Dynamics of Widespread Extreme Precipitation Events and the Associated Large-Scale Environment Using AMeDAS and JRA-55 Data

RYOSUKE SHIBUYA,^a YUKARI TAKAYABU,^a AND HIROTAKA KAMAHORI^a

^a *Atmosphere and Ocean Research Institute, The University of Tokyo, Kashiwa, Japan*

(Manuscript received 23 January 2021, in final form 29 July 2021)

ABSTRACT: This study examines disastrous historical precipitation cases that generate extreme precipitation simultaneously over a wide area in Japan (as in July 2018), defined as widespread extreme precipitation events. A statistically significant large-scale environment conducive for widespread extreme precipitation events over western Japan is investigated based on composite analysis. During a widespread precipitation event, a zonally elongated positive anomaly of the column-integrated water vapor extends from East China to western Japan. In the lower troposphere, a dipole of a geopotential height anomaly exists with positive and negative values at the east and west of the precipitation area, respectively. It is found that the negative geopotential anomaly is enhanced over East China at 2 days before the event and moves toward the precipitating area mainly due to the potential vorticity (PV) production term by diabatic heating, analogous to a diabatic Rossby wave. The temporal evolution of the dynamical forced vertical velocity is well in phase with that the PV production term, suggesting the importance of the coupling between the dynamical forced motion and diabatic heating. This result provides a physical understanding of the reason why both the background moisture and the baroclinicity are essential in the composited atmospheric fields and another view to the importance of the feedback parameter between the dynamical motion and diabatic heating.

KEYWORDS: Atmospheric circulation; Vertical motion; Extratropical cyclones; Extreme events; Precipitation

1. Introduction

During early summer in East Asia, a climatological frontal zone with a strong equivalent potential temperature gradient (θ_e) extends from the southern part of China to Japan (Ninomiya and Murakami 1987; Sampe and Xie 2010). This quasi-stationary front is called the mei-yu front (Chinese), the changma front (Korean), and the baiu front (Japanese). Ninomiya (1984) suggested that the mei-yu-baiu front is a subtropical front characterized by a deep, moist neutral layer and a sharp θ_e gradient. The rainband associated with the baiu front sometimes produces a torrential rainfall, which causes disastrous flooding. For example, the heavy rain event of July 2018 (e.g., Shimpō et al. 2019) associated with the baiu front generated considerably large amounts of precipitation, especially over western Japan and the Tokai region. During this event, many in situ ground-based observational stations simultaneously measured their highest recorded precipitation amount, and this condition led to widespread disasters.

By examining the historical records from 1979 to 2020, this study reveals that the heavy rainfall event of July 2018 is not the only case in which extreme precipitation was generated over a

wide area all at once, although the event of July 2018 is found to be the most extreme occurrence. In section 2, we define this type of rain event as a “widespread extreme precipitation” event. Since the widespread rainfall in these events is likely related to synoptic- or large-scale dynamics (Horinouchi 2014; Yokoyama et al. 2017; Horinouchi and Hayashi 2017) with a different scale from that of a single/multiple mesoscale system (Ninomiya and Akiyama 1992; Kato 2020), it is expected that such widespread extreme precipitation might accompany coherent atmospheric circulation anomalies over East Asia. The purpose of this study is to elucidate the characteristics of widespread extreme precipitation events and their associated atmospheric patterns from observations and reanalysis data.

Recent studies have extensively examined the atmospheric circulation anomalies associated with extreme precipitation over East China (Zhou and Yu 2005; Wei et al. 2014; Chen and Zhai 2016; Oh et al. 2018; Hu et al. 2019; Zhao et al. 2020) and over Japan (Hirota et al. 2016; Hamada and Takayabu 2018; Tsuji et al. 2020; Sugimoto 2020; Harada et al. 2020). It was reported that extreme precipitation cases over Japan were related to several key mechanisms: anomalous moisture transport, the westerly jet, the enhancement of upper-level and low-level vortices, and Rossby wave propagation associated with upstream blocking. Kamae et al. (2017) showed that heavy rainfall events over Japan during warm season are often related to the narrow and intense moisture transport as a form of the atmospheric river, such as those in the United States (Neiman et al. 2008; Demirdjian et al. 2020; Moore et al. 2020, 2021).

However, because a detailed dynamical analysis was only conducted as a case study of individual extreme precipitation (Shimpō et al. 2019; Sekizawa et al. 2019; Yokoyama et al. 2020), it is still not clear what mechanisms significantly contribute to the occurrence of extreme precipitation over Japan.

Denotes content that is immediately available upon publication as open access.

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-21-0064.s1>.

Corresponding author: Ryosuke Shibuya, shibuyar@ori.u-tokyo.ac.jp

DOI: 10.1175/JCLI-D-21-0064.1

© 2021 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the [AMS Copyright Policy \(www.ametsoc.org/PUBSReuseLicenses\)](#).

FIG. 1. (a) Horizontal map of 99th-percentile thresholds of the 12-h-averaged precipitation amount on the available AMeDAS stations in June and July from 1979 to 2020. Extreme precipitation amount in the 1st and 42nd case of the widespread extreme precipitation cases recorded on (b) 1200 UTC 6 Jul 2018 and (c) 1200 UTC 30 Jun 1979.

For example, [Tochimoto and Kawano \(2017\)](#) suggested that the development process of a baiu-frontal depression is consistent with a diabatic Rossby wave ([Parker and Thorpe 1995](#)). Although [Boettcher and Wernli \(2013\)](#) showed that diabatic Rossby waves are frequently detected around Japan during the baiu season for the years of 2001–10, a fuller picture as to whether diabatic Rossby waves were related to the past extreme precipitation cases has been still missing.

One of the theoretical perspectives that links large-scale circulation and local precipitation is the omega equation in the quasigeostrophic (QG) framework ([Holton 2004](#); [Horinouchi 2014](#); [Nie and Sobel 2016](#); [Horinouchi and Hayashi 2017](#); [Yokoyama et al. 2020](#)). The main advantage of the QG omega equation is that contributions from dynamic forcing and diabatic heating can be linearly decomposed, which leads to quantitative discussions of the role of the large-scale dynamics. [Nie and Fan \(2019\)](#) introduced the “dynamic forcing–diabatic feedback” perspective for the development of extreme precipitation in which dynamically forced vertical motions act as a forcing, and convection responds as a feedback. This view could offer further profound insights into the mechanism of extreme precipitation over Japan in terms of large-scale dynamics, moisture, and diabatic heating associated with convection. The present article is organized as follows. The methodology and data used herein is described in [section 2](#). The results of the composite analysis are given in [section 3](#).

A discussion is presented in [section 4](#), and [section 5](#) summarizes the results and provides concluding remarks.

2. Data and methods

a. Datasets and event selection

A rain gauge observation network has been installed across Japan in 1977 as the Automated Meteorological Data Acquisition System (AMeDAS) managed by the Japan Meteorological Agency (JMA). The amount of precipitation is recorded at a 1-h temporal resolution at each AMeDAS station. In this study, we use rain gauge data collected at 691 AMeDAS stations that have been continuously operating from 1979 to 2019; in particular, 241 of these AMeDAS stations are located in western Japan. [Figure 1a](#) shows the top 99th percentile thresholds of the 12-h-averaged precipitation amount for the AMeDAS stations. The AMeDAS stations are widely and quite uniformly distributed over western Japan with relatively higher precipitation threshold over the Kyushu region.

In this study, an extreme precipitation event is defined as a top-99th-percentile precipitation event over a 12-h duration in June and July at each AMeDAS station. The peak time of an extreme precipitation event is defined as the time at which each AMeDAS rain gauge records the highest precipitation amount

TABLE 1. Dates of the widespread extreme precipitation. The first column denotes the rank of the widespread extreme precipitation, and the third column denotes the number of the stations that recorded the extreme precipitation within their historical precipitation amount.

Rank	Time and date	No. of stations	Rank	Time and date	No. of stations
1	1200 UTC 6 Jul 2018	182	24	1200 UTC 20 Jun 1983	61
2	0000 UTC 6 Jul 2018	158	25	0000 UTC 28 Jun 1985	59
3	1200 UTC 29 Jun 1999	144	26	0000 UTC 5 Jul 2018	58
4	1200 UTC 3 Jul 1995	139	27	1200 UTC 7 Jul 2020	57
5	0000 UTC 19 Jun 2001	116	27	0000 UTC 7 Jul 2020	57
6	1200 UTC 27 Jun 1979	115	27	1200 UTC 2 Jul 1993	57
7	1200 UTC 5 Jul 2018	113	30	1200 UTC 2 Jul 1993	56
8	0000 UTC 7 Jul 2018	105	30	0000 UTC 14 Jul 2010	56
9	0000 UTC 3 Jul 1995	103	30	0000 UTC 27 Jun 1985	56
10	1200 UTC 2 Jun 1988	92	33	1200 UTC 13 Jun 2020	55
10	1200 UTC 25 Jun 1985	92	33	1200 UTC 19 Jun 2013	55
12	1200 UTC 29 Jun 1979	84	33	0000 UTC 24 Jul 1982	55
13	0000 UTC 29 Jun 1979	83	36	1200 UTC 26 Jun 2013	54
14	1200 UTC 24 Jul 1982	81	36	0000 UTC 20 Jun 1983	54
15	1200 UTC 19 Jun 2001	82	38	0000 UTC 18 Jun 2020	53
16	1200 UTC 7 Jul 2018	77	38	1200 UTC 20 Jun 2018	53
17	0000 UTC 27 Jun 1979	72	40	0000 UTC 4 Jul 1991	52
18	1200 UTC 21 Jun 2012	70	41	0000 UTC 20 Jul 1996	51
19	0000 UTC 19 Jun 2013	67	42	1200 UTC 24 Jun 2012	49
19	0000 UTC 19 Jul 2006	67	42	1200 UTC 3 Jun 1988	49
21	0000 UTC 25 Jun 1985	66	42	1200 UTC 30 Jun 1979	49
22	1200 UTC 6 Jul 2020	65			
23	1200 UTC 2 Jul 1990	64			

during the 12-h extreme precipitation event. During the analysis, precipitation records within 6 h before and after peak times detected in the higher ranks of the extreme precipitation are removed from the gauge record. Then, the peak time is classified into either 0000–1200 or 1200–2400 UTC for the composite analysis. To distinguish extreme precipitation events from those directly caused by tropical cyclones (TCs), we exclude extreme precipitation events during which any AMeDAS station was located within 1000 km of a TC center, where the locations of TCs were derived from the International Best Track Archive for Climate Stewardship (IBTrACS) dataset (Knapp et al. 2010). This is because extreme precipitation events directly caused by TCs likely have different atmospheric patterns from those related to the baiu front. Please note that remote effects of TCs (e.g., Byun and Lee 2012) more than 1000 km apart from the AMeDAS stations are not eliminated in this analysis. However, it was confirmed that the atmospheric circulation pattern does not strongly depend on the existence of the remote effect of TCs, which is discussed in Text S1 in the online supplemental material in detail.

Next, the number of western Japan AMeDAS stations at which extreme precipitation occurred simultaneously on a given day at either 0000–1200 or 1200–2400 UTC between June and July from 1979 to 2020 is examined. In this study, the top 99th percentiles for the number of AMeDAS stations that experienced extreme precipitation events are defined as widespread extreme precipitation events. Table 1 summarizes the dates of these widespread extreme precipitation events, some of which were analyzed as case studies by the previous studies (Ninomiya et al. 1984; Murakami and Huang 1984; Ogura et al. 1985). Figures 1b and 1c show the 12-h-averaged precipitation during

the 1st and 42nd widespread extreme precipitation events that occurred at 1200 UTC 6 July 2018 and at 1200 UTC 30 June 1979, respectively. The precipitation amount is depicted only for the AMeDAS stations that experienced extreme precipitation during those events. For example, in the first widespread extreme precipitation event, 183 AMeDAS stations simultaneously recorded large amounts of precipitation exceeding their top 99th percentiles. All dates reported in Table 1 are used to obtain the coherent atmospheric circulation patterns associated with the widespread extreme precipitation. A composite analysis is conducted based on the Japanese 55-year Reanalysis (JRA-55) 6-hourly dataset at a $1.25^\circ \times 1.25^\circ$ spatial resolution (Kobayashi et al. 2015). In addition, the fifth-generation reanalysis implemented by ECMWF called ERA5 (Hersbach and Dee 2016) at a $0.25^\circ \times 0.25^\circ$ spatial resolution is also used for the analysis of the surface variable. Note that some of persistent heavy rainfall events are classified as an individual widespread extreme precipitation event as in Table 1 and thus the composite members are not completely statistically independent. Nevertheless, this leads to weighting patterns of such persistent heavy rainfall events on the composite results, because persistent heavy rainfall events could have large social and economic impacts due to the long duration and high intensity. It was confirmed that the main results in section 3 were not largely influenced by including any individual persistent heavy rainfall event as the event of July 2018.

b. The QG omega equation

In this study, the QG omega equation is used to quantify the contributions from large-scale dynamic forcing and diabatic

heating. The QG omega equation may be written as follows (e.g., [Holtan 2004](#), 164–168; the derivation is described in our online supplemental material):

$$\sigma \nabla^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial p^2} = -2 \nabla \cdot \mathbf{Q} + f_0 \beta \frac{\partial v_g}{\partial p} - \frac{R}{p} \nabla^2 H, \quad (1)$$

where $\sigma = -(RT/p) \partial_p \ln \theta$ denotes the dry static energy, p denotes pressure, T denotes temperature, ω denotes the vertical pressure velocity, H denotes the diabatic heating, v_g denotes the meridional geostrophic wind velocity, θ denotes the potential temperature, ∇ denotes an operator of the horizontal gradient, and f_0 and β denote the reference Coriolis parameter and its meridional gradient at 35°N, respectively. In this study, H is calculated as the sum of the convective heating rate and the large-scale condensation heating rate. The vector $\mathbf{Q}$ is the so-called Q vector, which describes the forcing of the vertical motion by its convergence:

$$\mathbf{Q} = \left(-\frac{R}{p} \frac{\partial \mathbf{u}_g}{\partial x} \cdot \nabla T, -\frac{R}{p} \frac{\partial \mathbf{u}_g}{\partial y} \cdot \nabla T \right), \quad (2)$$

where $\mathbf{u}_g$ denotes the horizontal component of the geostrophic wind velocity. Assuming that a disturbance has a wavenumber of $k = 2\pi/L_x$ ($l = 2\pi/L_y$) with a zonal (meridional) characteristic length L_x (L_y), the QG omega equation for a single wavenumber is written as

$$\frac{\partial^2 \omega}{\partial p^2} - \sigma \frac{k^2 + l^2}{f_0^2} \omega = -\frac{2 \nabla \cdot \mathbf{Q}}{f_0^2} + \frac{\beta}{f_0} \frac{\partial v_g}{\partial p} - \frac{R}{p f_0^2} \nabla^2 H. \quad (3)$$

The QG omega equation can be solved using the Green's function method ([Nie and Sobel 2016](#)). Let there be a Green's function $\hat{G}$ that satisfies

$$\partial_{pp} \hat{G} - \sigma \frac{k^2 + l^2}{f_0^2} \hat{G} = \delta(p - p'), \quad (4)$$

where δ denotes the Dirac delta function and σ is assumed to be constant. If we denote the vertical motion component corresponding to Q-vector convergence as ω_D , the diabatic heating as ω_H and the other term as ω_{res} , Eq. (1) means that the vertical motion can be decomposed due to its linearity as follows:

$$\omega = \omega_D + \omega_H + \omega_{\text{res}}, \quad (5)$$

where ω_D denotes the dynamically forced vertical motion, ω_H denotes the vertical motion balanced by diabatic heating, and ω_{res} denotes the residual component including the vertical motion caused by the β effect in Eq. (3). By solving the Green's function in Eq. (4), ω_D and ω_H are obtained by integrating Green's function over the whole pressure coordinate with the corresponding terms in Eq. (3) as

$$\omega_D(p) = \int_{p_s}^{p_t} \hat{G}(p, p') \left\{ -\frac{2 \nabla \cdot \mathbf{Q}(p')}{f_0^2} \right\} dp', \quad (6)$$

$$\omega_H(p) = \int_{p_s}^{p_t} \hat{G}(p, p') \left\{ -\frac{R}{p' f_0^2} \nabla^2 H(p') \right\} dp', \quad (7)$$

where p_s and p_t denote the pressures at the surface and top boundary, respectively. Please note that p' is a variable that

appears only within the integral. Under the Dirichlet condition of $\hat{G}(p_s) = \hat{G}(p_t) = 0$, the analytical solution of the Green's function in Eq. (4) is obtained as follows:

$$\hat{G}(p, p') = \frac{e^{-b(p+p')}}{2b(e^{2bp_t} - e^{2bp_s})} \times \begin{cases} e^{2b(p_s+p)} - e^{2b(p'+p)} + e^{2b(p_t+p')} - e^{2b(p_s+p_t)} & (p < p') \\ e^{2b(p_s+p')} - e^{2b(p'+p)} + e^{2b(p_t+p)} - e^{2b(p_s+p_t)} & (p \geq p') \end{cases}, \quad (8)$$

where $b^2 = \sigma(k^2 + l^2)/f_0^2$. Equations (6) and (7) are numerically calculated by using the Fourier transform in the horizontal directions and Green's function in the vertical direction as in [Horinouchi \(2014\)](#). The Fourier transform is performed using the entire longitudinal range in the zonal direction and from 10° to 80°N in the meridional direction. The tapering windows are applied from 10° to 25°N and from 65° to 80°N with a cosine function. In this study, we set $p_t = 100$ hPa and $p_s = 1025$ hPa by introducing an additional bottom level. The results confirm that the analytical Green's function with the above Dirichlet condition provides a better profile than that with the Neumann boundary condition of $\partial_p \hat{G}(p_s) = \partial_p \hat{G}(p_t) = 0$ (not shown).

3. Composite analysis of widespread extreme precipitation

a. Atmospheric circulation patterns of widespread extreme precipitation

In this section, we examine the composited anomalies of atmospheric variables associated with the widespread extreme precipitation. The time of the composited peak precipitation is denoted as day ∓ 0 , where a negative (positive) sign denotes a time prior to (after) the peak. The anomalies in this study are defined as deviations from the daily climatology calculated from 1979 to 2018. First, we show composites of a column-integrated water vapor (CWV') from 1000 to 100 hPa, surface mean sea level pressure and dewpoint temperature 2 m above the surface, geopotential height (Φ'), and horizontal wind velocity ($\mathbf{u}'$) at 850 and 250 hPa on day ∓ 0 in [Fig. 2](#). Note that since the map of the dewpoint temperature 2 m above the land is quite complicated, likely due to the fine topography in ERA5, only the values over the ocean are shown in this study. We choose the height level of 850 hPa for the composite analysis because the moisture transport in the zonal direction is maximized about at 850 hPa. A region of zonally elongated positive CWV' appears from East China to western Japan, while a dipole of Φ' at 850 hPa exists with positive and negative values at the east and west of the precipitation area, respectively. As a result, the anomaly of the northeastward horizontal wind velocity is enhanced over western Japan, leading to a large moisture flux from southwest of Japan. The surface front with stronger horizontal gradient of dewpoint temperature, denoted by yellow dots in [Fig. 2b](#), is found at the west of the local minimum of the sea level pressure (SLP). In [Fig. 2d](#), an upper-level trough exists over the northwest of the low-level cyclonic anomaly at 850 hPa, which is consistent with the case studies of extreme precipitation events ([Takemura et al. 2019](#);

FIG. 2. A horizontal map of the composite anomaly of (a) column-integrated water vapor (kg m^{-2}) and column-integrated moisture flux ($\text{kg m}^{-1} \text{s}^{-1}$; dark blue vector), (b) surface mean sea level pressure (hPa) and dewpoint temperature 2 m above the surface (K), and geopotential height (m) and horizontal wind components (m s^{-1} ; vectors) at (c) 850 and (d) 250 hPa with day ∓ 0 of the widespread extreme precipitation. The contour interval is 2 kg m^{-2} in (a), 2 K in (b), and 5 m in (c) and (d). The shaded area indicates that the anomalies are statistically significant at a 95% confidence level using the Student's t test, while the dotted area indicates that the anomalies are statistically significant at a 90% confidence level. Yellow dots in (b) represent grid points where the horizontal gradient of dewpoint temperature exceeds $3.0 \times 10^{-6} \text{ K m}^{-1}$.

Yokoyama et al. 2020). In addition, a positive Φ' value is observed in the upper troposphere to the east of Japan over the North Pacific subtropical high (NPSH), suggesting an equivalent barotropic structure of the positive Φ' .

Temporal evolutions of Φ' and $\mathbf{u}'$ at 250 and 850 hPa, SLP and dewpoint temperature at the surface, and CWV' from day -5 to day -1 are shown in Fig. 3. The positive Φ' of the NPSH persists from day -5 to day -1 , indicating that positive Φ' might be a precursor of widespread extreme precipitation. The persistent positive Φ' of the NPSH preceding the widespread extreme precipitation event might be attributable to incoming coherent Rossby wave packets, which is discussed in Text S2 in the online supplemental material. It is notable that,

on day -5 , cyclogenesis occurred at the south of Lake Baikal around 45°N , 100°E , which corresponds to the lee side of the Altai–Sayan massif. This region is known as the most frequent cyclogenesis region in East Asia and is referred to as the “Altai–Sayan cyclogenesis” region (Chen and Lazić 1990; Chen et al. 1991). Historically, previous studies about the lee cyclogenesis have focused on the Alpine lee cyclogenesis, with theoretical approaches based on a baroclinic instability modified by orography [De Marchi 1900; Bergeron 1928; Pettersen 1956; Speranza et al. 1985; Pierrehumbert 1985; Rotunno and Ferretti 2001; see also a recent review by Buzzi et al. (2020)]. On day -3 , the Altai–Sayan cyclone moves southeastward toward 40°N , 120°E and becomes slightly

FIG. 3. A time evolution of the horizontal maps of the composite anomaly of (top to bottom) geopotential height (m) and horizontal wind components (m s^{-1} ; vectors) at 250 and 850 hPa, sea level pressure (hPa) and dewpoint temperature 2 m above the surface (K), and column-integrated water vapor (kg m^{-2}) with days (left to right) -5, -3, and -1 of the widespread extreme precipitation. The contour interval is 8 m in the top row, 5 m in the second row, 2 K in the third row, and 2 kg m^{-2} in the bottom row. The shaded area indicates that the anomalies are statistically significant at a 95% confidence level using the Student's t test, while the dotted area indicates that the anomalies are statistically significant at a 90% confidence level. The magnitude of the unit vector is drawn below the color bar. Yellow dots in the third row represent grid points where the horizontal gradient of dewpoint temperature exceeds $3.0 \times 10^{-6} \text{ K m}^{-1}$.

FIG. 4. A time evolution of the horizontal maps of the Q-vector convergence ($\text{m s}^{-1} \text{kg}^{-1}$; shading), zonal wind component (m s^{-1} ; blue contours), temperature (K; dashed contours) and Q vector (vectors) at 500 hPa with days (left to right) -2 , -1 , and 0 of the widespread extreme precipitation. The contour interval is 1 K. The magnitude of the unit vector is drawn below the color bar.

weakened. On day -1 , the upper-level cyclonic anomaly almost stagnates over the Korean Peninsula, while a new lower-level cyclonic anomaly is enhanced at the south of the propagating Altai–Sayan cyclone over the East China. It should be noted that, reviewing the individual widespread extreme precipitation events, some of the composited low-level cyclones on day -1 are found to be generated over the eastern edge of the Tibetan Plateau without the existence of the Altai–Sayan cyclone and to move toward the East China (Sugimoto 2020; Lee et al. 2020). It should be noted that a negative Φ' at 850 hPa around Taiwan on day -5 is due to a composite of tropical cyclones, which are found in approximately 60% of all widespread extreme precipitation cases.

A positive moisture anomaly starts to be enhanced over East China approximately three days before the widespread precipitation event, leading to elongated positive q' on day 0 . The eastward extension of positive q' toward Japan on day -1 was associated with the enhancement of the cyclonic anomaly in the lower troposphere. The filamentary structure of the moisture anomaly is found from day -3 to day -1 like the phenomena of the atmospheric rivers (ARs). However, during the time evolution of the moisture anomaly, the filamentary structure is not parallel to the column-integrated moisture transport and is mainly caused by the northward moisture transport, which is perpendicular to the quasi-stationary front from day -5 to day -3 . Therefore, although some of the widespread extreme precipitation might be related to the phenomena of the ARs, the characteristics of composited moisture anomaly might not be consistent with a typical characteristic of the ARs (Neiman et al. 2008; Demirdjjan et al. 2020; Moore et al. 2020, 2021). The time series of the moisture anomaly will be closely examined in section 3b using the prognostic equation of the column-integrated specific humidity.

Next, the large-scale dynamic forcing of the vertical motion is investigated following the framework of the QG omega equation in Eq. (1). Figure 4 shows the time series of the composited Q vector and its convergence following Eq. (2) from day -2 to day 0 at 500 hPa. The isotherms seem to be almost parallel to

the latitudinal lines over Japan, and therefore the zonal (meridional) component of the Q vector is mainly associated with a shear deformation (a confluence/diffuence) effect (Horinouchi and Hayashi 2017). On day -2 , a band-shaped region of the Q-vector convergence appears on a relatively strong temperature gradient from East China to Japan with a negative local maximum over East China. The strong convergence over East China on day -2 is likely caused by the cyclonic anomaly of the Altai–Sayan cyclone from the north (Text S3 and Fig. S3 in the online supplemental material). From day -2 to day 0 , anomalous Q-vector convergence moves eastward from 120° to 135°E and a strong Q-vector convergence appears over the precipitation area. The temporal evolution of the Q vector is likely associated with the enhancement and eastward extension of the cyclonic anomaly (Fig. 3). The direction of the Q vector indicates that both effects by confluence and by shear deformation contribute to the Q-vector convergence. The confluence effect is attributable to a zonal wind anomaly induced by the cyclonic anomaly, which is superimposed onto the jet entrance region of the climatological jet stream in the lower and upper troposphere.

b. Temporal evolution of the moisture anomalies

The temporal evolution of the column-integrated water vapor in Fig. 3 is quantitatively estimated using a prognostic equation of precipitable water:

$$\begin{aligned} \frac{\partial \langle q' \rangle}{\partial t} = & -\langle \mathbf{u}' \cdot \nabla \bar{q} \rangle - \langle \bar{\mathbf{u}} \cdot \nabla q' \rangle - \langle \mathbf{u}' \cdot \nabla q' \rangle \\ & - \langle \bar{q} (\nabla \cdot \mathbf{u}') \rangle - \langle q' (\nabla \cdot \bar{\mathbf{u}}) \rangle - \langle q' (\nabla \cdot \mathbf{u}') \rangle \\ & - (\omega q')'_{p=p_s} - P' + E', \end{aligned} \quad (9)$$

where a prime symbol denotes a deviation from the daily climatology and an overbar denotes the daily climatology. In addition, P denotes precipitation and E denotes evaporation. Angled brackets indicate mass-weighted vertical integration over the troposphere. Note that the $-(\omega q')'_{p=p_s}$ term remains in Eq. (9) since the mass-weighted vertical integral of $\langle \partial_p (\omega q') \rangle$ is not exactly

FIG. 5. Time series of the terms in the prognostic equation of the column-integrated water vapor in Eq. (9) averaged over (a) East China (28°–34°N, 115°–125°E) and (b) western Japan (30°–36°N, 130°–136°E). The contributions from the terms are denoted by colored and solid or dashed contours that are represented in the key to the right of the panels. For example, the convergence of the climatological moisture by the anomalous horizontal wind [$-\langle \bar{q}(\nabla \cdot \mathbf{u}') \rangle$] is denoted by the red and solid contour.

zero with $p_s = 1000$ hPa. The three terms on the right-hand side in the first row of Eq. (9) represent the vertical integration of the advection terms, while the terms in the middle row represent the vertical integration of the moisture convergence terms.

First, the time series of the terms in Eq. (9) over East China (28°–34°N, 115°–125°E) are shown in Fig. 5a. The convergence of the climatological moisture by the anomalous horizontal wind mainly moistens the atmospheric column, while the precipitation almost cancels out this tendency. In addition, the advection term of anomalous moisture by the anomalous wind components results in a negative time tendency of precipitable water. It should be noted that the time tendency of $\partial \langle q' \rangle / \partial t$ seems to be correlated with the first two advection terms in Eq. (9), namely $-\langle \mathbf{u}' \cdot \nabla \bar{q} \rangle$ and $-\langle \bar{\mathbf{u}} \cdot \nabla q' \rangle$. In particular, the correlation coefficient between $\partial \langle q' \rangle / \partial t$ and $-\langle \bar{\mathbf{u}} \cdot \nabla q' \rangle$ is approximately 0.88 during this period. Thus, the temporal evolution of the positive q' over East China seen in Fig. 3 from day -5 to day -1 is likely attributable to the advection terms. In other words, the climatological southerlies advect the positive q' around Taiwan, while the climatological moisture field

is advected by the anomalous northwestward flow caused by the positive Φ' over the NPSH. Note that the time series of the evaporation term has been removed since the contribution of evaporation is found to be negligible.

Next, the time series in Eq. (9) over western Japan (130°–136°E, 30°–36°N) is shown in Fig. 5b. The large amount of precipitation is almost balanced by $-\langle \bar{q}(\nabla \cdot \mathbf{u}') \rangle$ around day ∓ 0 . Over western Japan, the time tendency of $\partial \langle q' \rangle / \partial t$ is well correlated with $-\langle \bar{\mathbf{u}} \cdot \nabla q' \rangle$ with a high correlation coefficient of 0.786, as shown in Fig. 5a. This strong correlation means that the advection of the positive q' over East China by the climatological jet stream is one of the most important factors responsible for the temporal evolution of q' over western Japan, which is consistent with the temporal evolution of the zonally elongated anomaly depicted in Fig. 2.

c. Quantitative analysis for the enhancement of the cyclonic anomaly

As shown in Figs. 2 and 3, the cyclonic anomaly near western Japan on day ∓ 0 is enhanced over East China and moves eastward. The temporal evolution of this cyclonic anomaly can

FIG. 6. A time evolution of the horizontal maps of the composite anomaly of potential vorticity (PVU) at 550 hPa with days -2 , -1.5 , -1 , and -0.5 of the widespread extreme precipitation. The contour interval is 0.03 PVU. The shaded area indicates that the anomalies are statistically significant at a 95% confidence level using the Student's t test, while the dotted area indicates that the anomalies are statistically significant at a 90% confidence level.

be quantitatively analyzed by the prognostic equation of the potential vorticity (PV) following Hoskins et al. (1985):

$$\frac{\partial PV'}{\partial t} = -(\mathbf{u} \cdot \nabla PV)' - \left(\omega \frac{\partial PV}{\partial p} \right)' - g \left(\boldsymbol{\eta}_h \cdot \nabla H + \eta \frac{\partial H}{\partial p} \right)', \quad (11)$$

where PV denotes the potential vorticity, and $\boldsymbol{\eta}_h$ and η denote a horizontal and vertical component of the absolute vorticity, respectively. The nonlinear terms are calculated in each time step of JRA-55 and then the composites are made for the widespread extreme precipitation events. Figure 6 shows the temporal evolution of PV' at 550 hPa from day -2 to day -0.5 . We have selected the vertical level of 550 hPa to analyze PV' since the PV production term as the third term on the rhs of Eq. (11) has a peak near this level. From day -2 to day -1 , a cyclonic anomaly with a positive PV' is enhanced over East China at 35°N above a cyclonic anomaly (likely due to tropical cyclones) at 20°N , and then moves eastward toward Japan. To examine the enhancement of

PV' over East China from day -2 to day -1 , a horizontal map of the PV' time tendency, the sum of the advection terms, and the PV production terms following Eq. (11) on day -2 are shown in Fig. 7. The time change of PV' is positive along 30°N with a peak at a coastal region of East China and over Japan. The positive time tendency of PV' could apparently be attributable to the positive PV' production term, which is closely discussed in section 4.

Next, to examine the eastward extension and the enhancement of PV', the same analysis as in Fig. 7 is performed on day -1.5 in Fig. 8. Since the positive tendency of PV' appears from East China to the Sea of Japan, one might think that this is simply due to advection term by the climatological zonal jet stream, similar to the time evolution of q' . However, the sum of the advection terms in Fig. 8b is negative even in front of the cyclonic anomaly, although the advection term by the zonal wind component is positive as expected (Fig. 8d). This is because the sum of the time tendency by the meridional advection of the climatological low PV from the equatorward by the anomalous

FIG. 7. Horizontal maps of the composite anomaly of (a) time change of the potential vorticity, (b) a budget by the advection term in Eq. (11), and (c) a budget by the PV production term in Eq. (11) at 550 hPa on day -2 . The contour interval is $0.05 \text{ PVU day}^{-1}$. The shaded area indicates that the anomalies are statistically significant at a 95% confidence level using the Student's t test, while the dotted area indicates that the anomalies are statistically significant at a 90% confidence level.

northward wind and that by the vertical advection of the low PV in the lower troposphere exceeds the time tendency by the zonal advection. Thus, the positive time tendency of PV' is mainly caused by the PV production term due to diabatic heating. The important point is that the eastward extent and the enhancement of the cyclonic anomaly on day -1.5 clearly differs from a typical developing baroclinic disturbance (e.g., Holton 2004) because the diabatic process dominated the time tendency of PV' . The fundamental question discussed in section 4 is what processes determine the time and spatial evolution of the diabatic heating in front of the cyclonic anomaly.

4. Discussion

This section will focus on the coupling process between the dynamic forcing and diabatic heating. Analyzing the Q vector in the horizontal plane at a single vertical level is not sufficient for this purpose since the dynamically forced vertical motion is determined by multilevel dynamic forcing due to the property of the Laplacian operator in Eq. (3). By calculating Eqs. (6) and (7), we obtain the vertical profiles of ω_D , ω_H , and ω_{res} . As an example, Figure S4 shows a profile of ω from the JRA-55 and the obtained profiles of ω_D , ω_H , and ω_{res} on day ± 0 averaged over western Japan ($30^\circ\text{--}36^\circ\text{N}$, $130^\circ\text{--}136^\circ\text{E}$). The vertical profiles of ω_D and ω_H have peaks at approximately 500 hPa, while ω_{res} has a bottom-heavy profile. Although ω_D has much smaller amplitude than ω_H , the diabatic heating is considered to be a response of the relatively weak dynamical forcing in the extreme precipitation cases (Nie and Sobel 2016; Nie and Fan 2019). Therefore, we have compared the time evolution of the anomaly of the PV production term as the source of ω_H and the anomaly of ω_D (ω'_D) from day -2 to day -1 .

Figure 9 shows the horizontal maps of the anomaly of PV production term and ω'_D at 550 hPa on day -2 to day -1 . As previously discussed, the positive PV production term is distributed as a band-like shape from East China to Japan on day -2 , which leads to the enhancement of a new cyclonic

anomaly over East China. The anomalous dynamically forced upward motion is also analyzed over East China, which is likely induced by the wind anomalies superimposed on the climatological jet entrance region (Fig. 3; see also Fig. S3). Note that a low-frequency positive PV production anomaly appears over Japan Island in the presence of the baiu front, even before day -2 with little anomalous dynamic forcing. When this dynamically forced vertical motion was intensified on day -1 in front of the cyclonic anomaly (Fig. 9e), the positive PV production rate also increased west of Japan (Fig. 9b). To distinguish the low-frequency component of positive PV production anomaly, the differences in the PV production term and ω'_D between day -2 and day -1 are shown in Figs. 9c and 9f. The spatial pattern is quite similar except for the coastal region of East China. In addition, the time series of $\partial PV'/\partial t$, anomalies of advection term, and PV production term averaged over western Japan ($33^\circ\text{--}39^\circ\text{N}$, $130^\circ\text{--}136^\circ\text{E}$) at 550 hPa are also compared to those of ω' , ω'_D , and ω'_H in Fig. 10. It is found that the time series of ω'_D is well in phase with that of ω' , ω'_H , and the PV production term, implying the importance of the coupling process between the dynamically forced vertical motion and diabatic heating (Nie and Sobel 2016; Nie and Fan 2019), not only in the spatial distribution (Fig. 9) but also in the temporal evolution (Fig. 10).

It is likely that the coupling process inherently involves an interaction between a thermodynamic background condition and a well-organized rainfall system (Yokoyama et al. 2017; Yokoyama et al. 2020): Dynamical forced motion moistens the midtroposphere, which is favorable for a development of deep cumulus convection and its organization (e.g., Sherwood 1999; Kuang and Bretherton 2006; Takayabu et al. 2010; Schiro and Neelin 2019; Demirdjian et al. 2020). Once deep cumulus condition is promoted under a condition of a lower-tropospheric convective instability, convection itself can also moisten the midtroposphere further through the diabatic ascent, and hence enhances the favorable condition for the organization of the rainfall system.

Why is the dynamically forced motion excited in front of the low-level cyclonic anomaly? As shown in Fig. 3, southerly wind

FIG. 8. Horizontal maps of the composite anomaly of (a) time change of the potential vorticity, (b) a budget by the advection term in Eq. (11), (c) a budget by the PV production term in Eq. (11), and a time tendency by the advection term due to (d) zonal, (e) meridional, and (f) vertical wind components at 550 hPa with day -1.5 . The contour interval is $0.05 \text{ PVU day}^{-1}$. The shaded area indicates that the anomalies are statistically significant at a 95% confidence level using the Student's t test, while the dotted area indicates that the anomalies are statistically significant at a 90% confidence level.

anomalies are induced between the new low-level cyclonic anomaly and the persistent positive Φ' over NPSH. Since the low-level wind anomalies are superimposed on the low-level jet stream and are simultaneously across the baiu front, the Q-vector convergence is then effectively induced in the presence of the cyclonic anomaly following Eq. (2) (Figs. 4 and 9). Therefore, both the persistent positive Φ' over NPSH and the baiu front would be the important preliminary factor for the enhancement of the new cyclonic anomaly over East China.

The dynamical structure of the evolution of PV' in this study is consistent with a behavior of a diabatic Rossby wave (Parker and Thorpe 1995; Boettcher and Wernli 2013): The diabatic Rossby wave has a low-level positive PV anomaly located in a moist and a baroclinic environment, without a strong dynamical coupling from an upper-level disturbance. A meridional wind gives a thermal advection at the downstream of PV' , giving an ageostrophic vertical motion and therefore a diabatic heating as a PV source. Thus, the diabatic Rossby wave could propagate eastward rapidly along the baroclinic zone by its continuous diabatic generation of PV at the downstream of its original position, which is consistent with our results. In these

cases, the climatological frontal zone in East Asia, the baiu frontal zone, would be a key component for the evolution of PV' like a diabatic Rossby wave. A numerical study by Parker and Thorpe (1995) shows that the diabatic Rossby wave has a solitary behavior in which a growth rate, a phase speed, and a structure of an unstable mode are independent of wavelength within its range of 1900–4000 km. Although it is only a rough estimation in Fig. 6, the phase speed of PV' , which moves from 120° to 130°E within days -1.5 to -0.5 , is approximately 10.5 m s^{-1} , which is the same order of a theoretical phase speed of an unstable mode of the diabatic Rossby wave ($\sim 14 \text{ m s}^{-1}$).

According to Parker and Thorpe (1995), a numerical behavior of the unstable mode in the moist-baroclinic environment depends on a nondimensional parameter ε controlling a parameterized heating amplitude H :

$$\frac{D\theta}{Dt} \equiv H = \theta_z \varepsilon h(z) w(z_b), \quad (12)$$

where z denotes height and z_b denotes a height where PV' is located, w denotes a vertical wind component, θ denotes a potential temperature, and h denotes a specified function of height.

FIG. 9. Horizontal maps of the composite anomaly of the PV production terms at (a) day -2 and (b) day -1 at 550 hPa, and (c) their difference. (d)–(f) As in (a)–(c), but for the anomaly of dynamical forced vertical motion (ω_D') at 550 hPa. The contour interval is $0.05 \text{ PVU day}^{-1}$ in (a)–(c) and $0.8 \times 10^{-2} \text{ Pa s}^{-1}$ in (d)–(f). The shaded area indicates that the anomalies are statistically significant at a 95% confidence level using the Student's t test, while the dotted area indicates that the anomalies are statistically significant at a 90% confidence level.

Thus, the parameter ε represents an intensity of the diabatic heating as a response to the vertical wind component, which is similar to a diabatic heating feedback parameter α (Nie and Sobel 2016; Nie and Fan 2019) defined as a ratio of precipitation amounts due to ω_H against that due to ω_D . For low ε , the unstable mode is consistent with a dry baroclinic wave, while for large ε that has a moist-dominated solution as a diabatic Rossby wave in which the diabatic energy conversion plays a key role rather than the baroclinic conversion. It is clear by Eqs. (7) and (12) that a stronger coupling between diabatic heating and ω_D with large α is associated with an unstable system with large ε . Since a sensitivity of moist convection strongly depends on environmental moisture (e.g., Derbyshire et al. 2004; Holloway and Neelin 2009; Schiro and Neelin 2019), the higher CWV in the mei-yu-baiu region is a favorable condition for large α (Nie and Fan 2019) and therefore for the unstable regime of diabatic Rossby waves. It is notable that, although a typical diabatic Rossby wave has an isolated wave structure in the lower troposphere, the upper-level trough on the subtropical jet is also statistically significant on day ∓ 0 (Fig. 2c). It is therefore indicated that the interaction from the upper-level trough would be important for the widespread extreme precipitation events (Takemura

et al. 2019; Yokoyama et al. 2020). Further work is needed to elucidate the statistical behaviors of the diabatic Rossby waves and their interaction with the upper-level trough in the baiu region.

5. Summary

This study reveals the large-scale environment under which widespread extreme precipitation events have historically formed over western Japan for the years 1979 to 2020. Figure 11 summarizes the environmental conditions under the widespread extreme precipitation events. Preceding the event before approximately a week, the equivalent barotropic structure of the positive anomaly of NPSH is maintained due to a low-frequency Rossby wave train propagation through the polar front jet from eastern Europe to Japan (Fig. S2). In addition, lee cyclogenesis occurs at Altai–Sayan region and moves southeastward toward Japan. On three days before the event, the cyclonic anomaly induces Q-vector convergence over East China by producing the anomalous jet entrance region, while the low-frequency anticyclonic anomaly induces the moisture transport toward East China. Under such a background condition, a new cyclonic anomaly is enhanced due

FIG. 10. Time series of (a) the terms in the prognostic equation of the potential vorticity in Eq. (11) and (b) of ω' , ω_D' , and ω_H' at 550 hPa averaged over western Japan (33°–39°N, 130°–136°E). The contributions from the terms are denoted by lines represented in the key to the right of the panels.

to diabatic heating induced by the dynamically forced motion under the rich moisture anomaly over East China.

At the peak of the extreme precipitation, the pair of the positive and negative geopotential anomaly induces the strong

moisture fluxes and Q-vector convergence over the precipitating area. The temporal and spatial patterns of the dynamically forced vertical velocity are well in phase with those of the diabatic heating term, which is consistent with the dynamics of

FIG. 11. A schematic for factors of the atmospheric circulation pattern behind the widespread extreme precipitation events.

the diabatic Rossby wave. This might imply the importance of the coupling process between the dynamical forced motion and the diabatic heating in the widespread extreme precipitation events. Such features are statistically significant for historical widespread extreme precipitation cases.

A new finding in this analysis is that the low-level cyclonic anomaly that causes the widespread extreme precipitation evolves over East China and extends eastward as a similar manner to the diabatic Rossby wave. This result provides a physical understanding of the reason why the background moisture and the baroclinicity are both essential in the composited atmospheric fields and of the importance of the feedback parameter between the dynamical motion and diabatic heating.

The regional atmospheric condition for background moisture, stability, and baroclinicity could change due to global warming, likely leading to variation of the intensity and frequency of the widespread extreme precipitations. Further research would be interesting to examine frequencies and future changes of such an atmospheric circulation pattern using phases 5 and 6 of the Coupled Model Intercomparison Project [CMIP5 (Taylor et al. 2012) and CMIP6 (Eyring et al. 2016)]. In addition, different climate models show large uncertainties of regional climate changes (Shepherd 2019). For the description of plausible futures, the storyline technique (e.g., Zappa and Shepherd 2017) is a useful approach to understand what factors characterize the uncertainty of regional atmospheric circulation changes in climate change responses.

Acknowledgments. This study is supported by the Environment Research and Technology Development Fund (JPMEERF20192004) of the Environmental Restoration and Conservation Agency of Japan, the University of Tokyo, through a project “Research hub for the big data analysis of global water cycle and precipitation in changing climate”, and the Japan Society for the Promotion of Science (JSPS) through Grants-in-Aid for Scientific Research JP19H05702. The authors thank Dr. Hisashi Nakamura, Dr. Yu Kosaka, Dr. Nagio Hirota, Dr. Takeshi Horinouchi, and Dr. Koutarou Takaya for their many useful comments and discussions. We are also grateful to Dr. Chie Yokoyama and Dr. Hiroki Tsuji for their advice and supports. The JRA-55 data used in this study were provided by the Japan Meteorological Agency (JMA). All figures shown in this paper were created using the Dennou Club Library (DCL).

REFERENCES

- Bergeron, T., 1928: *Ueber die dreidimensional verknuepfende Wetteranalyse*. Vol. 5. Geofysiske Publikasjoner, 99 pp.
- Boettcher, M., and H. Wernli, 2013: A 10-yr climatology of diabatic Rossby waves in the Northern Hemisphere. *Mon. Wea. Rev.*, **141**, 1139–1154, <https://doi.org/10.1175/MWR-D-12-00012.1>.
- Buzzi, A., S. Davolio, and M. Fantini, 2020: Cyclogenesis in the lee of the Alps: A review of theories. *Bull. Atmos. Sci. Technol.*, **1**, 433–457, <https://doi.org/10.1007/s42865-020-00021-6>.
- Byun, K.-Y., and T.-Y. Lee, 2012: Remote effects of tropical cyclones on heavy rainfall over the Korean Peninsula—Statistical and composite analysis. *Tellus*, **64A**, 14983, <https://doi.org/10.3402/tellusa.v64i0.14983>.
- Chen, S.-J., and L. Lazić, 1990: Numerical case study of the Altai-Sayan lee cyclogenesis over East Asia. *Meteor. Atmos. Phys.*, **42**, 221–229, <https://doi.org/10.1007/BF01314826>.
- , Y.-H. Kuo, P.-Z. Zhang, and Q.-F. Bai, 1991: Synoptic climatology of cyclogenesis over East Asia, 1958–1987. *Mon. Wea. Rev.*, **119**, 1407–1418, [https://doi.org/10.1175/1520-0493\(1991\)119<1407:SCOCOE>2.0.CO;2](https://doi.org/10.1175/1520-0493(1991)119<1407:SCOCOE>2.0.CO;2).
- Chen, Y., and P. Zhai, 2016: Mechanisms for concurrent low-latitude circulation anomalies responsible for persistent extreme precipitation in the Yangtze River Valley. *Climate Dyn.*, **47**, 989–1006, <https://doi.org/10.1007/s00382-015-2885-6>.
- De Marchi, L., 1900: Il secolo XIX nella vita e nella cultura dei popoli. *La Fisica Terrestre*, F. Vallardi, 227–257, <https://www.maremagnum.com/libri-antichi/il-secolo-xix-l-astronomia-la-fisica-terrestre/163552295>.
- Demirdjian, R., J. D. Doyle, C. A. Reynolds, J. R. Norris, A. C. Michaelis, and F. M. Ralph, 2020: A case study of the physical processes associated with the atmospheric river initial condition sensitivity from an adjoint model. *J. Atmos. Sci.*, **77**, 691–709, <https://doi.org/10.1175/JAS-D-19-0155.1>.
- Derbyshire, S. H., I. Beau, P. Bechtold, J.-Y. Grandpeix, J.-M. Piriou, J.-L. Redelsperger, and P. M. M. Soares, 2004: Sensitivity of moist convection to environmental humidity. *Quart. J. Roy. Meteor. Soc.*, **130**, 3055–3079, <https://doi.org/10.1256/qj.03.130>.
- Eyring, V., S. Bony, G. A. Meehl, C. A. Senior, B. Stevens, R. J. Stouffer, and K. E. Taylor, 2016: Overview of the Coupled Model Intercomparison Project Phase 6 (CMIP6) experimental design and organization. *Geosci. Model Dev.*, **9**, 1937–1958, <https://doi.org/10.5194/gmd-9-1937-2016>.
- Hamada, A., and Y. N. Takayabu, 2018: Large-scale environmental conditions related to midsummer extreme rainfall events around Japan in the TRMM region. *J. Climate*, **31**, 6933–6945, <https://doi.org/10.1175/JCLI-D-17-0632.1>.
- Harada, Y., H. Endo, and K. Takemura, 2020: Characteristics of large-scale atmospheric fields during heavy rainfall events in western Japan: Comparison with an extreme event in early July 2018. *J. Meteor. Soc. Japan*, **98**, 1207–1229, <https://doi.org/10.2151/jmsj.2020-062>.
- Hersbach, H. and D. Dee, 2016: ERA5 reanalysis is in production. *ECMWF Newsletter*, No. 147, ECMWF, Reading, United Kingdom, <https://www.ecmwf.int/en/newsletter/147/news/era5-reanalysis-production>.
- Hirota, N., Y. N. Takayabu, M. Kato, and S. Arakane, 2016: Roles of an atmospheric river and a cutoff low in the extreme precipitation event in Hiroshima on 19 August 2014. *Mon. Wea. Rev.*, **144**, 1145–1160, <https://doi.org/10.1175/MWR-D-15-0299.1>.
- Holloway, C. E., and J. D. Neelin, 2009: Moisture vertical structure, column water vapor, and tropical deep convection. *J. Atmos. Sci.*, **66**, 1665–1683, <https://doi.org/10.1175/2008JAS2806.1>.
- Holton, J., 2004: *An Introduction to Dynamic Meteorology*. 4th ed. Elsevier Academic Press, 535 pp.
- Horinouchi, T., 2014: Influence of upper tropospheric disturbances on the synoptic variability of precipitation and moisture transport over summertime East Asia and the northwestern Pacific. *J. Meteor. Soc. Japan*, **92**, 519–541, <https://doi.org/10.2151/jmsj.2014-602>.
- , and A. Hayashi, 2017: Meandering subtropical jet and precipitation over summertime East Asia and the northwestern Pacific. *J. Atmos. Sci.*, **74**, 1233–1247, <https://doi.org/10.1175/JAS-D-16-0252.1>.
- Hoskins, B. J., M. E. McIntyre, and A. W. Robertson, 1985: On the use and significance of isentropic potential vorticity maps.

- Quart. J. Roy. Meteor. Soc.*, **111**, 877–946, <https://doi.org/10.1002/qj.49711147002>.
- Hu, Y., Y. Deng, Z. M. Zhou, C. G. Cui, and X. Q. Dong, 2019: A statistical and dynamical characterization of large-scale circulation patterns associated with summer extreme precipitation over the middle reaches of the Yangtze River. *Climate Dyn.*, **2**, 6213–6288, <https://doi.org/10.1007/s00382-018-4501-z>.
- Kamae, Y., W. Mei, S.-P. Xie, M. Naoki, and H. Ueda, 2017: Atmospheric rivers over the northwestern Pacific: Climatology and interannual variability. *J. Climate*, **30**, 5605–5619, <https://doi.org/10.1175/JCLI-D-16-0875.1>.
- Kato, T., 2020: Quasi-stationary band-shaped precipitation systems, named “Senjo-Kousuitai”, causing localized heavy rainfall in Japan. *J. Meteor. Soc. Japan*, **98**, 485–509, <https://doi.org/10.2151/jmsj.2020-029>.
- Knapp, K. R., M. C. Kruk, D. H. Levinson, H. J. Diamond, and C. J. Neumann, 2010: The International Best Track Archive for Climate Stewardship (IBTrACS): Unifying tropical cyclone data. *Bull. Amer. Meteor. Soc.*, **91**, 363–376, <https://doi.org/10.1175/2009BAMS2755.1>.
- Kobayashi, S., and Coauthors, 2015: The JRA-55 Reanalysis: General specifications and basic characteristics. *J. Meteor. Soc. Japan*, **93**, 5–48, <https://doi.org/10.2151/jmsj.2015-001>.
- Kuang, Z., and C. S. Bretherton, 2006: A mass-flux scheme view of a high-resolution simulation of a transition from shallow to deep cumulus convection. *J. Atmos. Sci.*, **63**, 1895–1909, <https://doi.org/10.1175/JAS3723.1>.
- Lee, J., S.-W. Son, H.-O. Cho, J. Kim, D.-H. Cha, J. R. Gyakum, and D. Chen, 2020: Extratropical cyclones over East Asia: Climatology, seasonal cycle, and long-term trend. *Climate Dyn.*, **54**, 1131–1144, <https://doi.org/10.1007/s00382-019-05048-w>.
- Moore, B. J., A. B. White, D. J. Gottas, and P. J. Neiman, 2020: Extreme precipitation events in Northern California during winter 2016–17: Multiscale analysis and climatological perspective. *Mon. Wea. Rev.*, **148**, 1049–1074, <https://doi.org/10.1175/MWR-D-19-0242.1>.
- , —, and —, 2021: Characteristics of long-duration heavy precipitation events along the West Coast of the United States. *Mon. Wea. Rev.*, **149**, 2255–2277, <https://doi.org/10.1175/MWR-D-20-0336.1>.
- Murakami, T., and W.-G. Huang, 1984: Orographic effects of the Tibetan Plateau on the rainfall variations over central China during the 1979 summer. *J. Meteor. Soc. Japan*, **62**, 895–909, https://doi.org/10.2151/jmsj.1965.62.6_895.
- Neiman, P. J., F. M. Ralph, G. A. Wick, J. D. Lundquist, and M. D. Dettinger, 2008: Meteorological characteristics and overland precipitation impacts of atmospheric rivers affecting the West Coast of North America based on eight years of SSM/I satellite observations. *J. Hydrometeorol.*, **9**, 22–47, <https://doi.org/10.1175/2007JHM855.1>.
- Nie, J., and A. H. Sobel, 2016: Modeling the interaction between quasigeostrophic vertical motion and convection in a single column. *J. Atmos. Sci.*, **73**, 1101–1117, <https://doi.org/10.1175/JAS-D-15-0205.1>.
- , and B. Fan, 2019: Roles of dynamic forcings and diabatic heating in summer extreme precipitation in East China and the southeastern United States. *J. Climate*, **32**, 5815–5831, <https://doi.org/10.1175/JCLI-D-19-0188.1>.
- Ninomiya, K., 1984: Characteristics of Baiu front as a predominant subtropical front in the summer Northern Hemisphere. *J. Meteor. Soc. Japan*, **62**, 880–894, https://doi.org/10.2151/jmsj.1965.62.6_880.
- , and T. Murakami, 1987: The early summer rainy season (Baiu) over Japan. C.-P. Chang and T. N. Krishnamurti, Eds., *Monsoon Meteorology*, Oxford University Press, 93–121.
- , and T. Akiyama, 1992: Multi-scale features of Baiu, the summer monsoon over Japan and the East Asia. *J. Meteor. Soc. Japan*, **70**, 467–495, https://doi.org/10.2151/jmsj.1965.70.1B_467.
- , H. Koga, Y. Yamagishi, and Y. Tatsumi, 1984: Prediction experiment of extremely intense rainstorm by a very fine-mesh primitive equation model. *J. Meteor. Soc. Japan*, **62**, 273–295, https://doi.org/10.2151/jmsj.1965.62.2_273.
- Ogura, Y., T. Asai, and K. Dohi, 1985: A case study of heavy precipitation event along the Baiu front in northern Kyushu, 23 July 1982: Nagasaki heavy rainfall. *J. Meteor. Soc. Japan*, **63**, 883–900, https://doi.org/10.2151/jmsj.1965.63.5_883.
- Oh, H., K. J. Ha, and A. Timmermann, 2018: Disentangling impacts of dynamic and thermodynamic components on late summer rainfall anomalies in East Asia. *J. Geophys. Res. Atmos.*, **123**, 8623–8633, <https://doi.org/10.1029/2018JD028652>.
- Parker, D. J., and A. J. Thorpe, 1995: Conditional convective heating in a baroclinic atmosphere: A model of convective frontogenesis. *J. Atmos. Sci.*, **52**, 1699–1711, [https://doi.org/10.1175/1520-0469\(1995\)052<1699:CCHIAB>2.0.CO;2](https://doi.org/10.1175/1520-0469(1995)052<1699:CCHIAB>2.0.CO;2).
- Petterssen, S., 1956: *Weather Analysis and Forecasting*. Vol. 1. McGraw-Hill, 428 pp.
- Pierrehumbert, R. T., 1985: A theoretical model of orographically modified cyclogenesis. *J. Atmos. Sci.*, **42**, 1244–1258, [https://doi.org/10.1175/1520-0469\(1985\)042<1244:ATMOOM>2.0.CO;2](https://doi.org/10.1175/1520-0469(1985)042<1244:ATMOOM>2.0.CO;2).
- Rotunno, R., and R. Ferretti, 2001: Mechanisms of intense Alpine rainfall. *J. Atmos. Sci.*, **58**, 1732–1749, [https://doi.org/10.1175/1520-0469\(2001\)058<1732:MOIAR>2.0.CO;2](https://doi.org/10.1175/1520-0469(2001)058<1732:MOIAR>2.0.CO;2).
- Sampe, T., and S.-P. Xie, 2010: Large-scale dynamics of the meiyu-baiu rainband: Environmental forcing by the westerly jet. *J. Climate*, **23**, 113–134, <https://doi.org/10.1175/2009JCLI3128.1>.
- Schiro, K. A., and J. D. Neelin, 2019: Deep convective organization, moisture vertical structure, and convective transition using deep-inflow mixing. *J. Atmos. Sci.*, **76**, 965–987, <https://doi.org/10.1175/JAS-D-18-0122.1>.
- Sekizawa, S., T. Miyasaka, H. Nakamura, A. Shimpo, K. Takemura, and S. Maeda, 2019: Anomalous moisture transport and oceanic evaporation during a torrential rainfall event over western Japan in early July 2018. *SOLA*, **15A**, 25–30, <https://doi.org/10.2151/sola.15A-005>.
- Shepherd, T. G., 2019: Storyline approach to the construction of regional climate change information. *Proc. Roy. Soc.*, **475A**, 20190013, <https://doi.org/10.1098/rspa.2019.0013>.
- Sherwood, S. C., 1999: Convective precursors and predictability in the tropical western Pacific. *Mon. Wea. Rev.*, **127**, 2977–2991, [https://doi.org/10.1175/1520-0493\(1999\)127<2977:CPAPIT>2.0.CO;2](https://doi.org/10.1175/1520-0493(1999)127<2977:CPAPIT>2.0.CO;2).
- Shimpo, A., and Coauthors, 2019: Primary factors behind the heavy rain event of July 2018 and the subsequent heat wave in Japan. *SOLA*, **15A**, 13–18, <https://doi.org/10.2151/sola.15A-003>.
- Speranza, A., A. Buzzi, A. Trevisan, and P. Malguzzi, 1985: A theory of deep cyclogenesis in the lee of the Alps. Part I: Modifications of baroclinic instability by localized topography. *J. Atmos. Sci.*, **42**, 1535, [https://doi.org/10.1175/1520-0469\(1985\)042<1521:ATODCI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1985)042<1521:ATODCI>2.0.CO;2).
- Sugimoto, S., 2020: Heavy precipitation over southwestern Japan during the Baiu season due to abundant moisture transport from synoptic-scale atmospheric conditions. *SOLA*, **16**, 17–22, <https://doi.org/10.2151/sola.2020-004>.
- Takayabu, Y. N., S. Shige, W.-K. Tao, and N. Hirota, 2010: Shallow and deep latent heating modes over tropical oceans observed

- with TRMM PR spectral latent heating data. *J. Climate*, **23**, 2030–2046, <https://doi.org/10.1175/2009JCLI13110.1>.
- Takemura, K., S. Wakamatsu, H. Togawa, A. Shimpo, C. Kobayashi, S. Maeda, and H. Nakamura, 2019: Extreme moisture flux convergence over western Japan during the heavy rain event of July 2018. *SOLA*, **15A**, 49–54, <https://doi.org/10.2151/sola.15A-009>.
- Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012: An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498, <https://doi.org/10.1175/BAMS-D-11-00094.1>.
- Tochimoto, E., and T. Kawano, 2017: Numerical investigation of development processes of Baiu frontal depressions. Part I: Case studies. *J. Meteor. Soc. Japan*, **95**, 91–109, <https://doi.org/10.2151/jmsj.2017-005>.
- Tsuji, H., C. Yokoyama, and Y. N. Takayabu, 2020: Contrasting features of the July 2018 heavy rainfall event and the 2017 northern Kyushu rainfall event in Japan. *J. Meteor. Soc. Japan*, **98**, 859–876, <https://doi.org/10.2151/jmsj.2020-045>.
- Wei, W., R. Zhang, M. Wen, X. Rong, and T. Li, 2014: Impact of Indian summer monsoon on the South Asian high and its influence on summer rainfall over China. *Climate Dyn.*, **43**, 1257–1269, <https://doi.org/10.1007/s00382-013-1938-y>.
- Yokoyama, C., Y. N. Takayabu, and T. Horinouchi, 2017: Precipitation characteristics over East Asia in early summer: Effects of the subtropical jet and lower tropospheric convective instability. *J. Climate*, **30**, 8127–8147, <https://doi.org/10.1175/JCLI-D-16-0724.1>.
- , H. Tsuji, and Y. N. Takayabu, 2020: The effects of an upper-tropospheric trough on the heavy rainfall event in July 2018 over Japan. *J. Meteor. Soc. Japan*, **98**, 235–255, <https://doi.org/10.2151/jmsj.2020-013>.
- Zappa, G., and T. G. Shepherd, 2017: Storylines of atmospheric circulation change for European regional climate impact assessment. *J. Climate*, **30**, 6561–6577, <https://doi.org/10.1175/JCLI-D-16-0807.1>.
- Zhao, Y., D. Chen, J. Li, D. Chen, Y. Chang, J. Li, and Q. Rui, 2020: Enhancement of the summer extreme precipitation over North China by interactions between moisture convergence and topographic settings. *Climate Dyn.*, **54**, 2713–2730, <https://doi.org/10.1007/s00382-020-05139-z>.
- Zhou, T.-J., and R.-C. Yu, 2005: Atmospheric water vapor transport associated with typical anomalous summer rainfall patterns in China. *J. Geophys. Res.*, **110**, D08104, <https://doi.org/10.1029/2004JD005413>.