Role of Interactions between Aerosol Radiative Effect, Dynamics, and Cloud Microphysics on Transitions of Monsoon Intraseasonal Oscillations

ANUPAM HAZRA AND B. N. GOSWAMI
Indian Institute of Tropical Meteorology, Pune, India

JEN-PING CHEN
Department of Atmospheric Sciences, National Taiwan University, Taipei, Taiwan

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ABSTRACT

Extended-range prediction of monsoon intraseasonal oscillations (MISOs), crucial for agriculture and water management, is limited by their event-to-event variability. Here, the authors propose a hypothesis supported by a number of model simulations involving detailed cloud microphysical processes indicating that aerosols contribute significantly to the transitions from “break” to “active” phases of MISO. The role of aerosol indirect effect in the process of invigoration of precipitation is demonstrated with a high-resolution regional model for Indian summer monsoon breaks that are followed by an active condition (BFA) and contrasted with breaks that are not followed by an active condition (BNFA). The BFA are characterized by higher concentrations of absorbing aerosols that lead to a stronger north–south low-level temperature gradient and strong moisture convergence. Forced uplift beyond the freezing level initiates the cold-rain process involving mixed-phase microphysics and latent heat release at higher levels, thereby invigorating convection, enhancing precipitation, and resulting in an active condition. While more aerosols tend to reduce the cloud drop size and delay the warm rain, it is overcome by the higher moisture convergence during BFA and invigoration by cold-rain processes. The net production of rainfall is sensitive to cloud structure as it depends on the relative strength of the warm- and cold-rain initiation processes. The results indicate the importance of aerosols on transitions of MISO and a pathway by which they influence the transitions involving complex interactions between direct radiative forcing, large-scale dynamics, and cloud microphysics. Broader implications of these results in event-to-event variability of MISO and its predictability are also highlighted.

1. Introduction

As building blocks of the Indian summer monsoon (ISM), the subseasonal monsoon intraseasonal oscillations (MISOs) influence the seasonal mean monsoon and its interannual variability as well as cluster synoptic disturbances in its active phases [see Goswami (2012) for a review]. We may recall that MISOs have large spatial scale and are associated with repeated northward propagation of the tropical convergence zone (TCZ; Goswami 2012). Also during active (break) phases, it is associated with a dipolelike structure in the north–south with positive (negative) rainfall anomalies over central India (equatorial Indian Ocean). With large amplitude associated with them (Goswami et al. 2011), it has been demonstrated that MISOs have high potential predictability (Waliser et al. 2003; Goswami and Xavier 2003; Fu et al. 2007). Prediction of the wet active spells and the dry break spells of MISO more than 2 weeks in advance is of great importance to the farmers as it affects the agricultural planning and water resource management. While the underlying periodicity provides basis for predictability of MISOs, the event-to-event variability arising from nonlinearity and other feedbacks leads to a limit on predictability of MISO. Therefore, it is important to understand the feedbacks and pathways through which the event-to-event variability of MISO is generated.

The MISO is an instability arising from a feedback between radiation, large-scale dynamics, and convection (Goswami and Shukla 1984; Wang 2012). The existence of the fundamental oscillatory character, its periodicity,
and the northward-propagating character of the MISO may not need aerosols. However, can aerosols influence any of these characteristics of MISO through some feedback? During wet active spells buildup of aerosols in the atmosphere and associated direct radiative cooling/warming is restricted because of heavy washout. During a dry break phase, buildup of aerosols may take place and depending on whether they are scattering type or absorbing type, they could either cool the surface only or cool the surface and warm the lower atmosphere. When the northern Indian surface cools, it not only makes the atmosphere locally stable, it also inhibits surface moisture convergence from oceans as north–south surface temperature gradient is weakened. Both these processes would try to delay the transition to an active phase. In the other case of absorbing aerosols, while the surface may still cool, warming of the lower atmosphere can lead to a positive north–south temperature gradient attracting low-level convergence of moist air to central India and facilitate invigoration of convection and thereby transition to an active condition. In a recent diagnostic study investigating differences between breaks that are followed by an active condition (BFA) and breaks that are not followed by an active condition (BNFA), Manoj et al. (2011) found that the BFA cases were associated with a 3-times-larger amount of absorbing aerosols over central India than the BNFA cases. They also showed from reanalysis data that the BFA (BNFA) cases were associated with much stronger (weaker) lower-atmosphere north–south temperature gradient followed by much stronger (weaker) moisture convergence to central India. Their results indicate that the large-scale upward motion caused by moisture convergence overcomes the local stability because of low-level atmospheric warming and surface cooling and leads to transition to an active condition in the BFA cases while it is not possible in the BNFA cases. This study demonstrated that aerosols do influence the observed MISO and hence inclusion of aerosol transport may be important in real-time prediction models of MISO.

While Manoj et al. (2011) well documented the role of absorbing aerosols in modifying the large-scale circulation and moisture transport through direct radiative effect leading to transition from a break to an active condition, they did not address the role of indirect effect of the aerosols in this process of invigoration. Much higher aerosol concentrations during BFA cases compared to BNFA cases may lead to higher concentration of cloud condensation nuclei (CCN) and smaller cloud drops, reducing efficiency of collision coalescence process to form bigger rain drops and delay in warm-rain process (Tao et al. 2007; Altaratz et al. 2008; Kulkarni et al. 2012). Indeed, the delay in warm-rain formation is noted over “dirty” environments during the Cloud Aerosol Interaction and Precipitation Enhancement Experiment (CAIPEEX) over India (Kulkarni et al. 2012; Konwar et al. 2012). How does the invigoration of precipitation take place during BFA cases even though they are dirtier compared to BNFA cases? What is role of cloud microphysics in this process? A framework for invigoration of convection in a polluted environment has been described in Rosenfeld et al. (2008). Building on the conceptual framework proposed by Rosenfeld et al. (2008), in the present study, we propose a hypothesis regarding how invigoration takes place in BFA and not in BNFA in the background of large-scale monsoon environment and demonstrate the veracity of the hypothesis using a series of experiments with a high-resolution regional climate model.

The hypothesis is illustrated in the schematic Fig. 1. We recognize that all breaks are generally dry. However, higher concentration of aerosols make the environment dirty during BFA cases while lesser aerosol concentrations make the environment relatively “clean” during the BNFA cases. The key to our hypothesis is based on the observation by several studies (Andreae et al. 2004; Lin et al. 2006; Rosenfeld et al. 2008) that enhancement of precipitation can occur if “mixed-phase precipitation” is generated even in a dirty environment. Our hypothesis finds support from Moderate Resolution Imaging Spectroradiometer (MODIS) observations where we note that ice-phase processes were important in transitions from break to active during BFA cases as compared to BNFA cases. By following Manoj et al. (2011), the dates of BFA and BNFA cases were identified between 2000 and 2009 and composite cloud ice water path during BFA days and BNFA days were calculated from MODIS, the difference of which is shown in Fig. 2a. It is clear that there is significant difference in ice water path during the two cases over central India. Averaged over 15°–25°N, 72°–85°E this is 165.3 g m\(^{-2}\) during BFA compared to 130.1 g m\(^{-2}\) during BNFA. Thus, mixed-phase processes including the ice phase seem to be crucial in the transitions from break to active during BFA. Over the Bay of Bengal and near the Myanmar coast and in some parts of the Indian Ocean, more cloud ice has been noticed during BFA compared to BNFA. These total ice-phase hydrometeors further amplify the instability and large-scale dynamics through latent heat release. In the BFA cases, even though the cloud drop size is reduced (Fig. 2b), the large-scale convergence and increased instability because moisture convergence lifts these drops above the freezing level releasing latent heat at higher levels and leading to further invigoration of low-level convergence. A related parameter is cloud drop effective radius and MODIS observation (Fig. 2b)
indicates that the cloud drop effective radius is smaller during BFA cases, particularly over central India (15°–25°N, 72°–85°E), compared to BNFA cases (BFA – BNFA = −2.04 μm), consistent with higher cloud drop number density during BFA cases compared to the BNFA cases. Although there are several other factors like giant CCN, and ice nuclei (IN) that can also control the cloud drop size, the higher CCN concentration during BFA cases due to larger aerosol concentration seems to be the main reason in this case.

In the present study, we test this hypothesis using a high-resolution regional model over the Indian monsoon region. While such invigoration of convection by aerosols has been demonstrated using regional models (Fan et al. 2012) and observation (Bell et al. 2008) in other regions, such a modeling study has not been carried out over the Indian monsoon region. As shown by Fan et al. (2012) such invigoration can be significantly influenced by the vertical shear of background winds and the radiative forcing in the atmosphere, which could be significantly different from one region to another. The model used, design of experiments, and validation of the model against observations are described in section 2 and the results are described in section 3. The main conclusions are summarized in section 4.

2. Model framework

a. General description

To test our hypothesis on how aerosols influence transitions of MISO breaks, a regional model is needed that includes effect of aerosols on cloud microphysics and rainfall formation. As such a model is not readily available, we undertook a major activity to modify the nonhydrostatic fifth-generation Pennsylvania State University–National Center for Atmospheric Research Mesoscale Model (MM5) to include the two-moment warm-cloud microphysical process of Chen and Liu (2004, called the CL scheme). This scheme has been demonstrated to adequately reflect the effect of cloud condensation nuclei (CCN) by Cheng et al. (2007, 2010). It is always a challenge to simulate suppressed monsoon convection events with any mesoscale model. For example, Chakravarty et al. (2011) successfully simulated active monsoon convection using the Weather Research and Forecasting Model (WRF) but failed to simulate the suppressed monsoon convection event. Ratnam and Kumar (2005) examined the sensitivity of MM5 in simulating the Indian monsoon climate with different cumulus parameterization schemes when the parameterization schemes for all other physical processes remain unchanged. Two contrasting monsoon years [1987 (dry year) and 1988
(wet year) were simulated by them. For the initial and lateral boundary conditions for the model the 6-hourly National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data were used. They concluded that the model is able to simulate the large-scale features and interannual variability of the monsoon with reasonable accuracy with all cumulus schemes. This background helped us to select the regional model for our study.

The CL scheme consists of a set of bulk formulas for masses (the third moment) and number concentrations (the zeroth moment) of liquid water condensates. These formulas are derived based on statistical analyses of results from the binned cloud microphysics of Chen and Lamb (1994). In addition, the CL scheme also provides diagnostic equations to calculate the terminal velocities and the effective radius of condensates, which are critical to precipitation process and radiation heating/cooling, respectively. This scheme requires description of CCN, which are assumed to compose of ammonium sulfate and maintain a trimodal lognormal size distribution; and their activation into cloud drops follows the Köhler theory. The Köhler-curve critical radius, which depends on the degree of supersaturation, of the last time step is retrieved from prognostic aerosol mass in the air and in the condensates. When the Köhler-curve critical radius of the present time step is smaller than that of the previous time step, CCN with radius values in between are activated. The tracking of aerosol masses inside clouds and inside precipitation also allows for aerosol recycling from the evaporation of cloud drops. The CL scheme also considers the creation of rain embryos directly from giant CCN, so autoconversion is not the only mechanism for warm-rain production (cf. Cheng et al. 2007).

The CL scheme was further coupled with the ice-phase parameterizations of Reisner et al. (1998) by Cheng et al. (2010), and is named as the CLR scheme. The ice microphysics of Reisner et al. (1998) includes three ice classes: cloud ice, snow, and graupel/hail. Their initiation, growth by vapor deposition, and riming, as well as the interchange between classes, are fully described. Some of these processes, including ice nucleation and snow melting, as well as size-dependent riming efficiency, were modified from the original scheme (cf. Cheng et al. 2010). The equation of DeMott and Roger (1990), based on results of laboratory measurement, is used to calculate homogeneous-freezing nucleation rate at temperatures lower than $-34^\circ$C, allowing liquid cloud drops to freeze without the help of ice nuclei. Although the CLR scheme has the capability of simulating IN effects, we applied implicit heterogeneous nucleation formulas of Huffman (1973) and Cooper (1986) for simplicity so as to focus on CCN.

Recently a major field experiment called CAIPEEX Phase I has been conducted in India by the Indian Institute of Tropical Meteorology (IITM) during the period of May–September 2009 (http://www.tropmet.res.in/~caipeex/). CAIPEEX Phase I in situ measurements revealed that the concentration of aerosols was in the range of 1000–2000 cm$^{-3}$ in June at Hyderabad, smaller in number ($\sim$1000 cm$^{-3}$) in July at Bangalore, and high (2000 cm$^{-3}$) in August at Bareilly (Prabha et al. 2011; Kulkarni et al. 2012). For the sensitivity studies of CCN, we selected two types of aerosol trimodal size distribution (i.e., the nucleation mode, accumulation mode, and coarse mode) from Whitby (1978) and verified with CAIPEEX Phase I measurements (Prabha et al. 2011; Kulkarni et al. 2012): 1) clean continental (total number of aerosols is 1800 cm$^{-3}$) and 2) dirty (or urban) type (total number of aerosols is 13 800 cm$^{-3}$). Given the aerosol concentrations used in this present study, CCN concentrations are calculated in the model through activation. For CCN activation, the minimum size of dry aerosol to be activated depends on supersaturation according to Köhler equation [details are available in Cheng et al. (2007, 2010)]. Initial CCN are initialized
with these size distributions at the surface and its concentrations are assumed to decrease exponentially with a scale height of 3.57 km in the vertical (Cheng et al. 2010), except below 850 hPa or within the lowest three sigma levels where there is no vertical variation (Cheng et al. 2010).

In the regional model, four nested domains were employed with horizontal grid spacing of 81, 27, 9, and 3 km, respectively (Fig. 3). The NCEP Final Analyses (FNL) data in $1° \times 1°$ horizontal resolution are used to provide initial and lateral boundary conditions for our simulations. To avoid discrepancy in large-scale dynamics that may propagate into the target areas, a controlled simulation is first run for the two outer domains. The results are then used as boundary conditions to drive the two inner domains for sensitivity tests.

b. Design of experiments

The objective of the model experiments is to address the following question. Given the appropriate large-scale environment (boundary forcing for the regional model), does the local aerosol condition over central India lead to much larger precipitation in the BFA cases as compared to the BNFA cases? Following Manoj et al. (2011, see their Fig. 3), three BFA cases and three BNFA cases were identified (Table 1). For all the cases the model integrations are conducted for 5 days. To bring out the role of enhanced aerosol loading, for each case two integrations are carried out one with dirty environment and another with clean environment. List of all these experiments and physical processes are summarized in Table 1. Composite results of three BFA and BNFA cases are presented for 120 h (5 days).

![Fig. 3. The domain (four-level nested) used for simulating the mixed-phase cloud system during ISM breaks. Three grid resolutions are adopted: 81, 27, 9, and 3 km, respectively: innermost domain resolution: 3 km × 3 km. Domain 1: 10.3°–33.2°N, 68.8°–92.2°E; domain 2: 13.6°–26.4°N, 72.2°–86.2°E; domain 3: 17.5°–24.2°N, 75.8°–82.8°E; and domain 4: 19.5°–22.3°N, 78.2°–81.2°E.](http://journals.ametsoc.org/doi/pdf/10.1175/JAS-D-12-0179.1)

### Table 1. The physical parameterization schemes and options used for the present sensitivity experiments.

<table>
<thead>
<tr>
<th>Model aspect</th>
<th>Setting</th>
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<tbody>
<tr>
<td>Model’s name and grid</td>
<td>Meteorological model MM5</td>
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<td></td>
<td>Four nested domains were employed with horizontal grid spacing of 81 km (10.3°–33.2°N, 68.8°–92.2°E), 27 km (13.6°–26.4°N, 72.2°–86.2°E), 9 km (17.5°–24.2°N, 75.8°–82.8°E), and 3 km (19.5°–22.3°N, 78.2°–81.2°E) (Fig. 3). Innermost domain resolution: 3 km × 3 km, 31 vertical sigma levels.</td>
</tr>
<tr>
<td>Initialization</td>
<td>NCEP FNL data in $1° \times 1°$ horizontal resolution are used to provide initial and lateral boundary conditions for our simulations.</td>
</tr>
<tr>
<td>Simulation duration</td>
<td>120 h (5 days)</td>
</tr>
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<td></td>
<td>CL scheme: two-moment warm-cloud microphysical scheme (Chen and Liu 2004); cloud drop mass and number, raindrop mass and number. Includes aerosol effects: CCN activation into cloud drops, aerosol recycling, and giant CCN activation into rain embryo.</td>
</tr>
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<td></td>
<td>Ice microphysics of Reisner et al. (1998) with three ice classes: cloud ice, snow, and graupel/hail; modified ice nucleation; modified snow melting</td>
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<tr>
<td>Convection scheme</td>
<td>Grell cumulus parameterization (Grell 1993).</td>
</tr>
<tr>
<td>Radiation scheme</td>
<td>Cloud radiation scheme (Grell et al. 1994).</td>
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<td>Surface scheme</td>
<td>Five-layer soil model (Dudhia 1996).</td>
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<tr>
<td>PBL scheme</td>
<td>MRF (Hong and Pan 1996).</td>
</tr>
<tr>
<td>Aerosol prescription</td>
<td>1) Clean continental: total number of aerosols is 1800 cm$^{-3}$.</td>
</tr>
<tr>
<td></td>
<td>2) Dirty (or urban): total number of aerosols is 13 800 cm$^{-3}$.</td>
</tr>
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</table>
c. Model performance in simulating aerosol indirect effect

As aerosol indirect effect is key to our hypothesis of invigoration of precipitation during BFA cases, it is important to establish the fidelity of the model in simulating this aspect. To have some confidence in our model results, the performance of CLR scheme that deals explicitly with the aerosol–cloud interaction is evaluated with some observations during ISM breaks.

First, we examine simulation of some gross microphysical properties by the model. In Fig. 4, we present the observed cloud liquid water path from MODIS (King et al. 2003) averaged longitudinally between 78.2° and 81.2°E and compare with model-simulated vertically integrated cloud water during BFA and BNFA cases. During BNFA, the model simulates the meridional distribution of cloud liquid water path reasonably well with a minimum around 20.4°N and maximum in the northern side being higher than that in the southern side. However, the model has a tendency to produce higher cloud liquid water everywhere. During BFA cases, while the model simulates the minimum in the middle and maxima on northern and southern sides correctly, it simulates the northern maximum to be higher than the southern one in contrast to the observations. Considering the fact that the remotely sensed cloud liquid water derived from MODIS itself may have some uncertainty (King et al. 2013; Kittaka et al. 2011), the model simulates the distribution of cloud liquid water reasonably well. 

Next, we attempt to validate the simulated cloud drop number concentration (CDNC) with observations. For this purpose, the results of model-simulated CDNC are compared with that obtained during CAIPEEX campaigns over two sites in India (15 June 2009 event, around Hyderabad and 19 August 2009 event, around Bareilly) (Fig. 5). With the limited observations available, it is seen that the model simulates the number concentration and vertical distribution of cloud drops in the lower layers during the August event in Bareilly well (Fig. 5b). During the June event in Hyderabad, while the CDNC over the middle atmosphere is simulated reasonably well by the model, it simulates some vertical structure not seen in the specific event (Fig. 5a). To get a sense of how the model simulates the variability of cloud drop number concentration and size (effective radius) under different thermodynamic and aerosol conditions, we show in Figs. 5c and 5d respectively model simulations of CDNC and cloud drop effective radius in three different regions, namely (i) the event on 27 October 2010 over Bay of Bengal (BoB) (27102010-BoB), (ii) the event on 5 July 2009 over the Arabian Sea.
(05072009-Arabian Sea), and (iii) the event on 24 August 2009 over the Indo-Gangatic Plain (IGP) (24082009-IGP) over the northern part of India. While the model-simulated results are comparable to CAIPEEX observation (Konwar et al. 2012, see their Fig. 3), the spatial variability is also captured reasonably well. The cloud drop number concentration also has been captured well as compared with CAIPEEX (Nair et al. 2012, see their Fig. 2). To provide a quantitative measure of the model’s fidelity, simulated cloud drop effective radius is compared with observation for the case of 27 September 2010 (27092010-BoB) over the Bay of Bengal (Fig. 5e). This figure shows that the model does a good job simulating the cloud effective radius up to about 5 km while the model may have a bias of simulating smaller effective radius at higher heights. These findings show the model with CLR scheme can simulate these microphysical properties reasonably well.

To test the fidelity of the model in simulating the first and second indirect effects, we show in Figs. 6 and 7 simulated vertical profiles of domain-averaged cloud fields (cloud water mixing ratio, cloud drop number, and cloud drop size) as a function of time, under clean and dirty aerosol conditions during both the break events. It is noteworthy that the features of the first indirect effect (smaller cloud drop size) and second indirect effect (more cloud water; Twomey et al. 1984; Albrecht 1989) are well simulated in both BFA (Fig. 6) and BNFA (Fig. 7) cases. As the number of cloud drops increase, cloud drop size decrease from clean to urban (dirty) conditions owing to the difference in CCN number concentration (Figs. 6c–f and 7c–f). These results are also in good agreement with many observations (e.g., Rosenfeld 1999, 2000; Twomey et al. 1984; Albrecht 1989) are well simulated in both BFA (Fig. 6) and BNFA (Fig. 7) cases. Although the present bulk-microphysical scheme (relatively simpler than most detailed bin-resolving cloud schemes) cannot simulate the changes in the size distributions nor can it execute more precise collision calculations as in the bin-resolving cloud schemes, the
performance of the bulk-formula two-moment cloud scheme is quite good and certain features of CCN effects are reasonably simulated (e.g., Khain et al. 2005; Teller and Levin 2006; van den Heever and Cotton 2007; Cheng et al. 2007, 2010). For instance, a comparison of cloud properties by a bin-resolving cloud scheme and a two-moment warm-cloud microphysical scheme by Morrison and Grabowski (2007) indicated that the features of the indirect effects can be well simulated by the two-moment scheme although the magnitudes of the indirect effects may be different from that simulated by the bin-resolving cloud scheme. For example, the result of cloud water mixing ratio are consistent among the bulk and bin simulations (ranging between 2.5 and 3.33 g kg$^{-1}$) for PRISTINE case. On the other hand, for POLLUTED cloud, the cloud water has somewhat larger values in bulk compared to bin simulations. The rainwater mixing ratio is also similar between the bulk and bin simulations. The values of cloud drop effective radius are a little larger in the bulk than in the bin model. Considering the fact that the objective of our study is to examine whether bulk invigoration of rain process can take place during the BFA cases with higher concentration of aerosols (dirty), we believe that our two-moment model is adequate for the purpose.

3. Results and discussions

Having demonstrated the model’s ability to simulate the aerosol indirect effect reasonably well in the previous section, here we show that model simulates much higher precipitation during BFA dirty cases compared to BNFA cases (either dirty or clean). This is followed by a series of analyses of simulated hydrometeors providing compelling evidence that mixed-phase processes play an important role in the invigoration of clouds and rain in the BFA dirty cases.

a. Response to microphysical hydrometeors and precipitation

To examine whether the larger aerosol concentration during BFA would lead to enhancement or decrease of
precipitation during ISM breaks, the composite of the surface accumulated precipitation over two domains (innermost domain 4 as well as the larger domain 3) during BNFA and BFA with higher (dirty) and lower (clean) aerosol concentrations are shown in Fig. 8. It may be noted that in the dirty environment (high CCN concentration), the precipitation is enhanced in BFA compared to either dirty or clean BNFA cases. It is also interesting to note that the process of invigoration occurs on a large scale and the difference in accumulated precipitation between BFA and BNFA cases is more prominent in the larger domain. The difference between the solid curve and the dashed (dotted) curves indicate the amount of increase in precipitation by increased CCN in the BFA case compared to either dirty or clean BNFA cases. It is also interesting to note that the process of invigoration in BFA compared to either dirty or clean BNFA cases. It is also interesting to note that the process of invigoration in BFA compared to either dirty or clean BNFA cases.

To understand the microphysical reasons behind such differences in precipitation formation, the detailed microphysical structures are analyzed. Figure 9a shows a much higher amount of cloud water during BNFA dirty case compared to BFA dirty case both with higher CCN concentration. However, the behavior of rainwater is completely different in the two cases. It may be noted (Fig. 9b) that the rainwater mass during BFA dirty case is much higher than that during BNFA dirty case. We note that the BNFA cases produce more cloud water characterized with a much higher peak (Fig. 9a) compared to BFA cases for both types of background aerosol concentrations (either clean or dirty). Although BNFA depicts more cloud water with smaller size (Fig. 10) it cannot produce rainwater because of poor collision-coalescence efficiency (e.g., Squires and Twomey 1961;
Warner and Twomey 1967; Warner 1968; Rosenfeld 1999). On the other hand, in BFA cases cloud water mixing ratio is less (Fig. 9a) but rainwater mixing ratio (Fig. 9b) is higher. As we show below, the BFA dirty cases are associated with stronger low-level convergence providing necessary uplift to small cloud drops above freezing level, initiating mixed-phase and ice-phase cold-rain processes and leading to invigoration of the rain.

To understand what is responsible for the enhancement of precipitation in the BFA dirty cases, we examine the inner domain-averaged vertical profiles of the sizes of hydrometeors (i.e., cloud ice, cloud, and rain drops) in Fig. 10 and amount of mixed-phase hydrometeors in Fig. 11 for BFA and BNFA cases. It is clear that bigger (smaller) rain drops arise because of larger (smaller) CCN number in BFA (BNFA) composites (Fig. 10). It is also interesting to note that while higher CCN concentrations reduce the cloud drop sizes in both the cases (Twomey’s indirect effect), the dirty environment produce bigger cloud ice sizes in both the BFA and BNFA cases. This is likely due to enhanced Wegener–Bergeron–Findeisen (WBF) process with which the cloud ice can grow faster owing to faster cloud drop evaporation. Note that cloud ice particles are larger in the BFA cases than in BNFA cases. Could this contribute to more rainwater in BFA events (Fig. 9b)? This question leads us to investigate the role of ice processes in those events (where moisture convergence is strong) that may supply higher rainwater masses with larger sizes.

While the cloud drop sizes are roughly similar in both BFA and BNFA cases (Fig. 10) but they are smaller than the critical diameter of 12–14 μm required for warm-rain initiation (Kulkarni et al. 2012; Konwar et al. 2012), the enhancement of rain in the BFA dirty cases must be due to cold-rain processes. In an attempt to understand the role of ice-phase processes (i.e., cloud ice, snow, and graupel/hail) in production of increased precipitation in BFA dirty cases, the domain-averaged vertical profile of cloud ice, snow, and graupel/hail mixing ratio are examined for BFA and BNFA cases in dirty as well as clean environments (Fig. 11). It is clearly seen that BFA produces more ice-phase hydrometeors compared to the BNFA cases. These results also support our hypothesis of the importance of ice-phase in the transition of breaks to active condition. What is the role of higher CCN concentrations (i.e., dirty environment) in the formation of ice-phases processes? Very interestingly, one can see that although the cloud ice mixing ratio is more in dirty environment of all cases, there are no significant changes in snow and graupel/hail formation for BNFA cases. On the other hand, in BFA cases higher CCN concentration produces more snow and graupel/hail (Fig. 11) that manifest in the higher rainwater formation (Fig. 10). What is responsible for lifting the small cloud drops to above the freezing level to initiate the cold-rain process? As proposed in our hypothesis earlier, the low-level vertical velocity arising from increased convergence during BFA cases is much larger than during the BNFA cases (Fig. 11d). Higher low-level vertical velocity seems to be successful in overcoming the stability and lifting the cloud drops to initiate the cold-rain processes. Such different behaviors in the BFA versus BNFA cases are related to detailed microphysical processes as discussed below.

b. Analysis of detailed microphysical tendency

To support the role of cold-rain processes in the invigoration we turn our investigation to details of the
mixed-phase microphysics for the two different types of ISM breaks. Figure 12 shows the production terms of deposition, accretion/riming, and melting for the three ice-phase hydrometeors (i.e., cloud ice, snow, and graupel/hail) considered in the model. It may be noted that the response of microphysical tendencies to higher CCN concentration is notably different in the two types of ISM breaks (Fig. 12). The initiation processes directly influences the subsequent growth of cloud ice. Cloud ice first grows by vapor deposition. The water vapor they need comes either from a continuous cooling of air by lifting or by the evaporation of cloud drops. Stronger vapor deposition results from the enhanced WBF mechanism because smaller and more numerous cloud drops tend to evaporate faster to provide water vapor for ice particles to grow (Cheng et al. 2010; Tao et al. 2012). So the deposition growth of cloud ice is influenced indirectly by CCN. The cloud ice deposition is weak in BNFA composites. The vapor deposition of snow is controlled by the amount of snow and available water vapor (i.e., ice supersaturation) and increases at high aerosol concentration when there is sufficient vapor or liquid drops to provide vapor. Graupel/hail deposition is similar to vapor deposition of snow. Both the depositions (snow and graupel) reveal higher magnitude in BFA events compared to BNFA. This is because of higher ice supersaturation in BFA in contrast to BNFA (not shown). Rain initiation includes two warm-rain processes, rain embryo formation and autoconversion, and two cold-rain processes, snow melting and graupel/hail melting. Figure 12 reveals that the suppression of warm rain takes place in ISM breaks (either BFA or BNFA) because of more aerosol concentrations. However, cold rain is enhanced in BFA and suppressed in BNFA cases. Finally their combination produces total precipitation in the surface (Fig. 8) that reveals enhancement (suppression) of rainfall in ISM breaks like BFA (BNFA) cases.
c. Latent heating/cooling due to explicit moisture scheme

While the low-level convergence may initiate the cold-rain process, any latent heat released in the upper atmosphere can further enhance low-level convergence and strengthen the invigoration of the rain. If the cold-rain processes dominate the latent heating, it should be evident in the higher altitudes in the vertical structure of the latent heat. To check out the latent heat released during this process, we analyzed the latent heating calculated from the explicit moisture scheme (Reisner et al. 1998).

The vertical profiles of latent heating during two types of breaks and in dirty and clean environments are shown in Fig. 13. It shows significant differences in latent heating for the BFA and BNFA cases, particularly in the upper levels. In BFA cases, dirty environments (more CCN) produce large cooling in the middle level mainly and more heating in the upper level. But the heating difference is negligible for the BNFA cases. The enhanced midlevel cooling is mainly due to stronger cold-rain production aloft, which causes more falling ice hydrometeors and thus melting near the 0°C level. The upper-level heating is mainly caused by higher deposition and riming from cloud ice, snow, and graupel/hail. This stronger deposition is partially resulted from more cloud water retention in the updraft as a consequence of reduced warm-rain process. The difference is stronger in BFA cases compared to that in BNFA cases owing to more water vapor and cloud water supply associated with the stronger moisture convergence. In summary, the cloud ice mixing ratio increases in dirty environments owing to more CCN concentrations for mixed-phase clouds. This increasing cloud ice initiates enhancement of various ice processes, which in turn leads to more latent heat release to invigorate the convection. Such aerosol effects on ice-phase microphysics and cloud thermodynamics are stronger when there is more moisture supply as in the BFA cases. These simulated results are in good agreement with the observational findings of Niu and Li (2011) and also support the invigoration theory of Rosenfeld et al. (2008).
4. Conclusions

A strong case of the role of aerosols on transitions of MISO is made here and a pathway through which aerosols influence the transitions is unraveled involving complex interactions between direct radiative forcing, large-scale dynamics, and cloud microphysics. Manoj et al. (2011) classified monsoon breaks as BFA and BNFA classes, where strong moisture convergence with high aerosol concentration and weak moisture convergence with low aerosol concentration occur, respectively. The present study endeavors to understand the role of microphysical processes on transitions of those break classes due to higher levels of aerosol concentrations. First, we find that the ISM break with strong moisture convergence is influenced strongly by the more aerosol concentrations and leads to production of more ice-phase hydrometeors and finally more rain. The latent heating/cooling due to explicit moisture scheme is significantly different between the BFA and BNFA cases. A higher-CCN-concentration environment leads to more cloud water mass and number and reduction of cloud drop size (Twomey effect) for both BFA and BNFA cases, and this tends to delay the warm-rain production. But the warm-rain suppression effect is overwhelmed by the cold-rain production enhancement, especially for systems with higher moisture convergence (e.g., the BFA cases). Thus, mixed-phase processes play a key role in deciding enhancement or suppression of precipitation though cold-rain production from snow and graupel melting that is more in “dirty” conditions during BFA cases (enhance precipitation) and less in BNFA cases (suppress precipitation). The observational findings (Niu and Li 2011) confirm that aerosols enhance precipitation from mixed-phase clouds (cold rain) but suppress precipitation from liquid clouds (warm rain) owing to reduced cloud drop size (microphysical effect). Whether such aerosol influence on cloud microphysics
plays a role in the timing or duration of ISM breaks requires further investigation. As indicated in Manoj et al. (2011), all monsoon breaks are not identical and certain large-scale conditions determine whether accumulation of absorbing aerosols take place during a given break and, as demonstrated here, the aerosols determine the type of transitions. Thus, the aerosols contribute significantly to the event-to-event variability of MISO. Our results highlight the strong need to incorporate proper cloud microphysical processes in the models for extended-range prediction of MISO.

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