Title:

Upper-level circulation anomaly over Central Asia in early summer and its relationship to the Asian monsoon and mid-latitude wave train

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Abstract

A large intraseasonal variation in geopotential height over the Central Asia region, where the Asian subtropical jet is located, occurs between May and June, and the most dominant variation has a wave-like distribution. This variation in geopotential height influences precipitation across South and Southeast Asia.

We used composite analysis in an attempt to determine the causes of this intraseasonal variation over the Central Asia region. The wave train propagates from the northern Atlantic Ocean to the Central Asia region over a period of one week, and generates the anomaly in geopotential height over the Central Asia region. The tropical disturbance, which is similar to the Madden–Julian oscillation, appears a few days before the maximum of the anticyclonic anomaly over the Central Asia region, and is accompanied by active convection over the Indian Ocean and suppressed convection over Central America.

Results of numerical experiments using a linear baroclinic model show that the active convection over the northern Indian Ocean causes the anticyclonic anomaly over the Central Asia region. Due to the negative thermal forcing over Central America, the wave train from the northern Atlantic Ocean to the Central Asia region appears, which is a similar phase distribution to that observed in the composites. The Central Asia region is the region where the effects in the tropics and middle latitudes tend to be concentrated, and it is thought to be an important connection between the Asian monsoon and middle latitudes.
Keywords: Asian summer monsoon, intraseasonal variability, mid-latitude wave train, and the Madden–Julian oscillation.
1. Introduction

The Asian summer monsoon shows distinct intraseasonal variability, which often causes remarkable variations in weather conditions, such as precipitation and temperature. This intraseasonal variation is influenced by disturbances both in the tropics and at the higher latitudes; i.e., the middle latitudes and the subtropics. The Madden–Julian Oscillation (MJO) is one of the most important sources from the tropics (Madden and Julian 1994; Kemball-Cook and Wang 2001; Pai et al. 2011). As the active convective part of the MJO arrives in the Indian Ocean and moves eastwards into the maritime continent, precipitation in South and Southeast Asia tends to increase (Pai et al. 2011). Kemball-Cook and Wang (2001) investigated boreal summer intraseasonal variation during two separate periods (May–June and August–October) because of the pronounced differences in their climatologies. In both periods, convection over the Indian Ocean propagates along the equator and then moves poleward. This poleward shift of convection is caused by Rossby waves emitted by the equatorial convection. Convection in May–June shows continuous propagation along the Maritime continent, while convection in August–October transfers from the Indian Ocean to the western Pacific.

Previous studies have shown the relationship between the Asian monsoon and anomalies in the middle latitudes and subtropics, and indicate that the wave train at higher latitudes is connected to the variation of the Asian summer monsoon (Fujinami and Yasunari 2004; Ding and Wang 2009; Krishnan et al. 2009; Watanabe and Yamazaki 2012). These studies also found a marked anomaly over Central Asia and the
western Tibetan Plateau. Ding and Wang (2009) showed that a Rossby wave train across the Eurasian
continent, and the summer monsoon convection in the northwestern India and Pakistan, are coupled at an
intraseasonal time scale, and hypothesized a positive feedback between the Eurasian wave train and
Indian summer monsoon. Watanabe and Yamazaki (2012) showed that the marked anticyclonic
geopotential anomaly over the western Tibetan Plateau generates the low-level Rossby wave along the
strong westerly duct from the Arabian Sea to Southeast Asia in early summer. Subsequently, the
circulation at the lower levels changes in South and Southeast Asia. As shown in detail in Section 4, the
intraseasonal variability of geopotential height over Central Asia and the western Tibetan Plateau is large
in the early summer (Fig. 1). The above studies emphasize the influence of the middle latitudes upon the
tropical monsoon. However, tropical convection also affects the mid-latitude atmospheric circulation. For
instance, convection in the South and Southeast Asian monsoons influences the vertical motion over
Southwest and Central Asia in the subtropics (Rodwell and Hoskins 1995; Zhang et al. 2004). This
suggests that the anomaly extending from Central Asia to the western Tibetan Plateau is an important
connection point in the relationship between the subtropics and the Asian summer monsoon.

The basic condition of the atmosphere over Asia changes considerably with the seasonal change
during the summer monsoon season. The Asian subtropical jet migrates northwards between May and
June, and locates around 40°N between July and August (Schiemann et al. 2009). The low-level westerly
that blows from South to Southeast Asia begins to strengthen in May (Webster et al. 1998). At the onset of
the Asian summer monsoon, this westerly accelerates rapidly, and dominates the South and Southeast
Asian monsoon regions until October. The abrupt temperature increase over the Tibetan Plateau is
observed at the onset phase of Asian summer monsoon (Tamura et al. 2010). They showed the step-wise
increase of temperature is caused by not only diabatic heating from the surface of the Tibetan Plateau but
also adiabatic subsidence in the upper level. These abrupt changes of the atmospheric condition cause the
distinct change of precipitation in the Asian monsoon region (Ueda and Yasunari 1998; Fujinami and
Yasuanri 2001).

For investigating the intraseasonal variability of the Asian summer monsoon, the summer monsoon
season is preferable to be divided into distinct periods, as shown by Kemball-Cook and Wang (2001).
Early summer, between May and June, represents the transition season for the Asian summer monsoon
(Zhang and Wang 2008). Kajikawa et al. (2012) pointed out the advance of the Asian monsoon onset
during 1979-2008. They suggested the advanced Asia monsoon onset is attributed to the heat contrast
between the Asian landmass and the tropical Indian Ocean and the heating trend over the Asian landmass
primarily contributes to the heat contrast. The landmass heating over the Iranian Plateau and around the
western Tibetan Plateau in the subtropics is outstanding in their Figure 4. The relationship between the
Asian summer monsoon and the higher latitudes seems to be important for the intraseasonal variability of
the Asian summer monsoon in early summer.

This study aims to develop a better understanding of the relationship between the intraseasonal
variations of the Asian summer monsoon and variation at higher latitudes, and focuses on the variation of
geopotential height over the western Tibetan Plateau during early summer. The remainder of this
manuscript is organized as follows. Sections 2 and 3 describe the data and methods used, respectively.
Section 4 defines an index that represents the variation of geopotential height over the western Tibetan
Plateau and shows the results of the composite analysis involving this index. Section 5 discusses the
mechanism that drives the variation in geopotential height over the western Tibetan Plateau, and Section 6
describes the results of the numerical linear baroclinic model experiment. Finally, the results are
summarized in Section 7.

2. Data

Three meteorological datasets were used in this study.

1) The 6-hourly reanalysis data, provided by the European Centre for Medium-Range Weather
Forecasts (ECMWF) (ERA40; Uppala et al. 2005) on a 2.5° × 2.5° grid, were used for May and June
between 1958 and 2002. The analyzed variables were zonal and meridional wind velocities, temperature,
and geopotential height. Stream function and velocity potential were calculated by using the expansion
into the spherical harmonics function with truncation of T25. Daily averaged data were used to remove
the influence of diurnal variability.

2) The daily averaged outgoing longwave radiation (OLR) data, provided by the National Oceanic
and Atmospheric Administration/Climatic Diagnosis Centers (NOAA/CDC) on a 2.5° × 2.5° grid, were available over the period from 1975 to 2002 (Liebmann and Smith 1996).

3) Rain gauge data, in the form of the 0.5° gridded daily precipitation product developed by the Asian Precipitation Highly Resolved Observational Data Integration Towards Evaluation of water resources (APHRODITE) project (Yatagai et al. 2009), were available for monsoon Asia (60°–150°E, 0°–55°N) between 1961 and 2002.

The daily climatology was calculated by averaging the data for each day over the period of each dataset, and then smoothing these data using a 5-day running mean. Anomalies were defined as the deviations from this daily climatology. The anomalies don’t include the seasonal cycle, but include interannual variability.

3. Methods

We used a time-lagged composite analysis technique based on two indices defined in this study. One index represents the variation in geopotential height over the western Tibetan Plateau, while the second represents the wave train propagating from the northern Atlantic Ocean to the western Tibetan Plateau. Sets of key day for two indices were determined by using suitable criteria. Detailed descriptions of two indices are given in Section 4 and 5, respectively. Time-lagged composites are calculated from selected key days. The composite for a set of key days is represented as “day 0”. “day +/- X” represents a
composite for the set of days which are X days before/after key days, respectively. To remove the effect of seasonal cycle an anomaly from daily climatology is composited.

The two-sided Wilcoxon–Mann–Whitney rank sum test was used as the significance test for the precipitation and OLR data because the distributions of precipitation and OLR are non-Gaussian. The two-sided t-test was used as the significance test for the temperature, geopotential height, stream function, and velocity potential data. For horizontal wind, zonal and meridional winds are generally tested separately. In this study, the coordinate system was rotated to align the x-axis parallel to the direction that connects the means of the positive and negative samples. The two-sided t-test was applied to the new x-component data, and this provided a higher level of statistical significance than the test for the original zonal and meridional winds.

We used a dry linear baroclinic model (LBM), which was constructed based on a linearized AGCM (atmospheric general circulation model) developed at the Center for Climate System Research (CCSR), University of Tokyo, and the National Institute for Environmental Studies (NIES) (Watanabe and Kimoto 2000). The selected horizontal and vertical resolutions were T21 and 11 levels, respectively. The damping time scale was set at 1 day for the lowest and topmost levels, and 30 days for the other levels.

4. Results

4.1. Variation of geopotential height at 200 hPa in May and June
Figure 1 shows the root-mean-square of the modified anomaly of the geopotential height at 200 hPa from May to June in 45 years from 1958 to 2002. A modified time series is obtained by removing May-June average of the anomaly of geopotential height in a year from the time series of the anomaly in the year. It is noted that the root-mean-square does not include the seasonal cycle and the interannual but includes the intraseasonal and synoptic scale variation. A large variation of geopotential height is seen over the western Tibetan Plateau, Afghanistan, Tajikistan and other Central Asia countries in May and June, at which time the Asian subtropical jet is located around 35°N (Hereafter this region is called “Central Asia region” in this paper). This disturbance over the Central Asia region is able to propagate eastwards, along the subtropical jet, which acts as the Rossby waveguide (Ambrizzi and Hoskins 1993).

Figure 2 shows the first eigenvector for the above mentioned root-mean-square of modified anomaly of geopotential height at 200 hPa over the Central Asia region and surrounding regions. The eigenvector was obtained from the principal component analysis by applying the generalized spatial weighting matrix that was introduced by Baldwin et al. (2009). The first and second principal components account for 22% and 16% of total variance, respectively. The first eigenvector has its center of variation over the Central Asia region, and a wave-like distribution along the subtropical jet. The first eigenvector of the variance removed the synoptic scale variation by using 5-day-lowpass filter shows similar wave-like distribution (figure not show), which mean the variations on a short scale are not dominant in May-June.

Consequently, we defined the geopotential height index (GPH index) as the average anomaly over the
Central Asia region (30°–40°N, 60°–75°E; rectangle in Fig. 1) and normalized the index with its standard deviation in 45 years, as in Watanabe and Yamazaki (2012). For example the time series of the normalized GPH index in 1995 is shown in Fig. 3. The lagged auto-correlation of the GPH index becomes zero around 6 day lags, which indicates that the upper-level geopotential height anomaly over the Central Asia region shows the periodicity of monthly scale. The periodicity is seen from Fig. 3.

For the purpose of the investigation of the relation between variation of the upper-level geopotential height anomaly over the Central Asia region and variation of the Asian monsoon in early summer some composites based on the GPH index are calculated. Two sets of key day, which are positive and negative key days, are determined according to two criteria that the GPH index is more/less than +1.5/-1.5. In the process of determining key days the interval between each key day was set as more than 20 days. The interval is related with the periodicity of the GPH index. If there were several key days within a 20-day period, the larger one was chosen. The composite for positive/negative key days is called as positive/negative composite, respectively. Then positive/negative key days are selected 35/40 over a period of 45 years, respectively (Tables 1 and 2).

4.2. Composite analysis based on the GPH index

We now consider the results of the composite analysis based on the GPH index. To emphasize variations related to the upper-level geopotential height over the Central Asia region the difference between positive and negative composite at the same lag-day, which is hereafter called “composite
difference”, is shown.

Figure 4 shows the composite difference of geopotential height at 200 hPa. The wave train propagates from the northern Atlantic Ocean to the Central Asia region via western Russia from day –7 to day 0.

Each center of geopotential height anomaly along the propagation path attains a maximum at day –7 over the northern Atlantic Ocean (Fig. 4a), at day –2 over western Russia (Fig. 4f), and at day 0 over the Central Asia region (Fig. 4h). After day 0, the wave train propagates eastward from the Central Asia region along the subtropical jet.

Figure 5 shows the composite difference for OLR averaged over 5 days. Between day –9 and day –5, the first pentad centered at day –7, the negative OLR anomalies are seen over the northern Indian Ocean, and South and Southeast Asia (Fig. 5a). The positive OLR anomaly is seen over Central America. OLR anomalies are also distributed over North America. OLR anomalies with alternate signs are seen along the wave train propagation route from the northern Atlantic Ocean to the Central Asia region in the first and second pentads (Fig. 5a and b). The negative OLR anomalies of more than 20 W/m² over the northern Indian Ocean move northward, and extend across the Arabian Sea, the Indian subcontinent, the Bay of Bengal, Southeast Asia, and south China (Fig. 5b). The positive OLR anomaly over the Central Asia region is intensified simultaneously. Between day +1 and day +5; i.e., in the third pentad, the negative OLR anomaly persists from the northern Arabian Sea to south China (Fig. 5c), and a positive OLR anomaly appears over the central Indian Ocean.
The composite for precipitation is consistent with that for OLR over most of Asia, and shows a more detailed distribution of anomalies than OLR (Fig. 6). An increase in precipitation appears along the west coast of the Indian subcontinent, and this enhanced precipitation is retained there for about two weeks. A marked increase in precipitation is seen along the southern foothills of the Tibetan Plateau from day +1 to day +5 (Fig. 6c). At the same time, an increase in precipitation is seen along the western coast of Southeast Asia and Myanmar. The negative anomaly around the Yangtze River basin is intensified. Precipitation over South Asia increases about a week before, and after, day 0. Precipitation tends to increase over the southern part of South Asia (south to 20°N) before day 0, while precipitation increases over the northern part of South Asia after day 0. Precipitation over Southeast Asia increases sharply after day 0.

Figure 7 shows the composite difference for temperature at a height of 2 m averaged for 5 days. The positive and negative temperature anomalies are distributed from the northern Atlantic Ocean to the Central Asia region alternately (Fig. 7a and b). The black body emission can be estimated from the Stefan-Boltzmann law: $Q = \sigma T^4$, where $\sigma$ is Stefan’s constant and $T$ is the black body’s temperature. The estimated black body emission anomaly (not shown) calculated from the composite for temperature at a height of 2 m along the propagation route is comparable with the composites for OLR (Fig. 5a and b). The variation in temperature near the surface from the North Sea to the Central Asia region is due to the weather conditions under the high/low pressure systems. The OLR anomaly along the Rossby wave
propagation route seems to mainly reflect the surface temperature anomaly. The positive temperature anomaly from the Iranian plateau to the Central Asia region is seen between day $-4$ and day $+5$ (Fig. 7b and c). These regions are arid. The positive temperature anomaly is caused by the upper-level anticyclonic anomaly, and is connected with the intensification of the heat-low at low levels (Watanabe and Yamazaki 2012). The negative temperature anomaly from South Asia to Southeast Asia is retained from the second to the third pentads (Fig. 7b and c). At the same time, an increase in precipitation and decrease in OLR are seen (Figs 5b–c and 6b–c). The estimated black body emission over these regions is not at all consistent with the OLR anomaly. A large negative OLR anomaly indicates well developed convective cloud with a high cloud top, and the decrease in near-surface temperature is due to the associated precipitation and cloud cover.

Figure 8 shows the composite difference for horizontal wind at 850 hPa. A westerly anomaly from the Arabian Sea to the Bay of Bengal is seen during the first pentad (Fig. 8a). The anomalous westerlies turn northward over the Bay of Bengal, and a cyclonic anomaly appears over South Asia. This anomalous westerly moves northward from the first pentad to the second (Fig. 8b). The anomalous westerlies are seen to be accompanied with OLR anomaly over Indian Ocean (Fig. 5a and 5b). This future is typical for the intraseasonal variation in summer monsoon season (Yasunari 1979; Wheeler and Hendon 2004; Kemball-Cook and Wang 2001; Wang et al. 2006; Pai et al. 2011). The wave-like anomalous wind along 15°N, which is identified as a low-level Rossby wave by Watanabe and Yamazaki (2012), is evident from
the Arabian Sea to Southeast Asia during the third pentad (Fig. 8c). The anticyclonic anomaly over the
Central Asia region strengthens the heat-low below it. And the developed heat-low induces the low-level
Rossby wave (Watanabe and Yamazaki 2012). It is suggested the variation from South and Southeast Asia
due to wave-train from the mid-latitudes coincides with that due to the intraseasonal variation in the
tropics. At the same time, an anomalous southwesterly and northwesterly blew towards the southern and
northern sides of the western Tibetan Plateau, respectively, and the convergence of these anomalous
horizontal winds caused variations in precipitation (Figs 5c and 6c). The prominent increase in
precipitation along the western coast of Southeast Asia and Myanmar is due to these anomalous winds
accompanied by the low-level Rossby wave (Fig. 6c). The easterly anomaly over the topical Pacific
Ocean is seen from the first to the second pentad (Fig. 8a and b). During the third pentad, the easterly
anomaly is seen only over the eastern Pacific Ocean (Fig. 8c).

Figure 9 shows the composite difference for the stream function at 200 hPa. Also shown is the wave
activity computed from the composite difference for the stream function and the climatology in May and
June (Takaya and Nakamura 2001). The wave activity flux shows the propagation of the wave train from
the northern Atlantic Ocean to the Central Asia region during the first and second pentads, and the
eastward propagation along the subtropical jet from the Central Asia region in the third pentad. The scale
of the wave activity flux differs between the first pentad and the final two pentads.

Figure 10 shows the composite difference for velocity potential at 200 hPa. The anomalous divergent
wind is consistent with the variations in OLR and precipitation (Figs 5 and 6). The disturbance in the
tropics, which has the structure of wave number 1, is seen from the first to the last pentad, and migrates
eastward. During the first pentad, an area with active convection is centered on the Indian Ocean, and
another area with suppressed convection is centered on Central America (Fig. 10a). The active convection
over the northern Indian Ocean causes convergence over the Central Asia region where the developed
anticyclonic anomaly is located (Fig. 10a and b). As shown by Rodwell and Hoskins (1995) and Zhang et
al. (2004), the convection over South and Southeast Asia is related to the anomaly over the Central Asia
region. We will confirm this relationship using a numerical LBM experiment Section 6. Part of the active
convection moves to Southeast Asia and converges to the north between day +1 and day +5, which may
cause the decrease in precipitation over the Yangtze River basin (Figs 6c and 10e).

The results of our composite analysis based on the GHP index show that the variation in the
geopotential height anomaly over the Central Asia region is connected to the variation in the early
summer Asian monsoon, and is influenced by two factors. One is the propagation of the wave train from
the northern Atlantic Ocean to the Central Asia region via western Russia. The other is the convection
over the northern Indian Ocean, which is related to disturbance with a wave number of 1 in the tropics.
The disturbance in the tropics shows similar characteristics to the MJO. Referring to Wheeler and Hendon
(2004) and Pai et al. (2011), the composites between day −9 and day −5, and between day −4 and day 0,
seem to correspond to phases 3 and 4 of the MJO index when the active convection is over the Indian
Ocean (Wheeler and Hendon 2004). The composite between day +1 and day +5 also resembles phase 5 of
the MJO index, except the variation from South Asia to Southeast Asia.

5. Wave-train propagation to the western Tibetan Plateau

In this section, we consider the propagation of the wave train from the northern Atlantic Ocean to the
Central Asia region. Firstly, we define an index that represents the propagation of the wave train. We
selected three areas where the centers of variation of geopotential height along the propagation route are
seen: the northern Atlantic Ocean (10°W–10°E, 60°–70°N), western Russia (20°–50°E, 50°–60°N), and
the Central Asia region where the GPH index is defined. The average geopotential height anomaly in each
area was calculated, and the seasonal change and the interannual variation were removed. Finally, the
ARC (Atlantic, Russia, the Central Asia region) index was defined as a combination of the three
standardized area-averaged geopotential height anomalies:

ARC index(t) = northern Atlantic Ocean (t–6) – western Russia (t–2) + Central Asia region (t),

where each term is the regionally-averaged geopotential height anomaly, and the number in parentheses
denotes the time lag, in days, to the Central Asia region. This lag is determined by the propagating speed
(Fig. 4). The term for western Russia is multiplied by –1 because it is opposite in phase to the others. The
ARC index excludes interannual variations by removing the seasonal average; consequently, it represents the intraseasonal wave propagation. It is also standardized. The positive (negative) ARC index represents the propagation from the northern Atlantic Ocean to the Central Asia region via western Russia with a phase distribution of high–low–high (low–high–low). The absolute value of the ARC index reflects the amplitude of the wave train, and its sign shows the phase. The correlation coefficient between the ARC index and the time series of the principal component of the first eigenvector of the variance of geopotential height at 200 hPa over the Central Asia region and surrounding regions (Fig. 2b) is about 0.6. Although the definition of the ARC index is somewhat subjective, the propagation of the wave train from the northern Atlantic Ocean to the Central Asia region is connected to the dominant variation of geopotential height over the Central Asia region and surrounding regions. Composites involving the ARC index when it was more/less than $\pm 1.5$ were calculated using the same procedure as for the GPH index.

Figure 11 shows the difference between the positive and negative composites associated with the ARC index for geopotential height at 200 hPa, and the wave activity flux based on the composite difference for the stream function. The composite of the ARC index shows the propagation from the northern Atlantic Ocean to the Central Asia region via western Russia well. At day $-6$, the high and low anomalies are seen over northeast North America and to its east (Fig. 10a). The composite of the ARC index is used to trace the propagation of the wave train back, and to seek the source of the wave train.

The composite difference for geopotential height averaged between day $-12$ and day $-8$ shows a
positive anomaly over the northeast of North America (Fig. 12a). The wave activity flux shows the
propagation of the wave train from northeast North America to the northern Atlantic Ocean. Before day
-12 there is no significant signal over North America. The wave train propagates to the Central Asia
region in about two weeks (Fig. 10). After arriving on the Central Asia region, the wave train propagates
eastward along the Asian subtropical jet (not shown).

Figure 12b shows the composite difference for the velocity potential at 200 hPa averaged between day
–12 and day –8, and Figure 12c shows the same composite for OLR. Disturbance with a wave number of
1 is seen in the tropics. The center of active convection is over the Indian Ocean, and the center of
suppressed convection is over the central Pacific Ocean, Central America, and the Caribbean Sea. The
phase distribution of the disturbance is similar to that observed in the composite of the GPH index (Fig.
9). We suggest that the marked wave train from northeast North America to the Central Asia region is
related to the MJO-like disturbance.

Barlow and Salstein (2006) and Lorenz and Hartmann (2006) showed that precipitation over Central
America is strongly influenced by the MJO. It is most likely that OLR anomalies from the eastern
equatorial Pacific Ocean to Central America are related to the MJO-like disturbance. The tropical
convection may act as a Rossby-wave source (Sardeshmukh and Hoskins 1988). We suggest that the
anomalous divergence/convergence caused by the variation in convection over the Eastern Pacific Ocean
and Central America is the source of the wave train from northeast North America to the Central Asia
To determine whether the variation in convection over Central America and the northern Indian Ocean, expected from the OLR composites (Figs 5 and 12c), caused the variation in geopotential height over the Central Asia region, two numerical experiments based on the LBM were carried out. Because the LBM experiment in this study includes only the linear response we discuss results of LBM experiments qualitatively, not quantitatively. We used the same magnitudes of the heat source as the second experiment just for simplicity.

In the first experiment, an elliptic heat source, with a horizontal 10° radius of –8 K/day was placed over Central America (centered at 90°W, 20°N), with its maximum at around 500 hPa, and was sustained for the integration period. This thermal forcing mimics the suppressed convection over Central America (Fig. 12c). Atmospheric states were relaxed to the climatology in May and June. The results show the propagation of a wave train from Central America to the Central Asia region (Fig. 13a). The location of each high and low is consistent with composites for the GPH and ARC indexes, and show that variations caused by the tropical disturbance over Central America can contribute to the generation of an ARC wave train with the correct phase. However, anomalies over the northern Atlantic Ocean, western Russia, and the Central Asia region are not large enough to explain the observed values. It seems likely that other
dynamical processes, such as transient wave forcing and/or a wave train from another source, reinforces
the anomaly over the north Atlantic. It is suggested that elaborate models like GCMs are useful for
advanced research into these particulars in future study.

The second numerical experiment followed the same procedure as the first, except that the thermal
forcing was now located over the Indian Ocean (90°E, 5°N) to establish the relationship between the
gopotential height anomaly over the Central Asia region and the variation in convection over the Indian
Ocean in early summer. The results showed an anomalous high over the Central Asia region, and
eastward propagation of the wave train (Fig. 13b).

Consequently, the effect of disturbances in the tropics and middle latitudes tends to be concentrated on
the Central Asia region. When a large-scale disturbance develops in the tropics, such as the MJO, and the
associated active/suppressed convection is located over the Indian Ocean/Central America, a large
positive geopotential height anomaly is likely to develop over the Central Asia region.

7. Summary

This study investigated the relationship between the geopotential height anomaly over the Central
Asia region and variation in the Asian monsoon during the early summer, and has identified one of the
mechanisms associated with the variation of the geopotential height anomaly over the Central Asia region.
The GPH index is defined as an anomaly of geopotential height over the Central Asia region, where a
large intraseasonal variation in geopotential height is seen. Our approach was based primarily on the composite analysis of this GPH index, and the results are summarized below.

1) The marked geopotential height anomaly over the Central Asia region is related to the wave train in the extratropics. The wave train moves from the northern Atlantic Ocean via western Russia before arriving at the Central Asia region and generating the geopotential height anomaly there. From the Central Asia region, another wave train propagates eastward along the Asian subtropical jet.

2) Due to the marked positive/negative geopotential height anomaly over the Central Asia region, an increase/decrease in precipitation from South Asia to Southeast Asia occurs and is sustained for about a week after day 0.

3) The negative near-surface temperature anomaly that develops from South Asia to Southeast Asia, and persists from day −4 to day +5, is caused by the cloud cover and precipitation. The variation in temperature near the surface from the North Sea to the Central Asia region is due to the weather conditions under the high/low pressure systems of the wave train.

4) A negative OLR anomaly over the northern Indian Ocean precedes the development of the positive
geopotential anomaly over the Central Asia region. The composites of velocity potential show a disturbance with a wave number of 1 in the tropics. The region where active convection is associated with the disturbance occurs over the Indian Ocean. Convection is suppressed from the eastern Pacific Ocean to Central America.

5) The wave train that arrives on the Central Asia region can be traced back to northeastern North America at day −10. A disturbance with a wave number of 1 in the tropics is seen, and is accompanied by positive and negative OLR anomalies over Central America and the Indian Ocean, respectively. The OLR anomaly over Central America is thought to be the source of the wave train from northeastern North America.

6) Two numerical experiments based on the LBM were completed to assess the role of tropical heat sources in the variation of geopotential height over the Central Asia region. In the first experiment, the thermal forcing was located over Central America to generate similar wave-train propagation to the composites of the GPH and ARC indexes, although the geopotential height anomaly over the Central Asia region was weak. In the second experiment, the thermal forcing was located over the Indian Ocean, and this also produced a positive geopotential height anomaly over the Central Asia region. Consequently, tropical heat sources associated with MJO-like wavenumber 1 disturbances
constructively generate the height anomaly over the Central Asia region. The positive (negative) geopotential height anomaly over the Central Asia region develops when convection is active (suppressed) over the Indian Ocean, and suppressed (active) over the eastern Pacific Ocean and Central America.
References


Fujinami H, Yasunari T (2004) Submonthly variability of convection and circulation over and around the...
Tibetan Plateau during the boreal summer. J Meteor Soc Japan 82:1545-1564


Madden RA, Julian PR (1994) Observations of the 40-50-day tropical oscillation –A review. Mon Wea


Tamura T, Taniguchi K, Koike T (2010) Mechanism of upper tropospheric warming around the Tibetan...


Watanabe T, Yamazaki K (2012) Influence of the anticyclonic anomaly in the subtropical jet over the western Tibetan Plateau on the intraseasonal variability of the summer Asian monsoon in early summer. J Climate 25:1291-1303

processes, predictability, and the prospects for prediction. J Geophys Res 103:14,451-14,510


Table 1. List of key days for the positive composite.

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Table 2. List of key days for the negative composite.

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</table>
Figure Captions

Fig. 1  The root-mean-square of the modified anomaly of the geopotential height at 200 hPa from May to June in 45 years from 1958 to 2002. Contour interval is 10 m. The rectangle represents the region where the GHP index is defined.

Fig. 2  The first eigenvector for the root-mean-square of time series of modified anomaly of geopotential height at 200 hPa shown in Fig. 1. The eigenvector is non-dimensional.

Fig. 3  (a) the time series of area-averaged geopotential height over the Central Asia region in 1995. The black thin line represents the time series in 1995. The gray thick line represents the climatology smoothed with 5-day running mean. The Unit is m. (b) the time series of the normalized GPH index in 1995. 13 May is selected as a positive key day. 25 May and 22 June are selected as negative key days. Arrows denote selected key days.

Fig. 4  Difference between positive and negative composites of geopotential height at 200 hPa from (a) day –7 to (h) day 0. Contour interval is 40 m. Shading represents the 95% confidence level.
**Fig. 5** Composite difference for OLR averaged between (a) day −9 and day −5, (b) day −4 and day 0, and (c) day +1 and day +5. Contours represent OLR. Contour interval is 10 W m$^{-2}$. Shading represents the 95% confidence level.

**Fig. 6** As for Fig. 5, but for precipitation. Contour interval is 2 mm day$^{-1}$.

**Fig. 7** As for Fig. 5, but for temperature at a height of 2 m. Contour interval is 1 K.

**Fig. 8** As for Fig. 5, but for horizontal wind at 850 hPa. The vector units are m s$^{-1}$. Vectors are only plotted at grid points where the value exceeds the 95% significance level.

**Fig. 9** As for Fig. 5, but for the stream function at 200 hPa. Contours represent the stream function. Contour interval is $2 \times 10^6$ m$^2$ s$^{-1}$. Arrows represent the wave activity flux and are plotted at grid points where the magnitude of wave activity flux is greater than 1 m$^2$ s$^{-2}$.

**Fig. 10** As for Fig. 5, but for velocity potential at 200 hPa. Contours represent velocity potential (interval = $10^5$ m$^2$ s$^{-1}$). Arrows represent the divergence wind, and are plotted at grid points where the magnitude of the divergence wind is greater than 0.5 m s$^{-1}$. 
Fig. 11 Difference between two composites based on the ARC index of geopotential height at 200 hPa at (a) day –6, (b) day –2, and (c) day 0. Contour interval is 40 m. Arrows represent the wave activity flux, and are plotted at grid points where the magnitude of wave activity flux is greater than 1 m$^2$s$^{-2}$.

Fig. 12 As for Fig. 11, but for (a) geopotential height at 200 hPa, (b) velocity potential at 200 hPa, and (c) OLR averaged between day –12 and day –8. (b) Contour Interval is $5 \times 10^5$ m$^2$s$^{-1}$. Arrows represent the divergence wind, and are plotted at grid points where the magnitude of the divergence wind is greater than 0.5 m s$^{-1}$. (c) Contour interval is 10 W m$^{-2}$.

Fig. 13 Results from (a) the first, and (b) the second numerical LBM experiments showing the 30-day average from day 11 to day 40 of the integration. Contours represent the geopotential height at 200 hPa. Contour interval is 4 m.
Figure 3