoon precipitation over the past half millennium from A.D. 1470 to 1999, using a recently reconstructed gridded data set of May–September total precipitation [Feng et al., 2013]. Meanwhile, we evaluate the data set from simulations of paleo-climate circulation and precipitation by a paleo-general circulation model (GCM).
From comparisons between precipitation variations in these data sets over the monsoon region, we evaluate model-simulated multidecadal variation that also appears in the reconstructed precipitation. The simulated data from the paleo-GCM describing the variation in monsoon precipitation are analyzed to gain understanding of the physical processes attributing to the simulated multidecadal monsoon variation.

It is known that even if a GCM may be able to simulate some observed variation in precipitation, it remains questionable if the model processes describing the variation are the same with or a close resemblance to that suggested in the observed/reconstructed data or similar to that occurring in the real world. It is also known that deficiencies in circulation models have been reported in their simulations of the Indian summer monsoon, and they remain under investigation [e.g., Kumar et al., 2005]. On the other hand, however, in the absence of any observational data, the model-simulated data showing variations similar to the observed (even not so, given the reason previously described, still) allow us to investigate physical processes in the development of the variation. Such effort in understanding the variation may be repeated when observational data and/or more reliable models become available to produce better information. This iterative process will help advance our understanding of causal processes for circulation variations at multidecadal timescales.

2. Data and Methodologies

2.1. Reconstructed Precipitation Data

We use the reconstructed gridded data set of May–September total precipitation for the Asian continent (5°–55°N, 60°–135°E, Figure 1b) from A.D. 1470 to 1999 [Feng et al., 2013]. The reconstructed data have a spatial resolution of 0.5°×0.5° in latitude and longitude. Multiproxy records, including tree ring, ice core, documentary, and a few early instrumental observations dated back to the nineteenth century in the reconstruction region were used in the precipitation reconstruction. Details of those proxy records, the method, the reconstruction procedure, and the reliability of the data are described in Feng et al. [2013].

It is necessary to explain that the May–September total precipitation is not exactly the “Indian monsoon precipitation,” because the monsoon starts around June 12 each year and ends in late September (according to the modern records). We use this precipitation amount in this study because it is the reconstructed amount for the past half millennium. Moreover, the error from using May–September precipitation to represent the monsoon precipitation is small because, as stated earlier, outside the monsoon period the precipitation takes barely 20% of the total annual precipitation in India. Adding the rainfall of May out of those 8 months of scarce precipitation to May–September precipitation may only cause a small deviation of the latter from the monsoon precipitation.

This reconstructed data set is used in our analysis in the following procedure. First, the May–September precipitation at grids in the land area of the region from 5°–27°N and 67°–90°E, which includes most of India (the Area C in Figure 1), is averaged to obtain the Indian monsoon precipitation for each year from A.D. 1470 to 1999. (Note that this region excludes the northern tip of India to avoid the strong orographic effect on precipitation that could deviate the precipitation variation from its monsoonal nature [Meehl et al., 2012].) Second, after removing any linear trend in the reconstructed monsoon precipitation [Feng et al., 2013], we apply a power spectrum analysis to identify significant components in the monsoon variation from A.D. 1470 to 1999. From this spectrum, the multidecadal variations are identified. Third, a low-pass filter is
applied to the precipitation data, and the filtered data are examined for the multidecadal variations in the monsoon precipitation.

While these reconstructed data show the variations in the precipitation at multidecadal timescales, they only show a collective outcome of the atmospheric processes and their interactions that produced them. Those processes cannot be shown by the data, however. A process-oriented method, such as numerical models, must be used to complement the data result, help identify those physical processes and their interactions, and aid us to comprehend how they may have developed and led to the observed variations in the precipitation at the multidecadal timescale.

2.2. Data From Paleo-GCM Simulations

Paleo-GCM simulations of the climate in the past by research groups using the National Center for Atmospheric Research (NCAR) Community Climate System Model 4.0 (CCSM4) provide the opportunity for us to investigate simulated multidecadal variations that are similar to that in the reconstructed precipitation data. Again, it would be ideal if the model can simulate the observed variations. Meanwhile, as we stated in section 1, even if the model-simulated variations are off in certain aspects from the observed, the modeled data still provide the opportunity for us to conduct process-oriented studies and identify the physical processes and possible mechanism for the modeled variations.

The CCSM4 is a general circulation model and consists of atmosphere, land, ocean, and sea ice components, which are linked through a coupler that exchanges state information and fluxes between those components [Gent et al., 2011]. This model has been used in studies of interannual to multidecadal variability of the Asian monsoon and shown better performance than its previous versions [Meehl and Arblaster, 2011; Landrum et al., 2013; Islam et al., 2013]. Meehl and Arblaster [2011] show that the correlations between the Indian monsoon rainfall and the tropical Pacific SST are comparable to the observed. Although this model still has weaknesses in simulating monsoon, it has been greatly improved in capturing most of the Asian monsoon characteristics [Meehl et al., 2012].

Recently, the CCSM4 was used to complete the Last Millennium Simulation and its extensions [Landrum et al., 2013]. The goal of these simulations is to evaluate the model’s ability in capturing observed variability and to provide long-term model data for analysis to improve our understanding of the paleo-climate and climate variation. The Last Millennium Simulation is from A.D. 850 to 1850, and its two extended simulations are from A.D. 1850 to 2005. The outputs of these simulations have monthly time resolution and are at the spatial resolution of 0.9°×1.25° in latitude and longitude on 26 hybrid vertical levels.

In the Last Millennium Simulation, the boundary conditions and the forcing are defined per Paleoclimate Modelling Intercomparison Project Phase III protocol [Schmidt et al., 2012]. In the two extended simulations of the Last Millennium Simulation, the boundary conditions and forcing are the same as that in the Coupled Model Intercomparison Project Phase 5 CCSM twentieth century simulation [Gent et al., 2011]. For modeling details, the readers are referred to the cited references. Between the two extended simulations, the first is used in our study. The connection from the Last Millennium Simulation to that extended simulation was made consistent by the CCSM4 modeling group, and we used the outcome.

In processing the model simulation data, we first interpolate them from the 26 hybrid levels to 17 pressure levels in the vertical direction. Afterwards, we obtain the averaged values of May–September temperature, winds, and geopotential height on the 17 pressure levels, the SST, and the sea level pressure (SLP) at 0.9°×1.25° latitude and longitude grids covering the Eurasian continent. We also calculate average values of the SST and SLP for December–February in each model year. Precipitation is summed to obtain the total amount of May–September.

It is important to note the difference between the modeled and reconstructed Indian monsoon precipitation. As discussed in Meehl et al. [2012], CCSM4 has a bias of overestimating the monsoon precipitation. While considerable improvements have been made to correct it over the years, it remains an issue under investigation. Meanwhile, the reconstructed data suffer a bias of underestimating the precipitation because of saturation of biological processes to precipitation information [Feng et al., 2013]. While the difference between the simulated and reconstructed precipitation is from both the mean values and fluctuations of these data series, the fluctuations around their means are similar [Feng et al., 2013]. In other words, while the simulated and reconstructed precipitations may have slightly different mean values, their variations are comparable. To avoid the effect of the differences in the mean values between the reconstructed and
model-simulated precipitation on our analysis results, we focus on the fluctuations of the reconstructed and model-simulated precipitation.

2.3. Closeness of the Multidecadal Variations in the Two Precipitation Data Series

The CCSM4 simulated precipitation data series for the Indian monsoon region is processed with the same procedure defined in section 2.1 and used in processing the reconstructed monsoon precipitation data. After obtaining the low-pass filtered precipitation time series from the modeled and reconstructed monsoon precipitation, we use the Monte Carlo test to examine their closeness/resemblance. The Monte Carlo method is used because the low-pass filtering reduces the effective length of data series and their degrees of freedom substantially enough that the Student’s t test becomes invalid to apply to test the correlation and significance of the filtered data series. The Monte Carlo method [Hope, 1968; Neumann et al., 1977; New and Hulme, 2000; Viladomat et al., 2014] has the capability to test the closeness/correlation of two time series that have strong autocorrelations.

The procedure of this test in our study follows Kroese et al. [2011]. Our null hypothesis is that the two filtered precipitation series are not significantly correlated at the 1% confidence level. Assuming that each filtered time series has a sample size of n, we (1) calculate the Pearson’s correlation coefficient (r) between the two filtered time series, (2) combine the two time series into one, (3) randomly sample the combined time series from step 2, both of which contain n data elements, and then calculate Pearson’s correlation coefficient of the two resampled time series, (4) repeat step 3 for 10,000 times to generate the reference distribution of Pearson’s correlation coefficients, and (5) calculate the probability of r in the reference distribution and check if r is significantly different from 0. If the probability is less than 1% or in other words r is significantly different from 0 at the 1% confidence level, the null hypothesis is deemed false.

2.4. Analysis of Physical Processes Attributing to the Multidecadal Monsoon Variation

After using the Monte Carlo method and examining the similarity between the model-simulated and the reconstructed multidecadal precipitation variations, we analyze the model-simulated variation from the perspectives of the general circulation and energy transport and balance in the atmosphere. Atmospheric enthalpy, its meridional gradient, and eddy transport of heat and associated circulation are examined to gain understanding of the processes in the multidecadal monsoon precipitation variation. The multidecadal variations in simulated monsoon precipitation are further compared to the model-simulated variations in the SST at selected sensitive oceanic regions. In addition, the detected relationships are compared to the relationships derived from comparisons between the reconstructed Indian monsoon precipitation variation and variations of the reconstructed SST at the same oceanic regions.
3. Results and Discussions


The variation of the reconstructed Indian monsoon precipitation from A.D. 1470 to 1999 is shown in Figure 2a. A power spectrum analysis of the variation reveals significant components of variation at interannual to centennial frequencies (Figure 2b). At the multidecadal frequency, there are two major components at the frequency of 22–28 and 39–52 years per cycle. These statistically significant components in the monsoon variation are also suggested in several prior studies [Agnihotri et al., 2002; Yadava and Ramesh, 2007; Krishnamurthy and Krishnamurthy, 2014b]. These components are visible in the variation in Figure 2a, showing their roles weaving the observed variation in the reconstructed precipitation.

A similar set of multidecadal variations is also identified in the simulated Indian monsoon precipitation over the same period, shown in Figure 2c. Comparisons of Figures 2c with 2b suggest that the model-simulated variation has nearly the same variation component at the frequency of 40–50 years per cycle. Both spectra also show outstanding variations at the frequency of 15–17 and 22–28 years per cycle, albeit their strength or significance level differs slightly. These similarities suggest that the model has simulated the multidecadal monsoon precipitation variations comparable to the reconstructed precipitation variations in A.D. 1470–1999. In this study, we will focus on the variation at the frequency range of 39–52 (or 40–50) years per cycle.

To further examine the variation at this frequency range, we apply a 35 year low-pass filter to both the simulated and reconstructed precipitation (after the linear trend of the precipitation is removed). The two filtered monsoon precipitation time series are shown in Figure 3. Note that the filtering has truncated 17 years on both ends of the original data series from A.D. 1470 to 1999, reducing the data length to 496 years (from A.D. 1487 to 1982). The Monte Carlo test applied to the two filtered precipitation time series shows that they are (statistically) significantly correlated. This correlation is also reflected in Figure 3 by the fairly consistent in-phase relationship between the two variations in most of the years from A.D. 1487 to 1982, albeit there are years when the simulated and reconstructed variations are apart.

It is known that the reconstructed precipitation cannot capture the details of the actual precipitation variation because the proxy records contain effects from other processes, such as temperature variation. Meanwhile, the model simulation of the past climate has difficulties to capture those details as well because of lack of specific past information and/or (forcing) conditions for the model. With these limitations in mind it is reasonable to assume that the dominant climatic processes causing the multidecadal variation in the reconstructed data and the model simulation are similar to a degree that warrants an examination of the model output variables to gain understanding of the major circulation processes and possible mechanisms of the multidecadal variation in the monsoon precipitation. Or, in the perspective of the opposite extreme, the understanding or the mechanisms for the multidecadal variation gained from this analysis would only apply to the model. Even in such an extreme, the understanding remains useful for our knowledge of multidecadal variation in the monsoon precipitation.

To gain such understanding, we focus on the period when the multidecadal variations have the highest correlation between the simulated and the reconstructed precipitation. This period is from A.D. 1640 to 1762, marked by the two vertical dashed lines in Figure 3. The Monte Carlo test result of the reconstructed and modeled precipitation in that period is shown in Figure 4a, confirming their strong similarity. In this period, there are three cycles of the 39–52 year variation, and they dominate the total variation in the monsoon precipitation (Figures 3 versus 2a). The reconstructed and modeled precipitation variations are closely matched, although there exist slight differences in the phase of the second cycle and in amplitude among all three cycles.
The similarity of these two variations is further suggested by their highly correlated spatial patterns both in the Indian monsoon region and across most of the Asian continent (Figures 4b versus 4c). A pattern correlation of Figures 4b and 4c yields a correlation coefficient of 0.79.

3.2. A Mechanism for the 40–50 Year Variation in the Monsoon Precipitation

In the following, we examine these three cycles of the 39–52 (or 40–50) year variation in A.D. 1640–1762 using model simulation data and identify the physical and circulation processes that may have contributed to this variation. Before we discuss the results, it is necessary to justify why only these three cycles are examined and clarify what may be the representativeness and limitation of the results for this variation in the Indian monsoon precipitation.

Indeed, it is desirable to study as many as possible cycles of this particular variation. Yet, the perplex nature of the climate system and interactions of variations at wide ranges of frequencies as well as varying forcing limit such a possibility. Owing to the impacts of forcings and processes of different nature and timescales in the Earth system, the Earth’s climate varies in all frequencies. If all those forces/processes have equal weight in their influence on the climate (e.g., precipitation), it would be at a steady state. The fact that the observed precipitation shows prominent frequencies in its variations (e.g., Figure 3) indicates that the forces and processes at some frequencies affect the precipitation stronger than the forces/processes at other frequencies at certain time periods. This fact provides an opportunity for us to identify the time period when those forces/processes are strongly affecting the climate/precipitation and to study how they work to cause the observed variation in climate. We need to examine mechanisms dominating the variations of the climate or precipitation in different conditions and time periods. Furthermore, we need to examine how those variations would change/be modified, or be replaced, by other variations at different frequencies and operated by other mechanisms. Such changes or transitions would be reflected in changes of dominant variation component(s) (at different frequencies) in the climate variation. These mechanisms are fundamental blocks in our understanding of the climate variation. Only after they are identified and understood can we predict the climate variation.
In this study, those three outstanding cycles in the monsoon precipitation variation in the years from A.D. 1640 to 1762 suggest a dominance of the forces/processes driving the variation at that particular frequency. We therefore have the opportunity to focus on those decades and examine the driving mechanism. We do not extend our study beyond those decades because after, as well as before, those decades the mechanism was weak and this variation was not as outstanding. In those decades, other/different forces and processes would be playing stronger roles in driving the precipitation variation, and different variation pattern would replace this particular variation in the precipitation.

A caveat in this process is that the limited number of cycles of this variation poses a limitation on the robustness of the results from this study. Thus, repeated testing may be required to validate, to prove or disprove, the results from studying the limited number of cycles. With this understanding, the results from this study would serve as a preliminary step toward understanding this variation, and they need to be reiterated and verified when more and better data become available.

With this understanding, we now examine the mechanism of the 40–50 year variation in the Indian monsoon precipitation. The development of the Indian summer monsoon is associated with and, to a large extend, is a part of the warm season migration of the ITCZ in the eastern Northern Hemisphere [Chao and Chen, 2001; Gadgil, 2003; Fleitmann et al., 2007]. The ITCZ is the process that generates the atmospheric potential energy in the tropical region and, meanwhile, creates meridional transport of the energy, via the averaged Hadley cell, from the tropics to the midlatitude and high-latitude region in both the Northern and the Southern Hemispheres [e.g., Hartmann, 1994]. In boreal summer, the ITCZ resides along 15°N in latitude on average in the eastern Northern Hemisphere. It can advance northward to 30°N in some longitudes. Although the Hadley cell with its rising branch in this ITCZ transports large amount of heat and energy southward to the Southern Hemisphere, a smaller direct circulation on the north side of the ITCZ (a much weaker Hadley cell in the summer hemisphere; see Palmén and Newton [1969]) still transports both angular momentum and energy to the midlatitude and high latitude in the Northern Hemisphere. In the midlatitude, this northward energy transport, while smaller in magnitude in summer than in winter, is essential for balancing the net radiation loss, which remains strong in the high-latitude regions even in the summer hemisphere. Thus, it is reasonable to expect stronger monsoon and heavier monsoon precipitation when there is a stronger need for such a poleward transport of energy/heat and weaker monsoon when the need weakens.

To elaborate on this hypothesis, we examine the meridional gradient of the vertically integrated atmospheric enthalpy, \( \frac{\partial}{\partial \phi} \left( \int_{1000}^{1000} \rho C_p T g^{-1} dp \right) \), for the focus period from A.D. 1640 to 1762 (in this formula, \( T \) is the air temperature and a function of space and time, \( \rho \) the gravitational constant, \( C_p \) the heat capacity of air under constant pressure, \( p \) the atmospheric pressure, \( a \) the radius of the Earth, and \( \phi \) the latitude). To be consistent in our analysis, we apply a 35 year low-pass filter to the model simulation data before calculating the

![Figure 5](image-url)
atmospheric enthalpy (the same procedure is used for all model data analyzed and presented in this study unless stated otherwise). The variation in the anomaly of the meridional enthalpy gradient averaged in the region of 5°–55°N and 60°–100°E, relative to the average gradient of A.D. 1640–1762 in the same region, is shown by the solid line in Figure 5a. Compared to the Indian monsoon precipitation variation, replotted by the dashed line in Figure 5a, the meridional enthalpy gradient reaches its peak about 100° ahead in precipitation cycle 1, nearly in phase in cycle 2, and 15° ahead in cycle 3. Similarly, the minimum in the enthalpy gradient anomaly occurs, on average over the three cycles, 79° before the minimum of the precipitation anomaly. This general lead in phase of the variation in the meridional enthalpy gradient to that of the monsoon precipitation variation suggests that the enlarged meridional heat gradient may have demanded and subsequently driven an increase in monsoon precipitation. Strengthened meridional gradient of vertically integrated atmospheric enthalpy demands more heat to be transported from the low-latitude to the high-latitude regions in order to keep a stable meridional temperature gradient and baroclinic state in the atmosphere.

In order to meet this demand, two things must happen: one is the availability of energy in the tropical and subtropical atmosphere, and the other is effective poleward transport of that energy by atmospheric

Figure 6. (a) Wind (arrows, m s\(^{-1}\)) and geopotential height (color scale, m) at 850 hPa averaged over the period A.D. 1640–1762. Anomalies in 850 hPa (b) wind (arrows, m s\(^{-1}\)) and (c) geopotential height (color scale, m) averaged for the peak wet and dry years in the three precipitation cycles. (d–f) The same as Figures 6a–6c but at the 300 hPa level.
circulation. The strengthened Indian monsoon, reflected in increasing monsoon precipitation, is a means in the subtropical region to increase the available/potential energy in the atmosphere. Thus, the phase lag of the variation in monsoon precipitation to the meridional enthalpy gradient could be an indication of the response of the tropical-subtropical region to the demand for heat/energy in the high-latitude region.

The eddy meridional transport anomaly of the atmospheric enthalpy is shown by the solid line in Figure 5b, calculated according to Peixoto and Oort [1992]. Comparisons of the variation in Figure 5b to that of the meridional enthalpy gradient in Figure 5a show that, on average over the three cycles, the northward eddy transport of enthalpy increases and peaks about 80° in phase after the peak in the meridional enthalpy gradient anomaly. Further comparisons of the variations between the eddy transport and meridional gradient of the enthalpy show weakening in the gradient following the increase in meridional transport of the enthalpy. The poleward transport weakens and arrives at its minimum shortly after the enthalpy gradient relaxes to its minimum.

These phase relationships of the anomalies in the meridional enthalpy gradient, heat transport, and the monsoon precipitation suggest some atmospheric circulation processes behind this multidecadal variation in the monsoon precipitation. In essence, the variation in the precipitation shows, and is a part of, the transient process of poleward heat and energy transport in the atmosphere in the longitudinal section from 60° to 100°E. When net radiation cooling causes large loss of heat in the high-latitude and polar region, the meridional gradient of the enthalpy in the atmosphere strengthens. In response to the need for heat in the north, the atmospheric circulation adjusts to increase the potential energy in the atmosphere in low-latitude regions, by enhancing atmospheric convection which also results in increasing precipitation in the tropical and subtropical (monsoon) regions. The excessive potential energy generated in this process is transported to the high-latitude region by the northward eddy transport.

The increased transport of enthalpy compensates the loss of atmospheric heat in the high-latitude and polar region and weakens the meridional enthalpy gradient. Subsequently, the need for poleward energy transport softens. The reduced need triggers a relaxation of the energy production in the subtropical and tropical regions and a slowdown of the poleward eddy transport of energy. So the monsoon weakens and its precipitation decreases. While this relaxation is taking place, the weakening meridional heat transport again allows radiative cooling effect to enhance its effect on temperature in the high-latitude region. This shift of heating/cooling balance rejuvenates the process to increase the meridional gradient of enthalpy in the atmosphere and establishes the condition for the next cycle of strong monsoon and precipitation in low-latitude and poleward transport of heat.

The associated atmospheric circulation anomalies composited for the peak wet and peak dry anomaly years during the three cycles are shown in Figure 6. In the peak wet years (Figure 6b), the lower troposphere (850 hPa) is characterized with negative geopotential height anomalies between 15° and 45°N in latitude sandwiched by positive anomalies in the north and the south. In the upper troposphere (300 hPa) (Figure 6e), strong positive anomalies are shown between 25° and 45°N. On both the north and the south of that latitude band are weak positive anomalies. An exception of this strong baroclinic profile is observed over Mongolia and central Russia, where strong positive geopotential anomalies are found in both the lower and upper tropospheres. A reversed geopotential anomaly pattern is found in the peak dry years of the precipitation cycles (Figures 6c and 6f); strong positive (negative) anomalies are shown in the lower (upper) troposphere between 25° and 45°N, and weak negative anomalies spread outside that latitude band in both levels.

These composite anomalies in geopotential highlight some important circulation features in achieving the strong poleward transport of heat in the wet phase of the monsoon variation. First of all, according to the relationship between the variations in the monsoon precipitation and the poleward eddy transport of enthalpy shown in Figure 5b, the transport peaks about 50°, on average of the three cycles, after the peak positive monsoon precipitation anomaly. (We use the Fourier series to approximate the three cycles in the variation of precipitation and heat transport. We then average for the cycles to determine the phase difference between these fields.) This phase relationship indicates that the largest increase in meridional eddy transport of energy occurs shortly after the peak monsoon precipitation. Thus, the geopotential anomaly patterns in Figures 6b and 6e describe the circulation/geopotential anomalies near the most active phase in heat transport in the variation.
nearly unchanged temperature in the tropical and subtropical regions south of the 20°N, these warmer temperatures in the high latitudes would weaken the meridional gradient of atmospheric temperature and the temperature in the regions north of the 45°N latitude also is warmer than the mean. In contrast to the where, as previously discussed, there are active air mass exchange and strong vertical motion. Moreover, the 500 hPa temperature anomaly shows positive anomalies in the midlatitude region from 20° to 45°N as the difference from the mean temperature of A.D. 1640 air temperature composited for the peak wet years of the three precipitation cycles. (This change is de

The outcome of a period of strong poleward transport of heat is the increase in atmospheric temperature in the monsoon precipitation variation. These results indicate more rising air and intense convection activity and, hence, more conversion of the eddy potential to eddy kinetic energy in the wet phase than in the dry phase during the monsoon precipitation variation. The positive geopotential or pressure anomalies in the midlatitude and low-latitude regions during the peak dry years weaken the previously discussed processes which convert potential energy of the mean flow into the eddy potential and eddy kinetic energy to support poleward heat transport. The inactive exchange further allows the radiation cooling effect to dominate the temperature variation in the high latitudes.

Another measure of atmospheric stability for convection is the ratio of the flow going over versus around the Tibetan Plateau [Valdes and Hoskins, 1991], 

\[ r = \frac{L}{h} \]  

where \( L \) is the aspect ratio of the Tibetan Plateau. A larger \( r \) indicates a large fraction of the impinging westerly flow onto the west slope of the plateau to flow over the plateau, and a small \( r \) indicates a large fraction of the impinging westerly flow to flow around the plateau. Our results show a large \( r \) during the wet half of the precipitation cycles and a smaller \( r \) during the dry half of the cycles. These results indicate more rising air and intense convection activity and, hence, more conversion of the eddy potential to eddy kinetic energy in the wet phase than in the dry phase during the monsoon precipitation variation.

The outcome of a period of strong poleward transport of heat is the increase in atmospheric temperature in the high-latitude region. This increase is confirmed and shown in Figure 8b by changes in the 500 hPa air temperature composited for the peak wet years of the three precipitation cycles. (This change is defined as the difference from the mean temperature of A.D. 1640–1762, shown in Figure 8a.) The distribution of the 500 hPa temperature anomaly shows positive anomalies in the midlatitude region from 20° to 45°N where, as previously discussed, there are active air mass exchange and strong vertical motion. Moreover, the temperature in the regions north of the 45°N latitude also is warmer than the mean. In contrast to the nearly unchanged temperature in the tropical and subtropical regions south of the 20°N, these warmer temperatures in the high latitudes would weaken the meridional gradient of atmospheric temperature and

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**Figure 7.** Vertical gradient of potential temperature from 700 to 300 hPa, averaged in the region 20°–45°N and 60°–100°E (units: K km⁻¹).
enthalpy. Subsequently, the baroclinicity of the atmosphere would weaken. These changes would disfavor further development of disturbances in the midlatitude region. The reduced activity of disturbances that serve as the mechanism for mixing of air masses from the south and the north and, in doing so, transporting energy poleward would weaken the northward heat transport in the atmosphere. These changes may also be interpreted as consequences of the weakening needs for poleward heat transport when the temperatures in the high-latitude region become warmer.

Opposite processes occur in the peak dry years as shown in Figure 8c. After a period of weak exchange and transport of heat between the low-latitude and high-latitude regions, the temperature in the midlatitude and high-latitude regions decreases. This decrease is also fueled by net radiation loss. As the temperature further decreases in this situation, the meridional gradient of the atmospheric enthalpy strengthens, and the atmospheric circulation is being prepared to enter another period of intense monsoon and elevated poleward heat transport.

We now extend this explanation/mechanism for the 40–50 year variation in the monsoon precipitation based on the baroclinic theory and examine the meridional mass circulation in the eastern Northern Hemisphere during the variation. Figure 9 shows the averaged meridional mass circulation in 60°–100°E for May–September over the years from A.D. 1640 to 1762 and the anomalies in the meridional circulation composited for the peak wet and peak dry years in the three monsoon cycles. Figure 9a shows a fairly typical boreal summer meridional mass circulation in the monsoon region, with an averaged Hadley cell having its rising branch around 10°–15°N and sinking branch at 20°S in latitude. In the midlatitude Northern Hemisphere is the Ferrel cell with its northern boundary at about 45°N. This mean structure of the meridional circulation is distorted quite dramatically during either phase (peak wet or dry years) of this multidecadal variation, as shown in Figures 9b and 9c.

Figure 8. (a) Temperature at 500 hPa averaged for the three cycles of the precipitation variation (units: K). The temperature anomalies (units: K) at 500 hPa averaged over the peak (b) wet and (c) dry years of the three cycles.
In the peak wet years (Figure 9b), the anomalous meridional circulation in the monsoon region shows a substantial northward extension of the direct circulation with a strong rising branch extending to about 37°N (and sinking branch in high latitude in the Southern Hemisphere). Northward of the 37°N is an anomalous meridional circulation with rising motion between 37° and 45°N and strong sinking motion around 60°N. This anomalous meridional circulation pattern indicates enhanced rising motion, or weakened sinking motion, in the latitudes from 30° to 50°N. This region also has anomalous low pressure in the lower troposphere as shown in Figure 6b. Because a region with lower pressure in the lower troposphere and enhanced rising or weakened sinking motion has weaker convective stability, more disturbances and weather systems would develop there. Such disturbances would achieve the eddy transport of heat from the low-latitude to the high-latitude regions [Oort, 1971].

This transport is shown by the strong anomalies of subsidence motion from 60°N to the polar region (Figure 9b). The anomalously strong subsidence raises the atmospheric temperature through adiabatic heating, which is the most efficient way of raising atmospheric temperatures in the polar region. This warming in the polar region is indicated by the temperature change in the same phase shown in Figure 8b. Because this adiabatic heating is maintained by the midlatitude disturbances, those enhanced disturbances in this phase of the variation achieve transporting heat from the tropical and subtropical monsoon regions to the high-latitude and polar region.

Figure 9. (a) May–September mean meridional mass circulation in 60°–90°E averaged for A.D. 1640–1762 (units: 10¹⁰ kg s⁻¹). The anomalies of the May–September meridional mass circulation from the mean in Figure 9a averaged for the peak (b) wet and (c) dry years in the three cycles of the precipitation variation.
A direct consequence of the rising temperature in the high-latitude and polar region is, again, the weakening of the meridional gradient in air temperature and the atmospheric enthalpy. This effect has been shown in Figures 5a and 8b after the wet phase of the variation. As we have also discussed, this change starts the dry phase of the variation with weakening poleward energy transport and convective instability in the midlatitude atmosphere (Figure 7). In the meridional mass circulation, these changes are shown by the weakening in the mean Hadley cell (Figure 9c). In Figure 9c, an anomalously strong subsidence in the midlatitude region occurs at the peak dry phase of the precipitation cycles, extending to 37°N where there is strong anomalous rising motion in the wet phase of the precipitation cycles (Figure 9b). These anomalies in the meridional circulation are consistent with the high pressure anomalies in the lower troposphere during the peak dry phase of the precipitation cycles (Figure 6c). They cause reduction in the Indian monsoon precipitation (Figure 2a). Also during this time, the meridional circulation in the high-latitude and polar region has anomalously strong rising motion. The anomalous rising motion in the polar region would result in adiabatic cooling, reducing the air temperature on top of the radiation cooling.

These changes would recondition the atmosphere in the eastern Northern Hemisphere to the state with large meridional enthalpy and temperature gradient and enhanced baroclinic instability in the midlatitude region. Under such conditions, the Indian monsoon would enhance with increasing precipitation.

### 3.3. What May Have Determined the Timescale of This Variation?

Indeed, the atmosphere has a small thermal inertia that would tend to make the atmosphere to respond to and remove thermal imbalance in rather short timescales (days to weeks) in forms of mesoscale and synoptic weather processes. The small thermal inertia does not, however, have to result in only fast response or short timescale variations, especially at hemispheric or global scales. The strong nonlinearity of the atmosphere rising from interactions among a wide spectrum of processes in the atmosphere, as well as interactions with radiation and the oceans, can sustain slow timescale variations in circulation and climate. Some of those variations are summarized in Hartmann [1994].

In this multidecadal variation of the Indian monsoon precipitation, its timescale is determined by the hemispheric scale circulation processes that increase the transport of heat from the tropical and subtropical (monsoon) regions to the high-latitude and polar region during the wet phase of the variation (driven by the strengthened meridional gradient of atmospheric enthalpy) and the relaxation process in the following dry phase of the variation. While the poleward heat transport persists in boreal summer [Palmen and Newton, 1969; Oort, 1971], the transport is not at a constant rate or intensity in time. This is partially because the atmosphere is a rather inefficient heat engine and only about 0.05% of the potential energy in the system is converted to the kinetic energy of disturbances [e.g., Holton, 2004]. Because these disturbances are the effective means of transporting heat poleward, it is a slow process for the heat in the tropics and subtropics to be generated and transported to the high-latitude region. Moreover, the factors such as atmospheric moisture content, land cover condition, and heat capacity of the land cover are utterly different between the high latitudes and the tropics/subtropics. These differences affect the heat sources/sinks in the tropical/polar region and also the rate of the transport. We have shown that these contrasts between the heat inertia of the high-latitude region and the low-latitude areas and the inefficiency in heat generation in the atmosphere create time delays between the strong radiation cooling in high-latitude region, the corresponding heat generation in the tropical and subtropical (monsoon) regions, and the poleward heat transported by the atmospheric circulation. Specifically, the phase of the Indian monsoon precipitation variation lags that of the meridional enthalpy gradient variation but leads that of the intensity change in heat transport. We propose that these special phase relationships among the heat sink, source, and transport, and the chasing process by the transport of heat to meet the need for it driven by the meridional enthalpy gradient, characterize such a multidecadal timescale. A more quantitative evaluation/proof of the time scale from the work of these processes may only be done using an idealized modeling method, which is a separate subject from this study.

It is also possible that the timescale of the proposed mechanism for this multidecadal monsoon variation could be related to variations in the SST across the oceans, especially because the oceans have much large heat capacity and long memory. In the following, we will examine the SST relationship with multidecadal monsoon variations, including this particular variation component.
4. Relationship Between Variations in SST and the Monsoon Precipitation, A.D. 1470–1999

The geographical locations where some information of past SST variations was extracted from proxy data and is used in this study to examine their relationship with the multidecadal Indian monsoon precipitation variation are shown in Figure 10a. All the SST variations have been subjected to a 35 year low-pass filter. This filtering process suppresses the variations with frequency higher than 35 years per cycle and reserves the multidecadal timescale variations. (As an example, we show in Figures 10b and 10c the power spectra of both

Figure 10. (a) The oceanic regions where SST anomalies and their relationship with Indian monsoon precipitation variation are examined. Region A is the tropical Indian Ocean and Arabian Sea (oceanic area between north of 10°S and the Asian continent and between west of 77°E and the African continent). Region B is the North Pacific Ocean where the PDO index is derived. Region C is the North Atlantic Ocean where the AMO index is derived. (b) Power spectrum of the model-simulated SST variations in the North Atlantic Ocean (region C in Figure 10a), after a 35 year low-pass filter was applied to the SST time series. The thick solid line is the power, the thin solid line is the red noise function, and the thin dashed line is the 95% confidence level of the red noise function. (c) Same as Figure 10b but for the SST time series in region C from reconstructed SST data.
the model-simulated SST variation and the reconstructed SST variation [Gray et al., 2004] in the North Atlantic Ocean (region C in Figure 10a). Both the spectra show outstanding multidecadal variations, i.e., the Atlantic Multidecadal Oscillation (AMO).

The correlations of the model-simulated multidecadal SST variations in the North Atlantic and in the other regions marked in Figure 10a with the simulated monsoon precipitation are shown in Figures 11a–11c.

In the following, we discuss these results in the alphabetic order for oceanic regions marked in Figure 10a. Figure 11a shows fairly consistent variations at multidecadal timescale between the Indian monsoon precipitation and the SST in the tropical Indian Ocean and the Arabian Sea. The Pearson’s correlation coefficient is +0.27 and is significant at the 99% confidence level based on the Monte Carlo test. Moreover, our analyses of lagged correlations show no significant lead or lag between the variations, suggesting that the SST in that region and the monsoon precipitation have varied in phase at the multidecadal timescale. This result is consistent with that from modeling studies of Shukla [1975] and Washington et al. [1977], although they suggested that the Indian Ocean SST variation can affect the Indian monsoon precipitation through anomalies in surface evaporation.

Figure 11b shows a statistically significant positive correlation (+0.31) between the Pacific Decadal Oscillation (PDO) index and Indian monsoon precipitation variation. The correlation between the monsoon precipitation and the AMO is also positive but statistically insignificant (Figure 11c). Again, additional tests reveal no lead or lag between these variations. (Note that we did not compare these simulated relationships of the SST and the Indian monsoon precipitation variations in the paleo-time with the results from some previous studies on...
this subject that used the observational data in the twentieth century. This is because such comparisons are questionable according to our discussion in the beginning of section 3.3 and the reasons described later in this section.

We now compare these correlations derived from the modeled data against the correlations derived from the reconstructed SST in those same regions and the reconstructed monsoon precipitation. The latter are shown in Figures 12a–12c. Before discussing these results, we need to explain the process of obtaining them.

The reconstructed records of SST in the different oceanic regions (Figure 10a) vary in time and data length. The reconstructed SST values in the tropical Indian Ocean and Arabian Sea [Mann et al., 2009] cover the entire period of A.D. 1470–1999, the same as the period of reconstructed precipitation data. So, all these reconstructed SST data are used in our analysis. The reconstructed PDO index data are from A.D. 993–1996 [McDonald and Case, 2005], and we use the data from A.D. 1470–1996 (493 samples). The reconstructed AMO data are from A.D. 1567–1990 [Gray et al., 2004], and all the 424 data are used. All these time series of the SST data are subjected to the same 35 year filter.

Figure 12a shows a significant negative correlation of $-0.24$ between the reconstructed monsoon precipitation and reconstructed SST variation in the Tropical Indian Ocean and the Arabian Sea. This result is opposite to the modeled positive correlation in Figure 11a. Clark et al. [2000] has indicated that this negative correlation is statistically nonstationary in time. Figure 12b shows that monsoon precipitation and the PDO index have an insignificant negative correlation of $-0.02$, in contrast to $+0.31$ from the model-simulated relationship.
(Figure 11b). The correlation between the monsoon rainfall and the AMO index is $-0.04$ and is statistically insignificant (Figure 12c). This result also differs from the model result of a positive and insignificant correlation of $+0.032$ (Figure 11c). This negative correlation from the reconstructed results is also opposite to a positive correlation of the two variations in the twentieth century [Feng and Hu, 2008]. This difference is again suggesting a nonstationary relationship between the North Atlantic SST and the Indian monsoon precipitation variation.

Additional evidence showing varying relationships between the Indian monsoon precipitation and SST variations in other oceanic regions and timescales includes the change of the previously recognized strong negative correlation between the El Niño–Southern Oscillation (ENSO) and the Indian monsoon in recent decades [Kumar et al., 1999; Webster, 2006]. While these changes at broad time scales indicate unstable relationships between the SST (assumed forcing) and the Indian monsoon variation, they show supports to the notion that the Indian monsoon precipitation variation, at interannual and multidecadal timescales, has its own internal processes that would interact with but not necessarily be forced by the SST anomalies in the near and far oceans. Some of these processes could enhance their roles in, or dominate, the variations in the monsoon precipitation in some conditions (time). They fade away or are overwhelmed by other processes of different frequencies under different conditions/time. These changes describe the varying relationships of the monsoon and the SST variations.

5. Summary and Concluding Remarks

Both the reconstructed Indian monsoon precipitation from A.D. 1470–1999 and paleo-climate model-simulated monsoon precipitation in that same period show multidecadal variations at the frequency of 16–17, 22–28, and 39–52 years per cycle. In other words, these components of variation at the multidecadal timescale, along with variations across the time spectrum, interact and constitute the reconstructed and modeled monsoon precipitation variation. These variation components are outcomes of specific physical processes in the climate system. These processes are constantly interacting with each other and with other (external) forcings. In certain conditions and as results of the interactions, some of those processes would become strengthened enough that they dominate the variation of the monsoon precipitation. As the condition changes in time, other processes may amplify, overtaking the dominant role of the previous process or processes and charging the variation of the precipitation. By understanding those physical processes and the conditions under which they may change their roles, we can understand the precipitation variations.

In this context, we examined the processes causing the 39–52 (or 40–50) year variation in the Indian monsoon precipitation, after identifying the time period from A.D. 1640 to 1762 when this variation was outstanding (thus dominant) and also statistically coherent between the reconstructed and the CCSM4 simulated monsoon precipitation variations.

Major results show that this variation in the Indian monsoon precipitation is an outcome of the hemispheric scale circulation processes that transport heat from the tropical and subtropical (monsoon) regions to the high-latitude and polar region. The monsoon precipitation intensifies shortly after the meridional gradient of the atmospheric enthalpy strengthens, in response to the demand for more heat/energy generation in the low latitudes. The subsequently elevated poleward transport of enthalpy redistributes the heat generated from intensifying monsoon processes and precipitation. As a result, the meridional gradient of the enthalpy gradually weakens. The weakening meridional gradient of enthalpy relaxes the need for heat production and poleward transport. The monsoon precipitation remains relatively low in the following period of two decades or so, while the meridional gradient of the enthalpy rebuilds and the need for strong poleward transport rises again.

The mechanism proposed and elaborated in this study provides a new perspective for development of multidecadal variations in the Indian monsoon. That is, in addition to potential external forcing processes, e.g., the variation in solar irradiance, there can be circulation processes in the atmosphere that could sustain multidecadal variations in the monsoon. As illustrated in this particular variation, its processes involve circulation and radiation interactions that build the hemispheric scale meridional gradient of the atmospheric enthalpy and initiate and achieve the poleward transport of heat to meet the need of balancing the radiation loss in the high-latitude region. The timescale of this variation is suggested to be largely determined by the interaction of the circulation and the radiation loss in the high-latitude region.
and the time delay of the heat generation in the monsoon region and the poleward transport. This time delay is caused by the inefficient generation rate of heat in the tropical and subtropical regions, dictated by the nature of an inefficient heat engine of the atmosphere. Oceanic effect definitely plays a role in these processes and the timescale of the variation by influencing the surface temperature and moisture in the low-latitude region [e.g., Shukla, 1975; Washington et al., 1977].

Finally, same as all the other components of variations in the climate system, this variation changes its amplitude and wobbles within its frequency range over time, as shown in the original time series. When the mechanism behind it weakens or is overwhelmed by other processes at different frequencies, this particular variation would fade into the background variation. Any particular relationship between this variation and other apparent forcing, e.g., SST, would be interrupted or become insignificant. This may explain the observed time varying nature of the Indian summer monsoon with other processes/forces, such as ENSO, the solar irradiance, and the Eurasian snow cover. Understanding the processes resulting in the individual multidecadal, as well as interannual, variation components in the monsoon variation and further examining and understanding their interactions are needed to improve our knowledge of and ability in predicting the Indian monsoon variation.

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