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An objective definition of the Indian summer monsoon season and a new perspective on the ENSO–monsoon relationship

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ABSTRACT: The concept of an interannually varying Indian summer monsoon season is introduced here, considering that the duration of the primary driving of the Indian monsoon – the large-scale meridional gradient of the deep tropospheric heat source – may vary from one year to another. Onset (withdrawal) is defined as the day when the tropospheric heat source shifts from south to north (north to south). This physical principle leads to a new thermodynamic index of the seasonal mean monsoon. While the traditional measure of seasonal rainfall, averaged from 1 June to 30 September, indicates a breakdown of the ENSO–monsoon relationship in recent decades, it is argued that this breakdown is partly due to the inappropriate definition of a fixed monsoon season. With a new physically based definition of the seasonal mean, the ENSO–monsoon relationship has remained steady over the decades. El Niño (La Niña) events contract (expand) the season, and thus decrease (increase) the seasonal mean monsoon by setting up persistent negative (positive) tropospheric temperature (TT) anomalies over the southern Eurasian region. Thus, we propose a new pathway, whereby the Indian summer monsoon could be influenced by remote climatic phenomena via modification of TT over Eurasia.

Diagnostics of the onset and withdrawal processes suggest that onset delay is due to the enhanced adiabatic subsidence that inhibits vertical mixing of sensible heating from warm landmass during the pre-monsoon months. On the other hand, the major factor that determines whether the withdrawal is early or late is the horizontal advective cooling. Most of the late (early) onsets and early (late) withdrawals are associated with El Niño (La Niña). This link between the ENSO and the monsoon is realized through vertical and horizontal advections associated with the stationary waves in the upper troposphere set up by the tropical ENSO heating. Copyright © 2007 Royal Meteorological Society

KEY WORDS apparent heat source; apparent moisture sink; length of rainy season; tropospheric temperature

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1. Introduction

The droughts and floods associated with the interannual variation (IAV) of the seasonal mean Indian summer monsoon rainfall (often referred to as 'all India rainfall', or AIR) have devastating effects on the people, agriculture and economy of the region. Therefore there is a great need for long-range prediction of the seasonal mean AIR. However, neither the empirical nor the dynamical models currently have any useful skill in predicting AIR (Rajeevan *et al*., 2004; Kumar *et al*., 2005; Wang *et al*., 2005). Improved prediction depends on a better understanding of the factors influencing the IAV of AIR: namely, local air–sea interactions, and connections with remote climatic phenomena such as El Niño and the Southern Oscillation (ENSO), Eurasian snow cover, and the North Atlantic Oscillation. Proper delineation of the summer monsoon season is essential to defining and understanding the IAV of the AIR. The traditional definition of the Indian summer monsoon as comprising the months of June, July, August and September (JJAS), though convenient for operational purposes, is arbitrary. The Indian summer monsoon may be regarded as a result of the seasonal march of the tropical convergence zone (TCZ) beyond 10 °N (Gadgil, 2003). The physical monsoon season is thus delineated by the establishment of the TCZ around 10 °N.

The physical length of the Indian monsoon season (the 'length of the rainy season', or LRS) could, therefore, vary from year to year, as the internal and external forcings responsible for the onset and withdrawal of the monsoon may vary. In fact, it is known that even the traditionally defined onset date has large variability, having a standard deviation of 8–9 days, with the earliest and latest onsets differing by 46 days (Ananthakrishnan and Soman, 1988). The withdrawal date also has significant IAV. It may be recalled that the seasonal monsoon rainfall is essentially a statistical average of intraseasonal

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spells. See Goswami, 2005b for a review. Fixing the monsoon season as the period from 1 June to 30 September means that some important rain spells associated with the summer monsoon may be left out, and the IAV of AIR based on such a definition could misrepresent the actual IAV of the summer monsoon rainfall. So far almost all studies on the IAV of AIR and its connection with other climatic parameters have been with an AIR time series based on such a fixed season (e.g. Parthasarathy *et al*., 1995). This raises the possibility that the results of the teleconnection studies based on such a time series may be misleading regarding the actual physical processes involved. Of particular interest is the reported weakening of the ENSO–monsoon relationship in recent years (Kumar *et al*., 1999; Krishnamurthy and Goswami, 2000; Parthasarathy *et al*., 1991). It is conceivable that part of the apparent decrease in correlation between AIR and ENSO indices – such as the NINO3 index of sea surface temperature (SST) – is simply due to an incorrect definition of the rainy season. In the present study, we propose an objective definition of the monsoon season and re-examine the ENSO–monsoon connection. In doing so, we discover a new mechanism through which ENSO can influence the Indian monsoon.

An objective definition of the monsoon season depends on objective definitions of the first onset, or monsoon onset over Kerala, and of the last withdrawal, or withdrawal of the southwest monsoon from Kerala. There is already a host of such definitions of monsoon onset and withdrawal. A brief review of these is provided here. The most widely accepted onset definition (Ananthakrishnan *et al*., 1967) is based on precipitation from over 70 rain gauge stations over Kerala. Onset is the date of transition from light to heavy rainfall, sustained for more than 5 days above a threshold of 10 mm day^{-1}. The definition used by the India Meteorological Department is similar to that of Ananthakrishnan *et al*. (1967), taking into account the subjective judgement of forecasters on the evolution of the large-scale circulation. This depends on several factors, including an increase in the kinetic energy (KE) of the low-level jet. Since the onset is a distinct phase of the large-scale annual cycle of the monsoon, it is natural to define it as a transition from dry to wet conditions, but such a definition based on rainfall over a small region such as Kerala is sensitive to small-scale or higher-frequency fluctuations, such as synoptic disturbances and intraseasonal oscillations, and could lead to 'bogus' monsoon onsets (Flatau *et al*., 2003; Flatau *et al*., 2001).

Wang and Lin Ho (2002) introduced a criterion also based on a rainfall threshold but taking into account the seasonality of rainfall relative to the winter mean. The threshold they used was 5 mm day^{-1}. They showed that this criterion may be useful in defining the onset and withdrawal of the monsoon over southern as well as eastern Asia. However, the withdrawal from southern India is found to be later than normal, so the method is not suitable for defining the Indian summer monsoon season. This is because the rainfall criterion cannot distinguish between summer (or southwest) and winter (or northeast) monsoon rainfall during the withdrawal phase.

Zhang *et al*. (2002) used a similar rainfall criterion to define the onset of the summer monsoon over the Indochina peninsula. Zeng and Lu (2004) presented globally unified monsoon onset and retreat indices, making use of a normalized precipitable water index defined in terms of the 'golden ratio'. This index gives spatial variations of the onset and withdrawal dates, but does not correctly represent the climatological onset and withdrawal of the Indian summer monsoon. In an attempt to make the definition of the onset and withdrawal of the Indian monsoon less susceptible to small-scale processes, Fasullo and Webster (2003) defined an index of onset and withdrawal in terms of vertically-integrated moisture transport, averaged between surface and 300 hPa and over a large area over the Arabian Sea and part of the Indian subcontinent. By their definition, the withdrawal of the monsoon from the peninsula is rather too early. Ramesh *et al*. (1996) used net tropospheric (1000–300 hPa) moisture build-up, mean tropospheric (1000–100 hPa) temperature increase, and sharp increase of KE at 850 hPa over the Arabian Sea (0–19.5 °N, 55.5–75 °E) to define the onset of the monsoon. Some of these definitions of onset, however, still depend on a somewhat arbitrary precipitation threshold, for which there is no justification from physical principles. Similarly, most definitions of withdrawal from the Indian peninsular are based on rainfall (or related parameters). As rainfall from synoptic disturbances from the northeast monsoon seamlessly takes over activity over the southern peninsula from the southwest monsoon TCZ rainfall, definitions of withdrawal based on rainfall are unable to distinguish between the two. What is required is a way to identify when the southwest monsoon withdraws from the southern peninsula.

What are the primary physical processes responsible for the Indian monsoon onset? Why is the Indian summer monsoon onset so much more abrupt than the transition during withdrawal? The abruptness of the onset is due to explosive growth of organized deep convection. What are the dynamical and thermodynamical factors that are built up prior to the onset and that lead to this explosive convection at the time of the onset?

In this paper we identify a large-scale thermodynamic forcing that is responsible for preconditioning the onset of the Indian summer monsoon, and describe the sequence of events that leads up to the onset. We then propose an alternative large-scale definition of the Indian summer monsoon onset, which is related to the basic dynamic and thermodynamic forcing. We also develop a thermodynamic definition of the withdrawal. The usual definitions of onset and withdrawal of the monsoon bear no significant relationship with the ENSO. Fasullo and Webster (2003), using the hydrological definition of onset and withdrawal, were the first to indicate a relationship between the onset and the ENSO, which had previously been postulated (e.g. Joseph *et al*., 1994). However, there was no significant relationship between the withdrawal and the ENSO. With the new definition of the monsoon

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season, we present evidence of a significant relationship between the length of the monsoon season and the ENSO, and provide a better insight into the ENSO–monsoon connection. The conventional description of this connection is in terms of the large-scale east–west shift in the tropical Walker circulation. We discover here that there exists an extratropical ENSO–monsoon connection through wave influences on the tropospheric temperature anomalies over Eurasia, induced by heating of the tropical troposphere during an El Niño event. Some preliminary results, and a short account of the new definitions of onset, withdrawal and LRS, have been reported by Goswami and Xavier (2005). The present study is more comprehensive, and includes a better foundation for the objective definitions, a detailed diagnosis of IAV of the onset and withdrawal, and a new large-scale index of the seasonal mean monsoon.

Data and methodology are briefly discussed in the next section. The physical basis for the definitions of onset date (OD) and withdrawal date (WD) is described in Section 3. The climatology and IAV of OD, WD and LRS are calculated, and their robustness examined, in Section 4. Connections with ENSO and interdecadal trends of the LRS are investigated in Section 5. A new thermodynamic index of monsoon is introduced, and the ENSO–monsoon relationship is re-examined, in Section 6. In Section 7, diagnostics of late and early onset and withdrawal are performed using heat and moisture budget calculations. The results are summarized in Section 8.

2. Data and methods

NCEP/NCAR reanalysis (Kalnay *et al.*, 1996; Kistler *et al.*, 2001) daily data for the period 1950–2003 have been extensively used in this study. TT is defined as the air temperature averaged between the levels 600 hPa and 200 hPa. The temperature is averaged over this range – unlike in Goswami and Xavier (2005) – to eliminate any possible influences of ground temperature over high terrains. Thanks to the large-scale spatial coherence, this slight difference in the averaging has not introduced any significant changes in the statistics of OD and WD. The computation of apparent heat source Q_1 and moisture sink Q_2 uses air temperature, horizontal winds u and v and specific humidity at all pressure levels; surface pressure; surface air temperature (at 995 hPa); surface horizontal winds; and surface height.

He *et al.* (2003) compared the heat and moisture budgets obtained from the ERA data (Uppala, 2002) with those obtained from the NCEP/NCAR reanalysis, and found that the budgets from the two reanalyses are quite consistent. Their results are corroborated by the observed data.

It may be noted that the accuracy of Q_1 and Q_2 estimates depends crucially on the accuracy of pressure vertical velocity ω, which is a highly scheme-dependent quantity. So ω is recomputed here by vertically integrating the continuity equation with certain constraints. The method of calculating ω is very similar to that used by Li and Yanai (1996). Tropopause level data are used, including the tropopause level pressure and the tropopause level air temperature. The daily values are converted to 5-day averages for convenience of computation. The AIR (June–September rainfall) is taken from Parthasarathy *et al.* (1995), and updated data are obtained from the Indian Institute of Tropical Meteorology (http://www.tropmet.res.in/data.html). Computation of AIR during the LRS period requires daily rainfall values over the Indian region. The new daily gridded high-resolution ($1^\circ \times 1^\circ$ latitude/longitude) rain-gauge data over the Indian subcontinent for the period 1951–2003, prepared by the India Meteorological Department (Rajeevan *et al.*, 2006), are used for this purpose.

3. Definition of onset, withdrawal and LRS: physical basis

Our index of Indian summer monsoon onset and withdrawal is based on the understanding that the abrupt seasonal changes of the Asian monsoon are related to tropospheric seasonal warming over the Asian monsoon region (Krishnamurti, 1985; Murakami and Ding, 1982; He *et al.*, 1987; Yanai *et al.*, 1992). The importance of the Tibetan Plateau as an elevated heat source warming the upper troposphere, and thus inducing the abrupt seasonal transition, has been much discussed (Flohn, 1957; Flohn, 1960; Ye and Gao, 1979; Gao *et al.*, 1981; Ye, 1981; Luo and Yanai, 1984; Murakami, 1987; Yanai *et al.*, 1992; Yanai and Li, 1994). The classical hypothesis of Flohn (1957) and Flohn (1960) is that seasonal heating of the elevated surface of the Tibetan Plateau, and the consequent reversal of tropospheric temperature and pressure gradients south of 35 °N, trigger the large-scale change of the general circulation over Asia and the abrupt burst of the monsoon over the Indian subcontinent. Some later studies (Li and Yanai, 1996; Ueda and Yasunari, 1998; He *et al.*, 2003; Yanai and Li, 1994; Yanai *et al.*, 1992) have shown that the onset of the Indian monsoon is closely related to the change in sign of the meridional gradient of upper tropospheric temperature south of the Tibetan Plateau. The meridional gradient of upper tropospheric temperature (for example, averaged between 200 hPa and 600 hPa) is proportional to the meridional gradient of deep tropospheric heating, and could lead to acceleration of the deep tropospheric circulation. The classical land–ocean contrast theory of monsoon involving surface temperature, on the other hand, could only produce a shallow circulation (Schneider and Lindzen, 1976). Sensible heating over the Tibetan Plateau during the pre-monsoon period plays a crucial role in producing the deep heating in the northern areas (Yanai *et al.*, 1992; Li and Yanai, 1996; Yanai and Li, 1994) and is instrumental in reversing the meridional gradient of TT. Heat advection associated with quasi-stationary planetary waves can also influence the temperature of the northern areas, and hence the meridional gradient of TT.

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The following sequence of events is envisaged around the time of Indian monsoon onset. The surface temperature gradient during the pre-monsoon period (late April and May) sets up cross-equatorial flow near the surface, bringing warm moist air to the continent, but is capped by cold and dry air subsidence above the planetary boundary layer. This leads to high build-up of potential convective instability that cannot be realized because of inhibition of subsidence above the planetary boundary layer.

The change in sign of the meridional gradient of TT signals the setting up of the off-equatorial large-scale deep heat source. The atmospheric response to this heat source (Gill, 1980) leads to cross-equatorial flow and strengthening of the low-level southwesterlies above the planetary boundary layer. This leads to a shift of the zero-absolute-vorticity line in the lower atmosphere (say, 850 hPa) to about 5 °N, facilitating symmetric inertial instability (Tomas and Webster, 1997). The monsoon onset, or explosive development of organized off-equatorial convection over India and the Bay of Bengal, takes place when the frictional boundary layer convergence induced by symmetric instability overcomes the existing inhibition due to large-scale subsidence (Tomas and Webster, 1997; Krishnakumar and Lau, 1998; Krishnakumar and Lau, 1997). In a day or two, this leads to an explosive increase in the KE of the low-level winds over the Arabian Sea.

Our objective definition of 'first large-scale' monsoon onset and withdrawal is based on the reversal of TT (averaged between 200 hPa and 600 hPa) between a northern box (40–100 °E, 5–35 °N) and a southern box (40–100 °E, 15 °S–5 °N), denoted ΔTT. The onset date is defined as the date when ΔTT changes sign from negative to positive, and the withdrawal date is defined as the date when ΔTT changes sign from positive to negative. These two boxes are chosen so that ΔTT represents reasonably well the large-scale heating gradients driving the large-scale monsoon circulation. The TT over these two regions is spatially coherent, so that ΔTT does not change much for small changes in the averaging areas. The onset and withdrawal dates thus defined are physically based and do not depend on an arbitrary threshold of precipitation, as do most traditional definitions. While rainfall-based definitions of withdrawal can be ambiguous because of rainfall from the northeast monsoon, our definition marks an absolute end of the southwest monsoon season, as it indicates the transition of a heat source from north to south of 5 °N, with northward propagation of the TCZ (Sikka and Gadgil, 1980; Webster *et al.*, 1998; Gadgil, 2003) being inhibited after this date. As this definition is based on averages over large regions (and because ΔTT changes slowly), it is less susceptible to 'bogus' onsets (Flatau *et al.*, 2003; Flatau *et al.*, 2001).

As mentioned earlier, a characteristic of the large-scale onset of the Indian monsoon is an abrupt increase of KE of the low-level monsoon flow. The day of abrupt increase in the KE of 850 hPa winds averaged over a large region (40–100 °E, 5–15 °N) above a threshold value of 40 m^2 s^{-2} and persisting for 5 consecutive days may be referred to as *KE onset*. We will show that the thermodynamic ΔTT onset described here correlates well with the traditional definitions of South Asian monsoon onsets based on the KE of the low-level monsoon flow, as well as those based on precipitation.

4. Climatology and interannual variability of OD, WD and LRS

The climatological annual evolution of TT averaged between 40 °E and 100 °E is shown in Figure 1(a). It is evident that the meridional gradient of the climatological mean TT changes sign around the end of May. As mentioned above, OD is defined as the first day when ΔTT changes sign from negative to positive (Figure 2), suggesting a shift of heat source from the equatorial region to the warm subcontinent. Similarly, WD is defined as the first day when ΔTT switches from positive to negative values (Figure 2). LRS is defined as the difference between WD and OD. The meridional evolution of climatological precipitation averaged between the same longitudes (Figure 1(b)) shows an abrupt transition of equatorial TCZ from about 10 °S to 15 °N associated with the rainfall onset. These two transitions are separated

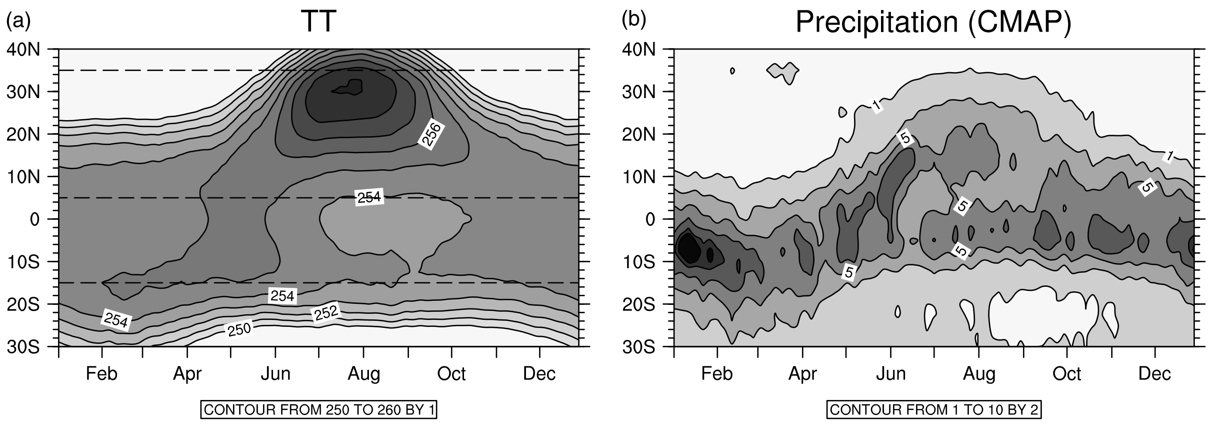

Figure 1. Time–latitude evolution of climatological mean (a) TT (K) and (b) precipitation (mm day^{-1}) averaged over the longitudes 40–100 °E. The dashed lines in (a) indicate the latitudinal zones for which averaging has been done in the computation of ΔTT.

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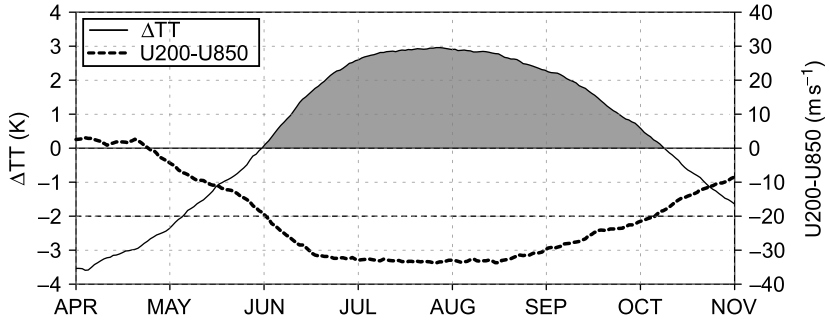

Figure 2. Evolution of climatological values of ΔTT (solid line, scale on the left) and of the climatological mean vertical shear of zonal winds (U_{200}–U_{850}) averaged over (50–95 °E, 0–15 °N) (dotted line, scale on the right). The shaded area under the ΔTT curve represents the climatological value of our thermodynamic index of the seasonal mean monsoon.

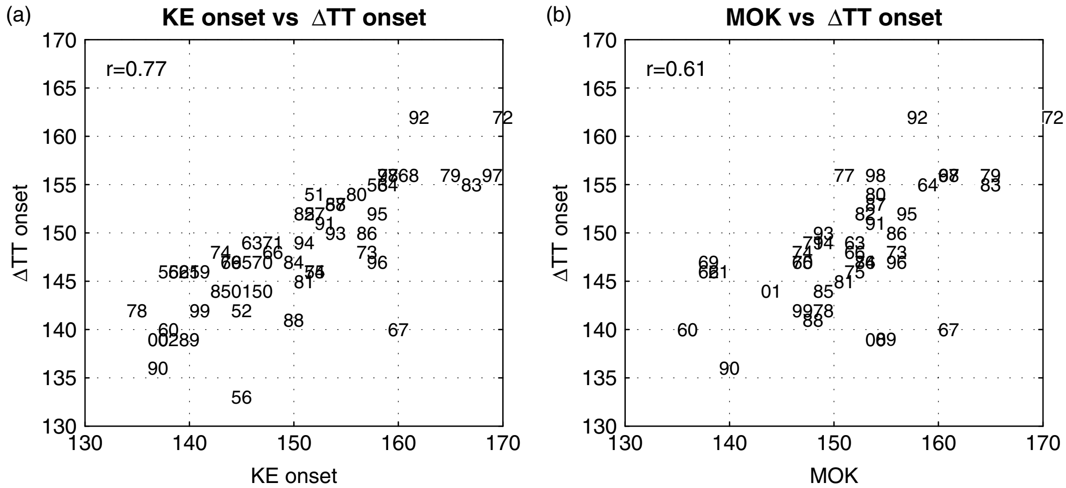

Figure 3. Scatter plot of OD defined from ΔTT with (a) KE onset, and (b) Monsoon onset over Kerala (MOK). Points are represented with the last two digits of the years. The correlation coefficient r is shown in each panel. Units are Julian days.

by a few days. It has recently been shown (Jiang *et al.*, 2004) that the northward propagation of TCZ over this region depends crucially on the strong easterly shear of zonal winds over the region. The climatological mean vertical shear of zonal winds (U_{200}–U_{850}) averaged over (50–95 °E, 0–15 °N) is also shown in Figure 2 (dotted line). It appears that easterly shear below a threshold of about -20 ms^{-1} sets up conditions for further northward movement of organized convection after the onset. The shear going below this critical value around the WD also sets up inhibition for northward propagation after this date. The OD, WD and LRS are noted from NCEP reanalysis for every year between 1950 and 2003.

The KE onset dates for all the years between 1950 and 2003 were also noted and compared with the ODs based on ΔTT (Figure 3(a)). A strong correlation ($r = 0.77$) between the two onsets may be noted. As expected, the KE onset generally occurs after the ΔTT onset. A few extremes (such as 1956 and 1967) are likely to be related to the ambiguity in selecting the KE onset because of sudden increase in KE due to small-scale processes. As the thermodynamic onset (OD) should lead to explosive organized convection and the monsoon onset over Kerala, the traditional definition of monsoon onset over Kerala should also be related to it. In fact, the OD and monsoon onset over Kerala defined by the India Meteorological Department (Joseph *et al.*, 1994) do correlate significantly (Figure 3(b), $r = 0.61$).

The time series of onset and withdrawal dates for 1950–2003 are shown in Figure 4, and the statistics of the OD, WD and LRS based on NCEP reanalysis datasets are presented in Table I. The figure shows a significant downward trend of WD, but no perceptible trend in OD. The climatological mean OD is 30 May and the climatological mean WD is 10 October, with standard deviations of 7 and 9 days respectively. There is a difference of almost a month between the earliest and latest OD, and similarly for WD. The table shows that the IAV of WD is larger than that of OD; this is consistent with the findings of Syroka and Toumi (2004),

Table I. Statistics of OD, WD and LRS.

	OD	WD	LRS (days)
Earliest/minimum	14 May	16 September	98
Latest/maximum	13 June	20 October	159
Mean	30 May	10 October	129
Standard deviation	7 days	9 days	12

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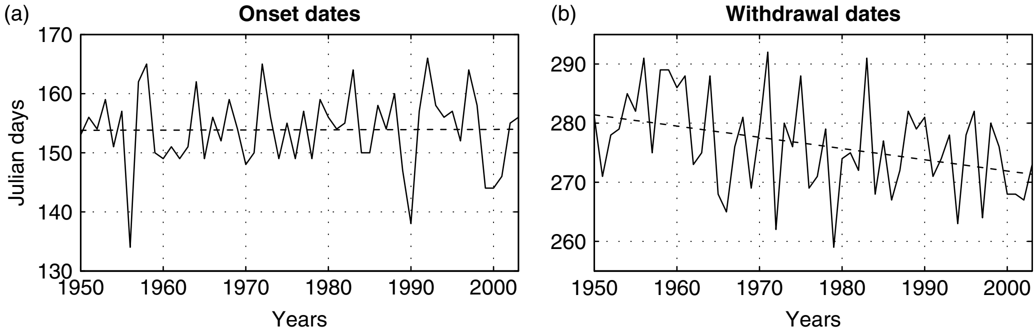

Figure 4. Interannual variations of (a) OD and (b) WD from 1950 to 2003. Dashed lines indicate the linear trends.

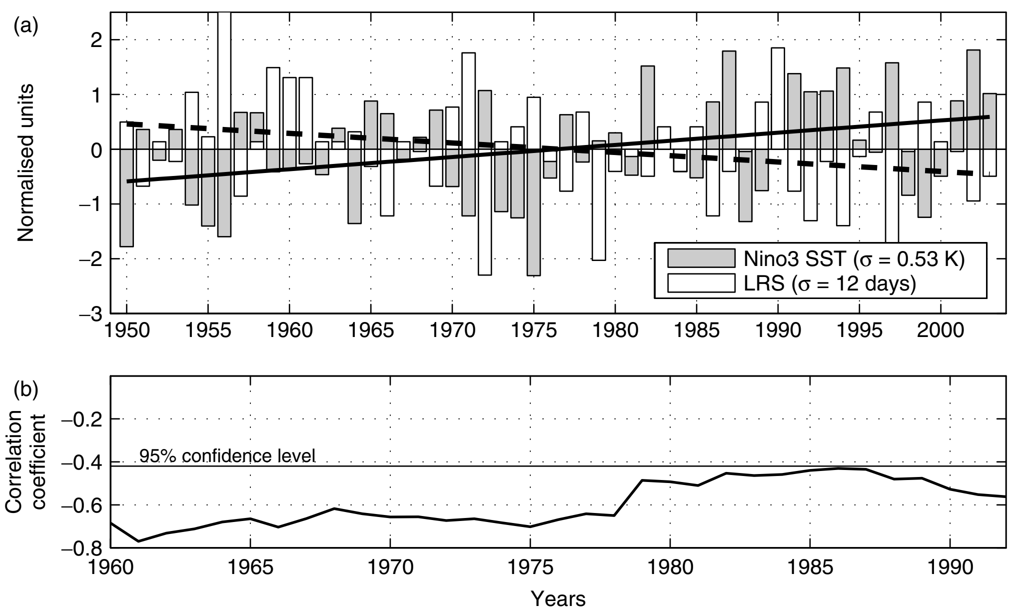

Figure 5. (a) Interannual variations of JJAS mean NINO3 SST anomalies and LRS. The positive trend in NINO3 SST and the negative trend in LRS are plotted as lines. The correlation coefficient between the two is −0.63. (b) 21-year running window correlation between JJAS mean NINO3 SST anomalies and LRS.

which are based on a daily circulation index. The average LRS is 129 days, with a standard deviation of 12 days. The extremes of LRS differ by about 60 days.

5. The ENSO–monsoon relationship

The large ranges in the OD, WD and LRS can significantly affect the seasonal mean rainfall by embracing more intraseasonal rain spells outside the JJAS season. This is very significant in relation to the recently-reported weakening of the ENSO–monsoon connection. We argue that part of this apparent weakening is a result of improper definition of the timing and length of the season, and thus exclusion of intraseasonal rain spells outside the JJAS season. It is interesting to note that the LRS is strongly correlated with NINO3 SST (Figure 5(a), $r = -0.63$). The interesting aspect of this ENSO–monsoon relationship is the fact that unlike the correlation between JJAS AIR and NINO3 SST, it has remained strong even during recent years (Figure 5(b)). The modulation of the Indian summer monsoon season by the ENSO occurs through modification of the meridional gradient of TT. ENSO regulates LRS through the atmospheric response to the diabatic heating associated with the ENSO SST, leading to a substantial reduction of the meridional gradient of TT over the Indian monsoon region from May to October. The composite TT anomalies (El Niño minus La Niña) averaged between 15 April and 15 May (Figure 6(a)) show that the negative TT anomaly over Eurasia during pre-monsoon months forced by El Niño heating delays the onset. The spatial pattern of ENSO-induced TT anomalies is consistent with the response of the tropical atmosphere to ENSO-SST-induced diabatic heating through tropical wave dynamics (Rodwell and Hoskins, 1996; Hui *et al*., 2003). The evolution of ΔTT composites for El Niño and La Niña years (Figure 6(b)) shows that El Niños shrink the monsoon season by delaying the onset and advancing the withdrawal. The evolution of ΔTT for strong and weak monsoon seasons (based on the seasonal mean rainfall) indicates that longer seasons are associated with stronger

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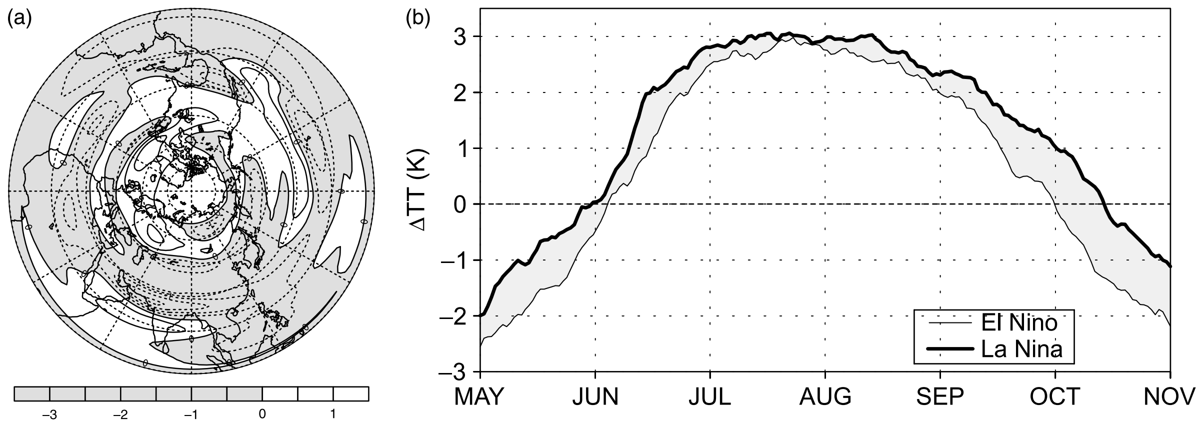

Figure 6. (a) Composite of TT (El Niño minus La Niña) (K) averaged over the period from 15 April to 15 May. (b) Evolution of ΔTT composites during El Niño (thin line) and La Niña (thick line).

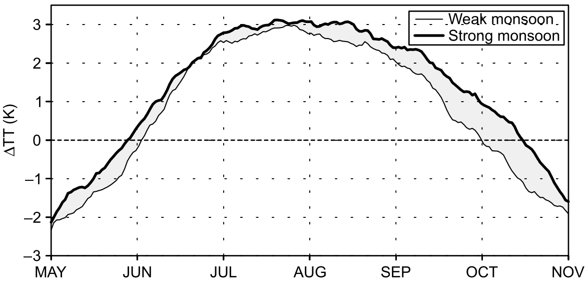

Figure 7. Evolution of ΔTT composites during strong (thick line) and weak (thin line) monsoon years. Strong (weak) monsoon years are defined as years when the seasonal mean rainfall is higher (lower) than its climatological values by 1 standard deviation of its interannual variability.

monsoons (Figure 7). This provides a basis from which to understand how the ENSO influences the monsoon through tropospheric heating.

It may be noted that the LRS (Figure 5(a), dashed line) has a negative trend. This is essentially a consequence of the negative trend of WD (Figure 4(b)). It turns out that over the same period, NINO3 SST has an increasing trend (Figure 5(a), solid line). The positive trend of tropical SST between 1950 and the present has been noted in many studies (Chelliah and Bell, 2004; Zhang *et al*., 1997). Thus, even on interdecadal time-scales, LRS and NINO3 SST are negatively correlated.

6. A thermodynamic index of the seasonal mean monsoon: reassessment of the ENSO–monsoon connection

The composite evolution of ΔTT during El Niño and La Niña years (Figure 6(b)) shows that, as well as the LRS being shorter, the area under the positive ΔTT region is significantly smaller during an El Niño than during a La Niña. Since El Niño (La Niña) is generally associated with weak (strong) Indian monsoons, it may be expected that the area under the positive ΔTT region during weak and strong monsoons may be similar to that in Figure 6(b). Indeed, the composite evolution of ΔTT for weak (strong) monsoons shows a very similar picture (Figure 7). This lends support to the idea that the area under the positive ΔTT curve for any given year may be related to the strength of the monsoon of that year. As the positive cross-equatorial pressure gradient (or positive ΔTT) is crucial for the existence of the Indian summer monsoon, it is reasonable to suppose that the cumulative positive ΔTT should be related to the strength of the seasonal mean Indian summer monsoon. Therefore, we define a thermodynamic index of the Indian summer monsoon (TISM) as the area under the positive part of the ΔTT curve (the shaded area in Figure 2). Using NCEP reanalysis data, the positive part of the ΔTT curve is integrated from the OD to the WD of each year to obtain the TISM. The interannual correlations of TISM with the seasonal rainfall from different sources are shown in Table II. The strong correlation between TISM and JJAS AIR establishes TISM as a good indicator of the strength of the Indian summer monsoon. It may be noted that the correlation between LRS and TISM is 0.73. This means that a significant proportion of the IAV of seasonal mean rainfall is governed by internal dynamics (in addition to the length of the season).

Figure 8(b) shows the stronger relationship between TISM and total AIR for the LRS period ($r = 0.76$), compared with that for the JJAS period (Figure 8(a), $r = 0.66$). The recently-reported regime shift in the ENSO–monsoon relationship is examined in Figure 8(c), which shows the 21-year running correlations between JJAS mean NINO3 SST and three other indices of the summer monsoon (total AIR for the JJAS season, total AIR for the LRS period, and TISM). The curve for

Table II. Correlations between TISM, AIR for the JJAS and LRS periods, and LRS.

	TISM	JJAS AIR	LRS AIR	LRS
TISM	1	0.67	0.75	0.73
JJAS AIR		1	0.94	0.49
LRS AIR			1	0.71
LRS				1

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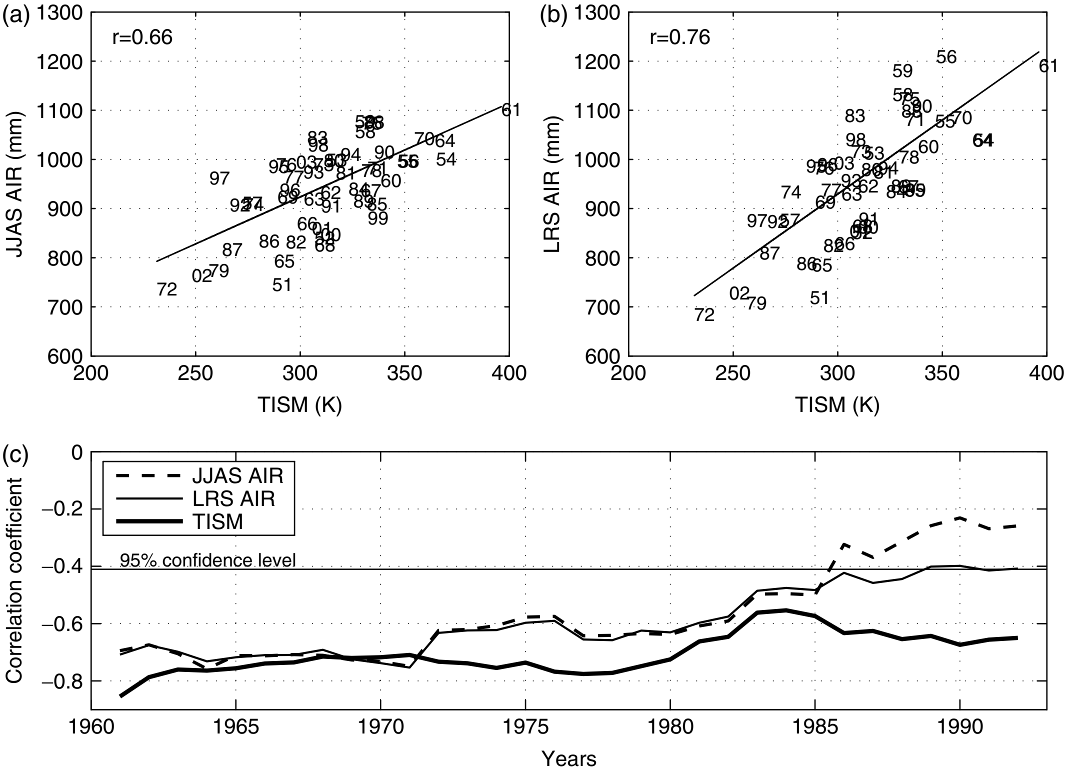

Figure 8. Scatter plot of TISM with the anomalies of total AIR for (a) the JJAS period and (b) the LRS period. Points are represented with the last two digits of the years. The correlation coefficient is shown in each panel. Linear least squares fits are also plotted. (c) 21-year running window correlation between JJAS NINO3 SST anomalies and anomalies of total AIR for the JJAS period, anomalies of total AIR for the LRS period, and TISM.

JJAS AIR shows a dramatic decay in the correlations after 1980. The correlations become insignificant in the recent couple of decades. This issue has been the subject of a number of studies (e.g. Kumar *et al.*, 1999; Krishnamurthy and Goswami, 2000). With the introduction of the new definition of the season in terms of LRS, the relationship of AIR with the ENSO tends to be more steady and significant.

However, the correlations between NINO3 and TISM are stronger than both the correlations mentioned above, and are remarkably steady over the decades, with an average value of −0.71. This suggests some new ideas regarding the ENSO–monsoon connection. Kumar *et al.* (1999) speculated that one of the reasons for the breakdown of the ENSO–monsoon relationship might be the southeastward shift in the Walker circulation anomalies associated with ENSO events, which might lead to reduced subsidence over the Indian region, thus favouring normal monsoon conditions despite the steady increase in the Pacific SST. Another hypothesis is that the increased surface temperatures over Eurasia in winter and spring, which are part of the midlatitude continental warming trend, might favour the enhanced land–ocean thermal gradient conducive to a strong monsoon. We show here that the apparent breakdown of the ENSO–monsoon relationship is partly a result of improper representation of the physical processes that regulate the initiation and termination of monsoon activity. The strong and steady relationship between ENSO and TISM suggests that a fundamental forcing of the monsoon by ENSO, through upper tropospheric temperature associated with wave activity, has remained steady over the years.

There are two teleconnection pathways through which monsoon rainfall may be affected: one is the extratropical connection described here, represented by the TISM; the other is the tropical connection via the shift of the Walker circulation and the modification of the local monsoon Hadley cell. Both these pathways may contribute to the total monsoon rainfall variability. The large-scale thermodynamic definition of monsoon (the TISM) seems to be less affected by internal variability, and the robustness and steadiness of the ENSO–TISM relationship seems to indicate that it represents the predictable component of monsoon variability. On the other hand, the rainfall-based definitions of monsoon, such as AIR, are more strongly influenced by the internal variability. The decrease in correlation between ENSO and AIR during recent years seems to be due to the relatively stronger role in recent years of internal variability (Goswami, 2004).

7. Diagnostics of variability of onset and withdrawal

As the variability of OD and WD determine that of LRS, which in turn is related to the IAV of the monsoon, a better understanding of the variability of OD and WD should lead to a better understanding of the IAV of the monsoon. Such an understanding may be achieved by examining the heat and moisture budgets during the onset and withdrawal phases. Yanai and Li (1994) examined the heat sources and moisture sinks before and after the onset of the summer rains over the Tibetan Plateau, from May to August 1979. Detailed diagnostics of the mechanism of reversal of TT gradient were derived by He *et al.* (2003), by examining the upper tropospheric heat budgets over the northern land areas in the monsoon regions. All these studies relating the heating of the Tibetan Plateau to the monsoon were based on climatological or monthly data for a few years (Li and Yanai, 1996; He *et al.*, 2003). Our objective here is to understand the IAV of onset and withdrawal and to infer the role of warming (or cooling) rates caused by horizontal temperature advection and the adiabatic heating related to the vertical motions. The role of boundary forcing in producing these differences in heating rates must also be elucidated. Since the amplitude of the IAV of OD and WD is of the order of 1 to 2 weeks, we work here with pentad data and examine the evolution of the heat and moisture budget terms during the transition periods.

Following Yanai *et al.* (1973), the *apparent heat source* Q_1 and *apparent moisture sink* Q_2 are defined as

$$Q_1 = C_p \left(\frac{p}{p_0}\right)^{\kappa} \left(\frac{\partial\theta}{\partial t} + \mathbf{V}\cdot\nabla\theta + \omega\frac{\partial\theta}{\partial p}\right) \quad (1)$$

and

$$Q_2 = -L\left(\frac{\partial q}{\partial t} + \mathbf{V}\cdot\nabla q + \omega\frac{\partial q}{\partial p}\right), \quad (2)$$

where θ is the potential temperature, q is the mixing ratio of water vapour, $\mathbf{V}$ is the horizontal velocity, ω is the vertical p-velocity, and p is the pressure. In Equation (1), $\kappa = R/C_p$, R and C_p are, respectively, the gas constant and the specific heat at constant pressure of dry air, $p_0 = 1000$ hPa, L is the latent heat of condensation, and ∇ is the isobaric gradient operator. Vertical integration of Equations (1) and (2) in the total atmospheric column yields

$$\langle Q_1\rangle = \langle Q_R\rangle + LP + S \quad (3)$$

and

$$\langle Q_2\rangle = L(P - E), \quad (4)$$

where

$$\langle\cdot\rangle = \frac{1}{g}\int_{p_{top}}^{p_s}(\cdot)\mathrm{d}p, \quad (5)$$

Q_R is the radiative heating rate, P is the precipitation rate, S is the sensible heat flux and E is the evaporation rate. In this study, the terms of Equation (1) are vertically integrated from 600 hPa to 200 hPa in order to understand the factors that contribute to the TT variations. Therefore we cannot expect a perfect balance for Equations (3) and (4). However, integrating Equation (1) from 600–200 hPa to give Equation (3) is expected to yield a close balance, because most of the latent heating occurs in the mid-troposphere. In addition, since the Tibetan Plateau acts as an elevated heat source, the vertical distribution of sensible heating in the northern box is greatest in the 600–200 hPa layer. For convenience, Equation (1) can be rewritten as

$$\langle Q_{11}\rangle = \langle Q_1\rangle - \langle Q_{12}\rangle - \langle Q_{13}\rangle, \quad (6)$$

where $\langle Q_1\rangle$, $\langle Q_{11}\rangle$, $\langle Q_{12}\rangle$ and $\langle Q_{13}\rangle$ are respectively the vertically integrated values of Q_1 and the three terms on the right-hand side of Equation (1). Hereafter the symbol $\langle\cdot\rangle$ refers to vertically integrated values in the 600–200 hPa layer.

Thus, the warming and cooling in the 600–200 hPa layer caused by horizontal temperature advection and adiabatic processes are represented by $-\langle Q_{12}\rangle$ and $-\langle Q_{13}\rangle$ respectively. Positive (negative) $-\langle Q_{12}\rangle$ corresponds to warm (cold) horizontal advection, and positive (negative) $-\langle Q_{13}\rangle$ corresponds to an adiabatic warming (cooling) due to downward (upward) motion in a stably stratified atmosphere. Our analysis is based on data spanning 54 years: a reasonably large sample for studying the IAV of monsoon. We shall first examine the different heat fluxes and heating rates and understand the climatology, before proceeding to study the extremes (early or late onset and withdrawal).

Figure 9(a) shows the horizontal distribution of the TT for the summer season, defined as 15 May to 15 October to take into account the IAV of OD and WD. A huge warm air mass is centred over southern Asia, with the maximum temperature over the southern Tibetan Plateau. This warm region is the primary source of strong temperature gradients in both the north–south and the east–west directions. This heating pattern is better understood by examining the distributions of $\langle Q_1\rangle$ and $\langle Q_2\rangle$ (Figure 9 (b) and (c)). There is intense heating over the Indian summer monsoon region, with a maximum over the Bay of Bengal, extending to the Pacific. Over the oceanic regions, large positive values of $\langle Q_1\rangle$ are accompanied by large values of $\langle Q_2\rangle$ (Figure 9(c)), indicating that the heating comes largely from the release of latent heat of condensation associated with deep cumulus convection. However, the larger values of $\langle Q_1\rangle$ compared to $\langle Q_2\rangle$ suggest the possible role of radiative, evaporative and sensible heat fluxes. Negative values of $\langle Q_1\rangle$ over the subtropical oceans are accompanied by negative $\langle Q_2\rangle$, suggesting the dominance of radiative cooling and moistening due to evaporation from the ocean surface. Li and Yanai (1996) noted that sensible heating is important over the Tibetan Plateau because large values of $\langle Q_1\rangle$ are accompanied by smaller values of $\langle Q_2\rangle$.

Most of the seasonal changes during the Indian summer monsoon occur to the north of 5 °N. To observe in detail the role and evolution of the components of the heat balance during the onset, vertically integrated values of the components are averaged over the northern box of the Indian summer monsoon region (40–100 °E, 5–35 °N). Figure 10 shows the climatological evolution of $\langle Q_1\rangle$, $\langle Q_2\rangle$, and the components $\langle Q_{11}\rangle$, $-\langle Q_{12}\rangle$ and $-\langle Q_{13}\rangle$,

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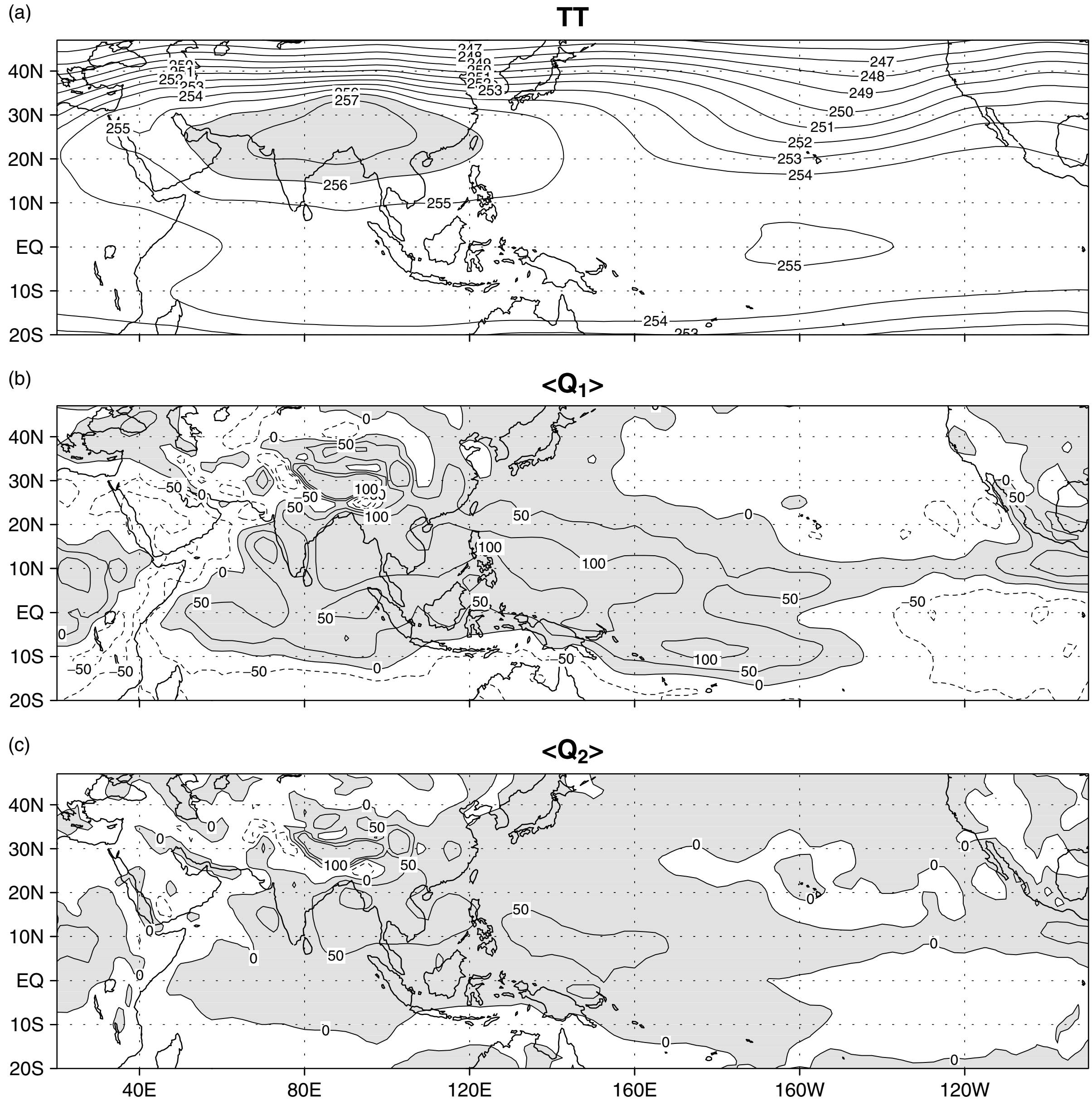

Figure 9. (a) Climatological mean TT (K), (b) vertically integrated heat source (Wm^{-2}), and (c) vertically integrated moisture sink (Wm^{-2}), for the summer monsoon season.

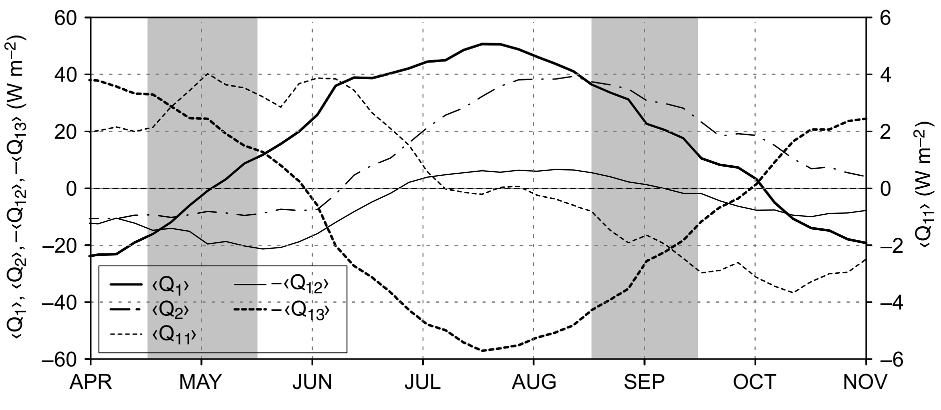

Figure 10. Climatological evolution of $\langle Q_1 \rangle$, $\langle Q_2 \rangle$, $-\langle Q_{12} \rangle$, $-\langle Q_{13} \rangle$ (scale on the left) and $\langle Q_{11} \rangle$ (scale on the right), averaged over the northern box (40–100 °E, 5–35 °N). The grey shaded areas indicate the time span considered for the budget calculations.

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over the northern box throughout the year. During winter and spring, $\langle Q_2\rangle$ is negative and slowly increasing, suggesting stronger evaporation than precipitation. Soon after the onset, it starts to increase faster, but the evaporation continues to dominate because of the warm landmass, and about 2 pentads after the onset, precipitation overcomes evaporation. So the moisture sink is not a leading process during the first onset or at the time of setting up of the Indian summer monsoon.

A remarkable feature is the cancellation between the heat source and the adiabatic cooling during the monsoon months. However, the significant temperature increase that is observed during the pre-monsoon arises largely from adiabatic warming, sensible heating over the Tibetan Plateau and vertical mixing. In the case of withdrawal, there is coherent evolution of the climatological values of $\langle Q_1\rangle$ and $\langle Q_2\rangle$. It may be noted that, unlike in the case of onset, there is no change of sign of $\langle Q_2\rangle$ during the withdrawal. This has some important implications, which justify our TT-based definition of the seasons. Moisture is constantly being precipitated out even after the withdrawal. There are several reasons for this. The onset of the northeast monsoon, which takes place in October, and the post-monsoon cyclone season, contribute to the precipitation. Moreover, the cold land surface after the summer monsoon slows down evaporation. Interestingly, during the withdrawal phase of the monsoon, the cooling trend of TT seems to originate from cold horizontal advection.

The climatological features discussed above should help us to understand the mechanism governing the interannual variations of OD and WD, by comparing the differences in the features associated with onset and withdrawal. The following sections discuss this.

7.1. Onset

In order to gain insight into the factors that lead to early and late onsets, Figure 11 compares composites of heat budget terms during the period from 15 April to 15 May for early and late onsets with those for the climatological onsets. The period from 15 April to 15 May is considered to be ahead of the onset, so that the heat-flux terms during this period are not influenced by the post-onset processes. During an early onset, $\langle Q_1\rangle$ is larger than the climatological value and is mostly a result of the sensible heating, as is evident from the small value of $\langle Q_2\rangle$ which suggest the absence of latent heating (Equation (4)). Reduced heating due to adiabatic subsidence, and proportionally reduced horizontal cold air advection, help the vertical mixing of sensible heating at the surface and thus lead to a large value of $\langle Q_1\rangle$ and a higher rate of heating during an early onset phase. On the other hand, a late onset has comparatively lower values of $\langle Q_1\rangle$ and $\langle Q_{11}\rangle$. Strong warming due to enhanced adiabatic subsidence during a late onset has the indirect effect of decreasing $\langle Q_{11}\rangle$ by suppressing the vertical mixing, even though cold horizontal advection is enhanced. The role of vertical mixing in regulating $\langle Q_1\rangle$ is shown in Figure 12, which plots the vertical profile of $\langle Q_1\rangle$ averaged over the same region, for the early- and late-onset composites. It is evident that during a late onset, the $\langle Q_1\rangle$ in the 600–200 hPa layer is nearly zero, whereas it is significant during an early onset. Vertical mixing, which is a major mechanism of heat transport to the upper troposphere during an early onset, is directly or indirectly suppressed by the adiabatic subsidence prevalent during a late onset.

7.2. Withdrawal

A similar analysis helps us to understand the features of early and late monsoon withdrawals. Figure 13 shows the heat budget terms for early and late withdrawals for the period from 15 August to 15 September. The processes that drive the withdrawal are quite different from those that control the onset. During the withdrawal, $\langle Q_1\rangle$ is largely driven by the latent heating (as indicated by the large values of $\langle Q_2\rangle$); it is nearly balanced by the adiabatic cooling due to vertical motion, and to some extent by cold horizontal advection. An early withdrawal is characterized by a weaker-than-usual $\langle Q_1\rangle$ and a more strongly decreasing tendency of TT ($\langle Q_{11}\rangle$), while a late withdrawal is characterized by a stronger-than-usual $\langle Q_1\rangle$ and a less strongly decreasing tendency of TT. The more strongly decreasing tendency of TT during an early withdrawal, compared with a late withdrawal, is primarily attributed to the stronger cold horizontal advection during early withdrawal and its near absence during a late withdrawal. Thus, unlike in the case of onset (Figure 11), the larger values of $\langle Q_2\rangle$ suggest the

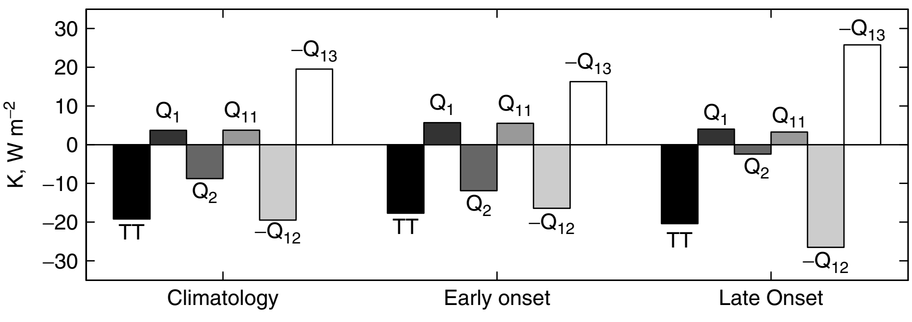

Figure 11. TT (in K) and $\langle Q_1\rangle$, $\langle Q_2\rangle$, $\langle Q_{11}\rangle$, $-\langle Q_{12}\rangle$ and $-\langle Q_{13}\rangle$ (in Wm^{-2}), averaged over (40–100 °E, 5–35 °N) and averaged from 15 April to 15 May, for climatology, early-onset and late-onset composites.

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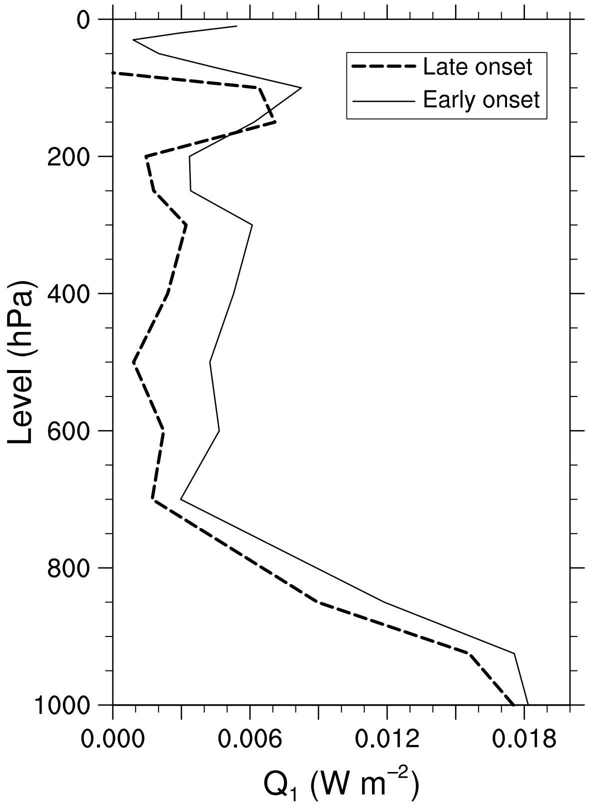

Figure 12. Vertical profiles of Q_1, averaged over (40–100 °E, 5–35 °N) and averaged from 15 April to 15 May, for early-onset and late- onset composites.

role played by the moisture during the withdrawal. In short, the run-up to the onset is a dry process, and its advancement or delay is determined mainly by the vertical advection or subsidence, whereas it is the moist processes that lead to the withdrawal, and the horizontal advection determines whether it is early or late.

Having diagnosed the features and the contribution of different sources of heating for late and early onset and withdrawal, our next goal is to understand the external factors that control the interannual variations of the onset and withdrawal.

7.3. ENSO influence on onset and withdrawal

The proposed modulation of the Indian summer monsoon by the ENSO is examined in the light of the factors that contribute to the observed variations of TT, which expand or shrink the monsoon season. Here we examine the role of ENSO in regulating the length of the season from the heat budget terms during eight El Niño years (1957, 1965, 1972, 1979, 1982, 1987, 1992 and 1997) and seven La Niña years (1955, 1964, 1970, 1973, 1975, 1988 and 1999) (Figure 14). The balance of terms for the period 15 April to 15 May for El Niño and La Niña composites are shown in Figure 14(a). There is a close similarity between the budgets of an early-onset composite (Figure 11) and the La Niña composite, with relatively lower adiabatic heating due to large-scale subsidence, favouring vertical mixing of the sensible heating at the surface, as discussed above. The striking similarity between an El Niño and a late-onset composite (Figure 11) is strong evidence for our proposed mechanism, where the reduced tropospheric heating is due to the large-scale subsidence that subdues vertical mixing. There are no large differences in the heating due to horizontal advection between El Niño and La Niña composites; this suggests that the influence of ENSO on $\langle Q_1\rangle$ and the TT tendency ($\langle Q_{11}\rangle$) during the onset phase is through the modulation of vertical advection.

Similarly, during the withdrawal phase, there is a strong correspondence between the budgets of the La Niña composite and the late-withdrawal composite, and between those of the El Niño composite and the early-withdrawal composite (Figures 14(b) and 13). The mechanism underlying the interannual variations of the withdrawal discussed above seems basically to arise from the influence of ENSO. During the El Niño, convective heating is subdued, and thus adiabatic cooling is reduced over the region. A stronger and persistent cold horizontal advection contributes to a faster decay of TT, and thus an early withdrawal of the monsoon. During La Niña, there is enhanced vertical motion over the region, and thus enhanced latent heating (due to greater availability of moisture), which maintains the heat source until the middle of October, contributing to a late withdrawal of the monsoon. As we have seen already for the late withdrawal, horizontal advection is almost negligible during La Niña. In short, El Niño delays the onset by inducing enhanced subsidence, persistent during the pre-onset phase, and advances the withdrawal by intensifying the horizontal cold air advection. Both these processes occur through the large-scale quasi-stationary planetary-scale waves triggered in the atmosphere by the tropical heating associated with the El Niño.

8. Discussions and conclusions

- Although the duration of the physical Indian summer monsoon season can vary from one year to another because of variation in the external and internal forcing factors, it has traditionally been defined as the period from 1 June to 30 September. The IAV of rainfall for a fixed season, and its connection with other climatic parameters, could misrepresent the real physical processes involved in the teleconnection. For the first time, an objective definition of the Indian summer monsoon season is provided. The definition stems from the recognition that the classical land–ocean surface temperature theory is inadequate to explain the deep baroclinic structure of the monsoon circulation and that the Indian summer monsoon is a result of a deep tropospheric off-equatorial heat source. The meridional gradient of tropospheric temperature represents the cross-equatorial pressure gradient that drives the monsoon circulation. The beginning (onset) and end (withdrawal) of the Indian summer monsoon season

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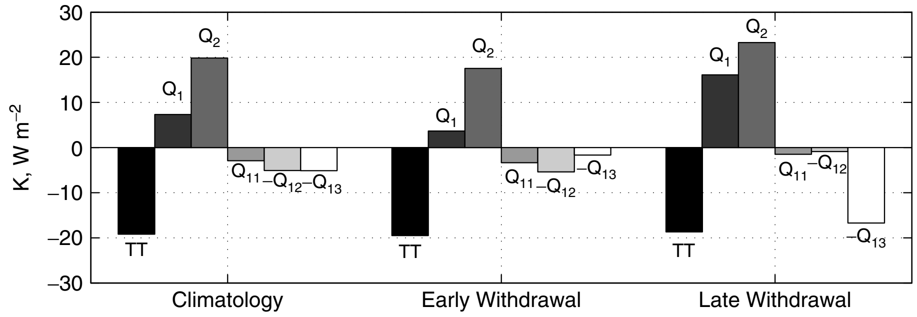

Figure 13. As Figure 11, but averaged from 15 August to 15 September, for climatology, early-withdrawal and late-withdrawal composites.

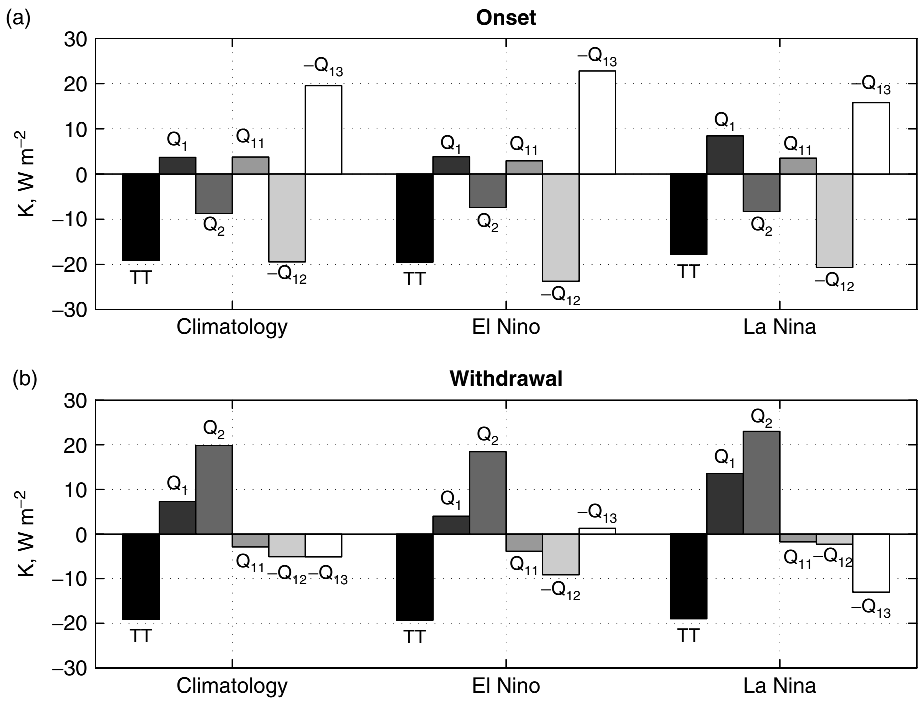

Figure 14. (a) TT (in K) and $\langle Q_1\rangle$, $\langle Q_2\rangle$, $\langle Q_{11}\rangle$, $-\langle Q_{12}\rangle$ and $-\langle Q_{13}\rangle$ (in Wm^{-2}), averaged over (40–100 °E, 5–35 °N) and averaged from 15 April to 15 May, for climatology, La Niña and El Niño composites. (b) As (a), but averaged from 15 August to 15 September.

are therefore associated with the times when the meridional gradient of TT (defined as temperature averaged between 600 hPa and 200 hPa) over the Indian summer monsoon region first becomes positive and later becomes negative. The onset and withdrawal dates are defined as the dates when the difference in TT (denoted ΔTT) between a northern box (40–100 °E, 5–35 °N) and a southern box (40–100 °E, 15 °S–5 °N) changes from negative to positive and from positive to negative, respectively. The thermodynamic conditions associated with the change of sign of ΔTT from negative to positive set up conditions for triggering symmetric instability, large-scale organized convection and the dynamic monsoon onset. A strong correlation between the OD obtained through our definition and one based on the KE of the low-level winds over the low-level jet region supports the proposed mechanism. These thermodynamic conditions also set up the required easterly shear of zonal winds for sustained northward propagation of the TCZ (the rain band). Similarly, the change of sign of ΔTT from positive to negative indicates the end of the southwest monsoon season, as the tropospheric heat source is then located south of 5 °N, with northeasterlies replacing southwesterlies at low levels. This also represents the time when the easterly shear falls below the critical level for supporting northward propagation of the rain band. The difference between WD and OD gives the LRS. Goswami and Xavier (2005) established the consistency of these onset and withdrawal definitions with ERA40. This confirms that the values of ΔTT during the monsoon onset period do not depend strongly on the physical parametrizations in the forecast models used in the reanalysis.

- The climatological LRS is about 10 days longer than the traditional JJAS definition of the season. The amplitude of the IAV of the LRS (12 days) is larger than that of either the OD (7 days) or the WD (9 days), and therefore contributes to the IAV of the seasonal mean monsoon rainfall. We find that the ENSO influences monsoon rainfall by modulating the

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LRS. By forcing a late onset and an early withdrawal, an El Niño shrinks LRS and monsoon rainfall; a La Niña does the opposite. The correlation between LRS and an index of ENSO has remained strong even during recent years, when the correlation between the AIR and ENSO indices has become insignificant. Thus, a part of the apparent recent weakening of the ENSO–monsoon relationship is simply due to keeping the season fixed. However, a part of the weakening may be real, caused by changes in the internal dynamics (for example, the intensity of rain spells within the season).

- As positive values of ΔTT define the Indian summer monsoon, it is reasonable to suppose that the cumulative value of positive ΔTT may represent the strength of the seasonal mean monsoon. The cumulative ΔTT thus provides a new thermodynamic index of the Indian summer monsoon, denoted by TISM. This index is found to correlate strongly with rainfall indices of the Indian summer monsoon, such as AIR. The correlation between TISM and ENSO indices is very strong in the earlier years of the dataset, and is fairly strong even in recent years. This representation of the Indian summer monsoon thus seems to be more predictable than rainfall representations such as AIR.
- An El Niño (La Niña) decreases (increases) the LRS and the strength of the Indian summer monsoon, through decrease (increase) of TISM, by setting up a persistent negative (positive) TT anomaly over northern India and the southern Eurasian region. Such a TT anomaly can result from stationary waves generated by tropical atmospheric heating over the Pacific during an El Niño (La Niña) event (Rodwell and Hoskins, 1996; Hui *et al.*, 2003). This is a hitherto unrecognized pathway through which ENSO can influence the Indian monsoon. This finding opens up the possibility that any phenomenon giving rise to a substantial TT anomaly over southern Eurasia could influence the Indian monsoon. For example, a positive North Atlantic Oscillation phase during northern summer may lead to a strengthening of the Indian monsoon through a positive TT anomaly over southern Eurasia, thanks to temperature advections caused by associated storm-track changes (Goswami *et al.*, 2006). Some connections between North Atlantic SST and the Indian monsoon noted in palaeoclimate proxies (Burns *et al.*, 2003; Gupta *et al.*, 2003) and on interdecadal time-scales (Goswami *et al.*, 2006) can now be understood through the mechanism described here.
- Diagnostic analysis of the heat and moisture budgets is carried out in order to gain some insight into the factors responsible for early and late onset and withdrawal of the Indian summer monsoon, which in turn determine the LRS. The latent heat release due to convection becomes an important factor in the heat budget only after the onset, during the summer monsoon season. While the heat budget during the run-up to the thermodynamic onset is essentially governed by dry processes, moist processes dominate the heat budget of the withdrawal phase. Vertical advection (associated with subsidence) plays a crucial role in determining whether the onset will be early or late. On the other hand, horizontal advective cooling plays an important role in determining whether the withdrawal will be early or late. It is found that most of the early (late) monsoon onsets are associated with La Niña (El Niño), while most of the early (late) withdrawals are associated with El Niño (La Niña). The vertical and horizontal advections associated with the stationary waves forced by El Niño (La Niña) over northern India and southern Eurasia influence the onset and withdrawal of the Indian summer monsoon.
- We note that our TISM has a downward trend between 1950 and 2003 (Figure 8(a)). Does this mean that the strength of the Indian summer monsoon has been decreasing since 1950? Alarm bells to this effect have been sounded in some recent studies (Joseph and Simon, 2005; Bingyi, 2005), where a downward trend

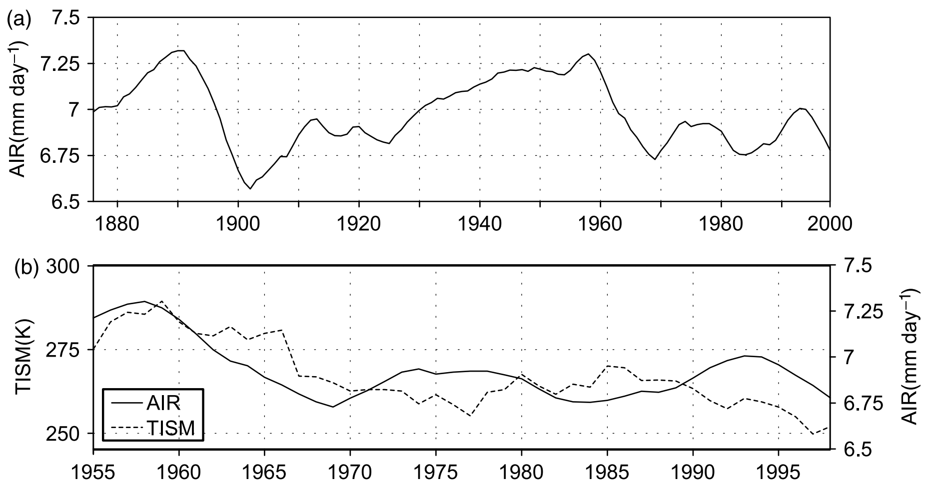

Figure 15. (a) 11-year running mean of JJAS AIR for the period 1876–2000. (b) 11-year running means of JJAS AIR (solid line) and TISM (dashed line) for the period 1955–1998.

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is observed between 1950 and 2003 in certain parameters correlated with the Indian summer monsoon. Over a longer period, the strength of the Indian summer monsoon as deduced from records of AIR has a multi-decadal oscillation with an approximate period of about 60 years, but no long-term trend (Figure 15(a); see also (Goswami, 2005a)). A positive phase of the multi-decadal oscillation between 1930 and 1965 has been followed by a negative phase during the recent three decades. The low-pass-filtered AIR and other indices of the Indian summer monsoon (such as TISM) are consistent between 1950 and 2003 (Figure 15(b)). Thus what appears as a trend when examined only between 1950 and 2003 is actually a part of a multi-decadal oscillation.

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