Quasi-Biweekly Oscillation over the Tibetan Plateau and Its Link with the Asian Summer Monsoon*

MEIRONG WANG

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics, Chinese Academy of Sciences, and University of the Chinese Academy of Sciences, Beijing, China

ANMIN DUAN

State Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics, Chinese Academy of Sciences, Beijing, China

(Manuscript received 26 September 2014, in final form 5 March 2015)

ABSTRACT

Intraseasonal variation (ISV) is especially prominent and unique in the Asian summer monsoon region. In this work, the dominant ISV mode over the Tibetan Plateau (TP) in the summer monsoon season (June–August), together with its structure and evolution, is identified using station observations, Global Precipitation Climatology Project precipitation data, and ERA-Interim during 1979–2011. Results indicate that quasi-biweekly oscillation (QBWO) is the dominant mode of ISV over the TP and is significant in terms of the circulation, precipitation, and diabatic heating fields. In particular, the QBWO is closely related to the onset and active/break phases of the TP summer monsoon. In most cases, the QBWO originates from the equatorial western Pacific and first propagates northwestward to the Bay of Bengal and northern India, then northward to the southeastern TP, and finally eastward to the East Asian area, showing a clockwise propagation pathway. Two main mechanisms are responsible for the northward propagation of the QBWO signals. The first, in operation when the QBWO signals are located to the south of 20°N, is the generation of barotropic vorticity induced by the easterly vertical shear, leading to the northward movement of the convection. The second mechanism, responsible for the propagation taking place farther north toward the TP, is a moisture advection effect that destabilizes the lower atmosphere ahead of the convection. Further analyses suggest that the QBWO plays a role in linking the ISV of the different subsystems of the Asian summer monsoon as a macroscale monsoon system.

1. Introduction

Intraseasonal variation (ISV) modes generally comprise the 30–60-day oscillation [also known as the Madden–Julian oscillation (MJO; Madden and Julian 1971) in the boreal winter and boreal summer intraseasonal oscillation (BSISO; Wang and Xie 1997)] and 10–20-day oscillation [also referred to as the quasi-biweekly

oscillation (QBWO)]. Madden and Julian (1971) were the first to detect the 30–60-day oscillation in the tropical Pacific and confirmed that it exists over the global tropics (Madden and Julian 1972). The QBWO was first found in the Indian Ocean and the western Pacific (Murakami and Frydrych 1974; Murakami 1975; Numaguti 1995). Subsequent studies indicated that these two dominant ISV modes are also evident over the subtropics (Blackmon 1976; Blackmon et al. 1977; Krishnamurti and Subrahmanyam 1982; Anderson and Rosen 1983; Krishnamurti and Gadgil 1985; Jeong et al. 2008; Mao et al. 2010; Wen et al. 2011) and even over the entire globe (Kikuchi and Wang 2009; Yang 2009).

It is well known that the active and break phases of the Asian summer monsoon are affected by ISV. Yasunari (1979, 1980) found that the 40-day fluctuation is closely

* Supplemental information related to this paper is available at the Journals Online website: <http://dx.doi.org/10.1175/JCLI-D-14-00658.s1>.

Corresponding author address: Dr. Anmin Duan, LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences, P.O. Box 9804, Beijing 100029, China.
E-mail: amduan@lasg.iap.ac.cn

associated with the major active and break cycles of the monsoon over the whole of the Asian monsoon areas. The QBWO also regulates the active and break phases of the Indian summer monsoon (Krishnamurti and Ardanuy 1980; Chen and Chen 1993), the South China Sea summer monsoon (Chen and Chen 1995; Chen et al. 2000; Chan et al. 2002), and the East Asian summer monsoon (EASM) (Mao and Wu 2006; Jiang and Lau 2008; Fujinami and Yasunari 2009; Jia and Yang 2013).

The elevated heating of the Tibetan Plateau (TP) plays a fundamental role in the formation and variation of the summer circulation, at least over Asia (Flohn 1957, 1960; Hahn and Manabe 1975; Luo and Yanai 1983, 1984; Wu and Zhang 1998; Duan and Wu 2005; Wu et al. 2012). Previous studies have demonstrated that the TP is a relatively independent subsystem of the Asian summer monsoon region (Tang and Reiter 1984; Duan et al. 2013), and the associated multiscale ISV modes in terms of wind, convection, precipitation, relative vorticity, and atmospheric diabatic heating have been recognized (Krishnamurti and Subrahmanyam 1982; Nitta 1983; Zhang et al. 2009; Wang et al. 2011; Zhang et al. 2014). Nitta (1983) demonstrated that the vertically integrated heat sources over the eastern TP present medium-term changes with 10–15- and 30-day periods, which is strongly associated with the South Asian high (or Tibetan high) and can be regarded as an ISV activity of the entire Asian summer monsoon system. He et al. (2006) found that, in summer, the southeastern TP was the most active ISV area of blackbody temperature. In particular, the ISV modes over the TP, to some extent, can influence local and even large-scale weather systems, including the plateau low vortex (Zhang et al. 2014), subtropical high (Li et al. 1991), and South Asian high (Liu and Lin 1991), as well as the precipitation anomaly in eastern China (Zhou et al. 2000). ISV signals over the TP may originate from either the lower latitudes, such as the Bay of Bengal and the southern rim of the TP (Krishnamurti and Subrahmanyam 1982; Zhang et al. 2009) or middle-to-high latitudes (Blackmon et al. 1984a,b; Xie et al. 1989). Other studies (Peng and Wu 1995; Zhou et al. 2000) have also argued that the TP itself may act as a source and a sink of ISV modes.

Compared to the large volume of literature concerning ISV over the East Asian and South Asian monsoon regions (Chen et al. 2000; Ko and Hsu 2006; Mao and Wu 2006; Fujinami and Yasunari 2009; Yang et al. 2010, 2014), the ISV over the TP monsoon region has received much less attention due mainly to insufficient observations. In addition, most studies were generally based on case studies. Therefore, it is necessary to conduct a systematic investigation of the ISV over the TP, including its characteristics and propagation path, by

analyzing long-term observational datasets. Furthermore, from the viewpoint of the Asian monsoon system, the role of the ISV modes during the onset, active, and break phases of the TP summer monsoon (TPSM) and its link to the ISV of the other two subsystems of the Asian summer monsoon [i.e., EASM and South Asian summer monsoon (SASM)] need to be fully explored.

In this paper, we perform an observational study for the period of 1979–2011 to gain insight into the structure and propagation characteristics of the dominant ISV mode over the TP and its link to monsoon activities in adjacent regions. In section 2, a brief description of the datasets and statistical methods is given. The characteristics of the ISV over the TP, including the dominant periodicities and propagations, are investigated in section 3. In section 4, the link between the QBWO and the onset, active, and break phases of the TPSM is investigated. Section 5 describes the specific propagation path and evolution of the QBWO in the whole Asian monsoon region. Finally, section 6 summarizes and discusses the key findings of the study.

2. Data and methodology

The following datasets are employed in this study:

- 1) Regular surface meteorological observations at 71 stations, with an initial quality control provided by the China Meteorological Administration: Data were gathered four times daily from 1979 to 2011, with variables including surface air temperature, ground surface temperature, 10-m wind speed, daily accumulated precipitation, and station pressure.
- 2) The European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) dataset with a horizontal resolution of 1.5° latitude by 1.5° longitude and 37 vertical levels (<http://apps.ecmwf.int/datasets/data/interim-full-daily/>): The analysis used here covers the period of 1979–2011 and the daily mean field is calculated by simply averaging the original 6-hourly data. We choose ERA-Interim because it shows significant improvement over the old generation of reanalysis datasets, such as the NCEP–NCAR and ERA-40 reanalyses, particularly with respect to the temperature and horizontal wind fields over the TP (Bao and Zhang 2013; Wang and Zeng 2012). Besides, ERA-Interim shows the highest skill among the several available reanalysis datasets in representing the climatology, long-term trend, and interannual variability of global monsoon precipitation (Lin et al. 2014).
- 3) Pentad Global Precipitation Climatology Project (GPCP) is a merged analysis that incorporates

FIG. 1. Mean power spectra of the (a) air temperature, (b) precipitation, (c) geopotential height, (d) surface sensible heat flux, (e) $\langle Q_1 \rangle$, (f) $\langle Q_2 \rangle$, (g) meridional wind at 500 hPa, and (h) 200-hPa ISV components by calculating the average of the individual power spectra for each of the 33 summer seasons (1979–2011) over the TP region. Red lines show Markov red noise spectrums; blue lines show a priori 99% confidence bounds; and green lines show a posteriori 99% confidence levels. The x axis represents the periodicity and the y axis denotes the variance.

precipitation estimates from satellite and gauge observations (Huffman et al. 1997), which is on $2.5^\circ \times 2.5^\circ$ grids over the whole globe and covers the period of 1979–2011 (ftp://ftp.cpc.ncep.noaa.gov/precip/GPCP_PEN/). Daily GPCP precipitation data with a horizontal resolution of 1° latitude by 1° longitude during 1997–2011 are also used. Ma et al. (2009) suggested that GPCP exhibits a smaller rainfall bias in most regions of China compared to the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997).

Following Duan and Wu (2008), the surface sensible heat flux at 71 stations over the central and eastern TP (CETP) is calculated by the bulk aerodynamic formula (Wang et al. 2012). The vertically integrated atmospheric apparent heat source ($\langle Q_1 \rangle$) and moisture sink ($\langle Q_2 \rangle$) are derived by residual budget analyses based on the thermodynamic and moisture equations (Yanai 1961; Yanai et al. 1973) and by using ERA-Interim.

The ISV component is extracted from the “raw” daily and pentad time series through first removing the climatology and then removing synoptic fluctuations by taking a 5-day running mean. The climatology is calculated by climatological 5-day running mean time series, following the same method as Yang et al. (2010). Hence, the ISV component to be examined in this study is the year-to-year varying part of the ISV, defined as

the transient intraseasonal oscillation (TISO; Yang et al. 2009).

3. Characteristics of the ISV over the TP

a. Dominant periodicities of the ISV over the TP

To obtain the dominant periodicities of the atmospheric variability over the TP, we analyze mean power spectra of the 33 summer seasons (June–August) of eight meteorological elements’ ISV components over the TP region from 1979 to 2011 (Fig. 1). It is clear that all the eight meteorological elements show a significant 10–20-day period (i.e., QBWO), including air temperature (Fig. 1a), precipitation (Fig. 1b), geopotential height (Fig. 1c), and surface sensible heat flux (Fig. 1d) averaged at 71 meteorological stations over the CETP. Similar results can be detected in the $\langle Q_1 \rangle$ (Fig. 1e) and $\langle Q_2 \rangle$ (Fig. 1f), and meridional wind at 500 (Fig. 1g) and 200 hPa (Fig. 1h), estimated by ERA-Interim in the whole TP region (25° – 40° N, 75° – 105° E). In addition, a 7–9-day period is also significant in almost all of these variables except surface sensible heat flux. For periods longer than 30 days, however, it is significant only in some variables such as geopotential height and meridional wind at 500 and 200 hPa.

To examine the interannual variability of the ISV over the TP, we choose precipitation as a representative

FIG. 2. The (a) interannual distribution of periods corresponding to marked spectra peaks of intraseasonal activity passing a posteriori 99% confidence level and (b) interannual variations of variance contribution percentage from the 10–20-, 20–30-, and 30–60-day oscillation to intraseasonal oscillation in the individual power spectra for the 71-station-averaged summer daily precipitation over the CESTP during 1979–2011.

variable, and spectral analyses for the precipitation field are conducted for each year from 1979 to 2011 (Fig. 2a). It is clear that the QBWO is significant in each individual year, although signals with periods longer than 20 days (e.g., 20–30 and 30–60 days) are also significant in some years.

In Fig. 2b, the year-by-year variance contribution percentage in individual power spectra from each ISV mode for observed daily precipitation time series averaged at 71 stations over the CESTP is plotted. It is apparent that the variance contribution from the QBWO is above 20% in most years and even exceeds 50% in some cases. For example, the QBWO explains nearly 70% of the variance in 1989. The 33-yr-averaged QBWO accounts for about 35% of the variance in the ISV of daily precipitation over the TP during 1979–2011, manifesting the dominance of the QBWO in the ISV of precipitation over the TP. In contrast, the 30–60-day oscillation

shows a much larger interannual variability, and its averaged variance contribution is about 24% during the 33 years. Meanwhile, the 20–30-day oscillation is typically less than 20%, except for three years (1981, 1982, and 2005), with a 9% average variance contribution over the 33 years. So, the 20–60-day oscillation accounts for about 33% of total variance, indicating that it also is significant in some certain years.

b. Characteristics of the QBWO propagation over the TP

The meridional propagation characteristics of the QBWO over the TP can be illustrated by a Hovmöller diagram of 10–20-day filtered daily $\langle Q_1 \rangle$ along 70°–100°E during 1979–2011 on a year-by-year basis. According to the QBWO propagation directions of $\langle Q_1 \rangle$, we can categorize them into three types: 1) propagating from the low latitudes to the TP (type I); 2) propagating

TABLE 1. The specific categories of propagation directions of filtered $\langle Q_1 \rangle$ and GPCP precipitation (parentheses) are listed with their occurrence year and total year statistics. The years in which all of the propagation directions of $\langle Q_1 \rangle$ occurred are in boldface.

Categories	Main years	Total years
From low latitudes to the TP (type I)	1979, 1980, 1983 , 1984, 1985, 1986, 1987 , 1988 , 1989, 1990, 1991, 1992, 1993, 1995, 1996, 1997, 1998, 1999, 2000, 2001 , 2002 , 2003 , 2004, 2005, 2008, and 2011 (1997, 1998, 1999, 2000, 2001, 2003, 2004, 2006, 2007, 2008, and 2011)	26 (11)
From high latitudes to the TP (type II)	1983 , 1984, 1987 , 1988 , 1989, 1990, 1995, 1997, 2001 , 2002 , 2003 , 2006, and 2009 (1997, 2000, 2001, 2003, and 2007)	13 (5)
From the TP to low and/or high latitudes (type III)	1979, 1981, 1983 , 1985, 1987 , 1988 , 1992, 1994, 1996, 1999, 2000, 2001 , 2002 , 2003 , 2007, 2008, 2009, 2010, and 2011 (1999, 2001, 2002, 2003, 2008, 2009, and 2010)	19 (7)

from the high latitudes to the TP (type II); and 3) propagating from the TP to high and/or low latitudes (type III). Statistical analyses show that type-I propagation is dominant in 26 years from 1979 to 2011, indicating the northward propagation from low latitudes toward the TP is the most pronounced and appears in almost all years. Meanwhile, the other two kinds are detected in 13 (type II) and 19 years (type III), respectively (Table 1). In the GPCP precipitation, these three kinds of propagation occur in 11, 5, and 7 years during 1997–2011, respectively. Most years are consistent with that for $\langle Q_1 \rangle$, especially in convection-dominant regions. Generally, there is more than one kind of propagation direction appearing in each year. Here, in Fig. 3, we only show the meridional propagation features of $\langle Q_1 \rangle$ for six years (boldface values in Table 1), in which all three kinds of propagation directions can be detected. The northward propagation from low latitudes is always much stronger and spans more latitudes than the other kinds of propagation; in some cases, the northward propagation even originates from the Southern Hemisphere, crosses the equator, and then spreads to the TP region. Moreover, the northward signals seem to be strengthened after reaching the TP in some years, such as 1983, 1988, 2002, and 2003. In contrast, the southward propagation from high latitudes is comparatively weak, although it also appears in many years. It is of interest to note that the propagation from low and high latitudes can spread to the TP region at the same time, denoting the phase-locking phenomenon. Another interesting feature is the propagation away from the TP itself; sometimes, it can propagate southward and northward simultaneously. Here, we only focus on the first kind of propagation (type I), and the other two kinds will be discussed in a separate paper.

To reveal the meridional and vertical structure of the QBWO, we take a composite approach by selecting 48 type-I propagation cases during 1979–2011. The composite is constructed as such that at day 0 the QBWO convection is located at exactly 27°N along 70°–100°E, where it is already over the TP. The temporal evolution

of the composite $\langle Q_1 \rangle$ and moisture conditions from day –18 to 0 (with a 2-day interval) is given in Figs. 4 and 5. The anomalous positive QBWO convection and corresponding moisture convergence can be found in the South China Sea near the equator at day –12. Then, the anomalous positive convection moves northwestward and intensifies gradually, in agreement with a cyclonic circulation and significant moisture convergence center. At day –4, the maximum $\langle Q_1 \rangle$ center arrives mainly in the Indian subcontinent and partly in the Bay of Bengal, with a somewhat weakened signal. It then turns northward and arrives at the southern slope of the TP at day –2; by this time, the anomalous convection has been intensified again. The strengthening of the QBWO in $\langle Q_1 \rangle$ could be related to the very strong moisture transport. This is because, from day –4, the cyclonic circulation gradually collapses and evolves into robust southwest airflow, which can transport plenty of moisture to the TP, coinciding with the large value of moisture convergence over the southern slope of the TP (Fig. 5). To confirm the results from $\langle Q_1 \rangle$, we also plot the composite QBWO evolution of GPCP daily precipitation (20 type-I cases) from day –18 to 0. Obviously, the results (contours in Fig. 4) are highly consistent with $\langle Q_1 \rangle$.

Of particular importance is the generation of an anomalous positive $\langle Q_1 \rangle$ center over the South China Sea at day –12. After that, this anomalous positive convection moves northwestward to the Indian subcontinent and then moves northward. It implies that such a kind of QBWO may originate from the tropical ocean.

The essential atmospheric dynamical processes that give rise to the northward propagation of the 10–90-day boreal summer intraseasonal oscillation have been examined in previous studies (e.g., Jiang et al. 2004; Abhik et al. 2013). Here, we conduct similar composite analyses of the meridional–vertical structure of 10–20-day filtered vorticity (Fig. 6a), divergence (Fig. 6b), vertical velocity (Fig. 6c), and specific humidity (Fig. 6d) fields with respect to the maximum convection center (MCC)

FIG. 3. Hovmöller diagrams of 10–20-day filtered $\langle Q_1 \rangle$ (W m^{-2}) along $70^\circ\text{--}100^\circ\text{E}$ for a sample of 6 yr. Positive $\langle Q_1 \rangle$ anomalies are shaded and negative anomalies are shown by dashed contour lines. The two vertical dashed gray lines and the horizontal dashed blue line represent the TP longitudinal area and the onset date of the TPSM, respectively. Black, green, and purple arrows represent the three kinds of propagation direction of 10–20-day QBWO. The blue arrows represent the propagation of the QBWO to the TP before the onset date of the TPSM.

at day -6 , when the MCC is located at 15°N . The MCC is defined by the latitude where the maximum $\langle Q_1 \rangle$ or precipitation is observed. Obviously, a barotropic structure of vertical velocity accompanies the MCC and coincides with the lower-level convergence and upper-tropospheric divergence. Similar to that for the

northward-propagating 15–90-day mode, as illustrated in Jiang et al. (2004), the most striking asymmetric structure with respect to the MCC associated with the northward QBWO is also noted in the vorticity field. With an equivalent barotropic structure, its positive center is situated roughly 200 km to the north of the

FIG. 4. Evolution of the 10–20-day filtered $\langle Q_1 \rangle$ (shaded; W m^{-2}) and GPCP precipitation (contours; mm day^{-1}) composite from day –18 to 0. Day 0 represents a reference time when the convection center moves to 27°N along 70° – 100°E . The shaded area and contours exceed the 90% confidence level. The black solid thick curves are drawn to indicate the TP area with height above 2000 m (the same below). The contour interval of GPCP is 0.5 mm day^{-1} . Blue solid dots in the top-left panel denote five stations used to calculate the TPSM index.

FIG. 5. As in Fig. 4, but for the vertically integrated water vapor flux (vectors; $\text{kg m}^{-1} \text{s}^{-2}$) and moisture flux divergence ($10^{-5} \text{ kg m}^{-2} \text{s}^{-2}$). The dotted areas and vectors exceed the 90% confidence level.

MCC. The barotropic vorticity and lower-level convergence in the boundary layer appear to help the lower-level moisture convergence to the north of the MCC, corresponding to the observed maximum specific humidity (Fig. 6d) appearing nearly 200 km to the north of

the MCC in the lower troposphere. In addition, at days -8 and -6 , a significant positive surface heating flux exists in northern India, located to the north of the MCC (figure not shown). Hence, surface sensible heating might also contribute to the northward propagation,

FIG. 6. Meridional-vertical structures of the northward-propagating QBWO cases for (a) vorticity (10^{-6} s^{-1}), (b) divergence (10^{-6} s^{-1}), (c) vertical velocity ($10^{-3} \text{ Pa s}^{-1}$), (d) specific humidity ($10^{-5} \text{ kg kg}^{-1}$), (e) horizontal moisture advection ($10^{-10} \text{ kg kg}^{-1} \text{ s}^{-1}$), and (f) equivalent potential temperature (K) from ERA-Interim at day -6 . Shaded areas exceed the 90% confidence level. The horizontal axis is the meridional distance (degrees) with respect to the maximum convection center. Positive (negative) values mean to the north (south) of the convection center, and the vertical axis is the pressure (hPa).

by making the lower troposphere less stable (Webster 1983; Hsu et al. 2004).

At day -4 , the MCC is located at approximately 20°N , and by the end of the following 2 days the convection has already been over the TP. Previous work (Jiang et al. 2004; Abhik et al. 2013) focused on the propagation of the 10–90-day oscillation in the area from the equator to around 20°N , without attention to the QBWO and whether and how the QBWO can continue to move northward to the TP. At day -4 (Fig. 7), the relatively weaker vertical structure of vorticity, convergence, and vertical velocity are similar to those at day -6 (Fig. 6). The main difference lies in the specific humidity field, which becomes much stronger than before. Actually, some studies have noted that a northward shift of

boundary layer moisture is associated with the northward-propagating boreal summer intraseasonal oscillation (Jiang et al. 2004; Fu et al. 2006) and the importance of negative horizontal moisture advection during the initiation of extended break spells of Indian summer monsoon (Prasanna and Annamalai 2012). Here, the 10–20-day filtered positive horizontal moisture advection maximum center leads the convection by 3° at low levels at day -4 (Fig. 7e), while it is negative at day -6 (Fig. 6e). The robust horizontal moisture advection will destabilize the lower atmosphere ahead of the convection by enhancing the equivalent potential temperature and shifting the specific humidity center to the north of the MCC. In Figs. 6f and 7f, we can see that the intensity of the equivalent potential temperature

FIG. 7. As in Fig. 6, but at day -4.

ahead of the convection has a certain gap. Therefore, at day -4, in the process of convection moving toward the TP, the very strong horizontal moisture advection, rather than the barotropic vorticity, plays a critical role by helping the lower-level moisture convergence move to the north of the MCC.

In brief, at day -6, the most striking asymmetric structures appear in the vorticity field and the homologous vertical motion and divergence fields, which suggests that the easterly vertical shear mechanism, as put forward by Jiang et al. (2004) to explain the northward propagation of the 15–90-day ISV mode over the Indian Ocean, may also play a role in the northward propagation of the QBWO identified in this study. The key process with this mechanism is the generation of barotropic vorticity due to the coupling between the free-atmosphere baroclinic and barotropic modes in the presence of the vertical shear of the mean flow. The

induced barotropic vorticity to the north of the MCC can lead to a northward shift of the moisture convergence in the boundary layer and thus a northward propagation of the convection. However, the easterly shear of the mean flow is very weak in the area from 20°N to the southern rim of the TP (see Fig. S1 of the supplementary material). After day -4, therefore, the reason that convection moves farther northward over the TP cannot be explained by the easterly vertical shear mechanism. Instead, the strong horizontal moisture advection destabilizes the lower atmosphere ahead of the MCC and shifts the specific humidity center to the north of the MCC. This then leads to the northward displacement of the convective heating, and thus the convection tends to move northward, which corresponds closely to the robust southwest moisture transport in Fig. 5 from day -4. Besides, the TP is the highest (average elevation: > 4000 m) plateau in the world, and the

TABLE 2. The onset dates of the TPSM.

Year	Date	Year	Date	Year	Date
1979	30 Apr	1990	18 May	2001	5 Jun
1980	4 Jun	1991	15 Apr	2002	16 May
1981	1 Jun	1992	16 May	2003	25 Apr
1982	1 Jun	1993	11 May	2004	21 Apr
1983	3 Jun	1994	3 May	2005	15 Apr
1984	5 Jun	1995	13 May	2006	26 May
1985	14 May	1996	20 Apr	2007	20 May
1986	10 May	1997	11 May	2008	22 Apr
1987	5 Jun	1998	5 May	2009	27 May
1988	26 May	1999	16 Apr	2010	16 Apr
1989	11 May	2000	2 Apr	2011	18 Apr

elevated strong heat source can strengthen the warm moist airflow from low latitudes toward the plateau (Liang et al. 2005). Thus, the TP itself also plays a role in the farther northward propagation of the QBWO by strengthening the moisture transport.

4. The relationship between QBWO and the onset and active/break phases of the TPSM

Over subtropical East Asia, the QBWO has been identified to be a very important mode of the ISV of precipitation and convection during summer (Fukutomi and Yasunari 1999; Chen et al. 2000; Ko and Hsu 2006; Mao and Wu 2006; Jiang and Lau 2008; Fujinami and Yasunari 2009; Yang et al. 2010, 2014). The activity of the QBWO can exert significant influence on the onset and active/break phases of the EASM and SASM and even the entire Northern Hemisphere summer monsoon (Yasunari 1979, 1980; Krishnamurti and Ardanuy 1980; Chen and Chen 1993; Lau and Yang 1996; Goswami et al. 2003; Zhou and Chan 2005). However, less effort has been devoted to examining the link between the QBWO and TPSM (Li and Duan 2011), due mainly to the limited availability of observational data. Following Tang (1995), we calculate the TPSM index from the geopotential height difference between the four boundary points (32.5°N, 80°E; 25°N, 90°E; 32.5°N, 100°E; and 40°N, 90°E) and the midpoint (32.5°N, 90°E) (shown in Fig. 4), which generally represents the activity of the surface heat low over the TP. Note that the geopotential height at five points is calculated from station pressures in terms of the barometric height formula. Following Bai et al. (2001), the onset date of the TPSM is defined as follows: when a particular day the TPSM index turns positive from negative, even if negative again, but the continuous number of days is less than 5, then that day will be defined as the start date of the TPSM. The onset dates of the TPSM for each year during 1979–2011 are listed in Table 2. The onset dates

of the TPSM exhibit large interannual variability. The earliest date is 2 April 2000, and the latest date is 5 June for 1984, 1987, and 2001. On average, the onset date appears around the middle of May.

To investigate the link between the QBWO and TPSM activity, 181 days (90 days before and after the onset date) of TPSM index and summer precipitation averaged at 71 stations are chosen for each year. The 33-summer-averaged Mexican hat wavelets (Torrence and Compo 1998) of TPSM index and precipitation are shown in Fig. 8. The QBWO signal in TPSM index strengthens markedly before monsoon onset. Moreover, the propagation of the QBWO is closely related to the TPSM evolution. Before and after the onset of TPSM, the QBWO signals exist over the TP in most years, accompanying the meridional propagation from low and/or high latitudes or the local QBWO over the TP region (Fig. 3). About one month after monsoon onset, the significant QBWO signal in precipitation is greatly enhanced and then reaches its maximum in the middle of July. Later, with the weakening and end of the TPSM, the precipitation over the TP reduces gradually and the QBWO signal also disappears at the end of August.

Composite analysis is performed to further characterize the heating and circulation features associated with the active and break phases of the QBWO in TPSM index. The active phase is defined as when the 10–20-day filtered TPSM index is greater than its standard deviation, and the break phase is defined as when the 10–20-day filtered TPSM index is smaller than its negative standard deviation. Based on this definition, 428 active and 430 inactive days are selected from the 33 summers (about 121 QBWO events). Then the active and break composites of filtered circulation, precipitation, moisture conditions, and diabatic heating are constructed in Fig. 9. When the QBWO signal is active, more precipitation occurs in the CETP and its downstream area, while less occurs in northern India, northern China, and eastern China. In the break phase, the conditions are opposite: that is, deficient precipitation is found in the CETP, whereas above-normal precipitation is found in northern India, northern China, and eastern China (Figs. 9c,d). This result is similar to that of Liu and Ding (2008), in which they investigated the interannual correlation in precipitation between the Indian summer monsoon region and northern China and found a teleconnection pattern characterized by consistent precipitation variations in the two regions. Moreover, the phase relationship between $\langle Q_1 \rangle$ (Figs. 9g,h) and precipitation indicates that the intraseasonal activity of the TPSM is closely associated with diabatic heating over the TP. In addition, regardless of whether we consider

FIG. 8. Mean wavelet power spectra of the (a) TPSM index and (b) summer precipitation by calculating the average of the individual wavelet power spectrum for each of the 33 summer seasons (1979–2011) over the TP region. Day 0 represents the TP summer monsoon onset day; day –90 indicates 90 days ahead of the onset day; and day 90 indicates 90 days after the onset day. Dotted areas exceed the 90% confidence level, and the thick dashed curve on either end indicates the edge effects.

precipitation or $\langle Q_1 \rangle$, the QBWO phases are opposite between the western and the CETP.

Precipitation anomalies are consistent with the large-scale circulation. Corresponding to the changes of the QBWO in the TPSM, the circulation and moisture conditions are depicted in Figs. 9a–f. During the active phase of the QBWO, the 500-hPa circulation is characterized by a cyclone over the TP. Correspondingly, a warm and wet southwesterly exists in the southeastern TP, and adequate water vapor supply from the low latitudes leads to a large value of water vapor convergence over the CETP, which favors above-normal precipitation. During the break phase, however, an anticyclone dominates the TP at 500 hPa, and most areas of the TP and its eastern side are controlled by a dry and cold northwesterly, corresponding to the water vapor divergence zones. Zhang et al. (2014) showed similar circulation and moisture conditions by analyzing TP vortices and the related 10–30-day intraseasonal oscillation. In both the active and break phases of the QBWO, the signal of the precipitation anomaly in northern India is opposite to that in the CETP.

At 200 hPa, the western and the CETP are under the control of different circulation types. In the summer climatology, the upper-tropospheric circulation over the TP is governed by the huge Tibetan high, so the anomalous anticyclone over the CETP in Fig. 9a can superimpose on the climatological anticyclone to strengthen itself favoring more precipitation over the CETP (Duan et al. 2008). On the contrary, the distribution in the break phase is shown in Fig. 9b, corresponding to less precipitation over the CETP. The changes of position and intensity of the Tibetan high during the active and break phases of the TPSM, in association with the QBWO of the Tibetan high, reflect the alternation of

active and break phases of the monsoon in the lower troposphere (Krishnamurti 1973).

In terms of precipitation occurrence, it is necessary to examine the high- and low-level configuration of divergence and vertical velocity (Fig. 10). In the active (break) phase of the QBWO, a significant negative (positive) center of velocity potential is situated in the CETP at 200 hPa; that is, high-level divergence and 500-hPa convergence become enhanced (weakened) over the CETP. When the QBWO is abnormally strong, the high- and low-level configuration of divergence is conducive to abundant precipitation over the CETP and vice versa. Correspondingly, in the height–longitude cross sections of vertical velocity averaged over 27°–40°N (Figs. 10e,f), we can see significant ascending (descending) motion taking place over the CETP in the active (break) phase, whereas the opposite situation happens in the western TP. Therefore, the positive and negative phases of the QBWO are closely related to the active and break phases of the TPSM.

5. Propagation of the QBWO in the Asian summer monsoon region

Considering the fact that the QBWO is the common feature for all three subsystems of the Asian summer monsoon, it is necessary to investigate whether the QBWO can link them together on intraseasonal time scales.

Figure 11 shows the meridional and zonal propagation characteristics of the QBWO in $\langle Q_1 \rangle$ over the Asian monsoon region. Similar features can also be noted based on the GPCP daily precipitation data (Fig. 11). Figures 11a,c clearly indicate that the QBWO signal over India can move northward into the TP area but

FIG. 9. Distribution of 10–20-day filtered circulation at (a),(b) 200 hPa; (c),(d) 500 hPa (vectors; m s^{-1}) and summer precipitation (shaded; mm day^{-1}); (e),(f) vertically integrated moisture flux (vectors; $\text{kg m}^{-1} \text{s}^{-2}$) and divergence (shaded; $10^{-5} \text{kg m}^{-2} \text{s}^{-2}$); and (g),(h) $\langle Q_1 \rangle$ (W m^{-2}) in (left) the active phase and (right) the break phase. Pink vectors and dotted areas exceed the 90% confidence level.

stops at about 30°N in most cases; the QBWO signal in some cases can also move to the midlatitudes (about 50°N) through the TP area, which can be seen in Fig. 3. This may be due to the fact that the TP itself could be an active ISV center (Peng and Wu 1995; Zhou et al. 2000). When the QBWO signal propagates from India into the TP area, it may become a part of the QBWO that originated in the TP local area, subsequently strengthening and becoming capable of propagating to the

middle-to-high latitudes. The QBWO, therefore, is likely to be a medium linking the SASM and TPSM together. From Fig. 11b, we can see that the significant QBWO signal in India may originate from the area around the South China Sea region (near 120°E), and undergoes westward movement toward the Indian region. Figure 11d indicates that the QBWO in the TP monsoon region exhibits an obvious eastward propagation and eventually arrives at eastern China longitudes

FIG. 10. The 10–20-day filtered velocity potential (contours; $10^6 \text{ m}^2 \text{ s}^{-1}$) and divergent winds (vectors; m s^{-1}) at (a), (b) 200 and (c), (d) 500 hPa and (e), (f) height–longitude cross sections of vertical velocity ($10^{-4} \text{ Pa s}^{-1}$) averaged over 27° – 40°N in (left) the active phase and (right) the break phase. Pink vectors and blue contours exceed the 90% confidence level.

(120°E). As a result, the QBWO may also serve to connect the TPSM with the EASM. Therefore, the QBWO over the TP is an important and efficient “bridge” linking the other two subsystems of the Asian summer monsoon (EASM and SASM) on the intraseasonal scale.

To further understand the physical nature and likely source area of the QBWO in the Asian summer monsoon area, we conduct an analysis of its life cycle to investigate its process of evolution. A better alternative for describing the complex spatiotemporal propagation features is to use the extended empirical orthogonal function (EEOF) technique (Weare and Nasstrom

1982). EEOF analysis with 2- and 4-day lags is used on the 10–20-day filtered $\langle Q_1 \rangle$ in the Asian monsoon area, and a pair of the first two leading EEOFs represents a half cycle of the QBWO mode (figure not shown). A similar pattern of the QBWO occurs in GPCP daily precipitation during 1997–2011, but the signals from midlatitudes are not so evident and more signals cover the East Asian area (see Fig. S2 of the supplementary material).

The strength and phase of the QBWO mode can be described by a combination of the first two coefficients [principal components (PCs)] for each EEOF mode, and we choose significant and well-defined QBWO events to

FIG. 11. (a) Meridional propagations of filtered $\langle Q_1 \rangle$ (shaded) and GPCP precipitation (contour) averaged along 60°–85°E and (b) zonal propagations of filtered $\langle Q_1 \rangle$ (shaded) and GPCP precipitation (contour) averaged along 10°–20°N calculated from the lead–lag correlation with respect to the Indian summer monsoon region. (c) Meridional propagations of filtered $\langle Q_1 \rangle$ (shaded) and GPCP precipitation (contour) averaged along 75°–100°E and (d) zonal propagations of filtered $\langle Q_1 \rangle$ (shaded) and GPCP precipitation (contour) averaged along 25°–40°N calculated from the lead–lag correlation with respect to the TP summer monsoon region. The units of the x axis in (left) and the y axis in (right) are days. Shading and pink contours denote correlation above the 90% confidence level. The two dashed gray lines in (c) represent the latitudinal range of the TP.

composite the life cycle of the QBWO. Here, the specific definition of a significant QBWO event follows that of Kikuchi et al. (2012). The composite life cycle of the QBWO (Fig. 12b) is constructed based on a phase space consisting of the first two PCs (Fig. 12a). A negative anomaly of convection associated with the QBWO appears first in the equatorial western Pacific in phase 1. It then moves northward to dominate the South China Sea

region and continues to spread westward to the South Asian region (phase 5). This northwestward propagation is in good agreement with previous studies, which suggest that the QBWO in the Indian monsoon region originates in the equatorial western Pacific and propagates westward to the Bay of Bengal (Krishnamurti and Bhalme 1976; Annamalai and Slingo 2001; Chatterjee and Goswami 2004). The westward phase speed of the type-I QBWO is

FIG. 12. (a) Phase space representation of the state of the QBWO (principal components PC1 and PC2) for the period 1979–2011. The state of the QBWO categorized as eight phases are used for the discussion. The approximate positions of the major negative convection areas in some phases are also denoted. Note that each PC is normalized by one standard deviation of the corresponding PCs during the period each EEOF analysis is performed to obtain the EEOFs. (b) Composite life cycle of the QBWO (phases 1–8): filtered $\langle Q_1 \rangle$ (shaded; W m^{-2}) and 500-hPa horizontal winds (vectors; m s^{-1}). Significant vectors and dotted areas are at the 90% confidence level.

about 6 m s^{-1} , with a baroclinic vertical structure (figure not shown), and is thus likely to be equatorial Rossby waves in the presence of monsoon mean flow and convective coupling (Chen and Chen 1993; Chatterjee and Goswami 2004; Kikuchi and Wang 2009).

Meanwhile, the convective anomaly in the South China Sea can also move northeastward, which generates a northeast–southwest band of convection (phases 4 and 5). In phase 5, we can see that there is a convective anomaly in northern China, which actually propagates from the midlatitudes through the northern rim of the TP. It then moves eastward to join the northeastern part of the convection band, ultimately reaching the western North Pacific (phases 5–8). So, for northern China, it is possible that the QBWO from the midlatitudes plays a leading role and also contributes to parts of the northern TP. These QBWO signals from the midlatitudes are mainly reflected in the upper layer and

so may be the reason why they can be seen more clearly in the vertically integrated atmospheric apparent heat source $\langle Q_1 \rangle$ than in the summer GPGP precipitation. At 200 hPa, the composite life cycle of the QBWO in terms of relative velocity shows that there is a wave train in the midlatitudes that spreads southward to the northern TP, northern China, eastern China, and the eastern oceans of Japan (see Fig. S3 of the supplementary material). This is in accordance with the findings of Yang et al. (2010), in which the origin of the biweekly variability over the Yangtze River basin was found to be primarily attributable to an upper-level anticyclonic anomaly, embedded in the southeastward-propagating wave trains along the midlatitude westerly jet stream. Similar results were also documented by Ding and Wang (2007), who explored the intraseasonal link between the summer Eurasian wave train and Indian summer monsoon. Furthermore, the origin of the midlatitude wave train is

believed to be the northeastern Atlantic in the exit region of the Atlantic jet stream (Simmons et al. 1983; Ding and Wang 2007). For the TP, this propagation path most likely corresponds to the type-II propagation from the midlatitudes to the TP.

The southwest part of the northeast–southwest band of convection reaches the Indian subcontinent and weakens gradually in phases 6 and 7. It then propagates northward and eventually moves into the TP area in phase 8. From phases 4–6, we can see that, when the anomalous positive convection arrives in the TP, it seems to extend downstream to the East Asian region, corresponding to the significant westerly flow. It may then converge with the signals from the midlatitudes in phases 7 and 8, which is consistent with the obvious eastward feature detected in Fig. 11d. Throughout the whole life cycle of the QBWO, the anomalous anticyclonic (cyclonic) circulation and anomalous negative (positive) convection are closely in phase. Additionally, the life cycle of the QBWO as denoted by GPCP precipitation presents a similar evolution (see Fig. S4 of the supplementary material), and the monsoon precipitation shows obvious positive anomalies when the QBWO is in active phase, which is in good agreement with the results presented in Fig. 9c. The whole process of the propagation and evolution of the QBWO in the Asian monsoon region is basically similar to that in the first kind of propagation over the TP (Fig. 4), demonstrating that this result mainly reflects the evolution of the first category and the most pronounced QBWO over the TP. For the other two types of QBWO signal over the TP, there will be some different evolution processes, but these are not discussed in detail in the present paper.

6. Summary and discussion

In this paper, by using station observations, GPCP precipitation data, and ERA-Interim during 1979–2011, we investigate the dominant ISV mode over the TP in terms of periodicity, spatial structure, and propagation characteristics and its link with monsoon activity in adjacent regions. The main conclusions can be summarized as follows:

- 1) The dominant ISV over the TP is the QBWO, which exists in precipitation, diabatic heating, and circulation from the surface to the upper troposphere. In some years, 20–60-day oscillation is also significant. The meridional propagation of the QBWO over the TP can be categorized into three types: propagating from the low latitudes to the TP; propagating from the high latitudes to the TP; and generated locally over the TP and then propagating to the high and/or low latitudes. Statistical analysis indicates that the

first kind of propagation of the QBWO accounts for most cases.

- 2) Diagnosis of the northward-propagating QBWO suggests that barotropic vorticity induced by vertical easterly shear may explain the northward shift of convection before it arrives at the TP. The farther northward movement of the convection toward the TP, however, could be primarily due to the much stronger horizontal moisture advection in the lower troposphere, which destabilizes the lower atmosphere ahead of the convection by enhancing the equivalent potential temperature, shifting the specific humidity center to the north of the MCC, and thus resulting in a farther northward movement of the convection.
- 3) The QBWO usually strengthens markedly before monsoon onset, implying that the QBWO may play an important role in triggering TPSM onset. Moreover, the propagation of the QBWO is closely related to TPSM evolution. In the active (break) phase of the QBWO in the TPSM, above (below)-normal precipitation occurs in the CETP and below (above)-normal precipitation occurs in northern India, northern China, and eastern China.
- 4) For the Asian monsoon system, the QBWO links the ISV of three monsoon subsystems (TPSM, EASM, and SASM). Typically, the QBWO over the TP may originate from the equatorial western Pacific and propagates northwestward first to the Bay of Bengal and northern India and then northward to the southeastern TP. The QBWO in the TP can also move eastward downstream to the East Asian region. The whole process shows a clockwise pathway. Consequently, the monsoon precipitation in each region shows an obvious positive anomaly when the active phase of the QBWO arrives.

The origins of the three different types of QBWO over the TP could be different and there is not a simple unified framework to clearly illuminate the origins and propagation characteristics of these three types of oscillation. For the type-I oscillation, its origin is over the equatorial western Pacific and the behavior of such westward propagation seems to be associated with equatorial Rossby wave. The type-II propagation, meanwhile, shows a southeastward propagation, and is likely to be an extratropical Rossby wave (Kikuchi and Wang 2009). Its origin may be in the northwest of Europe, and it propagates to the TP region along the midlatitude westerly jet (Ding and Wang 2007). Whether the TP is a source and/or a sink of ISV has been a controversial issue for a long time. According to the above analyses, the QBWO over the TP is a

complicated mix of the three types. The oscillation from northern India is usually strengthened over the TP area (Fig. 3). We still cannot answer the question of whether this oscillation is the original one from India, one regenerated in the TP local area, or a combination of the two. At present, overall agreement is yet to be reached on a given theory about the mechanisms and nature of the ISV's (10–20- and 30–60-day oscillation) generation and propagation (Waliser 2006).

In the Asian monsoon region, the QBWO is featured by apparent meridional propagation, and the TP exhibits obviously local QBWO signals (Fujinami and Yasunari 2004; Liu et al. 2007). In some years, the 30–60-day oscillation over the TP is also statistically significant and dominates the ISV over the TP. Therefore, the common characteristics and source of the 30–60-day oscillation over the TP also deserve attention in future work.

Acknowledgments. We thank Dr. Jiang X. A. for his constructive comments. This work was supported jointly by the National Natural Science Foundation Item of China (Grant 41175070), the Major Program of the National Natural Science Foundation of China (Grant 91337216), and the Special Fund for Tibetan Plateau research (Grant GYHY201406001).

REFERENCES

- Abhik, S., M. Halder, P. Mukhopadhyay, X. Jiang, and B. N. Goswami, 2013: A possible new mechanism for northward propagation of boreal summer intraseasonal oscillations based on TRMM and MERRA reanalysis. *Climate Dyn.*, **40**, 1611–1624, doi:10.1007/s00382-012-1425-x.
- Anderson, J. R., and R. D. Rosen, 1983: The latitude-height structure of 40–50 day variation in atmospheric angular momentum. *J. Atmos. Sci.*, **40**, 1584–1584, doi:10.1175/1520-0469(1983)040<1584:TLHSOD>2.0.CO;2.
- Annamalai, H., and J. M. Slingo, 2001: Active/break cycles: Diagnosis of the intraseasonal variability of the Asian summer monsoon. *Climate Dyn.*, **18**, 85–102, doi:10.1007/s003820100161.
- Bai, H. Z., J. N. Xie, and D. L. Li, 2001: The principal feature of Qinghai Xizang Plateau monsoon variation in 40 years (in Chinese). *Plateau Meteor.*, **20**, 22–27.
- Bao, X. H., and F. Q. Zhang, 2013: Evaluation of NCEP–CFR, NCEP–NCAR, ERA–Interim, and ERA–40 reanalysis datasets against independent sounding observations over the Tibetan Plateau. *J. Climate*, **26**, 206–214, doi:10.1175/JCLI-D-12-00056.1.
- Blackmon, M. L., 1976: A climatological spectral study of the 500 hPa geopotential height of the Northern Hemisphere. *J. Atmos. Sci.*, **33**, 1607–1623, doi:10.1175/1520-0469(1976)033<1607:ACSSOT>2.0.CO;2.
- , J. M. Wallace, N. C. Lau, and S. L. Mullen, 1977: An observational study of the Northern Hemisphere wintertime circulation. *J. Atmos. Sci.*, **34**, 1040–1053, doi:10.1175/1520-0469(1977)034<1040:AOSOTN>2.0.CO;2.
- , Y. H. Lee, and J. M. Wallace, 1984a: Horizontal structure of 500mb height fluctuations with long, intermediate and short time scales. *J. Atmos. Sci.*, **41**, 961–979, doi:10.1175/1520-0469(1984)041<0961:HSOMHF>2.0.CO;2.
- , —, —, and H. H. Hsu, 1984b: Time variation of 500mb height fluctuation with long, intermediate and short time scales as deduced from lag-correlation statistic. *J. Atmos. Sci.*, **41**, 981–991, doi:10.1175/1520-0469(1984)041<0981:TVOMHF>2.0.CO;2.
- Chan, J. C. L., W. X. Ai, and J. J. Xu, 2002: Mechanisms responsible for the maintenance of the 1998 South China Sea summer monsoon. *J. Meteor. Soc. Japan*, **80**, 1103–1113, doi:10.2151/jmsj.80.1103.
- Chatterjee, P., and B. N. Goswami, 2004: Structure, genesis and scale selection of the tropical quasi-biweekly mode. *Quart. J. Roy. Meteor. Soc.*, **130**, 1171–1194, doi:10.1256/qj.03.133.
- Chen, T. C., and J. M. Chen, 1993: The 10–20 day mode of the 1979 Indian monsoon: Its relation with time variation of monsoon rainfall. *Mon. Wea. Rev.*, **121**, 2465–2482, doi:10.1175/1520-0493(1993)121<2465:TDMOTI>2.0.CO;2.
- , and J. R. Chen, 1995: An observational study of the South China Sea monsoon during the 1979 summer: Onset and life cycle. *Mon. Wea. Rev.*, **123**, 2295–2318, doi:10.1175/1520-0493(1995)123<2295:AOSOTS>2.0.CO;2.
- , M. C. Yen, and S. P. Weng, 2000: Interaction between the summer monsoon in East Asia and the South China Sea: Intraseasonal monsoon modes. *J. Atmos. Sci.*, **57**, 1373–1392, doi:10.1175/1520-0469(2000)057<1373:IBTSMI>2.0.CO;2.
- Dee, D. P., and Coauthors, 2011: The ERA–Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, doi:10.1002/qj.828.
- Ding, Q. H., and B. Wang, 2007: Intraseasonal teleconnection between the Eurasian wave train and Indian monsoon. *J. Climate*, **20**, 3751–3767, doi:10.1175/JCLI4221.1.
- Duan, A. M., and G. X. Wu, 2005: Role of the Tibetan Plateau thermal forcing in the summer climate patterns over subtropical Asia. *Climate Dyn.*, **24**, 793–807, doi:10.1007/s00382-004-0488-8.
- , and —, 2008: Weakening trend in the atmospheric heat source over the Tibetan Plateau during recent decades. Part I: Observations. *J. Climate*, **21**, 3149–3164, doi:10.1175/2007JCLI1912.1.
- , —, and X. Y. Liang, 2008: Influence of the Tibetan Plateau on the summer climate patterns over Asia in the IAP/LASG SAMIL model. *Adv. Atmos. Sci.*, **25**, 518–528, doi:10.1007/s00376-008-0518-2.
- , J. Hu, and Z. X. Xiao, 2013: The Tibetan Plateau summer monsoon in the CMIP5 simulations. *J. Climate*, **26**, 7747–7766, doi:10.1175/JCLI-D-12-00685.1.
- Flohn, H., 1957: Large-scale aspects of the summer monsoon in South and East Asia. *J. Meteor. Soc. Japan*, **35**, 180–186.
- , 1960: Recent investigations on the mechanism of the “summer monsoon” of southern and eastern Asia. *Monsoons of the World*, N. Delhi, Ed., Hindu Union Press, 75–88.
- Fu, X. H., B. Wang, and L. Tao, 2006: Satellite data reveal the 3-D moisture structure of tropical intraseasonal oscillation and its coupling with underlying ocean. *Geophys. Res. Lett.*, **33**, L03705, doi:10.1029/2005GL025074.
- Fujinami, H., and T. Yasunari, 2004: Submonthly variability of convection and circulation over and around the Tibetan Plateau during the boreal summer. *J. Meteor. Soc. Japan*, **82**, 1545–1564, doi:10.2151/jmsj.82.1545.
- , and —, 2009: The effects of midlatitude waves over and around the Tibetan Plateau on submonthly variability of the East Asian summer monsoon. *Mon. Wea. Rev.*, **137**, 2286–2304, doi:10.1175/2009MWR2826.1.
- Fukutomi, Y., and T. Yasunari, 1999: 10–25 day intraseasonal variations of convection and circulation over East Asia and

- western North Pacific during early summer. *J. Meteor. Soc.*, **77**, 753–769.
- Goswami, B. N., R. S. Aiyyamohan, P. K. Xavier, and D. Sengupta, 2003: Clustering of synoptic activity by Indian summer monsoon intraseasonal oscillations. *Geophys. Res. Lett.*, **30**, 1431, doi:[10.1029/2002GL016734](https://doi.org/10.1029/2002GL016734).
- Hahn, D. G., and S. Manabe, 1975: The role of mountains in the South Asian monsoon circulation. *J. Atmos. Sci.*, **32**, 1515–1541, doi:[10.1175/1520-0469\(1975\)032<1515:TROMIT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1975)032<1515:TROMIT>2.0.CO;2).
- He, Y. H., C. Y. Li, Q. Jin, and X. L. Wang, 2006: Relationship between the low frequency oscillation of TBB in summer of Qinghai-Xizang Plateau and drought/flood in central China (in Chinese). *Plateau Meteor.*, **25**, 658–664.
- Hsu, H.-H., C.-H. Weng, and C.-H. Wu, 2004: Contrasting characteristic between the northward and eastward propagation of the intraseasonal oscillation during the boreal summer. *J. Climate*, **17**, 727–743, doi:[10.1175/1520-0442\(2004\)017<0727:CCBTNA>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<0727:CCBTNA>2.0.CO;2).
- Huffman, G. J., and Coauthors, 1997: The Global Precipitation Climatology Project (GPCP) combined precipitation datasets. *Bull. Amer. Meteor. Soc.*, **78**, 5–20, doi:[10.1175/1520-0477\(1997\)078<0005:TGPCPG>2.0.CO;2](https://doi.org/10.1175/1520-0477(1997)078<0005:TGPCPG>2.0.CO;2).
- Jeong, J. H., B. M. Kim, C. H. Ho, and Y. H. Noh, 2008: Systematic variation in wintertime precipitation in East Asia by MJO-induced extratropical vertical motion. *J. Climate*, **21**, 788–801, doi:[10.1175/2007JCLI1801.1](https://doi.org/10.1175/2007JCLI1801.1).
- Jia, X. L., and S. Yang, 2013: Impact of the quasi-biweekly oscillation over the western North Pacific on East Asian subtropical monsoon during early summer. *J. Geophys. Res.*, **118**, 4421–4434, doi:[10.1002/jgrd.50422](https://doi.org/10.1002/jgrd.50422).
- Jiang, X. A., and N. C. Lau, 2008: Intraseasonal teleconnection between North American and western North Pacific monsoons with 20-day time scale. *J. Climate*, **21**, 2664–2679, doi:[10.1175/2007JCLI2024.1](https://doi.org/10.1175/2007JCLI2024.1).
- , T. Li, and B. Wang, 2004: Structures and mechanisms of the northward propagating boreal summer intraseasonal oscillation. *J. Climate*, **17**, 1022–1039, doi:[10.1175/1520-0442\(2004\)017<1022:SAMOTN>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<1022:SAMOTN>2.0.CO;2).
- Kikuchi, K., and B. Wang, 2009: Global perspective of the quasi-biweekly oscillation. *J. Climate*, **22**, 1340–1359, doi:[10.1175/2008JCLI2368.1](https://doi.org/10.1175/2008JCLI2368.1).
- , —, and Y. Kajikawa, 2012: Bimodal representation of the tropical intraseasonal oscillation. *Climate Dyn.*, **38**, 1989–2000, doi:[10.1007/s00382-011-1159-1](https://doi.org/10.1007/s00382-011-1159-1).
- Ko, K. C., and H. H. Hsu, 2006: Sub-monthly circulation features associated with tropical cyclone tracks over the East Asian monsoon area during July–August season. *J. Meteor. Soc. Japan*, **84**, 871–889, doi:[10.2151/jmsj.84.871](https://doi.org/10.2151/jmsj.84.871).
- Krishnamurti, T. N., 1973: Tibetan high and upper tropical circulation during the northern summer. *Bull. Amer. Meteor. Soc.*, **54**, 1234–1239.
- , and H. N. Bhalme, 1976: Oscillations of a monsoon system. Part I: Observational aspects. *J. Atmos. Sci.*, **33**, 1937–1954, doi:[10.1175/1520-0469\(1976\)033<1937:OOAMSP>2.0.CO;2](https://doi.org/10.1175/1520-0469(1976)033<1937:OOAMSP>2.0.CO;2).
- , and P. Ardanuy, 1980: The 10–20-day westward propagating mode and breaks in the monsoons. *Tellus*, **32**, 15–26, doi:[10.1111/j.2153-3490.1980.tb01717.x](https://doi.org/10.1111/j.2153-3490.1980.tb01717.x).
- , and D. Subrahmanyam, 1982: The 30–50 day mode at 850 mb during MONEX. *J. Atmos. Sci.*, **39**, 2088–2095, doi:[10.1175/1520-0469\(1982\)039<2088:TDMAMD>2.0.CO;2](https://doi.org/10.1175/1520-0469(1982)039<2088:TDMAMD>2.0.CO;2).
- , and S. Gadgil, 1985: On the structure of the 30 to 50 day mode over the global during FGGE. *Tellus*, **37A**, 336–360, doi:[10.1111/j.1600-0870.1985.tb00432.x](https://doi.org/10.1111/j.1600-0870.1985.tb00432.x).
- Lau, K.-M., and S. Yang, 1996: Seasonal variation, abrupt transition, and intraseasonal variability associated with the Asian summer monsoon in the GLAGCM. *J. Climate*, **9**, 965–985, doi:[10.1175/1520-0442\(1996\)009<0965:SVATAI>2.0.CO;2](https://doi.org/10.1175/1520-0442(1996)009<0965:SVATAI>2.0.CO;2).
- Li, F., and A. M. Duan, 2011: Variation of the Tibetan Plateau summer monsoon and its effect on the rainfall and the circulation in Asia—A case study in 2008 (in Chinese). *Chin. J. Atmos. Sci.*, **35**, 694–706.
- Li, Y. K., Z. B. Sun, and J. J. Zhang, 1991: Relationship between the 30–60 day oscillations in and around the Qinghai-Xizang plateau during the summer half year and the atmospheric circulation (in Chinese). *J. Nanjing Inst. Meteor.*, **14**, 483–488.
- Liang, X. Y., Y. M. Liu, and G. X. Wu, 2005: The impact of Qinghai Xizang Plateau uplift on Asian general circulation in spring and summer (in Chinese). *Plateau Meteor.*, **24**, 837–845.
- Lin, R. P., T. J. Zhou, and Y. Qian, 2014: Evaluation of global monsoon precipitation changes based on five reanalysis datasets. *J. Climate*, **27**, 1271–1289, doi:[10.1175/JCLI-D-13-00215.1](https://doi.org/10.1175/JCLI-D-13-00215.1).
- Liu, F. M., and H. Lin, 1991: Relationships between the atmospheric low-frequency oscillation and the variation of the South Asian high (in Chinese). *Plateau Meteor.*, **10**, 61–69.
- Liu, Y. M., B. J. Hoskins, and M. Blackburn, 2007: Impact of Tibetan orography oscillation and heating on the summer flow over Asia. *J. Meteor. Soc. Japan*, **85**, 1–19, doi:[10.2151/jmsj.85.1](https://doi.org/10.2151/jmsj.85.1).
- Liu, Y. Y., and Y. H. Ding, 2008: Analysis and numerical simulation of the teleconnection between Indian summer monsoon and precipitation in north China. *Acta Meteor. Sin.*, **66**, 789–799.
- Luo, H. B., and M. Yanai, 1983: The large-scale circulation and heat sources over Tibetan Plateau and surrounding areas during the early summer of 1979. Part I: Precipitation and kinematic analyses. *Mon. Wea. Rev.*, **111**, 922–944, doi:[10.1175/1520-0493\(1983\)111<0922:TLSCAH>2.0.CO;2](https://doi.org/10.1175/1520-0493(1983)111<0922:TLSCAH>2.0.CO;2).
- , and —, 1984: The large-scale circulation and heat sources over Tibetan Plateau and surrounding areas during the early summer of 1979. Part II: Heat and moisture budgets. *Mon. Wea. Rev.*, **112**, 966–989, doi:[10.1175/1520-0493\(1984\)112<0966:TLSCAH>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<0966:TLSCAH>2.0.CO;2).
- Ma, L., T. Zhang, O. W. Frauenfeld, B. Ye, D. Yang, and D. Qin, 2009: Evaluation of precipitation from the ERA-40, NCEP-1, and NCEP-2 reanalyses and CMAP-1, CMAP-2, and GPCP-2 with ground-based measurements in China. *J. Geophys. Res.*, **114**, D09105, doi:[10.1029/2008JD011178](https://doi.org/10.1029/2008JD011178).
- Madden, R. A., and P. R. Julian, 1971: Detection of a 40–50 day oscillation in the zonal wind in the tropical Pacific. *J. Atmos. Sci.*, **28**, 702–708, doi:[10.1175/1520-0469\(1971\)028<0702:DOADOI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1971)028<0702:DOADOI>2.0.CO;2).
- , and —, 1972: Description of global-scale circulation cells in the tropics with a 40–50-day period. *J. Atmos. Sci.*, **29**, 1109–1123, doi:[10.1175/1520-0469\(1972\)029<1109:DOGSCC>2.0.CO;2](https://doi.org/10.1175/1520-0469(1972)029<1109:DOGSCC>2.0.CO;2).
- Mao, J. Y., and G. X. Wu, 2006: Intraseasonal variations of the Yangtze rainfall and its related atmospheric circulation features during the 1991 summer. *Climate Dyn.*, **27**, 815–830, doi:[10.1007/s00382-006-0164-2](https://doi.org/10.1007/s00382-006-0164-2).
- , Z. Sun, and G. X. Wu, 2010: 20–50-day oscillation of summer Yangtze rainfall in response to intraseasonal variations in the subtropical high over the western North Pacific and South China Sea. *Climate Dyn.*, **34**, 747–761, doi:[10.1007/s00382-009-0628-2](https://doi.org/10.1007/s00382-009-0628-2).
- Murakami, M., 1975: Cloudiness fluctuations during summer monsoon. *J. Meteor. Soc. Japan*, **54**, 175–181.
- Murakami, T., and M. Frydrych, 1974: On the preferred period of upper wind fluctuations during the summer monsoon. *J. Atmos.*

- Sci.*, **31**, 1549–1555, doi:[10.1175/1520-0469\(1974\)031<1549:OTPPOU>2.0.CO;2](https://doi.org/10.1175/1520-0469(1974)031<1549:OTPPOU>2.0.CO;2).
- Nitta, T., 1983: Observational study of heat sources over the eastern Tibetan Plateau during the summer monsoon. *J. Meteor. Soc. Japan*, **61**, 590–605.
- Numaguti, A., 1995: Characteristics of 4–20-day-period disturbances observed in the equatorial Pacific during the TOGACOARE IOP. *J. Meteor. Soc. Japan*, **73**, 353–377.
- Peng, R., and B. Y. Wu, 1995: The phase propagations and geographical features of the 10–20 day oscillation in summer 1982/1983 (in Chinese). *J. Appl. Meteor. Sci.*, **6**, 206–212.
- Prasanna, V., and H. Annamalai, 2012: Moist dynamics of extended monsoon breaks over South Asia. *J. Climate*, **25**, 3810–3831, doi:[10.1175/JCLI-D-11-00459.1](https://doi.org/10.1175/JCLI-D-11-00459.1).
- Simmons, A. J., J. M. Wallace, and G. W. Branstator, 1983: Barotropic wave propagation and instability, and atmospheric teleconnection patterns. *J. Atmos. Sci.*, **40**, 1363–1392, doi:[10.1175/1520-0469\(1983\)040<1363:BWPAA>2.0.CO;2](https://doi.org/10.1175/1520-0469(1983)040<1363:BWPAA>2.0.CO;2).
- Tang, M. C., 1995: Discussion on interdecadal oscillation of the plateau monsoon and its cause (in Chinese). *Sci. Meteor. Sin.*, **15**, 64–68.
- , and E. R. Reiter, 1984: Plateau monsoons of the Northern Hemisphere: A comparison between North America and Tibet. *Mon. Wea. Rev.*, **112**, 617–637, doi:[10.1175/1520-0493\(1984\)112<0617:PMOTNH>2.0.CO;2](https://doi.org/10.1175/1520-0493(1984)112<0617:PMOTNH>2.0.CO;2).
- Torrence, C., and G. P. Compo, 1998: A practical guide to wavelet analysis. *Bull. Amer. Meteor. Soc.*, **79**, 61–78, doi:[10.1175/1520-0477\(1998\)079<0061:APGTWA>2.0.CO;2](https://doi.org/10.1175/1520-0477(1998)079<0061:APGTWA>2.0.CO;2).
- Waliser, D. E., 2006: Intraseasonal variability. *The Asian Monsoon*, B. Wang, Ed., Springer Praxis, 203–257.
- Wang, A., and X. Zeng, 2012: Evaluation of multi reanalysis products with in situ observations over the Tibetan Plateau. *J. Geophys. Res.*, **117**, D05102, doi:[10.1029/2011JD016553](https://doi.org/10.1029/2011JD016553).
- Wang, B., and X. Xie, 1997: A model for the boreal summer intraseasonal oscillation. *J. Atmos. Sci.*, **54**, 72–86, doi:[10.1175/1520-0469\(1997\)054<0072:AMFTBS>2.0.CO;2](https://doi.org/10.1175/1520-0469(1997)054<0072:AMFTBS>2.0.CO;2).
- Wang, C. H., Y. Cui, and S. L. Jin, 2011: Oscillation propagation features of the atmosphere around the Qinghai-Xizang during the spring season of typical strong and weak monsoon years. *Sci. China*, **54**, 305–314, doi:[10.1007/s11430-010-4113-x](https://doi.org/10.1007/s11430-010-4113-x).
- Wang, M. R., S. W. Zhou, and A. M. Duan, 2012: Trend in the atmospheric heat source over the central and eastern Tibetan Plateau during recent decades: Comparison of observations and reanalysis data. *Chin. Sci. Bull.*, **57**, 548–557, doi:[10.1007/s11434-011-4838-8](https://doi.org/10.1007/s11434-011-4838-8).
- Weare, B. C., and J. S. Nasstrom, 1982: Examples of extended empirical orthogonal function analysis. *Mon. Wea. Rev.*, **110**, 481–485, doi:[10.1175/1520-0493\(1982\)110<0481:EEOEOF>2.0.CO;2](https://doi.org/10.1175/1520-0493(1982)110<0481:EEOEOF>2.0.CO;2).
- Webster, P. J., 1983: Mechanisms of monsoon low-frequency variability: Surface hydrological effect. *J. Atmos. Sci.*, **40**, 2110–2124, doi:[10.1175/1520-0469\(1983\)040<2110:MOMLFV>2.0.CO;2](https://doi.org/10.1175/1520-0469(1983)040<2110:MOMLFV>2.0.CO;2).
- Wen, M., S. Yang, W. Higgins, and R. Zhang, 2011: Characteristics of the dominant modes of atmospheric quasi-biweekly oscillation over tropical–subtropical Americas. *J. Climate*, **24**, 3956–3970, doi:[10.1175/2011JCLI3916.1](https://doi.org/10.1175/2011JCLI3916.1).
- Wu, G., and Y. Zhang, 1998: Tibetan Plateau forcing and the timing of the monsoon onset over South Asia and the South China Sea. *Mon. Wea. Rev.*, **126**, 913–927, doi:[10.1175/1520-0493\(1998\)126<0913:TPFATT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1998)126<0913:TPFATT>2.0.CO;2).
- , Y. Liu, B. He, Q. Bao, A. Duan, and F.-F. Jin, 2012: Thermal controls on the Asian summer monsoon. *Sci. Rep.*, **2**, 404, doi:[10.1038/srep00404](https://doi.org/10.1038/srep00404).
- Xie, A., Q. Ye, and L. X. Chen, 1989: The atmospheric oscillations over the Tibetan Plateau and surrounding areas as revealed from OLR data (in Chinese). *Acta Meteor. Sin.*, **47**, 272–278.
- Xie, P., and P. A. Arkin, 1997: Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates, and numerical model outputs. *Bull. Amer. Meteor. Soc.*, **78**, 2539–2558, doi:[10.1175/1520-0477\(1997\)078<2539:GPAYMA>2.0.CO;2](https://doi.org/10.1175/1520-0477(1997)078<2539:GPAYMA>2.0.CO;2).
- Yanai, M., 1961: A detailed analysis of typhoon formation. *J. Meteor. Soc. Japan*, **39**, 187–214.
- , S. Esbensen, and J. H. Chu, 1973: Determination of bulk properties of tropical cloud clusters from large-scale heat and moisture budget. *J. Atmos. Sci.*, **30**, 611–627, doi:[10.1175/1520-0469\(1973\)030<0611:DOBPOT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1973)030<0611:DOBPOT>2.0.CO;2).
- Yang, J., B. Wang, B. Wang, and L. J. Li, 2009: The East Asia–western North Pacific boreal summer intraseasonal oscillation simulated in GAMIL 1.1.1. *Adv. Atmos. Sci.*, **26**, 480–492, doi:[10.1007/s00376-009-0480-7](https://doi.org/10.1007/s00376-009-0480-7).
- , —, —, and Q. Bao, 2010: Biweekly and 21–30-day variations of the subtropical summer monsoon rainfall over the lower reach of the Yangtze River basin. *J. Climate*, **23**, 1146–1159, doi:[10.1175/2009JCLI3005.1](https://doi.org/10.1175/2009JCLI3005.1).
- , Q. Bao, B. Wang, D. Y. Gong, H. Z. He, and M. N. Gao, 2014: Distinct quasi-biweekly features of the subtropical East Asian monsoon during early and late summers. *Climate Dyn.*, **42**, 1469–1486, doi:[10.1007/s00382-013-1728-6](https://doi.org/10.1007/s00382-013-1728-6).
- Yang, Q. M., 2009: The 20–30-day oscillation of the global circulation and heavy precipitation over the lower reaches of the Yangtze River valley. *Sci. China*, **52**, 1485–1501, doi:[10.1007/s11430-009-0156-2](https://doi.org/10.1007/s11430-009-0156-2).
- Yasunari, T., 1979: Cloudiness fluctuation associated with the Northern Hemisphere summer monsoon. *J. Meteor. Soc. Japan*, **57**, 227–241.
- , 1980: A quasi-stationary appearance of 30–40 day period in the cloudiness fluctuations during the summer monsoon over India. *J. Meteor. Soc. Japan*, **58**, 225–229.
- Zhang, P. F., G. P. Li, X. H. Fu, Y. M. Liu, and L. F. Li, 2014: Clustering of Tibetan Plateau vortices by 10–30-day intraseasonal oscillation. *Mon. Wea. Rev.*, **142**, 290–300, doi:[10.1175/MWR-D-13-00137.1](https://doi.org/10.1175/MWR-D-13-00137.1).
- Zhang, Y., L. X. Chen, and J. H. He, 2009: A study of the characteristics of the low-frequency circulation over the Tibetan Plateau and its association with precipitation in the Yangtze River valley in 1998. *Acta Meteor. Sin.*, **23**, 175–190.
- Zhou, B., J. H. He, and H. M. Xu, 2000: LFO characteristics of meteorological elements over Tibetan Plateau and the relations with regional summer rainfall (in Chinese). *J. Nanjing Inst. Meteor.*, **23**, 93–100.
- Zhou, W., and C. L. Chan, 2005: Intraseasonal oscillations and the South China Sea summer monsoon onset. *Int. J. Climatol.*, **25**, 1585–1609, doi:[10.1002/joc.1209](https://doi.org/10.1002/joc.1209).