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The Indian drought of 2002—a sub-seasonal phenomenon?

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SUMMARY

A major drought occurred over India in the year 2002 with a seasonal rainfall deficit of 21.5%, a result of 56% below normal rainfall in the month of July. The largest anomalies occurred in the western parts of India, when an Indian monsoon field experiment was in progress there. The present study is based primarily on data collected from a research ship that was deployed 100–250 km off the west coast of India for the experiment. Surface and upper air observations made over the eastern Arabian Sea during July 2002 are presented. Sea-surface conditions were favourable for supporting deep convection over the study area. Strong atmospheric inversions around 800 hPa prevented the growth of cumulonimbus during the first half and towards the end of July. A second inversion above the melting level was also prominently present. The strength and persistence of the inversions are unusual. Back-trajectory analysis reveals a major change in the low-level circulation during July 2002 with frequent advection of dry air from over the deserts around the eastern Arabian Sea instead of marine air from across the equator.

KEYWORDS: Atmospheric inversion Field experiment Monsoon break Tropical convection

1. INTRODUCTION

The agriculture and economy of India depend on summer monsoon rainfall. Consequently, the performance of the monsoon has traditionally been linked to the seasonal rainfall amount in India. The area-weighted average of the daily rainfall at more than 300 rain-gauge stations spread across India is often used as an index of monsoon activity, and its cumulative value for June–September is called the all-India summer monsoon rainfall (ISMR) (Parthasarathy *et al.* 1995). For the period 1871–2002, the average value of ISMR is 849 mm and its standard deviation (σ) is close to 10% (values based on the *Monthly Rainfall Data for India*†). A year with ISMR within one σ of its long period average (LPA) is considered a normal monsoon year, and when the difference exceeds this, it is an ‘excess’ or a ‘drought’ year depending on the sign of the departure of ISMR from LPA. No drought occurred for 14 continuous years from 1988 to 2001, and then a major one struck with an ISMR deficit of 21.5% with the largest negative rainfall anomalies occurring in the western and north-western parts of India.

The drought of 2002 was the third largest in the past 100 years, after 1918 and 1972. An ISMR deficit of 2σ from the mean seems well within its natural variability. However, when we examine how this deficit came about in 2002, it turns out to be unique in the recorded history of the Indian monsoon. There were no major precursor signals suggesting a drought over India, except for signs of a developing El Niño (Kalsi *et al.* 2004). According to the standard ENSO indices, El Niño of 2002 was of moderate strength and comparable to those of 1986 and 1992 (McPhaden 2004). ISMR deficits were −12.5% and −7.8% in 1986 and 1992 respectively. Other large-scale features such as Eurasian and Himalayan snow cover (Mooley and Shukla 1987) and pre-monsoon sea-surface-temperature (SST) anomalies over the Indian Ocean (e.g., Clarke *et al.* 2000; Rajeevan *et al.* 2002) made for a normal ISMR, and all long-range prediction models, based on both empirical as well as numerical approaches, had predicted a normal summer monsoon rainfall over India and all predictions failed (Gadgil *et al.* 2002; Kalsi *et al.* 2004). Further, for the first time since systematic recording of rainfall started in India in 1871, no monsoon depression formed during the entire monsoon season of 2002 (Kalsi *et al.* 2004). Normally, when a drought occurs in India, the

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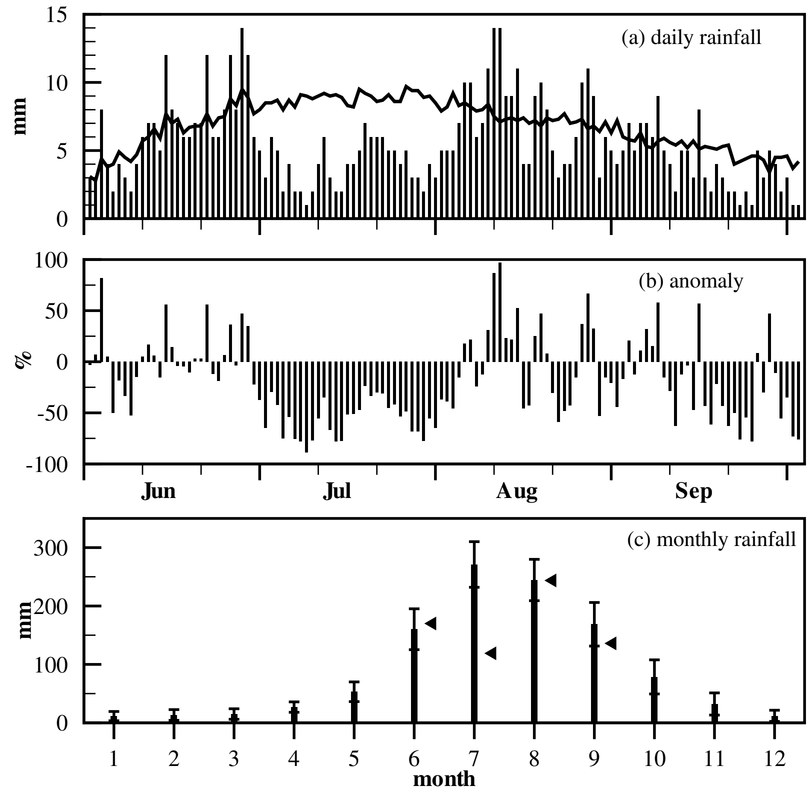

Figure 1. All-India rainfall: (a) daily average rainfall (line) (mm) and daily rainfall June–September 2002 (vertical bars) (mm); (b) June–September 2002 daily-rainfall anomaly (percentage difference from values shown by the line in (a)), and (c) average monthly rainfall throughout all twelve months of the year for the period 1900–2002. In (c) the error bar corresponds to $\pm\sigma$ for the month, and the arrow head indicates the monthly rainfall during 2002.

trends of deficit rains persist throughout the monsoon season, and all or three out of four months tend to be below-normal rainfall months during June–September (e.g., Table 3(a) of Webster *et al.* (1998)); the only exception before 2002 was 1911, when July and August months were 'deficit' months while rainfall in June and September was above average. In 2002, all-India daily rainfall was continuously below average for a period of 36 days with the largest deficits occurring in the first two weeks and in the last week of July (Fig. 1). Monthly rainfall was slightly above normal in June and August, and September rainfall was below normal but within one standard deviation from the mean for the month (Fig. 1(c)). It was in July that the monsoon rains failed with a deficit of 56% for the month (nearly 4σ departure from the normal for July). This is the largest deficit in July since 1871; the spectacular drought in 2002 was a result of continuously below-normal rainfall in the entire month of July.

Therefore, in order to understand the drought of 2002, we need to understand why July rainfall was scanty in that year. Periods of scanty rainfall within a monsoon season are referred to as monsoon breaks (see Gadgil and Joseph (2003) for various break definitions). The conditions that prevailed over the Indian region during July 2002 had many features that characterize monsoon breaks including low-level anticyclonic wind anomaly over India and weaker cross-equatorial flow (Sikka 2003; Saith and Slingo 2006). Long-lasting breaks in monsoons are large-scale phenomena and are affected by the propagation of low-frequency modes (10–20 and 30–50 day) and mid-latitude interactions (Krishnamurti and Surgi 1987; Annamalai and Slingo 2001). These are processes taking place on intra-seasonal time-scales, and the influence of intraseasonal events on ISMR is less understood than that of slowly varying (boundary) conditions such as El Niño (Rasmusson and Carpenter 1983) and winter-climate anomalies over Europe (Raman and Rao 1981). Gadgil *et al.* (2004) showed that major Indian droughts occurring in non-El-Niño years are associated with the negative phase of the equatorial Indian Ocean oscillation (EQUINOO). In the negative phase of EQUINOO, convection persists over the eastern equatorial Indian Ocean and deficit rainfall prevails over the Indian subcontinent (Gadgil *et al.* 2003). 2002 was a major negative EQUINOO year (Gadgil *et al.* 2004). Saith and Slingo (2006) on the other hand, attributed high Madden-Julian oscillation (MJO) activity in the equatorial western Pacific and the developing El Niño as the major cause of the 2002 drought. The MJO activity in the western and central Pacific was unusually high during May–July 2002 (McPhaden 2004; Saith and Slingo 2006).

July 2002 was also remarkable for the unusually high typhoon activity in the north-west Pacific (NWP). Typhoons started forming towards the end of June and, in total, seven storms formed or moved over NWP during July. The broad consensus among previous authors has been that high typhoon-activity in NWP decreases ISMR (e.g., Kumar and Krishnan 2005). However, there have been occasions where the remnants of NWP typhoons moved westwards, crossed over to the northern parts of the Bay of Bengal and provided the nucleus for the formation of monsoon depressions (Saha *et al.* 1981). Therefore, the path of NWP storm determines the favourable or adverse impact of storms on the monsoon rainfall of India. During July 2002, storms in the west Pacific initially moved north-westwards and then recurved and crossed 30°N between 120°E and 140°E, and no remnant of these entered the Bay (Kalsi *et al.* 2004). The initial large-scale circulation that favours a northward-recurving storm over the north-west Pacific has many features in common with that observed during monsoon breaks, including a monsoon trough—either very weak or displaced to the north (Harr and Elsberry 1991). In fact, after an active phase during mid-June, break-monsoon conditions prevailed over India during the last week of June 2002. When the Pacific typhoons move north of 30°N between 110°E and 140°E, a break-monsoon condition prevails over India (Raman 1955). Therefore, we should investigate if the initial circulation that prevailed over India and NWP towards the end of June was favourable for northward-recurving storm paths and if that circulation was further strengthened by the movement of typhoons themselves and so extended the break conditions over India during July.

Thus, several sub-seasonal events adverse to a normal monsoon rainfall prevailed over India during July 2002. Despite these complexities, Webster and Hoyos (2004) showed that both the phase and amplitude of rainfall activity even on sub-Indian spatial-scales were predictable 20 days in advance in 2002. Webster and Hoyos (2004) chose predictors from Indian Ocean longitudes only. The success of this scheme for 2002 suggests that the precursors of a dry July were present in the atmosphere–ocean system in the Indian region at least three weeks in advance. However, the central questions are difficult to answer: was there a primary mechanism/process that led to the changes in the circulation over the Indo–Pacific region during July 2002, or did the circulation evolve as a result of interactions among its various components? Basically, the South Asian monsoon-system involves complex feedbacks (e.g., Webster *et al.* 1998); finding cause-and-effect relationships is not simple. Often what we see are only associations. For example, the MJO events over the western Pacific during May–July 2002 had their origins over the Indian Ocean, with cloud bands originating in the tropical Indian Ocean propagating up to the Date Line; the associated events (westerly-wind bursts) helped in extending the 29 °C isotherm beyond the Date Line by the end of June (McPhaden 2004). This could have led to higher values of zonal winds in the equatorial Indian Ocean, and also facilitated the formation of typhoons east of 150°E during July (e.g., Chen *et al.* 1998). Over the Indian Ocean, the anomaly of the zonal component of surface wind averaged between 2.5°S and 2.5°N (whose average over longitudes 60°E–90°E constitutes the EQUINOO index) had greater positive values during May and June 2002 than in any other period or month that year (Gadgil *et al.* 2003). Thus, the strong MJO activity over the Pacific during May–July and high values of the EQUINOO index over the Indian Ocean during May–June were perhaps not independent but evolved together within the Indo–Pacific basin during the first half of 2002. In principle, some of the issues, such as the role of warmer SSTs in Niño 3 and Niño 3.4 regions on rainfall variability over India, can be tested using atmospheric general circulation models by specifying the observed sea-surface conditions. However, the present numerical models lack the skills and correct phase relationship with SST to simulate intraseasonal variations (including MJO) reasonably (e.g., Molteni *et al.* (2003) and Saith and Slingo (2006)). Thus, pure numerical experiments will have limited success in simulating the rainfall in years such as 2002 where intraseasonal events seem to have played a crucial role. Therefore, we need to explore the drought of 2002 taking different approaches, including detailed analysis of observational/reanalysis data and numerical modelling. The present study, based on the analysis of data from an observational experiment, is one such effort.

Indian scientists had planned a field campaign in the eastern Arabian Sea (AS) and the west coast of India during June–August 2002 to understand very heavy rainfall events (IREs) on the west coast of India. This experiment was called the Arabian Sea Monsoon Experiment (ARMEX), and extensive data over land and ocean were collected. However, not a single IRE formed within the main study area where at least a few were expected. The main objective of the present study is a detailed analysis of the surface and upper air data collected from a research ship during ARMEX with emphasis on 2002 July. We shall see that the state of the local sea-surface was conducive to deep convection, though an unusually strong inversion in the atmosphere around 800 hPa inhibited convection during the first half of July and towards the end of the month. Back-trajectory analysis shows that dry air from deserts was intruding into AS frequently and the low-level circulation in July 2002 was different from that in a normal monsoon year. Section 2 briefly describes ARMEX, section 3 the analysis of surface and upper-air properties over AS. Section 4 discusses the findings and section 5 concludes the paper.

2. ARMEX

ARMEX is the second field-experiment under the Indian Climate Research Programme (ICRP), the first being the Bay of Bengal Monsoon Experiment (BOBMEX) carried out over the bay during 1998–99 (Bhat *et al.* 2001). Details pertaining to the scientific background, experimental design and field phase of ARMEX, along with some preliminary results, can be found in a special issue of *Mausam* (Rao 2005). When monsoon is active, many places along the Indian west coast receive more than 200 mm rain in 24 hours (Rao 1976; Francis and Gadgil 2005), and such events were designated as IREs in ARMEX. A range of mountains (the Western Ghats) within 50 km from the coast with a mean altitude of 800 m run parallel to the west coast of India between latitudes 8.5°N and 21°N. The high rainfall over the west coast and the heavier rainfall over the Western Ghats are generally attributed to forced ascent over the Western Ghats and the blocking of wind by the Ghats (Grossman and Durran 1984). IREs occur during the onset of the monsoon in June and also during July–August when the monsoon revives and becomes active (Francis and Gadgil 2005). IREs have been associated with the tropical convergence zone (Francis and Gadgil 2005), offshore vortices (George 1956) and mid-tropospheric cyclones (Miller and Keshavamurthy 1968). Thus, IREs occur under synoptic conditions associated with deep convection, and local orographic features modulate the ascent of air resulting in intense precipitation in certain locations. Objectives of ARMEX included understanding the IREs and associated offshore convective systems.

There have been several earlier observational experiments conducted over AS, e.g., the International Indian Ocean Expedition (IIOE) of the mid-1960s (Colon 1964), the Indo-Soviet Monsoon Experiments of 1973 (ISMEX-73) (Pant 1982), the monsoon experiment of 1977 (MONSOON-77) (e.g., Pant 1982) and the major international monsoon experiment of 1979 (MONEX-79) (Fein and Kuettner 1980). In contrast to the earlier monsoon experiments, where a broad area over the AS was explored, the focus in ARMEX was the region within 250 km of the west coast of India. Furthermore, earlier experiments covered the period from mid-May to early July whereas ARMEX covered mid-June to late August with emphasis on IREs, and the revival and maintenance of the monsoon on the west coast after the onset phase was over.

The present study is primarily based on data collected by the author and his colleagues from the Indian research ship ORV *Sagar Kanya*, deployed 21 June–15 August with a break for a port call 14–17 July. Sections, parallel and normal to the coast, and time-series (TS) observations were carried out (Fig. 2). The ship was positioned 150–200 km off the Indian west coast for TS—at 16.9°N, 71.2°E from 30 June to 10 July, and at 15.5°N, 72.2°E from 22 July to 5 August 2002. The TS locations were chosen to be near the belt of heaviest rainfall on the west coast where the probability of occurrence of IREs was high. The experimental arrangement and equipment/sensors for basic meteorological and oceanographic measurements on ORV *Sagar Kanya* during ARMEX were nearly identical to those used during BOBMEX (Bhat *et al.* 2001). The surface variables, including those required for surface energy balance (sensible, latent and radiative fluxes), were measured continuously. The meteorological sensors were calibrated before and after the field phase. Vaisala radiosondes (model RS80-15G) were launched (2–4 per day) to monitor the upper air. In order to reduce the dry-bias of Vaisala radiosondes (Wang *et al.* 2002), humidities measured by sonde were compared with those measured by a calibrated reference humidity probe (Rotronics make): the two were held next to each other for several minutes, and corrections applied to the radiosonde data so that they matched those from the reference probe.

3. RESULTS

(*a*) *Convection over the Indian region*

During July 2002, no offshore vortex formed and the maximum daily rainfall amount was below 200 mm in areas known to be subject to IREs. In order to get a broad view of the large scale convective activity over the Indian region, the outgoing long-wave radiation (OLR) anomaly for July 2002 is shown in Fig. 3(a). (OLR data are from US National Oceanic and Atmospheric Administration for the period 1979–2002 and have a 2.5° resolution.) OLR anomalies were positive over the equatorial western Pacific Ocean, southern Bay of Bengal, west and north-west India and eastern AS, whereas the western Pacific Ocean north of 10°N, eastern equatorial Indian Ocean and head of the Bay of Bengal had negative OLR anomalies. The west coast of India and the eastern AS, the main areas of focus during ARMEX, were among the areas with least convection.

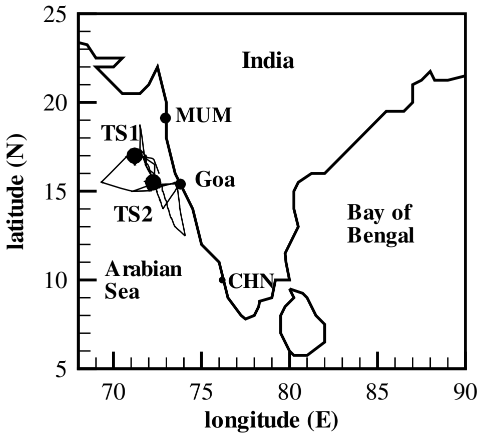

Figure 2. Cruise track and time-series positions of ORV *Sagar Kanya*. The time-series (TS) positions are: TS1, 16.9°N, 71.2°E, 30 June–10 July 2002; TS2, 15.5°N, 72.2°E, 22 July–5 August 2002. MUM denotes Mumbai (Bombay) and CHN Cochin.

The OLR anomalies westward of 150°E resemble the 'quadrapole' structure associated with the 40-day mode of the Indian monsoon described by Annamalai and Slingo (2001). 850 hPa wind anomalies (Fig. 3(a)) show weaker cross-equatorial flow into AS, stronger southward currents at the northern end of AS, and northerly and north-westerly anomalous winds in the head of the Bay of Bengal. (Wind data are from the European Centre for Medium-Range Weather Forecasts (ECMWF) re-analysis product ERA-40 (Simmons and Gibson 2000) and anomalies are based on 1979–2002 climatology.) Westerly winds in the region of excess convection over the western Pacific were strongly enhanced. The OLR and wind anomalies seen in Fig. 3(a) over the Indian region have many features in common with the composite monsoon-break anomalies corresponding to +1 and +2 triads (Krishnan *et al.* 2000) which correspond to the period of two to seven days after the start of the break. This suggests that the whole of July was rather like a prolonged monsoon-break. Figure 3(a) also suggests a strong correlation between wind and atmospheric heating anomalies. Whether the large wind-anomalies over the Arabian Sea were a local response to reduced convection over the Indian region, or a remotely forced Rossby-wave-like response to enhanced convection over the western Pacific, needs to be explored further as this will bring out the teleconnections involved in the monsoon circulation. A study similar to that carried out by Annamalai and Sperber (2005), with atmospheric heat sources and sinks representative of July 2002, is required to understand this.

Figure 3(b) shows spatially averaged daily OLR between 10°N and 20°N over AS and the west coast (70°E–75°E), the Bay of Bengal (85°E–90°E) and a major portion of the Indian land mass (72°E–82°E). (Henceforth these areas are referred to as ASWC, BoB and LAND, respectively). Monsoon onset over Kerala was three days ahead of the normal onset date of 1 June and the monsoon was active over India till the last week of June (Fig. 1(a)). From 10 to 25 June, OLR values were below 200 Wm^{-2}. OLR increased rapidly from 25 June onwards over BoB; LAND and ASWC followed suit with a delay of about two days, indicating a monsoon break in the region. OLR values remained high between 210–270 Wm^{-2} until July 28 over ASWC. OLR values over BoB were lower than that over ASWC during July and reduced below 175 Wm^{-2} on three occasions. OLR pattern over LAND was more like that over ASWC during the first

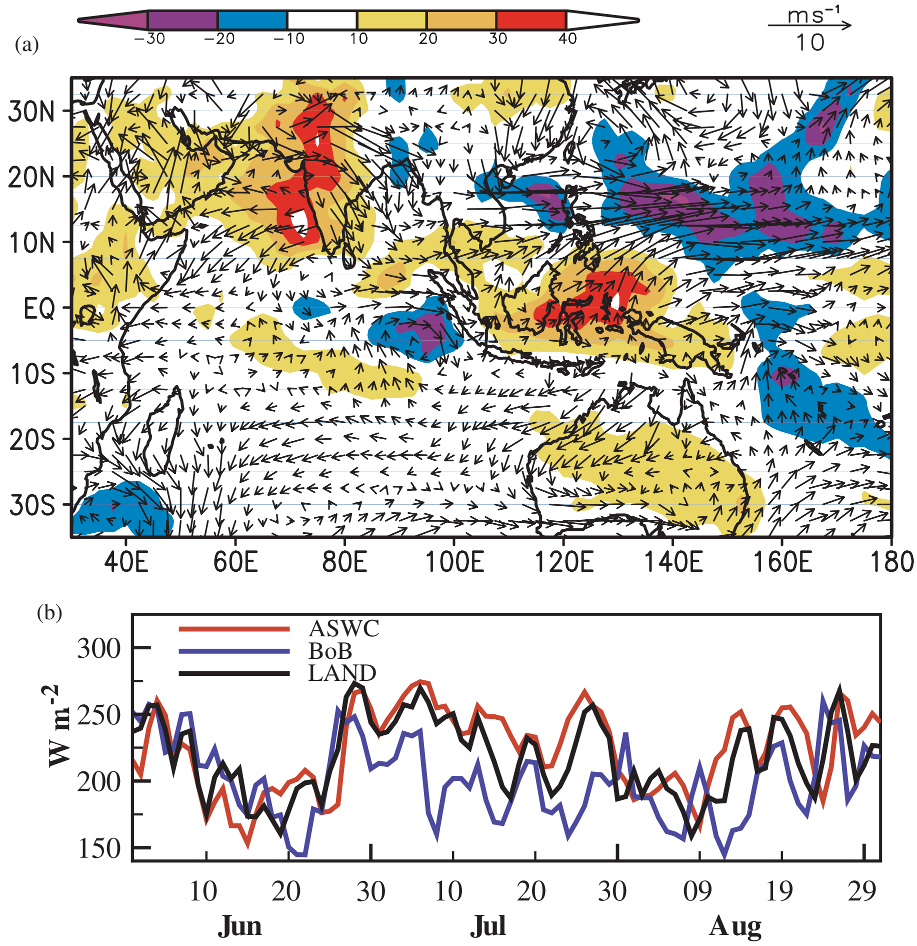

Figure 3. Outgoing long-wave radiation (OLR) (W m^{-2}): (a) anomalies for July 2002, with corresponding low-level wind anomalies (m s^{-1} using scale shown), and (b) daily average OLR from three regions of India between 10°N and 20°N (ASWC denotes the eastern Arabian Sea and the west coast of India (70°E–75°E), BoB denotes the Bay of Bengal and LAND denotes a major portion of the Indian land mass (72°E–82°E)). In (a), negative OLR anomalies indicate above-normal deep convective activity, and vice versa.

half of July and in between ASWC and BoB trends during July second half. Figure 3 suggests that convection was near normal over BoB and satellite images revealed that many cloud systems formed there during July 2002. However, the cloud systems that moved over land from BoB dissipated eastwards of 80°E during the first half of July, thus limiting the rainfall activity mainly to the eastern parts of India.

(*b*) *Surface conditions over the Arabian Sea*

The temporal variations of rainfall, surface pressure and temperature, wind speed, wind direction and equivalent potential temperature (θ_e) measured from the ship over the open sea are shown in Fig. 4. The total amount of rainfall measured on board the ship over AS during the first two weeks of July was 13 mm. That during the second half was even less so that the total for the month was less than 20 mm. (High OLR values over the ASWC box in Fig. 3(a) suggest that it is unlikely that any significant rainfall occurred at the ship's observing position (∼17°N, 72°E), during its port call.)

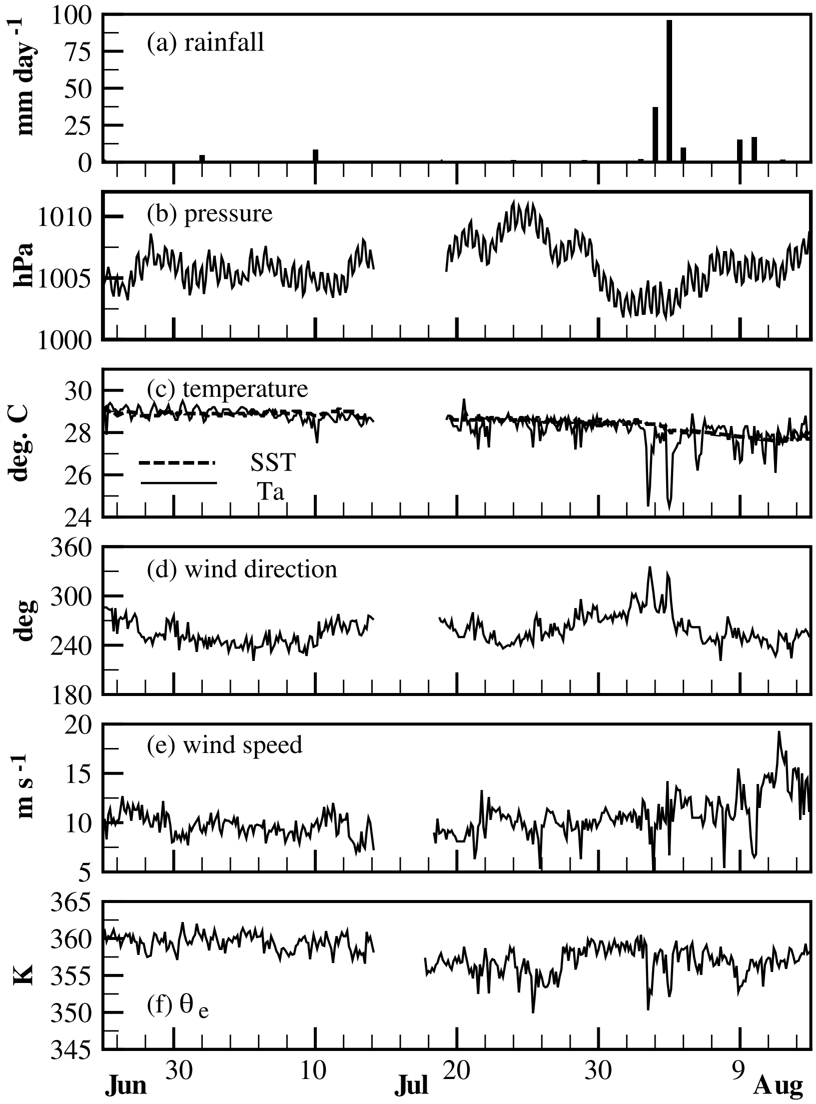

Figure 4. Time series of near-surface quantities measured on board ORV *Sagar Kanya*, June to August 2002: (a) daily rainfall (mm day^{-1}); (b) air pressure (hPa); (c) sea surface temperature (°C) and air temperature (°C); (d) wind direction (deg from north); (e) wind speed (m s^{-1}), and (f) equivalent potential temperature, θ_e (K). See Fig. 2 for the position of the ship during the two series.

In the absence of similar direct measurements in this area during other years, the rainfall deficit in July 2002 is estimated from the Global Precipitation Climatology Project (GPCP) rainfall dataset (Huffman *et al.* 1997). For the grid box where the ship was positioned, the average (1979–2002) July rainfall was 271 mm; for July 2002 it was 47 mm—a July 2002 deficit of 83%. The previous minimum rainfall for the month of July for this grid box was 94 mm, in 1987, also a drought year. The changes in the surface pressure were small during 26 June–10 July (Fig. 4(b)); the second half of July saw larger variations with a rise 19–25 July followed by a rapid fall. SST was between 28.5 °C and 29 °C throughout July, i.e. above the convection threshold value of 28 °C for the Indian Ocean (Gadgil *et al.* 1984). Air temperature (T_a) and SST were very close to each other except when it rained during which T_a decreased by 1–4 degC and recovered rapidly after rain stopped. Diurnal variation in SST is small while that in T_a is clearly discernible, especially during the first half of July. Winds were south-westerly during July with speeds 8–12 m s^{-1} except on a few occasions. This wind-speed range compares well with those measured during break-monsoon conditions during earlier field experiments over AS (e.g., Holt and Raman 1987). No marked increase in the average wind speed took place 1–3 August as dry weather gave way to wet; wind direction, however, showed a gradual but clear trend and became north-westerly 3 and 4 August (Fig. 4(d)). Increased wind-speed (Fig. 4(e)) coincided with the revival of the monsoon over the land in the first week of August (Fig. 1(a)). Equivalent potential temperature θ_e varied in a narrow range around 360 K during the first half of July and between 350 K and 360 K during the second half (Fig. 4(f)). Values of θ_e were above 355 K more than 85% of the time, and values in the present study are comparable to those observed over the western Pacific warm pool (Kingsmill and Houze 1999) and over the northern part of BoB during BOBMEX (Bhat 2001). Values of θ_e above 345–350 K are normally sufficient to support deep convection in the tropical atmosphere.

(*c*) *Upper air*

Representative vertical profiles of temperature, humidity, wind speed and wind direction are shown in Fig. 5. All the three July temperature- and humidity-profiles are similar and show several distinct layers. The lowest 500–1000 m layer was well mixed (except for a shallow superadiabatic surface layer) with a lapse rate close to the dry adiabatic value; the top of this layer (located between 950 hPa and 900 hPa) is marked by a decrease in relative humidity. Above this, near 800 hPa, is a shallow, but sharp, inversion through which relative humidity decreases from more than 85% to less than 30%. Temperature and humidity structures in this inversion layer are akin to soundings in the dry (air) intrusion category observed during TOGA COARE (Lucas and Zipser 2000). Further up, the lapse rate between 600 hPa and 500 hPa levels resembles those normally found in the atmospheric mixed layer. This layer was capped by an inversion with its base about 50 hPa above the melting level. The profiles shown are not isolated cases and two inversions were generally present during the first half of July (Bhat 2005). Inversions disappeared and the moisture levels in the lower troposphere increased just before the monsoon revived in the first week of August. Humidity profiles for 2 and 4 August resemble the convective cases discussed by Lucas and Zipser (2000). In the July soundings, westerly winds in the boundary layer gradually turned to north-westerly above 800 hPa, and north-westerly/northerly winds prevailed up to 400 hPa, with easterlies above 350 hPa. During the revival of the monsoon in August, winds were north-westerly in the boundary layer and the easterly winds moved downwards to 600 hPa. Wind speed at low levels strengthened on 4 August, showing the development of a low-level jet.

(*d*) *Inversion characteristics*

The base of the lower inversion was located between 1.7 km and 2.5 km above the surface and its strength (measured by the increase in the potential temperature θ across the inversion layer) was in 2 K to 9 K range with an average around 7 K during the first half of July (Bhat 2005). During 17–25 July, inversions were either absent or very weak. They reappeared again 27 July–1 August, but weaker and higher (Bhat 2005). In order to see the temporal variation of thermal stratification over AS and the west coast

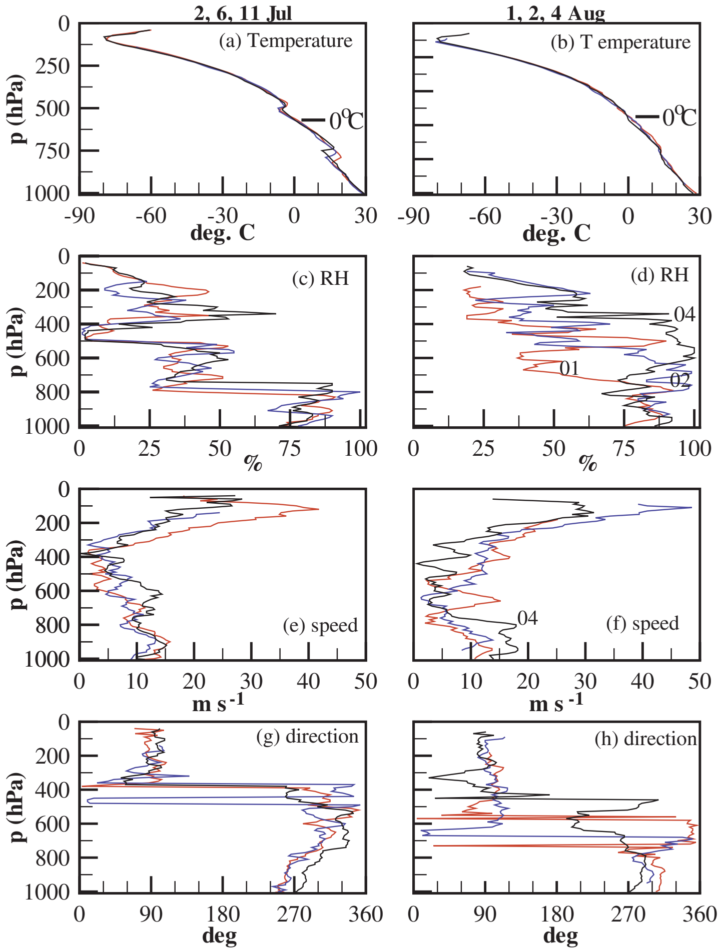

Figure 5. Vertical profiles (linear pressure (hPa)) measured from ORV in the eastern Arabian Sea in 2002: (a, b) air temperature (°C); (c, d) relative humidity (%); (e, f) wind speed (m s^{-1}), and (g, h) wind direction (deg from north). Left-hand panels show measurements made from TS1 (16.9°N, 71.2°E) (thin line, 2 July; thick line 6 July, and dotted line 11 July), and right-hand panels from TS2 (15.5°N, 72.2°E) (thin line, 1 August; thick line, 2 August, and dotted line, 4 August). See Fig. 2.

of India, values of the difference in θ at 700 hPa and 850 hPa ($\Delta\theta$) during June, July and August 2002 are shown in Fig. 6(a), from soundings made on board ORV *Sagar Kanya* (SK), and at the coastal stations of Goa and Mumbai (MUM) (Fig. 2). The Indian Navy launched Vaisala radiosondes at Goa and Mumbai radiosonde data are from the India Meteorological Department (IMD). Since the radiosonde data provided by IMD and the Indian Navy do not have a high vertical resolution, signatures of inversions are sought here using $\Delta\theta$ across the layer between 850 hPa and 700 hPa. Values are typically 7–9 K during a convective season in the tropics, and when $\Delta\theta$ exceeds 10 K it implies stronger than normal thermal stratification. From Fig. 6(a) we see that the broad features of $\Delta\theta$ are similar for the three soundings with higher values during the first half of July and also towards the end of that month. There is a general decrease in $\Delta\theta$ from north to south with stronger stratification at Mumbai than at Goa. In situ observations are not available over other parts of AS to show us the spatial extent of the inversion during July 2002. Based on an analysis of satellite-sounder-derived products, Narayanan *et al.*

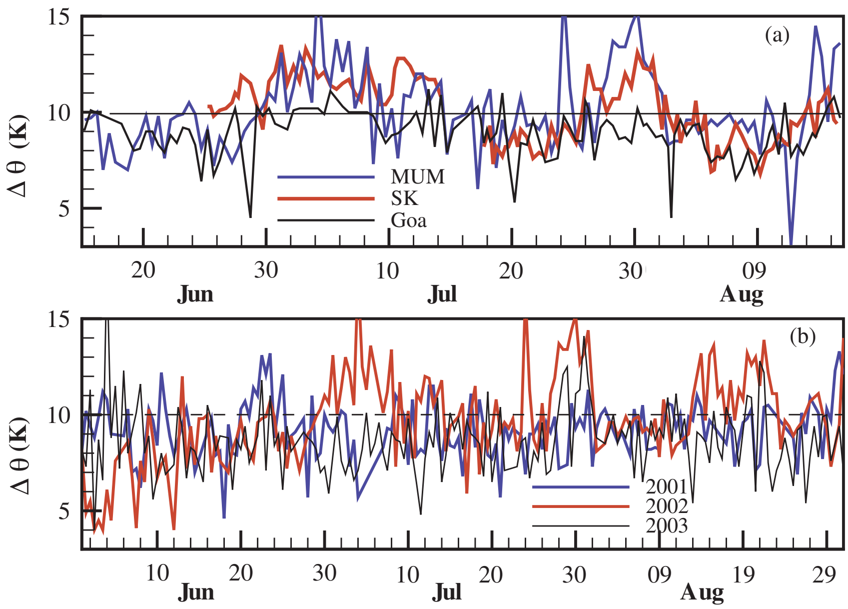

Figure 6. The increase in potential temperature between 850 hPa and 700 hPa, $\Delta\theta$ (K): (a) near Mumbai (MUM), Goa and the locations of ORV *Sagar Kanya* in the year 2002, and (b) Mumbai during the years 2001 (thin line), 2002 (thick line) and 2003 (dotted line).

(2004) showed that thermal stratification unfavourable to deep convection extended all over the AS westwards of 70°E during July 2002.

Temperature inversions have also been observed over the AS near 800 hPa during earlier observational experiments. In the western AS near the African coast, strong low-level temperature inversions occur when warm continental air from neighbouring deserts flows over the cool maritime air of AS (Colon 1964). As this air moves eastwards, the inversion weakens and rises, and vanishes beyond 70°E (Colon 1964). On some occasions however, inversions extend up to the west coast of India (Fig. 6(a)). Since upper-air data over the AS are very limited in space and time, it is not possible to infer if the strong thermal stratification (inversion) observed over the eastern AS and Indian west coast during ARMEX is unique to July 2002 or not. In earlier field experiments over this area (e.g., Holt and Raman 1987; Grossman and Durran 1984) the thermal stratifications were weaker than in July 2002 (Bhat 2005). Figure 6(b) shows the temporal variations of $\Delta\theta$ for the years 2001, 2002 and 2003 for Mumbai. ISMR values were 6% below the LPA in 2001 and 2% above LPA in 2003. Episodes of strong thermal stratification did occur in 2001 and 2003, but less frequently and for shorter durations than in 2002. Thus, while the low level inversion is not unique to July 2002, its persistence, strength and spatial extent appear to be unprecedented.

The second inversion with its base near 500 hPa level, i.e., about 50 hPa above the melting level, is prominent in the soundings made during the first half of July (Fig. 5(a)). At the top of this inversion, relative humidity (RH) often decreased below 10% (Fig. 5(c)), i.e. the inversion air was warm and dry. The largest value of the temperature gradient (dT/dz) across a layer of 20 hPa thickness within the inversion exceeded +9 degC km^{-1}, and the averaged value across the inversion is more than +5 degC km^{-1}. A stable layer near the melting level appears to be common in the tropics (e.g., Johnson *et al.* 1996). Johnson *et al.* (1996), for instance, observed stable layers

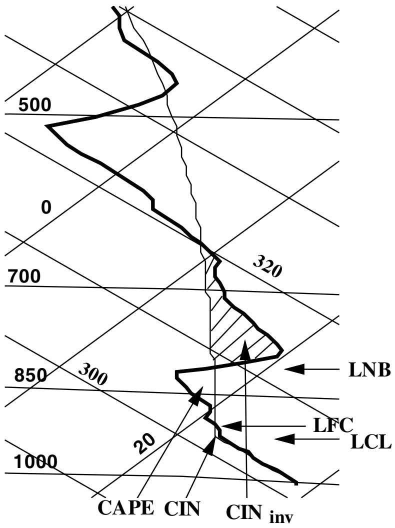

Figure 7. Skew T–log p (°C–hPa) plot of the sounding (thick line) made on board ORV *Sagar Kanya* (SK) at TS1, 16.9°N, 71.2°E, on 4 July 2002, and the corresponding surface-air pseudo-moist adiabat (thin line): LFC denotes the level of free convection, LCL the lifting condensation level, LNB the level of neutral buoyancy, CAPE the area representing the convective available potential energy, CIN the 'convection inhibition' or area representing the energy barrier a parcel has to overcome before it can ascend on its own, and CIN_{inv} the energy inhibiting convection in the inversion-related stable layer.

within ∼100 hPa layer centred about the melting level in more than half the soundings taken over the western Pacific warm pool. The stable stratification near the melting level is much stronger in the present AS profiles (Fig. 4) than in their warm- pool profiles (Fig. 18 of Johnson *et al.* 1996) and it is also higher (50–100 hPa above the melting level). Johnson *et al.* (1996) attribute the formation of warm and dry (stable) layers near the melting level to the intrusion of mid-latitude air. In section 4, we report on a back-trajectory analysis carried out to find the source of the AS inversion air.

(*e*) *Convective instability*

Figure 7 shows the sounding made on board ORV *Sagar Kanya* (SK) on 4 July 2002 and the corresponding surface-air pseudo-moist adiabat: the level of free convection (LFC) and lifting condensation level (LCL) are not far apart. The level above the LFC where the parcel and ambient air temperatures become equal again is the level of neutral buoyancy (LNB) and conventionally in forecasting practice defines the main cloud-top height. It is observed from Fig. 7 that this is near 800 hPa, although a parcel forced through the negative buoyancy layer up to about 650 hPa would again attain a positive buoyancy and remain buoyant for a considerable height. The convection inhibition (CIN) which represents the energy barrier that a rising parcel at LFC would have to overcome before it could ascend on its own, and the convective available potential energy (CAPE), a measure of the vertical instability of the atmosphere under moist convection, are shown in Fig. 8. Here CAPE and CIN are calculated from the integral equations given by Williams and Renno (1993), i.e.,

$$\text{CAPE} = -\int_{\text{LFC}}^{\text{LNB}} (T_{\text{vp}} - T_{\text{ve}}) R_{\text{D}}\, \text{d}(\ln p), \quad \text{CIN} = \int_{p_{\text{i}}}^{\text{LFC}} (T_{\text{vp}} - T_{\text{ve}}) R_{\text{D}}\, \text{d}(\ln p) \quad (1)$$

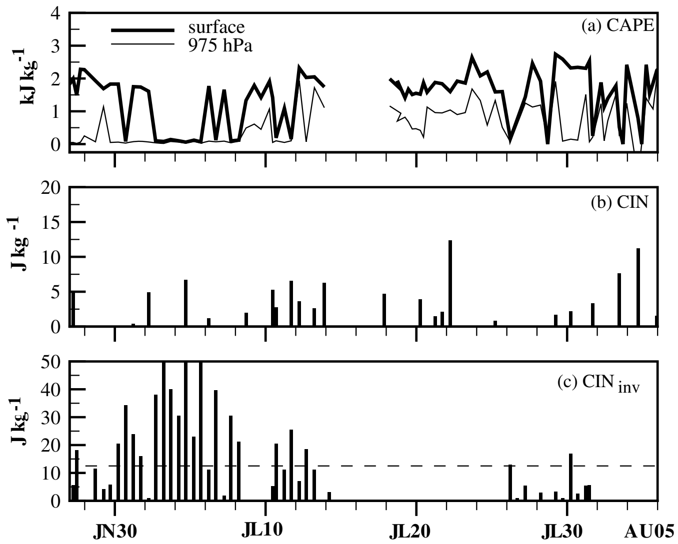

Figure 8. Time series for the surface air at TS1 and TS2 in 2002: (a) CAPE (thick line); (b) CIN and (c) $\mathrm{CIN_{inv}}$. Also shown in (a) is CAPE for a parcel originating at 975 hPa (thin line), approximately the midpoint of the atmospheric mixed layer for the July 2002 soundings over the Arabian Sea. In (c), the pecked line denotes $\mathrm{CIN_{inv}} = 12.5\ \mathrm{J\ kg^{-1}}$, a limiting value of $\mathrm{CIN_{inv}}$ based on an undiluted parcel assumption. Typically, tropical convective clouds do not penetrate such inversions. See text or captions of Figs. 2 and 7 for definitions.

where R_D is the gas constant of dry air, T_vp and T_ve are respectively the virtual temperatures of the parcel and the environment at pressure p, and p_i is parcel starting level pressure. Following Williams and Renno (1993), a pseudo-moist adiabatic process is used here for the computation of CAPE. However, the pseudo-moist adiabatic process is not appropriate for estimating CIN since CIN involves parcel buoyancy near LCL where all the condensed water is likely to be retained in the parcel. Therefore, T_vp based on a reversible moist adiabatic process is appropriate for estimating CIN, and present values of CIN are based on this process. It is observed from Fig. 7 that the parcel buoyancy is positive for a short distance above LFC and then turns negative inside the inversion. The inversion not only increases the vertical stability of the atmosphere but may completely inhibit the development of deep convective clouds. The (negative) area inside the inversion enclosed between the environmental sounding and the moist adiabat represents the energy barrier that the parcel has to overcome in order to be able to penetrate the inversion. To quantify the inhibition to cloud growth by the lower inversion, the inhibition energy due to the inversion ($\mathrm{CIN_{inv}}$) is defined here as

$$\mathrm{CIN_{inv}} = -\int_{p_1}^{p_2} (T_\mathrm{vp} - T_\mathrm{ve}) R_\mathrm{D}\, \mathrm{d}(\ln p), \tag{2}$$

where p_1 (= LNB) and p_2 are the lower and upper limits within the inversion-related stable layer where the parcel buoyancy is negative. Since the inversion is within a kilometre or two of the cloud base, it is not unrealistic to assume that the liquid water is carried with the parcel. This assumption is a key element in the reversible moist adiabatic process which is the basis for integrating Eq. (2).

Figure 8(a) shows that on several days in July, CAPE values were less than $0.1\ \mathrm{kJ\ kg^{-1}}$. These were days with a strong inversion. If these days are not considered, then the average value of CAPE of the surface air is about $1.7\ \mathrm{kJ\ kg^{-1}}$. Considering that this CAPE corresponds to that of the surface air over a warm sea with SST in the range

28.5 °C to 29 °C under weak convective conditions (with θ_e around 360 K), values about 1.7 kJ kg^{-1} may be considered small. Consider, for example, that average values of CAPE of 2–3 kJ kg^{-1} over the tropical West Pacific and Atlantic oceans (Williams and Renno 1993), and around 3 kJ kg^{-1} over the BoB during a break monsoon at comparable values of SST (Bhat 2001). CIN (Fig. 8(b)) is small and rather compared to the BoB (Bhat 2001) and TOGA-COARE atmospheres (Kingsmill and Houze 1999). $\mathrm{CIN_{inv}}$ was more than 10 J kg^{-1} on a majority of the days during the first half of July (Fig. 8(c)). Note that $\mathrm{CIN_{inv}}$ was either zero or comparatively small during the second half of July. A tentative critical value for $\mathrm{CIN_{inv}}$ above which a cumulus cloud cannot grow past the inversion can be estimated as follows. At an altitude of ~2 km above the surface over the tropical oceans, the only source of energy available for an ascending cloud-parcel to overcome $\mathrm{CIN_{inv}}$ is its kinetic energy. In the lowest two kilometres, vertical velocities inside tropical convective clouds are less than 5 m s^{-1} (typically 2–3 ms^{-1}) (e.g., Lucas *et al.* 1994)—the corresponding maximum value of kinetic energy is less than 12.5 J kg^{-1}. Therefore, when the $\mathrm{CIN_{inv}}$ is more than 12.5 J kg^{-1}, such clouds do not penetrate the inversion. This limiting value of $\mathrm{CIN_{inv}}$ assumes that the parcel is undiluted, an essential part of both the pseudo-moist adiabatic and the reversible moist adiabatic assumptions.

In the real atmosphere, mixing with the environmental air invariably takes place. Since the inversion layer air was very dry, evaporative cooling as it mixes with the environmental air in this region significantly reduces the parcel buoyancy. Consequently, 12.5 J kg^{-1} may be viewed as an upper limit and we may expect cloud growth to be prevented at even lower values of $\mathrm{CIN_{inv}}$.

4. Discussion

The results presented in section 3 show that both SST and surface air θ_e were well above the respective convection threshold values over AS during July 2002. Absence of convection in the presence of a positive CAPE is generally attributed to high values of CIN (e.g., Emanuel 1994, p.171). However, during the first half of July 2002, CIN did not suppress deep cloud activity over AS: values of CIN there were less than those over other equally warm tropical oceans where atmospheric convection did occur. From the local thermodynamics point of view, it was a strong and persistent low level inversion that prevented the development of deep convective clouds over the study area during most days in the first half of July 2002. This argument does not explain why rains failed during the second half of July, particularly 18–24 July when inversions were either weak or absent, CIN was low and CAPE values were around 2 kJ kg^{-1}. During this period (20–25 July in particular), surface pressure increased rapidly (Fig. 4(b)) which, in this part of the world, tends to be associated with reduced large-scale convection. During 28 July–1 August, values of CIN were nearly zero and CAPE around 2 kJ kg^{-1}, however the low level inversion reappeared and $\mathrm{CIN_{inv}}$ was nonzero. Thus, when viewed locally, strong low-level inversions and a high-pressure regime together made conditions unfavourable for the development of deep convection during most of July (Fig. 9). There were narrow windows when local conditions favoured convection but deep clouds did not develop. Basically, vertical instability (thermodynamic) constraints such as positive CAPE and small CIN are necessary but not sufficient conditions for organized deep convection. The large-scale dynamics/circulation also needs to be conducive to deep convection.

Warm and dry air in the lower inversion could have come from the boundary layer (BL) over the deserts that surround AS to the west and north, or have subsided from

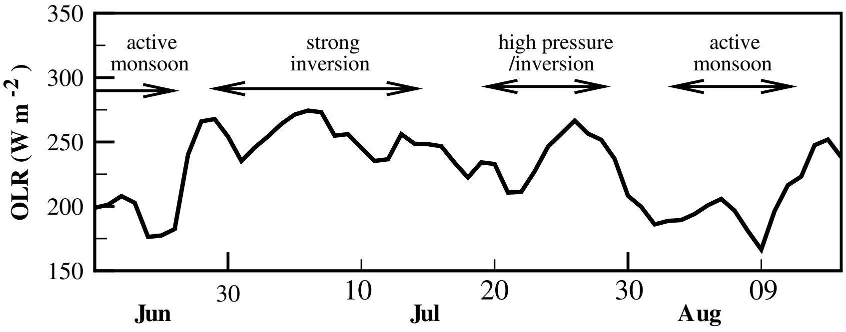

Figure 9. Time series of outgoing long-wave radiation (OLR) (W m^{-2}) over the ASWC box, along with the atmospheric/surface conditions that prevailed there during the period June–August 2002. The box covers the eastern Arabian Sea and the west coast of India between 10°N–20°N and 70°E–75°E.

above. The inversion in the western AS near the African coast, for example, has been attributed to the advection of BL air from the neighbouring deserts (Colon 1964). While advection cannot be ruled out at the ARMEX ship location (distance about 1500 km from the deserts), subsidence may also have contributed. Subsiding air, when not of BL origin, is an outflow from a convective system and saturated at that instant. If this air does not mix with air from other sources, its specific humidity q is conserved. This information can be used to locate the initial height of the inversion air. Figure 10 shows the vertical profile of air temperature and specific humidity q at the ship location on 4 July 2002. Values of q fall as one ascends through the stable layer and are a minimum at, or just above the top of each inversion. This suggests a laminated atmosphere in which the air in stable layers resists vertical exchange and that air immediately above each inversion had a source different from (and probably from a higher level than) that of the air in the layer below it. Figure 10 also shows saturation specific humidities q_S corresponding to temperatures representative of a convecting tropical atmosphere (corresponding to a moist adiabat of a surface-air parcel with an initial temperature of 27.5 °C and 80% RH), and of an average of three mid-latitude soundings taken on 25 and 26 June 2002 at Divnoe Observatory (45.91°N, 43.35°E). The mid-latitude sounding is considered here because back-trajectory analysis (to be discussed next) suggested that, on some days, inversion air came from north of 30°N. In Fig. 10 we may see that if the lower inversion layer air subsided from a convective system, then, it descended from 6–8 km to 2 km level, i.e., the originating level was at least a kilometre above the melting level. Air in the upper inversion (located near 7 km) came from between 11 and 12 km.

To explore the possible geographic location of the inversion air, three-dimensional back-trajectory analysis has been carried out using 6-hourly ERA-40 winds. The horizontal locations of the air parcels during the eight days before they arrived at 800 hPa above the ship's position are shown in Fig. 11. Figure 11(a) shows that, during the first half of July 2002, air came frequently from over the desert regions to the north and north-west of AS. During the first half of August (Fig. 11(b)), this trend reduced, but not completely. The vertical paths (not shown) show the prevalence of subsidence (Bhat 2005). Going further back in time takes the trajectories north of 30°N, and the warm dry air of the lower inversion seems to be of mid-latitude origin. In the corresponding periods in 2001 (a normal monsoon year), by contrast, air occupying the region of the present study was of marine origin (Fig. 11(c) and (d)). Consequently, we may conclude that the low-level large-scale circulation in the AS region during July and August 2002 was significantly different from that of a normal monsoon year.

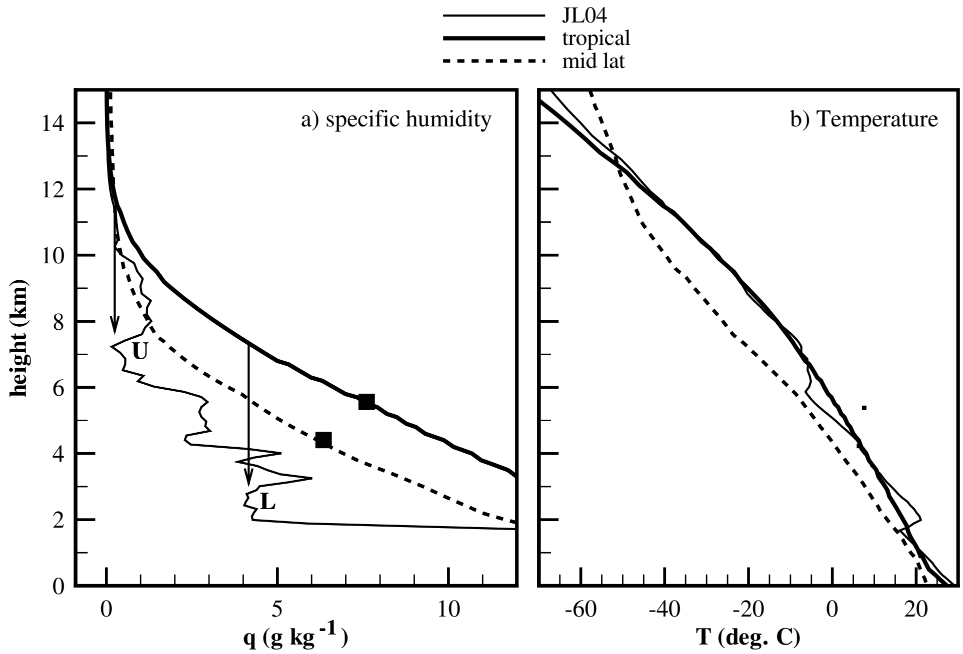

Figure 10. Vertical profiles (thin lines) at the ship location TS1 (16.9°N, 71.2°E) on 4 July 2002, and typical values for tropical (thick lines) and mid-latitude atmospheres (dashed lines): (a) specific humidity, q (g kg^{-1}), and (b) air temperature (°C). In (a) the typical tropical and mid-latitude values are of saturation specific humidity, q_s (g kg^{-1}); black squares indicate the melting level; 'L' and 'U' denote the dry layers above the lower and upper inversions and the downward pointing arrow at top left denotes the path on the diagram of air subsiding from mid-latitudes. Comparison of (a) with (b), level for level, shows that values of q on 4 July 2002 fell with height through the stable layer and were a minimum at, or just above, the top of each inversion.

5. CONCLUSIONS

1. The Indian drought of 2002 was atypical. While the ISMR deficit was 21.5%, there was no seasonal trend in the monthly-rainfall anomaly. The drought was primarily a consequence of below-normal rainfall for a period of 36 continuous days centred around July.

2. There were no clear precursor signals to a major drought before June 2002. Several factors, unfavourable for an active monsoon over India developed/prevailed over the Indo-Pacific region during June–July 2002. They include the negative phase of EQUINOO in the equatorial Indian Ocean, high MJO and typhoon activity in the central and west Pacific.

3. ARMEX observations provided information on the conditions that prevailed over the eastern Arabian Sea during the driest July over India in recorded history. Total rainfall measured on board the ship was less than 20 mm during July 2002. The sea surface conditions were conducive to deep convection. CAPE values were 1.5–2.5 kJ kg^{-1} on many days and CIN values were low. A persistent and strong inversion near 800 hPa prevented the growth of convective clouds during the first half of July and also towards its end.

4. A second inversion above the melting level was frequently observed.

5. Whilst the presence of an inversion over the eastern Arabian Sea and the west coast of India is not unique to July 2002, its strength and persistence were.

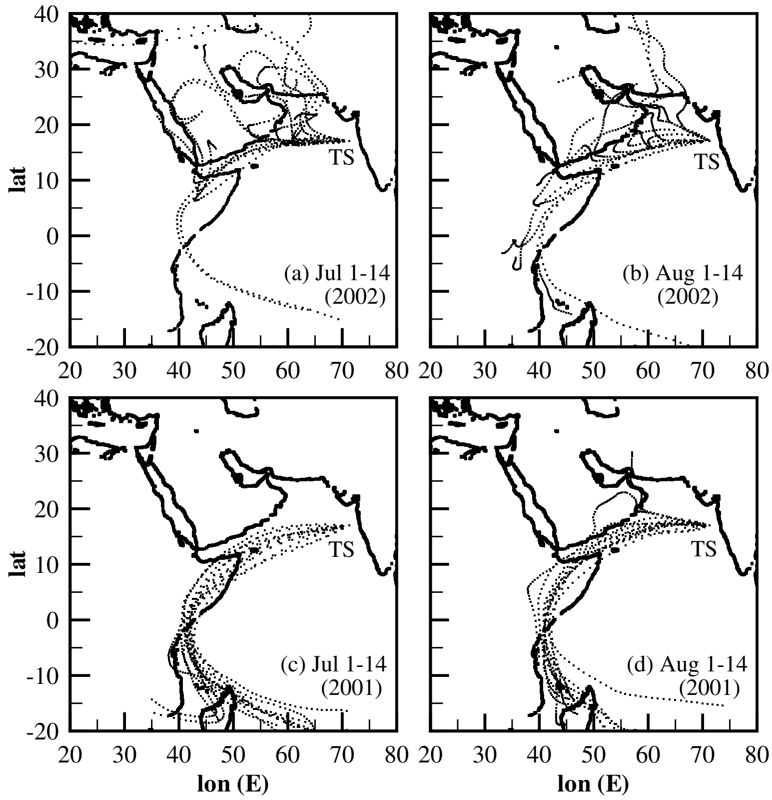

Figure 11. Horizontal projections (dotted lines) of three-dimensional back-trajectories from 800 hPa at location TS1 (16.9°N, 71.2°E) during two periods in 2002 and the corresponding periods in 2001. The three-dimensional path for each day was retraced for the previous eight days, using ERA-40 data from the European Centre for Medium-Range Weather Forecasts.

6. Back-trajectory analysis shows the prevalence of air from over the deserts at 800 hPa over the eastern Arabian Sea during July and August 2002 in contrast to the marine air from the southern Indian Ocean in a normal year.

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REFERENCES

Annamalai, H. and Slingo, J. M. 2001 Active/break cycles: diagnosis of the intraseasonal variability of the Asian summer monsoon. *Clim. Dyn.*, **18,** 85–102

Annamalai, H. and Sperber, K. R. 2005 Regional heat sources and the active and break phases of boreal summer intraseasonal (30–50 day) variability. *J. Atmos. Sci.*, **62,** 2726–2748

Bhat, G. S. 2001 Near surface atmospheric characteristics over the north Bay of Bengal during the Indian summer monsoon. *Geophys. Res. Lett.*, **28,** 987–990

2005 Convection inhibition energy of the inversion and the suppressed rainfall over the Arabian Sea during July 2002. *Mausam*, **56,** 89–96

Bhat, G. S., Gadgil, S., Hareesh Kumar, P. V., Kalsi, S. R., Madhusoodanan, P., Murty, V. S. N., Prasada Rao, C. V. K., Ramesh Babu, V., Rao, L. V. G., Rao, R. R., Ravichandran, M., Reddy, K. G., Sanjeera Rao, P., Sengupta, D., Sikka, D. R., Swain, J. and Vinayachandran, P. N. 2001 BOBMEX — the Bay of Bengal Monsoon Experiment. *Bull. Am. Meteorol. Soc.*, **82,** 2217–2243

Chen, T.-C., Weng, S.-P., Yamazaki, N. and Kiehne, S. 1998 Interannual variation in the tropical cyclone formation over the western north Pacific. *Mon. Weather Rev.*, **126,** 1080–1090

Clark, C. O., Cole, J. E. and Webster, P. J. 2000 Indian Ocean SST and Indian summer monsoon rainfall: Predictive relationships and their decadal variability. *J. Climate*, **13,** 2503–2519

Colon, J. A. 1964 On interactions between the southwest monsoon current and the sea surface over the Arabian Sea. *Indian J. Meteorol. Geophys.*, **15,** 183–200

Emanuel, K. A. 1994 *Atmospheric convection.* Oxford University Press, New York, NY, USA

Fein, J. S. and Kuettner, J. P. 1980 Report on the summer MONEX field phase. *Bull. Am. Meteorol. Soc.*, **61,** 461–474

Francis, P. A. and Gadgil, S. 2006 Intense rainfall events over the west coast of India. *Meteorol. Atmos. Phys.*, doi: 10.1007/s00703-005-0167-2

Gadgil, S. and Joseph, P. V. 2003 On breaks of the Indian monsoon. *Proc. Indian Acad. Sci. (Earth Planet Sci).*, **112,** 529–558

Gadgil, S., Joseph, P. V. and Joshi, N. V. 1984 Ocean-atmosphere coupling over monsoonal regions. *Nature*, **312,** 141–143

Gadgil, S., Srinivasan, J., Nanjundiah, R. S., Krishna Kumar, K., Munnot, A. A. and Rupa Kumar, K. 2002 On forecasting the Indian summer monsoon: the intriguing season of 2002. *Current Science*, **83,** 394–403

Gadgil, S., Vinayachandran, P. N. and Francis, P. A. 2003 Droughts of the Indian summer monsoon: Role of clouds over the Indian Ocean. *Current Science*, **85,** 1713–1719

Gadgil, S., Vinayachandran, P. N., Francis, P. A. and Gadgil, S. 2004 Extremes of the Indian summer monsoon rainfall, ENSO and equatorial Indian Ocean oscillation. *Geophys. Res. Lett.*, **31,** L12213

George, P. A. 1956 Effect of off shore vortices on rainfall along the west coast of India. *Indian J. Meteorol. Geophys.*, **7,** 225–240

Grossman, R. L. and Durran, D. R. 1984 Interaction of low-level flow with the Western Ghat mountains and offshore convection in the summer monsoon. *Mon. Weather Rev.*, **112,** 652–672

Harr, P. A. and Elsberry, R. L. 1991 Tropical cyclone track characteristics as a function of large-scale circulation anomalies. *Mon. Weather Rev.*, **119,** 1448–1468

Holt, T. and Raman, S. 1987 A study of mean boundary-layer structures over the Arabian Sea and the Bay of Bengal during active and break monsoon periods. *Boundary-Layer Meteorol.*, **38,** 73–94

Huffman, G. J., Adler, R. F., Arkin, P., Chang, A., Ferraro, R., Gruber, A., Janowiak, J., McNab, A., Rudolf, B. and Schneider, U. 1997 The global precipitation climatology project (GPCP) combined precipitation dataset. *Bull. Am. Meteorol. Soc.*, **78,** 5–20

Johnson, R. H., Ciesielski, P. E. and Hart, K. A. 1996 Tropical inversions near the 0 °C level. *J. Atmos. Sci*., **53,** 1838–1855

Kalsi, S. R., Hatwar, H. R., Jayanthi, N., Subramanian, S. K., Shyamala, B., Rajeevan, M. and Jenamani, R. K. 2004 'Various aspects of unusual behaviour of monsoon 2002'. *IMD Met. Monograph Synoptic Meteorology No. 2/2004*, National Climate Centre, India Meteorology Department, Lodi Road, New Delhi, India*

Kingsmill, D. E. and Houze, Jr., R. A. 1999 Thermodynamic characteristics of air flowing into and out of precipitating convection over the West Pacific warm pool. *Q. J. R. Meteorol. Soc*., **125,** 1209–1229

Krishnamurti, T. N. and Surgi, N. 1987 'Observational aspects of summer monsoon'. In *Monsoon Meteorology*, C.-P. Chang and T. N. Krishnamurti (Eds.), Oxford University Press, New York, NY, USA Dynamics of breaks in the Indian summer monsoon. *J. Atmos. Sci.*, **57,** 1354–1372

Kumar, V. and Krishnan, R. 2005 On the association between the Indian summer monsoon and tropical cyclone activity over north-west Pacific. *Current Science*, **88,** 602–612

Lucas, C. and Zipser, E. J. 2000 Environmental variability during TOGA COARE. *J. Atmos. Sci.*, **57,** 2333–2350

Lucas, C., Zipser, E. J. and Lemone, M. A. 1994 Vertical velocity in oceanic convection off tropical Australia *J. Atmos. Sci.*, **51,** 3183–3193

McPhaden, M. J. 2004 Evolution of the 2002/03 El Niño. *Bull. Am. Meteorol. Soc.*, **85,** 677–695

Miller, F. R. and Keshavamurthy, R. N. 1968 'Structure of an Arabian Sea summer monsoon system'. *I. I. O. E. Meteorological Monograph 1*, East–West Centre Press, Honolulu, Hawaii, USA*

Molteni, F., Corti, S., Ferranti, L. and Slingo, J. M. 2003 Predictability experiments for the Asian summer monsoon: Impact of SST anomalies on interannual and intraseasonal variability. *J. Climate*, **16,** 4001–4021

Mooley, D. A. and Shukla, J. 1987 'Variability and forecasting of the summer monsoon rainfall over India'. Pp. 26–59 in *Monsoon Meteorology*. Eds. C.-P. Chang and T. N. Krishnamurti. Oxford University Press, New York, NY, USA

Narayanan, M. S., Rao, B. M., Shah, S., Prasad, V. S. and Bhat, G. S. 2004 Role of atmospheric stability over the Arabian Sea and the unprecedented failure of monsoon 2002. *Current Science*, **86,** 101–111

Pant, M. C. 1982 Some characteristic features of the low-level jet field over the Arabian Sea during the Indian summer monsoon. *Mausam*, **33,** 85–90

Parthasarathy, B., Munot, A. A. and Kothawale, D. R. 1995 'Monthly and seasonal rainfall series for all-India homogenous regions and meteorological subdivisions: 1871–1994'. *IITM Research Report-RR065*, Indian Institute of Tropical Meteorology, Pune, India*

Rajeevan, M., Pai, D. S. and Thapliyal, V. 2002 Predictive relationships between Indian Ocean sea surface temperatures and Indian summer monsoon rainfall. *Mausam*, **53,** 337–348

Raman, C. R. V. 1955 Breaks in Indian southwest monsoon and typhoons in southwest Pacific. *Current Science*, **24,** 219–220

Raman, C. R. V. and Rao, Y. P. 1981 Blocking highs over Asia and droughts over India. *Nature*, **289,** 271–273

Rao, P. S. 2005 Arabian Sea monsoon experiment: An overview. *Mausam*, **56,** 1–6

Rao, Y. P. 1976 'Southwest monsoon'. *Meteorological Monograph Synoptic Meteorology, No. 1*, India Meteorological Department, Lodi Road, New Delhi, India*

Rasmusson, E. M. and Carpenter, T. H. 1983 The relationship between eastern equatorial Pacific sea surface temperatures and rainfall over India and Sri Lanka. *Mon. Weather Rev*., **111,** 517–528

* Reports are normally available from the originating institution.

Saha, K., Sanders, F. and Shukla, J. 1981 Westward propagating predecessors of monsoon depressions. *Mon. Weather Rev*., **109,** 330–343

Saith, N. and Slingo, J. M. 2006 The role of the Madden-Julian oscillation in the Indian drought of 2002. *Int. J. Climatol.* **26,** 1361– 1378

Sikka, D. R. 2003 Evaluation of monitoring and forecasting of summer monsoon over India and a review of monsoon drought of 2002. *Proc. Indian Natl. Acad*., **69A,** 479–504

Simmons, A. J. and Gibson, J. K. 2000 'The ERA-40 project plan'. *ERA-40 Project Report Series No. 1*, European Centre for Medium-Range Weather Forecasts, Shinfield Park, Reading, UK*

Wang, J., Cole, H. C., Carlson, D. J., Miller, E. R., Beierle, K., Paukkunen, A. and Laine, T. K. 2002 Corrections of humidity measurement errors from the Väisälä RS80 radiosonde – Application to TOGA COARE data. *J. Atmos. Oceanic Technol*., **19,** 981–1002

Webster, P. J., Magana, V. O., Palmer, T. N., Shukla, J., Tomas, R. A., Yanai, M. and Yasunari, T. 1998 Monsoons: Processes, predictability, and prospects for prediction. *J. Geophys. Res*., **103,** 14451–14510

Webster, P. J. and Hoyos, C. 2004 Prediction of monsoon rainfall and river discharge on 15–30-day time-scales. *Bull. Am. Meteorol. Soc*., **85,** 1745–1765

Williams, E. and Renno, N. 1993 An analysis of the conditional instability of the tropical atmosphere. *Mon. Weather Rev.*, **121,** 21–36