SUMMARY

Using the ECMWF re-analysis, OLR and precipitation data during 1979-93, this paper examines major factors contributing to the onset of the Australian summer monsoon. The low-level (850 hPa) westerly wind and convective activity (OLR, precipitation) over a region in northern Australia (NAU) are used to determine the onset dates. Composite results are then made based on these dates chosen in 1979-93. Daily Q_1 (apparent heat source) and Q_2 (apparent moisture sink) are obtained as residuals of the large-scale heat and moisture budgets for clarifying the roles of various heating processes in the onset. Four major factors contributing to the onset are identified: (1) land-sea thermal contrast, (2) barotropic instability, (3) arrival of the Madden-Julian oscillation (MJO), and (4) intrusion of midlatitude troughs. The thermal contrast due to differential heating between land and sea acts as a seasonal preconditioning for the onset. The sensible heating over the continent leads to a reversal of meridional temperature gradient between the Australian continent and the Arafura Sea in a layer below 800 hPa in September-March, and sets up a thermally induced meridional-vertical circulation which helps to transport low-level moist air inland. The criterion of the barotropic instability is often met at 850 hPa in NAU several days prior to the onset. The sudden onset is then triggered by the arriving MJO and at times by the intrusion of a midlatitude trough.

KEYWORDS: Heating processes over Australia, Land-sea thermal contrast, Barotropic instability, Madden-Julian oscillation

1. Introduction

It has been believed for many centuries that the land-sea thermal contrast is important for monsoons (e.g. Webster 1987; Webster et al. 1998), although some authors (e.g. Chao 2000; Chao and Chen 2001) challenged this view and argue that the existence of land is not a necessary condition for the presence of monsoons. Basically, landmass has two major effects on the large-scale atmospheric motion: (1) mechanical forcing due to topography, and (2) thermal forcing due to land-sea contrast. To isolate clearly the role of the thermal effect, it is desirable to find a monsoon region without high lands. The Australian summer monsoon seems an ideal candidate for examining the thermal effect in a pure form, because flat land surface of Australia has no complexity of topography.

The Australian summer monsoon usually begins in December and ends in March. Its mature stage is characterized by heavy precipitation and low-level westerly wind in northern Australia and the Arafura Sea (McBride 1987; Manton and McBride 1992). This monsoonal nature of northern Australia climate (the seasonal precipitation and reversals of wind directions) was recognized in the early 20th century (Hunt et al. 1913). In 1960s, several studies (e.g. Berson 1961) showed that the Australian summer monsoon has similarities to the Asian summer monsoon, such as its abrupt poleward shift, sudden onset over a large region and intraseasonal active-break periods.

There were not many works on the Australian monsoon in literature until the First Global Atmospheric Research Program (GARP) Global Experiment (FGGE) in 1978-79 and the Australian Monsoon Experiment (AMEX) in 1986-87. The FGGE data made diagnostic analyses possible for the large-scale circulation associated with the Australian summer monsoon (e.g. Murakami and Sumi 1982a, b; Davidson et al. 1983). The AMEX observations, on the other hand, provided high-density synoptic and mesoscale data over northern Australia. The AMEX phase II field work lasted 36 days from 10 January 1987 which was 4 days before the monsoon onset (Hendon et al. 1989). The success of the AMEX led to several detailed observational results, such as the vertical heating profiles in the northern Australian monsoon region (Frank and McBride 1989).

Besides these results from the special experiments, studies with longer records were made by several authors. Hendon and Liebmann (1990a, b) suggested that the Madden-Julian oscillation (MJO; Madden and Julian 1971, 1972) acts as a trigger for the monsoon onset based on a composite study from station data obtained at Darwin in 1957-87. On the other hand, Davidson et al. (1983) studied the synoptic situations prior to the monsoon onset using operational analyses, and argued that the midlatitude events plays a role in the onset.

Despite these studies, the Australian summer monsoon has received relatively little attention compared to the Asian summer monsoon. Several gaps in our existing knowledge need to be addressed:

(i) The AMEX happened to be in an El Nino year. A late onset and large-scale sinking anomalies were observed in northern Australia. We feel that it is necessary to study large-scale atmospheric conditions for many years and use the composite method to describe more general situations.

(ii) The composite studies from previous works were mainly from station data. The ”re-analysis” datasets are now available for us to study 3-dimensional features of the Australian monsoon. The classic sea-breeze model associated with the land/sea thermal contrast can be re-examined. It is also possible to apply the dynamic instability theories to the monsoon onset.

(iii) In previous studies, the large-scale heat and moisture budgets have not been seriously examined in the Australian region. With the re-analysis datasets, heat and moisture budgets can be calculated for a more systematic examination of thermodynamic effects. For example, the role of sensible heating over the Australian continent can be studied.

(iv) Previous observational works focus on different onset processes separately (e.g. MJO triggering, mid-latitude events). It is desirable to present an unified view of the onset mechanisms and identify major factors contributing to the onset with composite results.

In this paper, we used the re-analysis dataset to calculate large-scale heat and moisture budgets and the role of the land-sea thermal contrast in the Australian summer monsoon is re-examined. In section 2, the datasets, the definition of the onset and the composite method are described. In section 3, the composite features showing the heating profiles before/after the monsoon onset are presented. In section 4, four major factors contributing to the onset are discussed. Finally, section 5 presents conclusions.

2. Data and the definition of the onset

a. Data

The primary data used for this work is the 15-year (1979-93) European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA; Gibson et al. 1997). The daily outgoing longwave radiation (OLR) measurements and monthly Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997) for the same period are also used. In addition, large-scale heat and moisture budget residuals, *apparent heat source* Q_1 and *apparent moisture sink* Q_2 (Yanai et al. 1973), are defined and computed by:

$$Q_1 \equiv c_p(\frac{p}{p_o})^{\kappa} \left(\frac{\partial \overline{\theta}}{\partial t} + \overline{\mathbf{v}} \cdot \nabla \overline{\theta} + \overline{\omega} \frac{\partial \overline{\theta}}{\partial p} \right) \tag{1}$$

and

$$Q_2 \equiv -L \left(\frac{\partial \overline{q}}{\partial t} + \overline{\mathbf{v}} \cdot \nabla \overline{q} + \overline{\omega} \frac{\partial \overline{q}}{\partial p} \right) \tag{2}$$

where θ is the potential temperature, q the mixing ratio, $\mathbf{v}$ the horizontal velocity, ω the vertical p-velocity. L is the latent heat of vaporization, $\kappa = R/c_p$ with R the gas constant and c_p the specific heat capacity at constant pressure of dry air, and p_0 = 1000 hPa. The overbar denotes the running horizontal average over a large-scale area.

In (1) and (2), the horizontal wind components, potential temperature, and mixing ratio are directly obtained from the ERA products. However, for the accuracy of Q_1 and Q_2, the vertical p-velocity ω is recomputed following the method described in Tung et al. (1999) and Tung and Yanai (2002). The 6-hourly estimates of Q_1 and Q_2 are averaged into daily mean values for the composite study, and a constant number, c_p^{-1}, is multiplied for Q_1 and Q_2 in order to express them in units of equivalent warming/cooling rate (K d^{-1}).

The Q_1 and Q_2 are interpreted by:

$$Q_1 = Q_R + L(\overline{c} - \overline{e}) - \nabla \cdot \overline{s'\mathbf{v}'} - \frac{\partial \overline{s'\omega'}}{\partial p} \tag{3}$$

and

$$Q_2 = L(\overline{c} - \overline{e}) + L\nabla \cdot \overline{q'\mathbf{v}'} + L\frac{\partial \overline{q'\omega'}}{\partial p} \tag{4}$$

where Q_R is radiative heating rate, c and e are condensation and evaporation per unit mass of air respectively, s the dry static energy, and the prime denotes the deviation from the running average. Equation (3) represents the total effect of radiative heating, latent heat released by net condensation $L(\overline{c} - \overline{e})$, and the horizontal and vertical convergence of fluxes of sensible heat due to subgrid-scale eddies such as cumulus convection and turbulence. Equation (4), on the other hand, represents the total effect of net condensation and divergence of eddy moisture fluxes due to subgrid-scale eddies. It is customary to ignore the contribution of the eddy horizontal transport terms $\nabla \cdot \overline{s'\mathbf{v}'}$ in Eq. (3) and $\nabla \cdot \overline{q'\mathbf{v}'}$ in Eq. (4) in convection-large-scale interaction (e.g. Yanai and Johnson 1993).

b. Definition of the onset

Traditionally, the onset of the Australian summer monsoon has been defined using records at Darwin (12°S, 130°E) or stations in its vicinity (e.g. Troup 1961; Nicholls et al. 1982; Hendon and Liebmann 1990a). In the present study, in order to capture the large-scale circulation changes before and after the onset, low-level (850 hPa) westerly wind and convective activity (OLR, precipitation) over a region including northern Australia and the Arafura Sea (2-15°S, 115-150°E; hereafter NAU; see Fig. 1) are used to determine the onset dates.

The first day with the average 850 hPa zonal wind exceeding 2 m s^{-1} in NAU is chosen as the onset day when the westerly wind is sustained longer than 10 days and the OLR is lower than 210 W m^{-2} for at least several days during the 10-day period. Figure 2 shows an example of mean zonal winds for NAU in the 1979-80 Australian summer monsoon season. The monsoon westerlies abruptly increased in a layer below 500 hPa during late December - early January over NAU (Fig. 2a). This low level westerlies appeared with strong upper level easterlies. The value of average 850 hPa zonal wind (Fig. 2b) along with the OLR data (not shown) then defined the onset date as 28 December for this case. In this example, two monsoon active periods can be recognized in December - February, and a break (850 hPa zonal wind lower than 2 m s^{-1}) is observed for 16 - 24 January.

Based on the same definition, the onset dates for monsoon seasons in 1979-93 were chosen and listed in Table 1. The onset dates selected by the present study are close to previous results. The 14-event mean onset date is 25 December which is the same as the result by Hendon and Liebmann (1990a) for 30 years (1957-87). The standard deviation of the onset date obtained in the present study is 14 days, which is close to previous results (16 days by Hendon and Liebmann 1990a; 15 days by Holland 1986).

The 14-event composite of all variables were made relative to the onset dates. The ”day 0” denotes the onset day, and positive (negative) values of dates represent the days after (before) the onset. Figure 3 is an example showing the composite of zonal wind component. It is clearly seen in NAU that the lower level westerlies abruptly increased after day 0, because we define the onset day mainly based on the 850 hPa zonal wind (Fig. 3a). However, the composite does show some other interesting features. The lower level westerlies appear with upper level easterlies which are strongest at 100 hPa. The first monsoon active period lasts about 30 days. After a monsoon break (about 6 days), the second active monsoon period starts and remains slightly longer than the first one. This is the composite result, so it only shows the general features and does not mean that every year has the same active and break periods. However, the standard error of zonal wind at every level is calculated for the composite mean, and the value of it is about 1 m s^{-1} for layers lower than 500 hPa. Therefore, we may conclude that there are about two active periods in a monsoon season. We examine all the 14 monsoon cases, and 11 of them have two monsoon active periods within 90 days from the onset day.

In order to compare the atmospheric structure in NAU and that over the Australian continent, we define another region, the Australian continent (AUC), along the same longitudinal boundaries as NAU, but shifted poleward to cover the whole continent (15-35°S, 115-150°E; see Fig. 1). The composite zonal wind field over the AUC shows that the westerly jet is at the level of about 200 hPa (Fig. 3b). It moves poleward to a place outside of the AUC as season progresses from spring to summer, so the averaged zonal wind over the AUC gets weak before the onset. The poleward movement of the jet has been documented in an early study by Radok and Grant (1957).

3. Composite features of heat sources and moisture sinks

It is well-known that the Tibetan Plateau acts as an elevated heat source in NH spring-summer (e.g. Yanai et al. 1992). The heating leads to a reversal of the meridional temperature gradient south of the plateau in the upper troposphere (500-200 hPa), and the onset of the Asian summer monsoon is concurrent with it (e.g. Flohn 1957; Li and Yanai 1996). In contrast, the upper troposphere does not exhibit similar structure in SH spring-summer in the Australian summer monsoon region. The Australian continent does not have such high lands to produce large-scale elevated heating, and the sensible heating over the AUC is observed only in the lower level (below 750 hPa).

The vertical distributions of heating over the Australia have been examined by previous studies. Using data from the Wangara experiment (from 15 July to 27 August, 1967; Clarke et al. 1971), surface heat flux over southeastern Australian continent has been studied (e.g. Yamada and Mellor

1975). Schaack et al. (1990) used the ECMWF Global Weather Experiment (GWE) data from December 1978 to November 1979 to calculate the 3-dimensional heating rate through a vertical integration of the isentropic mass continuity equation. They found that the maximum Q_1 over northern Australia in January is about 2.3 K d^{-1} and its peak is located between 500-400 hPa. With their method, however, Q_2 is not obtainable. Frank and McBride (1989) used the data obtained from the AMEX (13 January - 14 February 1987) rawinsonde network to calculate the heat and moisture budgets at 6-hour intervals. The network is located in the Gulf of Carpentaria, so the results mostly represent the vertical heating distribution over northern Australia.

Our results provide a more general heating structure by making composites from data in 1979-93. The vertical profile of Q_1 over NAU shows the maximum heating between 500-400 hPa after the onset, although there is relatively weak heating prior to the onset in about the same layer (Fig. 4a). The maximum Q_1 does not move to other layers in vertical after the onset. This agrees with Frank and McBride (1989) that the maximum Q_1 does not change too much in height during different stages of convective systems.

After the onset, the vertical Q_2 profile over NAU (Fig. 4b) has a peak in a layer between 800 and 650 hPa which is lower than the peak of Q_1 in height. This suggests the contributions from eddy vertical transports due to deep cumulus convection in the Australian summer monsoon region. The well-separated Q_1 and Q_2 peaks in vertical are typical features of a deep convective atmosphere (Yanai et al., 1973). Besides this lower Q_2 maximum, the distribution exhibits a second peak of Q_2 in a layer between 500-400 hPa in some periods. This second peak is located at about the same pressure level as the Q_1 peak. It indicates the presence of stratiform rain in NAU during the monsoon season.

In contrast, the vertical distribution of Q_1 over the AUC has a very different structure. Significant heating confined in a shallow layer (surface-750 hPa) is observed (Fig. 5a). Over the AUC, Q_2 (Fig. 5b) is negative (about -0.5 K d^{-1}) below 700 hPa. The negative Q_2 value implies the moistening due to net evaportation near surface. All the values of Q_1 below 750 hPa ($\sim$ 2.5

K d^{-1}) before and after onset are larger than Q_2, and the maximum Q_1 is near surface. With the evidence of a warm surface (e.g. Yamada and Mellor 1975), the strong heating observed in the shallow layer is concluded to be caused by sensible heat flux from ground surface. Actually, the sensible heating in a layer below 750 hPa over the Australian continent starts in September prior to the monsoon onset on the monthly time scale, so the sensible heating is constantly observed from the composite result within the period shown in Fig. 5a.

In order to get a clearer view of the heating structure, the meridional-vertical distributions of Q_1 and Q_2 profiles for a pre-onset (day -7 to -1) and a post-onset (day 0 to 6) periods are averaged over the Australian sector (115-150°E) and shown in Fig. 6 and Fig. 7. The heating over the Australian continent is constantly observed below 750 hPa in both periods, but the location of deep convection near the equator shows a sudden poleward shift after the onset. This deep convective heating centered at 5°S in the pre-onset period (Fig. 6) is associated with the migrating intertropical convergence zone (ITCZ). It suddenly jumps poleward 5-10 degrees and covers a large region after the onset (Fig. 7). The maximum of Q_1 in the ITCZ is always at 500-400 hPa, but the profile of Q_2 with one lower peak changes to double peaks at 800-650 hPa and 500-400 hPa after the onset. The vertical profiles of Q_1 and Q_2 after the onset are similar to the major mode found by Tung et al. (1999) during Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean Atmosphere Response Experiment (COARE) Intensive Observing Period (IOP). They indicate that this structure is a typical mode during the convective phase of the MJO.

4. Factors contributing to the onset

a. Land-sea thermal contrast

With the contribution of the sensible heating from the surface, a reversal of meridional temperature gradient between the Australian continent and the Arafura Sea in a layer below 800 hPa occurs from September to March. The composite 1000 hPa temperature (Fig. 8) shows a band of very warm air caused by surface heating at 15-30°S in the Australian sector (115-150°E). Before the monsoon onset, the maximum temperature over the continent exceeds 304 K which is about 2 to 4 K higher than the air temperature over the ocean near the equator. This warming starts from 2-3 months prior to the onset. It makes the meridional temperature gradient $\partial T/\partial y$ negative between land and sea, because the air temperature over the Arafura Sea has very small change.

The latitude-height distribution of temperature along the Australian sector indicates that such reversal of the meridional temperature gradient exists from surface to about 800 hPa before the onset (Fig. 9a). The distribution of relative humidity (RH; Fig. 9b) shows that the air over the continent is dry (RH $\sim$ 40-50%), and the moist air (RH $\sim$ 80%) is confined in the equatorial region.

The sensible heating over Australia sets up a thermally-induced meridional-vertical circulation, as seen from the outflow above 700 hPa and low-level inflow below 925 hPa over the continent (Fig. 10a). Originally, this meridional-vertical circulation is a separate system from the circulation associated with the ITCZ near the equator. The low-level inflow contributes to transport moist air inland and intensify the monsoon circulation in NAU. A narrow region of downward motion near 15°S between the ascending branch of the ITCZ and the upward motion induced by sensible heating can be clearly seen in Fig. 10a. However, these two regions with upward motion merge together, and the narrow area with downward motion disappears after the onset of the monsoon in NAU (Fig. 10b). The role of the subtropical heating for inducing the meridional circulation was discussed by Plumb and Hou (1992) with an idealized zonally symmetric model. The importance of the thermal heat low in creating a convectively unstable condition for the onset of the Australian summer monsoon was suggested by Kawamura et al. (2003).

b. Barotropic instability

Several authors applied dynamic instability theories to the monsoon onset (e.g. *inertial instability*: Tomas and Webster 1997; *moist baroclinic instability*: Moorthi and Arakawa 1985, Xie and Saiki 1999). Over the Australian sector, the monsoon westerlies occur north of 15°S, where the atmosphere is nearly barotropic (see Fig. 9a). Here, we examine the possible role of barotropic instability in the onset of the Australian summer monsoon. The stability of barotropic zonal current was discussed by Kuo (1949) for the westerly zonal current and by Nitta and Yanai (1969) for the easterly basic flow. The changing sign of the meridional gradient of the absolute vorticity is a necessary condition for the barotropic instability in zonal currents.

In order to see the relevance of barotropic instability to the onset of the Australian monsoon, we calculate the meridional gradients of absolute vorticity at 850 hPa in a zone, 10-15°S, where the monsoon westerlies usually occur and negative meridional gradients appear before the onset. The 850 hPa level is chosen because the zonal wind there is used to define the monsoon onset in Section 2. Figure 11a shows the meridional gradients in this zone averaged over 115-150°E in time-latitude sections from 19 November to 18 January of each year. Nine of the total fourteen cases in 1979-93 show negative meridional gradients of absolute vorticity at least 3 days before the monsoon onset (the onset days are indicated by thick lines in Fig. 11a). Five exceptional cases are the monsoon seasons in 1979-80, 1982-83, 1988-89, 1990-91 and 1991-92. In 1982-83 and 1991-92, the Australian summer monsoon is very weak due to the El Nino. In 1988-89 (a La Nina year), an unusual deep trough from subtropical region is involved for the onset that occurred outside of the 10-15°S zone. In 1979-80 and 1990-91, the onsets took place in the western and eastern parts of the NAU region, respectively, so the onset area is not centered in the Australian section (115-150°E).

A 14-event composite of the mean meridional gradients of absolute vorticity was made relative to the onset dates to obtain an overall view of the role of the barotropic instability (Fig. 11b). The composite time-latitude distribution of the mean meridional gradient of the absolute vorticity over 115-150°E shows that the negative gradient in the 10-15°S zone appears 7 days before the onset and disappears after the onset. The horizontal distribution of the meridional gradient of the absolute vorticity averaged from 6 days to 1 day before the onset (Fig. 12a) shows a negative center in NAU. This center (130-150°E) moves to northeastern Australia and does not remain in NAU after the onset (Fig. 12b). These results show that the barotropic instability criterion is often met prior to the onset, and provides a condition favorable for the monsoon to occur.

c. The arrival of the MJO

Previously, Hendon and Liebmann (1990a and b) suggested that the MJO is the major trigger for the Australian summer monsoon onset based on a composite study. They found that the onset coincides with the arrival of the eastward propagating MJO which can be traced back to the Indian Ocean for several days.

The time-longitude section of the OLR composite shows the eastward propagation of two MJO events (Fig. 13a). The deep convection associated with the first MJO event suddenly intensifies in the Australian sector, and it can be traced back to the Indian Ocean for about 10 days prior to the onset. The land-sea thermal contrast clearly plays an important role in intensifying the vertical circulation in NAU section, so the convection with the MJO system in the Australian sector is stronger than that in the Indian Ocean. Although the deep convection associated with the MJO continues to move eastward to the date line, the monsoon system remains in the Australian sector until the monsoon break. The second MJO event after the break can also be traced back to the Indian Ocean.

The time-longitude section of the 850 hPa zonal wind component (Fig. 13b) shows an eastward propagating feature similar to that of OLR, but the center of active convection observed from OLR leads the westerly wind bursts (WWBs) for several days. However, the maximum upward velocity (not shown) in a layer (500-400 hPa) where the maximum Q_1 is observed coincides with the center of deep convection. A detailed case study of MJOs during the TOGA COARE IOP especially for the phase difference between the zonal wind oscillation and convection was provided by Yanai et al. (2000).

d. The intrusion of midlatitude trough

Although the arrival of the MJO is usually concurrent with the onset of the Australian summer monsoon, some years (1981-82, 1988-89, 1990-91) show other disturbences acting as triggers for the onset. For example, in 1990-91, a tropical cyclone located at 5-20°S moving from 170°E to 150°E during the week before the monsoon onset. It triggered the onset in the eastern part of NAU, resulting an unusual onset situation. In 1981-82 and 1988-89 monsoon seasons, the propagation of the MJOs was slowed down when it reached the Australian region. An unusually deep trough extending from midlatitude was observed before the onset in both cases.

Davidson et al. (1983) has emphasized that the midlatitude events (troughs and ridges) play a substantial role in the monsoon onset. They showed several typical synoptic sequences including a trough in the western coast of Australia prior to the onset. We examined the synoptic sequences for the 14 monsoon cases, and found that only one case (1992-93) had no midlatitude trough involved in the onset. For most cases, the MJO was the major trigger, but the midlatitude trough played a secondary role for the onset. However, when the MJO is weak or absent, it is the time when the midlatitude trough plays the major role.

The 14-year composite streamlines and temperature field at 850 hPa show the following synoptic sequences. Starting from about one week before the onset, temperature over the Australia continent increases significantly, while a trough in the midlatitude westerlies extends to the western coast of Australia (Fig. 14a). This trough gradually moves into the continent, and a cyclone is generated in the western part of Australia. When the trough moves further to the central Australia, the cyclone merges with the MJO system in NAU. A commonly observed horizontal structure of Australian summer monsoon flow then establishes: a monsoon low stands in NAU, while anticyclones cover the western and eastern sides of Australia (Fig. 14b). At this stage, the strong low-level westerlies starts and the monsoon onset occurs.

5. Conclusions

a. Heating processes in the Australian region

In this paper, we used the ERA data of 1979-93 and the products of Q_1 and Q_2 calculated from heat and moisture budgets to study the onset of the Australian summer monsoon. A 14-event composite relative to the onset dates is used to reveal the onset mechanism. Major findings from Q_1 and Q_2 diagnosis can be summerized as following:

(1) Q_1 and Q_2 in northern Australia-Arafura Sea (NAU)

The layer of maximum Q_1 located at 500-400 hPa before and after the onset, but the largest peak of Q_2 after the onset is in a layer between 800-650 hPa. The well-separated Q_1 and Q_2 peaks in vertical imply the presence of deep convection in this area. However, Q_2 has another weaker peak at 500-400 hPa after the onset. This indicates that the stratiform rain is also present.

(2) Q_1 and Q_2 in the Australian continent (AUC)

The maximum Q_1 ($\sim$ 2.5 K d^{-1}) is confined in a shallow layer near the surface, while Q_2 is negative below 700 hPa. The positive Q_1 suggests the sensible heating from ground surface. This heating starts in September which is several months before the onset, and contributes to create the land-sea thermal contrast in the Australian sector (115-150°E).

b. Major factors contributing to the onset

Four major factors contributing to the onset of the Australian summer monsoon are summarized as following:

(1) Land-sea thermal contrast

This factor acts as a seasonal preconditioning for the onset. The sensible heating over the

Australian continent leads to a reversal of meridional temperature gradient between the Arafura Sea and the Australian continent in a layer below 800 hPa in September-March, and sets up a thermally induced meridional-vertical circulation which helps to transport low-level moist air inland to intensify the monsoon circulation in NAU.

(2) Barotropic instability

Between 15°S and the equator in the Australian sector, the atmosphere is nearly barotropic before the onset. We examine the condition of barotropic instability at 850 hPa where the zonal wind is used to define the monsoon onset. A region of negative meridional gradient of absolute vorticity often appears in NAU prior to the onset and disappears after the onset. This suggests that the barotropic instability is an additional preconditioning for the monsoon onset.

(3) The arrival of the MJO

The onset of the Australian summer monsoon coincides with the arrival of the eastward propagating MJO for most cases. The deep convection (observed from OLR), 850 hPa westerlies and ascending motion in the middle troposphere (500-400 hPa) associated with MJOs can be traced back to the Indian Ocean about 10 days prior to the onset. The center of active deep convection (low OLR) coincides with the vertical velocity at 500-400 hPa, but leads the 850 hPa westerlies for several days.

(4) The intrusion of midlatitude trough

When the MJO is weak or absent, the onset is triggered by the midlatitude trough or other disturbances. Two cases in 1979-93 have the midlatitude trough playing a major role in the onset and only one case in the 14 events has no trough involved for the onset. The composite synoptic sequences suggest that the intrusion of the midlatitude trough into the western Australia plays a supplemental role in the onset, but its importance varies case by case.

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Fig. 10: 1979-93 latitude-height composites of the meridional circulation (v, ω; arrows) and values of ω (Pa s^{-1}) averaged over 115-150°E: (a) from 7 days to 1 day before the onset, and (b) from the onset day to 6 days later. Areas of upward motion (negative ω) are shaded.

Fig. 11: (a) Time-latitude sections of the meridional gradient of absolute vorticity (contour interval 5×10^{-12} m^{-1} s^{-1}) at 850 hPa averaged over 115-150°E (only negative values are shown). The zones between 10-15°S are shown from 19 November to 18 January for each year in 1979-93. The thick lines indicate the onset days. (b) 1979-93 time-latitude composites of the meridional gradient of 850 hPa absolute vorticity (contour interval 5×10^{-12} m^{-1} s^{-1}).

Fig. 12: 1979-93 composites of the meridional gradient of absolute vorticity (10^{-12} m^{-1} s^{-1}) at 850 hPa: (a) from 7 days to 1 day before the onset, and (b) from the onset day to 6 days later.

Fig. 13: 1979-93 time-longitude composites of (a) OLR (contour interval 10 W m^{-2}; values lower than 220 W m^{-2} are shaded), (b) 850 hPa u (contour interval 1 m s^{-1} for positive and 2 m s^{-1} for negative values). The thickest lines indicate the 200 W m^{-2} contours of OLR.

Fig. 14: 1979-93 composites of 850 hPa streamlines: (a) 6 days before the onset, and (b) 1 day after the onset. The gray shading is the 1000 hPa temperature (K).

List of table

Table 1: Onset dates (year-month-day) of the Australian summer monsoon obtained in the present study. The dates given in Hendon and Liebmann (1990a) are listed for references.