Role of the Kuroshio Current on the Diurnal Variation of the Early-Summer Meiyu-Baiu Rainband

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Research Article

Keywords: diurnal cycle of precipitation, Meiyu-Baiu rainband, Kuroshio Current, convective instability, frontogenesis

Posted Date: August 23rd, 2021

DOI: <https://doi.org/10.21203/rs.3.rs-828776/v1>

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**Role of the Kuroshio Current on the Diurnal Variation of the Early-Summer
Meiyu-Baiu Rainband**

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Submitted to: Climate Dynamics

Submission date: August 19, 2021

18 **Abstract**

19 Mechanism of the strong diurnal cycle of precipitation over the Kuroshio Current
20 (KC) during mid-June is investigated, when the climatological location of the
21 Meiyu-Baiu front overlaps the KC. Heating from the KC intensifies in the morning
22 when the temperature difference between the sea surface and the surface air (TDF)
23 maximizes. The diurnal cycle of precipitation, on the other hand, peaks in the
24 afternoon, consistent with previous studies. It is revealed that convective
25 precipitation (CP) due to convective instability is in phase with TDF, whereas
26 large-scale precipitation (LSP) caused by the cross-frontal circulation matures
27 later. Intensified convective instability via enhanced heating from the KC in the
28 morning hours (03–12 LST) increases the mean amount of CP as well as the
29 probability of stronger CP. Surface wind convergence is also strengthened during
30 the morning hours and helps sustain the convection. The diurnal cycle of LSP,
31 which peaks in the afternoon hours (12–15LST), covaries with the intensity of the
32 Meiyu-Baiu front and the associated cross-frontal circulation. The wind
33 convergence and deformation anomalies associated with the intensified thermal
34 heating over the KC during the morning hours intensifies the frontogenesis
35 function, which leads to the maximization of the frontal intensity in the afternoon.
36 The direct contribution of diabatic heating to the frontogenesis is relatively weak.

37

38 *Keywords:* diurnal cycle of precipitation; Meiyu-Baiu rainband; Kuroshio Current;
39 convective instability; frontogenesis

1. Introduction

The early-summer East Asian rainy season is characterized by a quasi-stationary rainband elongated from the eastern China to Japan. This rainband, known as Meiyu in China and Baiu in Japan, are frequently accompanied by heavy rainfall events and exerts an enormous impact on local communities. In particular, intense precipitation exceeding twice the amount in other months is observed during the early summer (June) when the Meiyu-Baiu (MB) rainband is located over the East China Sea (ECS) and is affected from beneath by warm sea surface temperature (SST) (Sasaki et al. 2012).

The Meiyu-Baiu Front (MBF) draws a boundary between mid-latitude and subtropical air masses and is characterized by a strong gradient of equivalent potential temperature (Ninomiya 1984; Ninomiya and Akiyama 1992; Kodama 1993). The MBF exhibits multi-scale characteristics from meso- to large-scale spatial distribution (Akiyama 1989; Akiyama 1990a, Ninomiya and Akiyama 1992; Ninomiya 2000) and diurnal to several-day temporal distribution (Akiyama 1990a; Ninomiya 2004). The MBF entails the meso- α to γ scale clouds along the frontal zone, constituting a cloud system family with a length scale of ~ 2000 km. The type of precipitation depends on the position relative to the MBF. The southern side is characterized by a higher ratio of convective rainfall, with more intense and deep precipitation (Yokoyama et al. 2014). Convergence of equivalent potential temperature flux accompanying the low-level jet stream in the MBF zone produces convective instability, which plays a role of maintaining moist-neutral stratification against the stabilizing effect of convection (Kodama 1992; Akiyama 1979; Yokoyama et al. 2017; Ninomiya 2004).

High SST and SST gradient (SSTG) accompanying oceanic mesoscale features, including eddies and the western boundary currents (WBC), are known to exert a significant influence on local climatology. The effects of WBC and mesoscale eddies on the atmosphere near the sea surface are generally understood in terms of pressure adjustment mechanisms (PAM; Lindzen and Nigam 1987) and vertical mixing mechanisms (VMM; Hayes et al. 1989). The PAM and VMM relate warm SST with localized low pressure and high surface wind speed, respectively. Based on this, correlation between SST and near-surface atmospheric conditions in the worldwide oceans has been reviewed (Chelton et al. 2001; Nonaka and Xie 2003; Vecchi et al. 2004; Tokinaga et al. 2005; O'Neil et al. 2005, Xu et al. 2010; Frenger et al. 2013).

Atmospheric variables over the ECS are also significantly affected by the Kuroshio Current (KC) seasonally. Many studies discussed atmospheric response to the KC during winter and/or spring, during which high wind speed, strong wind convergence, and frequent cumulus convection accompanied by deep clouds are observed (Xie et al. 2002; Nonaka and Xie 2003; Xu et al. 2011; Liu et al. 2013; Liu et al. 2016). Similar atmospheric responses are also observed in early summer. Even though temperature difference between the ocean and the atmosphere is relatively small, presence of the MBF and rainband over the KC makes air-sea interaction interesting during early summer.

Sasaki et al. (2012) analyzed the effect of the KC on the Baiu rainband based on satellite measurements, reanalysis fields, and a regional atmospheric model. The average precipitation pattern over the ECS during June shows locally enhanced fine-scale features along the KC warm tongue embedded in the large-

scale Baiu rainband. Warm SST along the KC is also accompanied by stronger surface wind, evaporation, and wind convergence. Furthermore, it is shown that the KC can anchor ascending motions and diabatic heating extending to upper troposphere. Xu et al. (2018) analyzed the effect of the KC on the climatological evolution of the Meiyu rainband by dividing the period when it is located over the warm and cold side of the SST front based on high-resolution satellite and reanalysis data. More frequent precipitation events are observed when the Meiyu rainband is located on the warm side of the SST front (May 21–June 19). In the post-front stage (June 25–July 24), the rainband moves to the cold side of the front. During this stage, rain intensity does not change significantly but occurrence frequency decreases, resulting in a reduction in the strength of climatological rainband. During the period of the rainband moving from the warm to the cold sides of the front, surface wind convergence over the KC diminishes. In contrast, large-scale environmental conditions such as precipitable water and atmospheric stratification become more conducive to precipitation, emphasizing the role of SST-induced surface wind convergence.

Heating of the atmosphere by the ocean is largely determined by the temperature difference between them. Since the atmosphere exhibits a large diurnal variation of temperature due to its relatively low specific heat, it is expected that heating of the atmosphere by the ocean and corresponding response will also show a significant diurnal change. In fact, precipitation and high clouds over the KC and the Gulf Streams (GS) exhibit substantial diurnal variation during the Northern Hemisphere rainy season (Minobe and Takebayashi 2018). Diurnal variation of precipitation in the region of WBC is higher than that in the

112 Intertropical Convergence Zone or South Pacific Convergence Zone. Features of
113 diurnal variation over the two WBC regions are slightly different, possibly due to
114 the presence of the MB rainband over the KC. A precipitation maximum is seen
115 from early to late morning over the GS, whereas precipitation peaks later over the
116 KC. The relative amplitude of the diurnal cycle over the KC compared to the mean
117 is smaller due to the MB rainband than that over the GS. Monthly change of the
118 diurnal cycle is also distinct. The diurnal component is persistent throughout the
119 summer over the GS, but it is prominent only in June and weak in May and July
120 over the KC. The role of SST to the diurnal cycle over the KC is emphasized, since
121 the area of large diurnal range of precipitation is confined to the KC even though
122 the MB rainband is much broader.

123 Despite the interesting findings earlier, mechanism of the intensified
124 diurnal cycle of precipitation over the KC during early summer remains unclear.
125 This study attempts to explain the intensified diurnal cycle in terms of frontal
126 circulation and convective process specifically when the climatological positions of
127 the two frontal systems (i.e., the KC and the MBF) overlap. For that purpose,
128 precipitation is decomposed into the convective and the large-scale components
129 with distinct diurnal variations. The rest of the paper is organized as follows. In
130 Section 2, the data used and the equation for frontogenesis will be addressed.
131 Physical mechanisms responsible for the diurnal variation of the convective and
132 the large-scale precipitation will be described and discussed in Section 3, followed
133 by a summary in Section 4.

2. Data and Method of Analysis

2.1. Data

To evaluate the diurnal cycle of rain, 3-hourly precipitation on a $0.25^\circ \times 0.25^\circ$ resolution from the Tropical Rainfall Measuring Mission (TRMM) 3B42 product is used. The optimally interpolated SST version 2.1 product (OISST v2.1) archived at National Centers for Environmental Information, National Oceanic and Atmospheric Administration (<https://www.ncei.noaa.gov/oisst>) is also used. The spatial resolution of the OISST is $0.25^\circ \times 0.25^\circ$ with a daily time step. Analysis of the atmospheric fields is conducted using the latest reanalysis data, ERA5 (Hersbach et al. 2020), provided by the European Centre for Medium-Range Weather Forecasts (ECMWF). The ERA5 reanalysis product is provided at an hourly interval with a spatial resolution of $0.25^\circ \times 0.25^\circ$ at 37 pressure levels. Several improvements over the ERA-interim product are detailed in Hersbach and Dee (2016). The most significant improvement for our study is that the horizontal resolution is high enough to resolve mesoscale oceanic features. Precipitation data from the ERA5 product is also used after a comparison with the TRMM 3B42 data in order to classify precipitation into convective and large-scale types. East Asia (109° – 145° E, 20° – 45° N; area shown in Fig. 1a) is the target domain of analysis during the 21 years of summer from 1998 to 2018, when the TRMM dataset is available.

2.2. Method

The frontogenesis (FG) function is used to investigate diurnal variation of the MBF. A strong gradient of specific humidity characterizes the MBF zone. Temperature

gradient is relatively weak due to cooling in the frontal precipitation zone (Ninomiya 2004). Therefore, using equivalent potential temperature is more appropriate than potential temperature to define the front. Thus, the equation of FG suggested by Ninomiya (1984) is used. It can be written as follows:

$$FG \equiv \frac{d|\nabla\theta_e|^2}{dt} = FG1 + FG2 + FG3 + FG4 + FG5, \quad (1)$$

where

$$FG1 = 2\nabla\theta_e \cdot \nabla J^*, \quad (2a)$$

$$FG2 = -|\nabla\theta_e|^2 \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right), \quad (2b)$$

$$FG3 = \left\{ \left(\frac{\partial\theta_e}{\partial x} \right)^2 - \left(\frac{\partial\theta_e}{\partial y} \right)^2 \right\} \times \left(\frac{\partial u}{\partial x} - \frac{\partial v}{\partial y} \right), \quad (2c)$$

$$FG4 = \left\{ -2 \frac{\partial\theta_e}{\partial x} \frac{\partial\theta_e}{\partial y} \right\} \times \left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right), \quad (2d)$$

$$FG5 = -2 \frac{\partial\theta_e}{\partial p} \left(\frac{\partial\theta_e}{\partial x} \frac{\partial\omega}{\partial x} + \frac{\partial\theta_e}{\partial y} \frac{\partial\omega}{\partial y} \right). \quad (2e)$$

Here, θ_e represents equivalent potential temperature. The FG equation in Ninomiya (1984) is multiplied with $|\nabla\theta_e|$ to come up with the final equation in (1). $J^*(\equiv D\theta_e/Dt)$ is diabatic heating excluding latent heat release so that J^* does not account for the contribution of heating due to condensation. Near the surface over the KC (say, 975 hPa), diabatic heating is primarily in the form of sensible heating, part of which is offset by a radiative process (Sasaki et al. 2012). Therefore, $FG1$ term near the surface can be understood as the effect of direct heating by the KC. Frontogenesis by surface wind convergence produced by the KC is manifested in $FG2$, which will be called the indirect effect of the KC. $FG3$ and $FG4$ terms represent respectively the effect of stretching deformation (confluence) and

179 shearing deformation. These deformation terms are usually most prominent in
180 large-scale processes (Ninomiya 1984). *FG5* term denotes the tilting effect of the
181 front. The tilting effect tends to oppose changes due to deformations to maintain
182 the thermal wind balance (Lackmann 2012).

3. Results and Discussion

Shading in Fig. 1a depicts the TRMM precipitation (TRMM PR) pattern averaged in June 7–23. This averaging period is determined by selecting when precipitation is particularly intense over the KC in a climatological sense (see Fig. 1c). Extending the target period to the entire June or shortening to 10 days centered on June 15, however, does not significantly affect our main conclusions. The primary reason for the intense precipitation that extends from southeast China to the south of Japan is the large-scale environmental forcing, as analyzed in several previous studies (e.g., Sampe and Xie 2009; Horinouchi 2014). However, high-resolution TRMM PR data (Fig. 1a) shows that the precipitation pattern is not uniformly distributed along the southern flank of the upper-level jet entrance where large-scale upward motions are expected but rather elongated along the narrow KC warm tongue.

Figure 1c shows the sub-seasonal variation of precipitation and the MBF along the lines crossing the KC. This result is obtained by interpolating variables along each cross frontal line (dashed lines in Fig. 1a) as a function of time and then taking an average of across the lines. The ordinate of Fig. 1c and d represents the latitude of the thick dashed line (Fig. 1a) as a reference. The red semi-transparent shading depicts the location of the atmospheric front defined by the gradient of equivalent potential temperature exceeding 5 K/100 km. During early summer, location of precipitation band migrates northward with the front, crossing the KC in mid-June (semi-transparent red shading in Fig. 1c). Mid-tropospheric warm advection or Q-vector convergence, which represents a large-scale forcing, explains the seasonal migration of upward motion and, henceforth, precipitation

following the front (Xu et al. 2018; Kim and Kim 2019). These environmental factors, however, do not fully explain the intensity of precipitation, particularly in mid-June, during which precipitation intensity is notably stronger than in other periods. Mid-tropospheric warm advection is intensified in early July (Xu et al. 2018), and Q-vector convergence at the end of June (Kim and Kim 2019), both of which develop later than the intense precipitation. Surface wind convergence, on the other hand, has a direct link with heating by the KC and covariates with the precipitation (blue contours in Fig. 1c). Therefore, the effect of the KC must also be considered to understand the complete picture of the Meiyu-Baiu precipitation.

Figure 1b and d shows the ERA5 total precipitation (TP) for the same period. Precipitation patterns in Southeast China, East China, Taiwan, and the East China Sea in the TRMM data are reasonably reproduced in the ERA5 product. In contrast, noticeable difference is seen near Kyushu; TRMM PR shows precipitation around Kyushu, but ERA5 TP is shifted slightly to the north. Despite some differences, the two datasets commonly show localized peaks over the KC, although TP significantly underestimates TRMM PR. Also, the seasonal march of the precipitation zone during June seems adequately represented (Fig. 1d).

Figure 2 shows the diurnal cycle of rainfall (shading) and the temperature difference between the sea surface and surface (10 m) air (TDF; contours) averaged during the target period (June 7–23) along the cross frontal lines in Fig. 1a. With respect to the latitudinal position, TDF is maximized over the KC and shows positive (ocean is warmer) values to the south and negative (ocean is colder) values to the north. TDF is maximized over the KC at 09 LST, because atmospheric temperature is significantly low in the morning while ocean temperature remains

nearly the same during the course of a day. TRMM PR over the KC gradually increases from 03 LST and shows a maximum around the noontime and then decreases (Fig. 2a). Precipitation signal starts to migrate at 06 LST from the north to the south of the KC in the afternoon. This southeastward migration signal is consistent with that in Minobe and Takebayashi (2015). The daily change of TP (Fig. 2b) over the KC shows a similar pattern to the TRMM PR in that it gradually intensifies from 03 LST and exhibits a weak precipitation pattern near midnight. While the southeastward movement can also be found in the TP, its speed is slower than that of the TRMM PR, taking three hours longer to reach the KC.

Although the southeastward propagation speed of TP is slightly lower and the intensity over the KC is somewhat underestimated, it seems that the diurnal variation of the TP is reasonable compared to that of the TRMM PR. Dividing TP into the convective (CP) and the large-scale (LSP) components yields a better understanding of the physical processes contributing to precipitation. The diurnal variation of CP over the KC (Fig. 2c) is different from that of LSP (Fig. 2d). Since the diurnal variation of CP is in phase with TDF, it seems that the air parcel near the surface gains buoyancy due to intense direct heating from the warm KC, resulting in increased convective instability and ultimately CP. A more detailed statistical relationship between the two variables will be shown after the diurnal variation of LSP is described. LSP, which lags TDF by about 6 hours, accounts for more precipitation than CP. The southeastward propagating signal is also explained primarily by LSP. Therefore, the intense precipitation that starts in the morning hours (Fig. 2a and d) seems associated with atmospheric circulation rather than a

convective process. Two types of precipitation are analyzed separately since their mechanisms seem different.

The probability distribution of CP near the KC is illustrated in Fig. 3. This joint probability distribution function (PDF) is obtained as follows. First, to minimize the effect of synoptic disturbances, which depends more on the propagation of individual systems than heating from the SST warm tongue, CP and TDF are averaged over the target period (June 7–23) for every three hours (the temporal resolution of TRMM PR) each year. Then, 136 (8 times $\times$ 17 years) samples are obtained for the entire period at each of the 1087 grid points in the red box in Fig. 1b. As a result, a total of 147,832 pairs of CP and TDF are produced and are binned at 0.016 mm hr^{-1} and 0.07 K interval, respectively (see Fig. 3). Since the KC warm tongue area is set as the target, average TDF is $\sim 0.4 \text{ K}$, meaning that heat flux, on average, is from the ocean to the atmosphere.

Figure 3 shows that CP tends to intensify as TDF increases. Both the upper and lower 10 percentiles and the median of CP increase almost linearly with TDF when TDF is in the range of -0.5 K and 1.2 K . TDF lower than -0.5 K does not result in much weaker precipitation. When TDF is higher than 1.7 K , CP tends to decrease. The upper 10% line is steeper than the others, indicating that not only the mean amount of CP but also the probability of stronger CP are enhanced as TDF increases. To examine the diurnal dependency of CP, joint PDFs are obtained by separating the morning (03–12 LST) and afternoon (15–00 LST) hours. The color contours in Fig. 3 show the differences between the two PDFs. The red contours show that CP are more intense in the morning and the blue contours depict a weakening of CP in the afternoon; the joint PDFs indicate a clear difference in the

CP intensity between the morning and the afternoon hours. Also, intensity of CP in the morning tends to be more widely spread with increased mean and extreme values compared with that in the afternoon.

The oceanic front is known to cause surface wind convergence through the mechanisms mentioned earlier: PAM and VMM. During June, PAM is a dominant mechanism causing surface wind convergence above the KC (Sasaki et al. 2012; Xu et al. 2018). VMM is not a primary cause of surface wind convergence in this season, since convergence is produced over the region of strongest SST gradient slightly off the KC warm tongue (Liu et al. 2013). Meanwhile, PAM develops localized surface low-pressure just above the KC warm tongue, which results in wind convergence by friction in the marine atmospheric boundary layer (MABL). Correlation between wind convergence and pressure can be shown by applying the Laplacian operator to surface pressure, which acts as a high-pass filter removing large-scale structures such as monsoon or synoptic storms (Minobe et al. 2008).

Figure 4 shows the daily variation of (a) 1000-hPa wind convergence and (b) the Laplacian of SLP. The overall structure of the diurnal variation of surface wind convergence resembles that of CP with a maximum at 09 LST, which is coherent with the enhanced surface heating and is nearly twice the amount of CP at 18 LST. Surface wind over the KC responds quickly to changes in heating from the surface. Southeastward propagation of the core is also seen in wind convergence. The diurnal variation of the Laplacian of SLP seems to partly explain surface wind convergence above the KC but no propagation of the core is observed. Thus, it can be interpreted that low pressure develops above the KC in response to enhanced

surface heating during the early morning. Then, surface wind convergence due to friction in the MABL can mechanically raise air triggering condensation.

The interpretation above based on PAM is not entirely satisfactory for the following reasons. Theoretical models for PAM suggested by Minobe et al. (2008) and others (e.g., Takatama et al. 2012; Takatama et al. 2015) assume stationarity. In reality, however, pressure gradient, Coriolis force, and frictional force will change in time within the MABL even if all other physical processes are neglected. This model is adequate when it is applied to fields averaged over a month or longer (Takatama et al. 2015). Further, the relative amplitude of the diurnal cycle of surface wind convergence is stronger than that of SLP Laplacian, which implies that other mechanisms, possibly including moist process, should also be considered. If PAM, instead of convective instability, is the leading cause of surface wind convergence and the onset of precipitation, LSP rather than CP is expected to correlate closely with surface wind convergence. This, however, is not true because the diurnal variation of wind convergence is strongly in phase with CP but not with LSP (Fig. 2). As will be described later, variation of LSP is found to coincide with front and frontal circulation rather than convective instability. Thus, surface wind convergence over the KC seems to have occurred to compensate for the rising air due to convective instability rather than friction associated with cyclonic circulation. It is suggested that surface wind convergence over the KC represents a response to convection rather than serving as a mechanical cause for precipitation, at least during the morning hours.

The two precipitation types also show markedly distinct spatial patterns (see supplementary Fig. S1). Contrary to CP, which shows a local maximum over

the KC warm tongue, the distribution of LSP is much broader along the MBF. As shown in Fig. 2d, LSP explains a larger fraction of the MB precipitation than CP does. In order to understand total precipitation, therefore, it is necessary to analyze the diurnal variation of LSP in addition to that of CP. The intensity of the front is chosen as a crucial variable explaining the diurnal variation of LSP. Many previous studies explained precipitation through cross-frontal ageostrophic circulation accompanying a large-scale atmospheric motion (e.g., Ninomiya 1984, 2004). There are several recent studies showing an intensification of the front by the WBC (e.g., Parfitt et al. 2016; Reeder et al. 2021). Therefore, diurnal variation of the frontal strength, frontogenesis, and ageostrophic circulation are examined here.

Figure 5 shows the diurnal variation of the front intensity and the *FGs* at 975 hPa (see the Method section for definitions). During mid-June, the MBF is located about 3 degrees north of the KC (Fig. 5a, and shading in Fig. 1c). The strength of the front is not uniform throughout the day but is maximized at 15 LST; this is coherent with the diurnal variation of LSP. An intense front accompanies ageostrophic circulation with an upward motion to the south of the front. *FG1* is maximized about three hours earlier than the front at the same latitude. Quantitatively, however, *FG2* and *FG3* explain a much larger fraction of frontal variation than the diabatic component (*FG1*). Frontogenesis driven by *FG2* and *FG3* continues until the frontal intensity is maximized in the afternoon (Fig. 5c and d). It should be noted that *FG2* and *FG3* are located about 2–300 km south of the front. Since the equation for the tendency of the frontal intensity is given in the form of the total derivative, frontogenesis terms do not coincide directly with the local

change in frontal intensity, as shown by a comparison between Fig. 5a and Fig. 5c or 5d. Distance between the location of frontogenesis and the front (see Fig. 5) seems to be due to advection by prevailing southwesterlies, which is the main direction of surface wind during mid-June. *FG2* is the product of wind convergence and frontal intensity. The local maximum of *FG2* results from the enhanced surface wind convergence in the morning (Fig. 4a), which in turn is the imprint of enhanced heating over the KC. Thus, adiabatic process (*FG2*) is at least as important as diabatic process (*FG1*) in the discussion of the influence of the KC on the atmospheric frontogenesis. *FG3* also shows a daily variation similar to that of *FG2*. *FG3* is mainly controlled by the confluent deformation (see also supplementary Fig. S2), which shows a diurnal variation similar to that of surface wind convergence.

Shading in Fig. 6 shows the vertical structure of anomalous patterns of front intensity (Fig. 6a) and frontogenesis functions averaged along the cross frontal lines (Fig. 6b-f) in the morning hours (06–09 LST). Contours in Fig. 6a represent anomalous cloud cover (CC). Wind vectors represent ageostrophic circulation with the vertical component exaggerated by a factor of 20. Anomalous patterns are obtained by subtracting daily averages. The gray shading shows the location of the KC warm tongue. In the morning, the location of the positive front anomaly at the 700-hPa level is about 2° north of the KC. Ageostrophic circulation develops around the positive anomaly located around 28.5°N, with a corresponding upward motion to the south. Local CC maxima are found at three distinct levels: just above the KC, south of the front, and further north of the KC near the surface. The local peak of CC above the KC appears to be the result of convection caused by direct

heating from the surface, which matches the local maximum of CP shown in Fig. 2c. The lower-tropospheric cloud to the north of the KC seems like fog generated by radiative cooling in the morning. This cloud covers a large area but does not accompany precipitation (Fig. 2a). Cloud located to the south of the front appears to be caused by the frontal circulation. The altitude of this cloud anomaly is higher than those at the other two locations. The TRMM PR (Fig. 2a) and LSP (Fig. 2d), located further north of the KC in the morning, appear to be caused by this frontal circulation.

The vertical structure shows that the contribution of *FG2* to frontogenesis is most significant, followed by *FG3*. *FG2* shows a peak below 900 hPa just above the KC, which is apparently due to low-level wind convergence arising from a direct influence of the KC as discussed earlier. Near 29°N, north of the KC, the vertical structure of *FG2* is characterized by frontogenesis at 850 hPa and frontolysis at 500 hPa. They are associated respectively with convergence and divergence of wind, which occurs due to the mid-tropospheric ascending branch of the frontal circulation to satisfy the mass continuity. Similar to *FG2*, *FG3* also results in frontogenesis at a slightly higher latitude than the KC. Meanwhile, *FG1* causes frontolysis despite that diabatic heating from the surface is vigorous in the morning, which indicates that the directions of gradient of equivalent temperature ($\nabla\theta_e$) and diabatic heating (∇J^*) are opposite to each other. The *FG4* is smallest in magnitude. *FG5* causes frontolysis near the positive front anomaly. The negative relation between *FG5* and the front can also be found in the climatological mean (Fig. S3).

397 In the afternoon (Fig. 7; 12–15 LST), vertical distribution of the frontal
398 activity clearly shows the effect of frontogenesis in the morning. The strength of
399 the front intensifies compared to the morning, showing positive anomalies from
400 the surface to the upper troposphere. The upward branch of the ageostrophic
401 circulation moves over the KC with a higher altitude, making a taller cloud than in
402 the morning (red contours in Fig. 7a). Although the height of the cloud is tall,
403 precipitation at this time is classified as LSP (Fig. 5d) rather than CP (Fig. 5c),
404 because the role of frontal circulation is more important than convective process
405 near the surface. Besides, the fact that the afternoon TDF is weaker than in the
406 morning supports that this deep cloud is triggered by a dynamic forcing. To the
407 north of the KC, fog near the surface seems to have disappeared by strong solar
408 insolation in the afternoon. Although *FG1*, *FG2*, and *FG3* all contribute to the
409 frontogenesis, frontolysis by *FG5* is more significant in the afternoon. Thus, the
410 intensity of the front weakens gradually until midnight.

4. Summary

In this study, causes of the strong diurnal cycle of precipitation over the KC in mid-June were analyzed. Precipitation peaks in the afternoon, as addressed in many earlier studies. Heating from the KC, however, is maximized during the morning. This temporal gap was explained by separating the convective and large-scale components of ERA5 total precipitation. Although there are some differences from the satellite observation (TRMM PR), the ERA5 product reasonably reproduces the essential features of precipitation such as the localized peak over the KC, afternoon maximization, and the seasonal migration of the Meiyu-Baiu rainband.

The convective component (CP) is in phase with temperature difference between the sea surface and the atmosphere 10 m above (TDF), which is explained by the intensified convective instability triggered by enhanced heating from the KC during the morning hours (03–12 LST). The intensity of CP in the morning tends to be more widely distributed with increased mean and extreme values compared to that in the afternoon. The enhanced surface wind convergence is also observed during the morning hours, which sustains the convection. The diurnal cycle of large-scale component (LSP), however, is quite different from that of TDF, and shows a peak in the afternoon. The vertical structure near the front reveals that the diurnal cycle and the location of LSP are well explained in terms of the ageostrophic circulation encircling the front in the meridional direction. An intense front and an accompanying upward motion to the south, which trigger LSP, are observed during the afternoon. Calculation of the frontogenetic functions reveals that the contribution of the adiabatic terms ($FG2-FG5$) exceeds the diabatic term ($FG1$). In particular, terms associated with wind convergence ($FG2$) and shear

435 deformation (*FG3*) show the largest contributions. Positive *FG2* anomaly develops
436 from the early morning until the afternoon, when the front intensity is maximized.
437 The vertical structure of *FG2* shows a strong signal, which is confined below 850
438 hPa over the KC. The spatio-temporal distribution of *FG2* is dominated by the near-
439 surface wind convergence, which is an imprint of the enhanced heating in the
440 morning over the KC.

441 This study focused on the diurnal cycle of precipitation over the KC.
442 However, the Gulf Stream (GS) is another region of a strong diurnal signal. The
443 diurnal cycle over the GS peaks several hours earlier than the KC (Minobe and
444 Takebayashi 2015). Hence, the lag between TDF and precipitation is expected to be
445 smaller. Since a salient front is not seen over the GS, it is suggested that the MBF,
446 which takes several hours to intensify fully, is a possible explanation for the
447 difference.

448 ***Data Availability***

449 The data (ERA5) employed in the present study is publicly available at the ECMWF
450 website (ecmwf.int).

451

452 ***Acknowledgments***

453 This work was supported by the National Research Foundation of Korea grant
454 funded by the Korean government (NRF-2020R1A2B5B01002600).

455

456 ***Author Contributions***

457 HJP conceived the idea in this study. HJP and KYK carried out the analysis, interpreted the
458 results, and wrote the manuscript.

459

460 ***Additional Information***

461 ***Competing Interests:*** The authors declare no competing interests.

462 ***Ethical Approval:*** We do not have any experiments involving animals.

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Figure Captions

Figure 1. Shading in the top row represents the pattern of (a) TRMM PR and (b) TP averaged from June 7 to 23. Black and blue contours represent respectively SST (K) and 200-hPa zonal wind (at the contour levels 30, 33, 36, and 39 m s^{-1}). Shading in (c) and (d) shows sub-seasonal variation of the TRMM PR and TP along the dashed lines crossing the KC in Fig. 1a. The red semi-transparent shading represents the location of atmospheric front defined by the gradient of equivalent potential temperature exceeding 5 K/100 km. Blue contours in (c) represent 1000-hPa wind convergence (at $2 \times 10^{-6} \text{ s}^{-1}$ interval from $\pm 1 \times 10^{-6} \text{ s}^{-1}$) with dashed contours for negative and solid contours for positive values.

Figure 2. The diurnal cycle (shading; mm hr^{-1}) of (a) TRMM PR, (b) TP, (c) CP, and (d) LSP during the target period (June 7–23) along the cross frontal lines shown in Fig 1a. Red contours represent temperature difference (K) between the sea surface and 10 m air above it.

Figure 3. The joint PDF of CP and TDF near the KC (shading). The thin black lines represent the 10 and 90 percentiles of CP with the median in thick black line for each TDF bin. The color contours represent differences in the joint PDF between the morning (03–12 LST) and the afternoon (15–00 LST). The red contours show that the cases are more frequent in the morning and the blue contours in the afternoon. Details for the calculation of the joint PDF can be found in the main text.

Figure 4. Diurnal cycle (shading) of (a) 1000-hPa wind convergence (10^{-6} s^{-1}), and (b) Laplacian of sea level pressure ($10^{-10} \text{ Pa m}^{-2}$). Red contours show temperature difference (K) between the sea surface and the air 10 m above it.

Figure 5. (a) The diurnal cycle of $|\nabla\theta_e|$ along the cross frontal lines (shading). Unit for $|\nabla\theta_e|$ is 10^5 K m^{-1} (upper left color bar). The diurnal cycle of (b) $FG1$, (c) $FG2$, (d) $FG3$, (e) $FG4$, and (f) $FG5$ terms of the frontogenesis functions with the unit of $10^{14} \text{ K}^2 \text{ m}^{-2} \text{ s}^{-1}$ (upper right color bar). All the panels are for the 975-hPa level. Definition for each term can be found in the Method section of the text. Red contours represent temperature difference (K) between the sea surface and 10 m air above it.

Figure 6. The vertical structure (shading) of anomalous patterns of (a) $|\nabla\theta_e|$ (in units of 10^5 K m^{-1} ; upper left color bar) and (b)–(f) individual terms ($FG1$ – $FG5$) of the frontogenesis function (in units of $10^{14} \text{ K}^2 \text{ m}^{-2} \text{ s}^{-1}$; upper right color bar) averaged along the cross frontal lines (Fig. 1a) in the morning (06–09 LST). Contours in panel (a) represent anomalous cloud cover (CC) at the interval of 1% from $\pm 1\%$ with blue contours for negative values and red contours for positive values. Wind vectors describe the ageostrophic circulation with the vertical component exaggerated by a factor of 20. Anomalies are defined as departures from corresponding daily averages. The gray shading denotes the location of the KC warm tongue.

Figure 7. Same as Fig. 6, except that the patterns represent the anomalous conditions in the afternoon (12–15 LST).

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