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 the northern Australia (NAU) are used to determine the onset date. 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 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 met at 850 hPa in NAU several days prior to the onset. The sudden onset is then triggered by the arriving MJOs and at times by the intrusion of midlatitude troughs. To isolate the role of land in monsoons, the differences and similarities of the Australian summer monsoon and the ITCZ over the Indian Ocean are examined.

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) 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 (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 1982; 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 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) 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:

(1) 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.

(2) 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.

(3) In previous study, 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 from the Australian continent can be studied.

(4) Previous observational works focus on different onset processes (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, the onset mechanism associated with triggering systems are discussed. In section 5, in order to examine the role of land in monsoons, the comparison of structures and time evolutions between Australian summer monsoon and the Intertropical Convergence Zone (ITCZ) over the Indian Ocean at the same latitude are shown. Finally, section 6 presents summary and discussion.

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 specific 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, and the prime denotes the deviation from the average.

The horizontal wind components, potential temperature, and mixing ratio are directly obtained from 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 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 can be 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, and s the dry static energy (Yanai and Johnson 1993). 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 flux 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 eqution (4) in convection-large-scale interaction (e.g. Wu 1994).

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 the vicinity of it (Troup 1961; 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 in the northern Australia (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 the 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 January - February, and a break is observed on 16 - 24 January.

Based on the same definition, the onset dates for other monsoon seasons were chosen and listed in Table 1. The onset dates selected by the present study are close to previous results. The 15-year 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 shown 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 informations. 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 levels is calculated for the composite mean, and the value of it is about 1 m s^{-1} for layers lower than 500 hPa. This indicates that there are about 2 active periods in a monsoon season in general. We examine all the 14 monsoon cases, and 11 of them have 2 monsoon active periods within 90 days started from the onset day.

In order to compare the atmospheric structure in NAU and that over the Australian continent, we define another region, Australian continent (AUC), along the same longitudinal boundaries for NAU, but shift 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. Because the Australian continent does not have a plateau as the Asian continent does, a sudden jump of the jet influenced by the mechanical forcing due to topography (Trenberth and Chen 1988) is not found. However, the westerlies at 200hPa over the AUC become weak after the monsoon onset reflecting the poleward shift of the westerly jet in the upper level, concurrent with the sudden onset of the monsoon circulation.

3. Composite features of heat sources and moisture sinks

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

The vertical distrubitions 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 the Northern Australia in January is about 2.3 K d^{-1} and the peak of it is located between 500-400 hPa, but with their method, Q_2 is not available. 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 the northern Australia.

Our results provide a more general heating structure by making composites from 15-year data. 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 in the AMEX.

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 structures for 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 that the presence of stratiform rain in NAU during monsoon season.

In contrast, the vertical distribution of Q_1 over AUC has a very different structure. Significant heating in the lower level (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 warm surface (e.g. Yamada and Mellor 1975), the strong heating observed at the lower level is concluded to be mainly 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 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 convection centered at 5°S in the pre-onset period is associated with the migrating ITCZ (Fig. 6). 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

Using the composite method, four major factors contributing to the onset of the Australian summer monsoon are recognized as following:

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 warming over the continent determines the meridional temperature gradient, because the temperature over the Arafura Sea has very small change.

Prior to the onset of the Australian summer monsoon, the reversal of meridional temperature gradient is already observed (Fig. 8a). The air over the continent is dry. The relative humidity (RH) on the continent is only about 40-50%, and the moist air (RH $\sim$ 80%) is confined in the equatorial region (Fig. 8b). The sensible heating 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. 9a).

Originally, this meridional-vertical circulation is a separated system from the ITCZ near the equator. The low-level inflow has contributions to transport moist air inland and intensify the monsoon circulation in NAU. A narrow region of downward motion near 15°S between the upward branch of the ITCZ and upward motion induced by the sensible heating can be clearly seen in Fig. 9a. 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. 9b).

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, easterlies in the lower level are present south of 20°S, because they are balanced by the thermal wind relationship. On the other hand, the westerlies occur north of 15°S, where the atmosphere is nearly barotropic (see Fig. 8a). 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 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 where the zonal wind is used to define the onset in section 2.

Figure 10 shows an example of the onset of the Australian summer monsoon in 1980-81. The time-latitude section is averaged over 115-150°E. Values of the meridional gradients are almost positive everywhere, but a negative area between 11-16°S appears just before the monsoon onset (the onset day: January 4 1980) and disappears later. Similar situations can be found in other years, too. Therefore, a 14-event composite was made relative to the onset to obtain a more general view. The horizontal distribution averaged from 6 days to 1 day before the onset (Fig. 11a) 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. 11b). The presence of this negative center implies the barotropic instability exists prior to the onset, and acts as a precontioning of the onset. The composite time-latitude profile of the meridional gradient of the absolute vorticity over 130-150°E shows that the negative gradient appears before the onset, but the strongest negative value is observed about 7 days before the onset (Fig. 12) just like the example in Fig. 10. Although the atmosphere is not purely barotropic, the result here suggests that the instability plays a role in the monsoon onset.

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 this first MJO event suddenly intensifying 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 constrast 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. Detailed comparisons between the structure and circulation with the Australian monsoon and the ITCZ over the Indian Ocean will be given in the next section.

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 OLR composite result (Fig. 13a) shows that there are two MJO events corresponding to the presences of two active periods in a monsoon season on average as we discussed in section 2. The second MJO event 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 (Fig. 13a). 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). The results indicate that the zonal component $\partial u/\partial x$ alone is not sufficient to describe the horizontal divergence, and the meridional term $\partial v/\partial y$ is also important (Mathews et al. 1999).

d. The intrusion of midlatitude trough

Although the arrival of MJO is usually concurrent with the onset of the Australian summer monsoon, some years (1981-82, 1988-89, 1990-91) show that the onset actually occured more than one week after the arrival of the westerlies associated with MJO in NAU. During these years, after arriving northwestern Australia, the MJO stalls there for about 7-10 days until a sudden onset of monsoon in NAU. This implies that the MJO alone is not sufficient to trigger the onset in these specific years.

For these cases, the delaying time from the arrival of MJO to the onset is about the time interval of arriving midlatitude systems. Davidson et al. (1983) emphasized the midlatitude events (troughs and ridges) play a substantial role in the monsoon onset. They show several typical synoptic sequences including a trough in the western coast of Australia prior to the onset.

In all three delaying cases mentioned above, we also observe a trough extenting to the Australian continent. This suggests that the intrusion of the midlatitude trough is a factor contributing to the monsoon onset.

In fact, the 14-case composite results show similar interesting feature of the midlatitude systems. The difference is that the delay is smoothed out by adding more cases. 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 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. We examined these synoptic sequences for 14 monsoon cases, and found that only one case (1992-93) has no midlatitude trough involved in the onset.

5. Differences and similarities of the Australian summer monsoon and the ITCZ over the Indian Ocean

In recent papers by Chao (2000), and Chao and Chen (2001a, b), the monsoons are interpreted as an off-equatorial ITCZ. Because the existence of the ITCZ does not require land, these authors suggested that ”The existence of landmass is not considered a necessary condition for monsoons” (Chao and Chen 2001b). We examine the differences and similarities between the Australian summer monsoon and the ITCZ over the Indian Ocean to clarify the role of land-sea thermal contrast in monsoons.

Using CMAP and ERA 850 hPa wind data, the rainfall band is observed to be connected all the way from the Indian Ocean to NAU in the southern summer (Fig. 1). The rainfall in the equatorial Indian Ocean (60-90°E) associated with the low level confluent flow is called the Indian Ocean ITCZ. On the other hand, the rainfall in NAU and the circulation associated with a thermal heat low belongs to the Australian summer monsoon. In NAU, the rainfall has a very significant seasonal variation, and there is almost no precipitation in seasons other than summer (December - March). However, the Indian Ocean ITCZ always resides in the equatorial region throughout the year.

The composite 1000 hPa temperature shows a band of very warm air caused by surface heating at 15-30°S in the Australian sector, 115-150°E (Fig. 15a). Before the monsoon onset, the maximum temperature over the continent is warmer than 304 K which is about 2 to 4 K greater than the air over the ocean near the equator. This warming starts from 2-3 months prior to the onset, and makes the meridional temperature gradient $\partial T/\partial y$ negative between land and sea. The negative value of $\partial T/\partial y$ does not appear at the same latitude in the Indian Ocean sector, 60-90°E (Fig. 15b), because the ocean in this region can not provide similar sensible heating from surface as the Australian continent does.

It is clear to see the seasonal march of the warm temperature band (warmer than 300 K) migrating from the Northern Hemisphere (NH) to the Southern Hemisphere (SH) in Fig. 15, especially for the ocean region north of the equator in NAU sector (Fig. 15a) and south of the equator in the Indian Ocean ITCZ sector (Fig. 15b). The clearest difference between these two sectors is the warm temperature caused by sensible heating over land. Due to the different heat capacity of land and sea, the continent provides a strong sensible heat at the latitudes where land is located. The strength of this heating increases with time when the season progresses from southern spring to summer, but the location of the heating does not change with the seasonal variation of solar angle in latitudes.

The composite 850 hPa zonal wind component (Fig. 16) shows the similar seasonal migration of westerlies from the NH to the SH Several westerly wind maxima associated with MJOs can be recognized. Because the composite is based on the onset of the Australian summer monsoon, the westerly maxima in the Indian Ocean ITCZ sector (Fig. 16b) appear earlier than that in NAU sector (Fig. 16a) (MJOs propagate eastward.) An interesting different feature between these two sectors is the sudden increase and poleward extension in NAU sector after the onset (day 0). The poleward extension is one of the differences between ITCZs and monsoons. Webster and Chou (1980) used a simple model including ocean and atmosphere to show that the existence of continent results the poleward extension of the monsoon circulation and the location of precipitation peak. The composite 850 hPa vorticity shows such sudden poleward jump in NAU sector very clear (Fig. 17a), but the jump does not appear in the Indian Ocean ITCZ sector (Fig. 17b).

To summerize, the migrating ITCZs are very similar in both sectors before the monsoon onset. However, when the ITCZ approaches the continent, the poleward jump of the convection occures and a thermal heat low builds up. On the other hand, the migrating ITCZ moves poleward following the seasonal path in the Indian Ocean sector, but it remains in lower latitudes and no thermal heat low is built. A very significant warming due to the sensible heating over the continent contributes to the poleward extension of circulation. The heating located in the subtropical region can help to drive the meridional circulation (Plumb and Hou 1992) and intensify the monsoon system. Lacking of the continent, the circulation is only an off-equatorial ITCZ, and does not have the poleward extension and a sudden onset which we usually recognized as important features of monsoons.

6. Summary and discussion

a. Heating processes in the Australian region

In this paper, we used the ERA data 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 15-year (1979-93) 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 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 AUC

The maximum Q_1 ($\sim$ 2.5 K d^{-1}) is near 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 have contribution to the land-sea thermal contrast in the Australian sector.

b. Major factors contributing to the onset

Based on the composite method, four major factors contributing to the onset of the Australian summer monsoon can be 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 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 negative region of meridional gradient of absolute vorticity appears in NAU just prior to the onset, but is not observed after it. This suggests that the barotropic instability is a preconditioning for the monsoon onset.

(3) The arrival of the MJO

In NAU, the low-level westerlies are introduced by the MJO. The onset 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. The phase difference between the westerlies and the ascending motion suggests the importance of the meridional component of the horizontal divergence and the zonal component alone is not sufficient.

(4) The intrusion of midlatitude trough

In some years, the onset occured more than one week after the arrival of the MJO in NAU. This suggests that the MJO alone is not enough to trigger the onset for these cases. The period of the delaying time for the onset is about the time interval of midlatitude systems (troughs and ridges) passing through Australia. The composite synoptic sequences suggest that the intrusion of the midlatitude trough into the western Australia plays a supplemental role, but its importance varies case by case.

c. discussion

In order to examine the role of land in monsoons, we studied the differences and similarities of the Australian summer monsoon and the ITCZ over the Indian Ocean. The Australian summer monsoon sector in the pre-onset stage is very similar to the Indian Ocean ITCZ. However, the poleward jump of the convection and the presence of the thermal heat low during and after the onset are the unique feautures of monsoons which the ITCZs don't have. Although the condition of the land-sea thermal contrast and barotropic instability are not a direct trigger for the monsoon onset, it is necessary to set up a preferred location for monsoons. The heating located in the subtropical region is necessary to drive the meridional circulation (Plumb and Hou 1992). The importance of the thermal heat low in the onset of the Australian summer monsoon is also suggested by Kawamura et al. (2002)

In the Australian sector, the meridional circulation is mainly driven by the land-sea thermal contrast. This circulation due to the sensible heating from the ground surface brings moist air inland and can intensify the circulation itself. This mechanism provides a seasonal preconditioning for the monsoon. In NAU, the low-level westerlies are then introduced by the arriving MJOs, and the onset is triggered by it.

Several case studies have been done to reveal the onset process of the Australian summer monsoon (e.g. Murakami and Sumi 1982; Davidson et al. 1983; Hendon et al. 1987). Our study here is aimed to focus on the general features of the onset and obtain the major factors contributing to it. We conduct a composite method for this purpose, so the actual onset for any specific year is unlikely to be the same. This weakness is also pointed out by other composite studies (e.g. Hendon and Liebmann 1990a). Nevertheless, the important factors presented here provide the dominant and general features of the onset. Any detailed analysis for the specific onset should be left for the case studies which is not the main purpose of this work.

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Figures

Fig. 1. 1979-93 averaged 850 hPa streamlines and precipitation rate (mm d^{-1}) for December-February. The northern and southern rectangular boxes show the regions of northern Australia (NAU) and Australian continent (AUC) respectively (see text).

Fig. 2. (a) Time-height section of u (contour interval 1 m s^{-1} for positive and 2 m s^{-1} for negative values), and (b) time series of u at 850 hPa (solid line) and 200 hPa (dashed line) in NAU from September 1979 to April 1980. The line on 28 December indicates the onset day in this case.

Fig. 3. 1979-93 time-height composites of u in (a) NAU (contour interval 1 m s^{-1} for positive and 2 m s^{-1} for negative values) and (b) AUC relative to the onset day (contour interval 5 m s^{-1}).

Fig. 4. 1979-93 time-height composites of (a) Q_1 (contour interval 0.5 K d^{-1}) and (b) Q_2 (contour interval 0.25 K d^{-1}) in NAU.

Fig. 5. Similar to Fig. 4, except for the AUC.

Fig. 6. 1979-93 latitude-height composites of (a) Q_1 and (b) Q_2 (contour interval 1 K d^{-1}) averaged over 115-150°E from 7 days to 1 day before the onset.

Fig. 7. Similar to Fig. 6, except for composites from the onset day to 6 days later.

Fig. 8. Similar to Fig. 6, except for (a) temperature (K; values greater than 290K are shaded) and (b) relative humidity (%; values greater than 60 are shaded).

Fig. 9. 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. 10. Time-latitude section of the meridional gradient of 850 hPa absolute vorticity (10^{-12} m^{-1} s^{-1}) averaged over 115-150°E in December 1980 and January 1981. The line on January 4 indicates the onset day for this case.

Fig. 11. 1979-93 composites of the meridional gradient of 850 hPa absolute vorticity (10^{-12} m^{-1} s^{-1}): (a) from 7 days to 1 day before the onset, and (b) from the onset day to 6 days later.

Fig. 12. 1979-93 time-latitude composites of the meridional gradient of 850 hPa absolute vorticity (10^{-12} m^{-1} s^{-1}).

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 red shading is the 1000 hPa temperature (K).

Fig. 15. 1979-93 time-latitude composites of 1000 hPa temperature (K) averaged over (a) 115-150°E and (b) 60-90°E.

Fig. 16. Similar to Fig. 15, except for u (m s^{-1}) at 850 hPa.

Fig. 17. Similar to Fig. 15, except for vorticity (10^{-6} s^{-1}) at 850 hPa.