A change of mean physical quantity in the monsoon and pre-monsoon.

A. Watanabe[1], Y. Tachibana[2], S. Ogino[3], Y. Shibagaki[7], T. Ohsawa[3],Y., M. Yamanaka[3], T. Oki[6], T. Satomura[4], A. Sumi[5], K. Mushiake[1]
(1Fukushima University, 2Tokai University, 3Kobe University 4Kyoto University, ５CCSR University of Tokyo, ６University of Tokyo, ７Osaka Electro-Communication University,)
Fukushima University, Japan,960-1296, Kanayagawa, Fukushima, Japan. e-mail:may@sss.fukushima-u.ac.jp

Abstract

It is very important to understand a difference of the atmospheric circulation between the pre-monsoon and the monsoon periods. We are able to divide into 3 classifications on Apr. 15 to May 13, May 14 to Jun. 15, and Aug. 15 to Sep. 15, to understand the fluctuation of the monsoon. And we calculated the mean field of the meteorological element every 3 classification. The pressure decreases about 2hPa under the 5km in the monsoon than the pre-monsoon. But, the pressure increases about 3hPa over the 5km in the monsoon than the pre-monsoon. Furthermore, the temperature decreases about 2℃ under the 3km, and the temperature increases about 4℃ over the 3km in monsoon than the pre-monsoon. Accordingly, we are unable to explain the pressure decrease of the lower layer with the temperature change. We are conceivable that this mean field was made with a convective activity. It became clear that decreasing of pressure of the lower layer strengthens the south wind, and convective instability is strengthened by advection of water vapor by the southerly wind.

Keyword: monsoon onset, rawinsonde observation, convective activity, Thailand

1.Introduction

We carried out rawinsonde observation to understand the atmospheric circulation in the pre-monsoon and the monsoon periods over Thailand. We observed the upper atmospheric structure by rawinsonde four times a day from Apr. 15 to Jun.15 and Aug.15 to Sep.15 1998 at Chiang Mai, Ubon Ratchathani and Bangkok in Thailand. It is very important to understand a difference atmospheric circulation between the pre-monsoon and the monsoon periods. The reason is that we can understand the monsoon onset mechanism due to the difference of the atmospheric circulation. In particular, we clarify the difference of the mean field of meteorological element to compare the pre-monsoon period with the monsoon period.

Many investigators have reported the trigger of monsoon onset. Orgill (1967) and Eguchi (1996) have indicated that many of the monsoons have started by the northing of onset vortex. And, Wu and Zhang (1998) pointed out that the trigger of monsoon onset is overlapping of phase of Madden-Julian oscillation and three-week oscillation. Further more, Xu and Chan (2001) have indicated the trigger of monsoon in 1998 is the northing of cyclone. Although the trigger of the monsoon is pointed out various as above there is little research of observation. We calculated the mean value with Bellamy method by using the rawinsonde observation data that we observed for every 6 hours at three stations in Thailand.

Our purpose is to indicate the difference of each meteorological element between the pre-monsoon and the monsoon periods, and to understand the trigger of monsoon onset in the Indochina Peninsula.

2. Analytical method

We obtained the meteorological data of upper atmosphere by rawinsonde four times a day from Apr. 15 to Jul. 15, and from Aug. 15 to Sep 15 in 1998 at Chiang Mai, Ubon Ratchathani and Bangkok in Thailand. We have calculated the average of rawinsonde data every 100m from surface to 15km above the sea level. And we obtain time series meteorological data every 100m during each period. Each missing data are calculated by Kriging method. As we calculate the average every 6 hours at each level, we can understand the situation of the mean field during each period. On the other hand, it is possible that we calculate the physical quantity using average data of the 3 stations.

Here, u and v are eastward and northward component of wind at the centroid of three stations. We can express the wind component of eastward and northward by following equations.

$$u_i = ax_i + by_i + c \qquad (1)$$
$$V_i = dx_i + ey_i + f$$

Here, indicate each station. We can calculate the constant (a, b, c, d, e, f) of equation (1) by substituting observation data of three stations. Then, we can obtain the wind component (u, v) at the centroid by substituting coordinate x, y of the centroid.

$$u = ax + by + c \qquad (2)$$
$$V = dx + ey + f$$

And it is possible to calculate the physical quantity by differentiating equations (2) in x or y. Similarly, it is possible that we calculate the zenith component of the divergence (D) and vorticity (ς) as following equations.

$$D = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \qquad (3)$$

$$\varsigma = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \qquad (4)$$

In addition, it is possible to calculate the vertical velocity (w) using the continuity equation as following equations.

$$w = -\int \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) dz \qquad (5)$$

Here, we calculated the vertical velocity as w=0 with z=0, and vertical velocity is integrated every 100m.

3. Analytical result

We have calculated the each physical quantity, and we consider the difference of mean field of meteorological elements between the pre-monsoon and monsoon periods.

Figure 1 shows the time-lag correlation of meteorological elements at Chiang Mai that was calculated by Watanabe (2002). We can find that the lag-correlation of the meteorological elements shows the negative value from May 14. Accordingly, We are able to divide into 3 classifications on Apr. 15 to May 13 (pre-monsoon), May 14 to Jun. 15 (monsoon), and Aug. 15 to Sep. 15 (mature-monsoon). We average each meteorological element every 3 classification. We consider the trigger of the monsoon onset from the characteristic of the mean value.

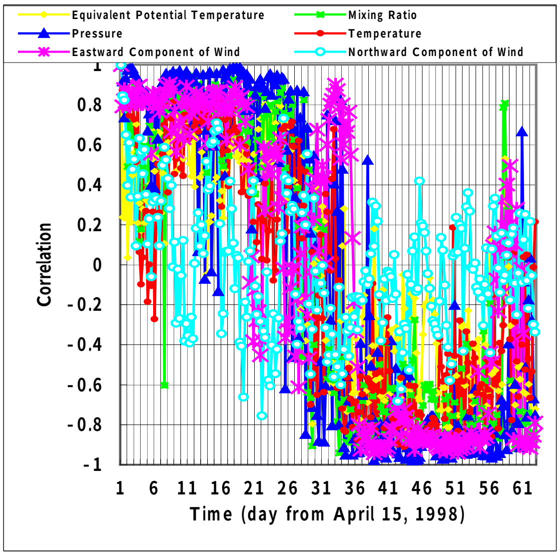

Fig.1 Time-lag correlation of Meteorological elements at Chiang Mai. The coefficient of lag correlation is calculated with the vertical distribution of the deviation from each height mean value.

Figure 2 shows the vertical profile of temperature at the centroid. The temperature in three periods is in

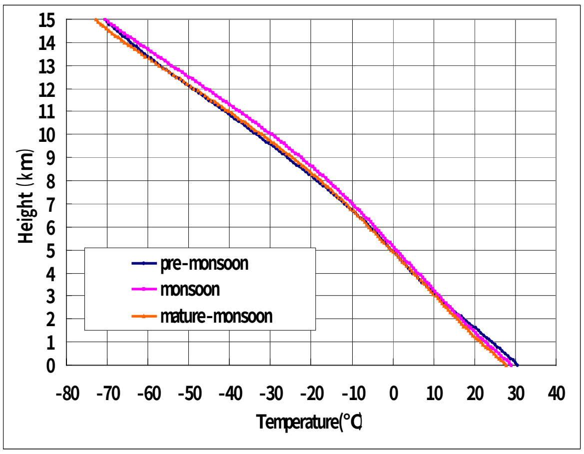

Fig.2 Vertical profiles of temperature at the centroid in the each period.

accord with about 10℃ at 3km above the sea level. It is the highest temperature in the lower layer than 3km at the pre-monsoon in the three periods. And, it is the lowest temperature in the higher layer than 3km at the pre-monsoon in the three periods. Namely, the cloud does not develop in the pre-monsoon; therefore the temperature is high relatively near the surface. However, the temperature in the lower layer than 3km in the monsoon and mature-monsoon is lower than the pre-monsoon. And, the temperature in the higher layer than 3km in the monsoon and mature-monsoon is higher than the pre-monsoon. The temperature doesn't rise when it enters into the monsoon. The solar radiation is interrupted by the cloud in the monsoon.

Figure 3 shows the vertical profile of pressure difference of Chiang Mai and Ubon Ratchathani from Apr. 15 to Jun. 15, 1998. We can find that the pressure

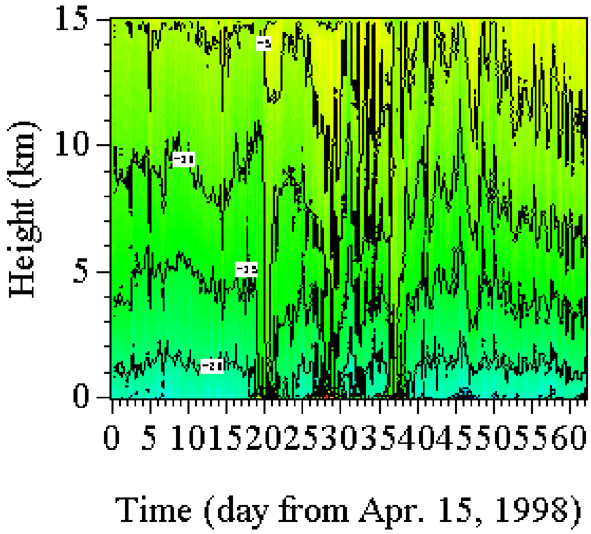

Fig.3 The vertical profile of pressure difference (hPa) of Chiang Mai and Ubon Ratchathani from Apr. 15 to Jun. 15, 1998.

difference of the two stations don't change largely in the lower layer than 2km. However, the pressure difference becomes conspicuously small in the upper layer than 10km after the monsoon onset. It is depend on the pressure rise at Chiang Mai after monsoon onset. Xu and Chan (2001) indicated that the reversion of the temperature gradient has not occurred in the Indian Ocean and Tibetan highland of the upper part troposphere. But, it is clearly that the pressure difference of the north and south decrease after monsoon onset. This is corresponding well with the rise of the temperature of the upper layer. It is clear that the decrease of the air pressure gradient in the upper layer is the trigger of a monsoon onset.

The vertical profile of the eastward component of wind is shown in Figure 4, and Figure 5 shows the vertical profile of northward component of wind. The southwesterly wind is excelling with the lower layer than 5.5km, and the northwesterly wind is excelling between 5.5km and 13km in the pre-monsoon. However, the southwesterly wind is excelling from surface to 10km above the sea level in the monsoon. The wind of the lower layer does not change largely with the monsoon and pre-monsoon. The wind of upper layer over 5.5 km is conspicuously different. The decrease of the pressure difference of the south and the north in the upper layer are gradually weakening the west component of wind in the upper layer clearly. The northeasterly wind excels in the layer over 7 km in the mature-monsoon and the pressure difference of the south and the north must reverse. The vertical distribution of the horizontal advection of water vapor in each period is shown in Figure 6. We consider the horizontal water vapor

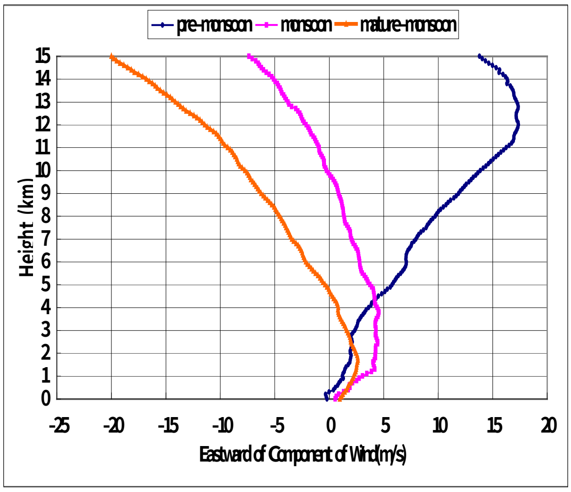

Fig.4 Vertical profiles of eastward component of wind in each period.

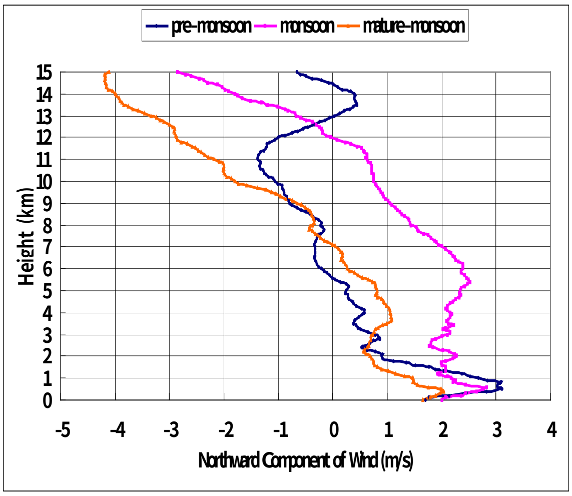

Fig.5 Vertical profiles of northward component of wind in each period.

advection difference by the difference of the direction of the wind.

Figure 6 shows the vertical profile of horizontal advection of water vapor mixing ratio in 12 hours. We

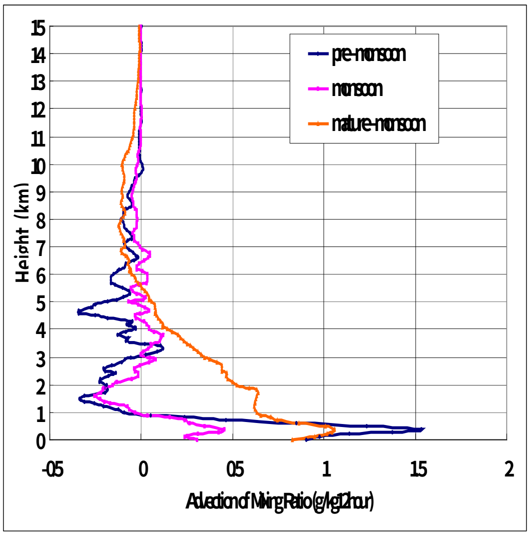

Fig.6 Vertical profiles of horizontal advection of water vapor mixing ratio in 12 hours

find that the horizontal water vapor advection in the layer of lower than 1km is obvious in the all period. It is corresponding with the strength of the southerly wind of the layer. However, the horizontal water vapor advection between 1 km and 3 km shows the value of negative and the correspondence with the direction of the wind is not necessarily evident. We find that the horizontal advection of water vapor is very small, although the value of negative is shown in the upper layer than 3 km. It is the positive horizontal advection of water vapor in the southwest wind range of the lower layer than 5 km, and it is negative horizontal advection of water vapor in the northeast wind range of the upper layer than 5 km in the mature monsoon. As for the distribution of the vapor of the lower layer the monsoon period is largest and the surface temperature is low. Accordingly it can do solidification easily with a little convergence in the lower layer than 1km. We are conceivable that the dried air advection between 1km and 3km is doing the work that continues solidification.

The distribution of such wind influences vorticity production largely. Figure 7 shows the vertical profile of vorticity. The vorticity with the entire layer is negative in the pre-monsoon. Also, the vorticity is

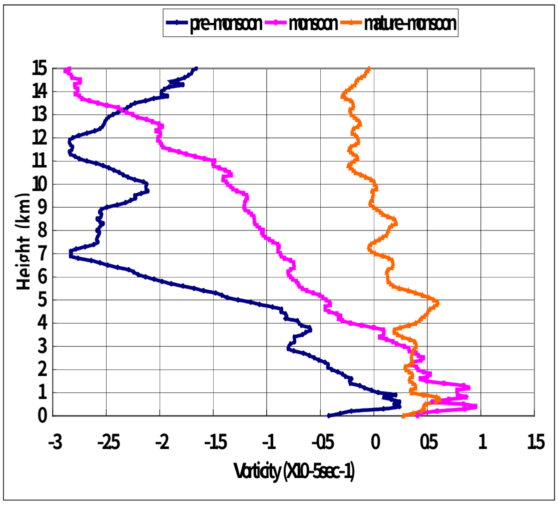

Fig.7 Vertical profiles of mean vorticity in each period.

positive in the lower layer than 4km and the vorticity is negative in the upper layer than 4km in the monsoon. Furthermore, it is the positive vorticity with all the layers in the mature-monsoon. The northwesterly wind of the pre-monsoon is strong in the north of the Indochina Peninsula, be weak the south. The southwest wind of the mature-monsoon is strong in the south of the Indochina Peninsula, be weak with north. The distribution of such wind is deciding the vorticity.

Figure 8 shows the vertical profiles of horizontal temperature advection. The cold air advection is obvious with a lower layer than 1 km in the pre-monsoon. However, the temperature advection in the monsoon is showing the value of positive with all of lower layers than 12 km. Also, the horizontal temperature advection of the mature-monsoon indicates the value of negative with the upper layer than 5km, it of positive with a lower layer than 5 km. The negative temperature advection region correspond with the region of the southward component of wind, and positive temperature advection region correspond with the region of the northward component of wind except for the region of the negative temperature.

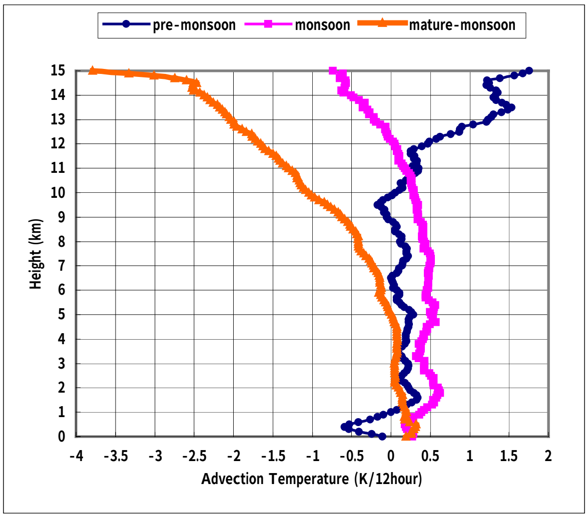

Fig.8 Vertical profiles of mean horizontal temperature advection in each period.

Figure 9 shows the profiles of mean vertical speed in each period. A relatively strong convergence region is appearing with the lower layer than 1km and this region become an ascent flow in the pre-monsoon and mature-monsoon. The descent flow appears between 1km and 3 km and weak ascent flow appears with a upper layer than the 3 km in the pre-monsoon.

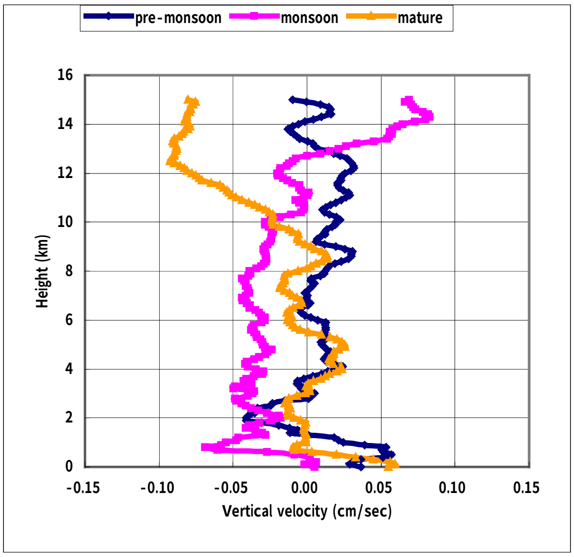

Fig.9 Vertical profiles of mean vertical speed in each period.

However, the convergence is weak with the lower layer than 1km, and it is divergence in the upper layer than 1km in the monsoon. Accordingly, although only the ascent flow appears in the lower layer as for the others, the descent flow is appearing in the monsoon. The mean quantity of convergence is not corresponding directly with the convection activity over the Indochina Peninsula. However, it is conceivable that the condition of the convection activity becomes lively it most be made in the pre-monsoon period, if the subsidence place of the lower layer is eliminated

4. Summary and discussion

Our purpose is to indicate the difference of each meteorological element between the pre-monsoon and the monsoon periods, and to understand the trigger of monsoon onset in the Indochina Peninsula.

We calculated the mean value from the data of rawinsonde observation in each period. We have clarified the characteristics of the mean field of the meteorological element in each period by using the mean value. And, we tried to clarify the cause of monsoon onset from the difference of the mean field. As a result the following became clear. The excellence of the northwest wind that makes a detour the Himalayas mountainous region generates the negative vorticity and be causing to the dry air advection in the pre-monsoon. However, the water vapor advection is large with the lower layer than 1 km, the southwest wind is excelling in the layer. Accordingly, the atmosphere instability is largest in the pre-monsoon, the convective activity is expected to become active in the period. But, the convective activity has not occurred in the period. It is reason that the drying air is subsiding with the layer between 1 km and 3 km. The stable layer is formed in the upper layer. The trigger of monsoon onset is the disappearance of this stable layer. A stable layer sometimes disappears with the irruption of cyclone, as Orgill (1967) and Eguchi (1996) are pointing out it. However, it is a basis that the intrusion of the drying air that makes a detour the Himalayas mountainous region is disappeared over the Indochina Peninsula. The reversion of the pressure gradient is necessary in the south and the north of the Indochina Peninsula by the temperature rising in the northern region. The temperature rise in the northern region, the irruption of the southwest wind strengthens and the irruption of humid air is become active over the region. This is the beginning of monsoon when was analyzed from the mean field. The north and south temperature gradient increases 2 ℃ in the monsoon period and pre-monsoon period in the upper layer than 2km, although it decreases with the lower layer than 2 km. The pressure gradient of the north and south decreases with the lower layer than 2 km, it is increases 1.5 hPa in the upper layer than 2km.

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