Roles of Rossby Waves, Rossby–Gravity Waves, and Gravity Waves Generated in the Middle Atmosphere for Interhemispheric Coupling

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ABSTRACT: It has often been reported that warming at high latitudes in the Southern Hemisphere (SH) summer mesosphere and lower thermosphere (MLT) appears during Arctic sudden stratospheric warming (SSW) events. This phenomenon, which is called “interhemispheric coupling” (IHC), has been thought to occur because of the modulation of mesospheric meridional circulation driven by forcing of gravity waves (GWs) originating in the troposphere. However, quasi-two-day waves (QTDWs) develop during SSWs and result in strong wave forcing in the SH mesosphere. Thus, this study revisits IHC following Arctic SSWs from the viewpoint of wave forcing, not only by GWs and Rossby waves (RWs) originating in the troposphere but also by GWs, RWs, and Rossby–gravity waves generated in situ in the middle atmosphere, and elucidates the causes of warm anomalies in the SH MLT region. During SSWs, westward wind anomaly forms because of cold equatorial stratosphere, GW forcing is then modulated, and barotropic/baroclinic and shear instabilities are strengthened in the SH mesosphere. These instabilities generate QTDWs and GWs, respectively, which cause significant anomalous westward wave forcing, forming a warm anomaly in the SH MLT region. The intraseasonal variation in QTDW activity can explain seasonal dependence of the time lag in IHC. Moreover, it is revealed that the cold equatorial stratosphere is formed by middle-atmosphere Hadley circulation, which is strengthened by wave forcing associated with stationary RW breaking leading to SSWs. The IHC mechanism revealed in this study indicates that waves generated in the middle atmosphere contribute significantly to the meridional circulation, especially during SSWs.

KEYWORDS: Atmospheric circulation; Gravity waves; Rossby waves; Teleconnections; Middle atmosphere

1. Introduction

Sudden stratospheric warming (SSW) events often occur in the polar region in Northern Hemisphere (NH) winter. This is a phenomenon in which a meridional circulation in the stratosphere is driven by stationary Rossby wave (RW) forcing, resulting in adiabatic heating in the polar stratosphere and adiabatic cooling in the equatorial stratosphere (e.g., Matsuno 1971). Interhemispheric coupling (IHC) of the middle atmosphere, which is described by an anticorrelation between the temperature anomaly in the winter stratosphere and that in the summer mesopause, was first reported by Becker et al. (2004). Gumbel and Karlsson (2011) showed a strong anticorrelation between the winter stratospheric temperature anomaly and the occurrence of summer mesosphere noctilucent clouds. Karlsson et al. (2009b) showed that the noctilucent cloud occurrence frequency in the Southern Hemispheric (SH) mesosphere responded to NH zonal winds at 60°S and 5 hPa with delays of +2 and +7 days in December 2007 and January 2008, respectively.

Additionally, multiple numerical simulations have been performed to examine IHC (e.g., Becker and Fritts 2006;

Karlsson et al. 2009a; Smith et al. 2020). Körnich and Becker (2010, hereafter referred to as KB10) proposed an IHC mechanism in which modification of the meridional circulation of the mesosphere in both hemispheres is driven by gravity wave (GW) forcing. This scenario argues the following. 1) A westward RW forcing anomaly occurs in the winter stratosphere, resulting in a westward wind anomaly. 2) More GWs having eastward phase velocities propagate from the winter troposphere into the winter mesosphere, and the total westward GW forcing in the mesosphere is reduced. Thus, the GW forcing anomaly is eastward and causes an equatorward meridional circulation anomaly (i.e., the circulation toward the North Pole weakens) in the winter hemisphere. This anomalous circulation forms a warm anomaly in the tropical mesosphere. In the summer mesosphere, an eastward wind anomaly appears responding to the tropical warm anomaly via the thermal wind balance. 3) The eastward wind anomaly (i.e., the weakened westward wind) induces a downward shift in the eastward GW forcing, which causes a westward GW forcing anomaly at the original level of the eastward GW forcing. As a consequence, the meridional circulation anomaly in the upper mesosphere is poleward (i.e., the circulation toward the equator weakens) and induces a positive temperature anomaly at the winter pole side of the westward GW forcing anomaly.

This scenario was confirmed using an axisymmetric version of the Kühlungsborn Mechanistic General Circulation Model (KMCM) with GW parameterizations. However, it appears that the altitude of the warm anomaly in the SH mesosphere and lower-thermosphere (MLT) region simulated by KMCM

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is lower than that in the real atmosphere and the response time lag simulated by KMCM (~ 4 days) is shorter than the observations (~ 5 – 10 days; [Karlsson et al. 2009a](#)). Conversely, [Smith et al. \(2020\)](#) argued that wave forcing in the SH is not necessarily important for IHC and the response in the SH is due to circulation caused by the zonal-mean mass balance of the atmosphere.

The residual-mean meridional circulation of the middle atmosphere has been thought to be driven mainly by waves originating in the troposphere. However, it has recently been shown that the forcing of waves generated in the middle atmosphere is important in the MLT region. (e.g., [McLandress et al. 2006](#); [Sato and Nomoto 2015](#); [Becker and Vadas 2018](#)). The most remarkable waves in the summer MLT region are quasi-two-day waves (QTDWs), which are identified as westward-propagating Rossby-gravity normal modes with zonal wavenumbers of 2–4 and wave periods of 1.5–2.5 days (e.g., [Ern et al. 2013](#)). The seasonality of QTDWs is explained by barotropic (BT) and/or baroclinic (BC) instabilities of the summer westward jet (e.g., [Plumb 1983](#)). QTDWs play the second largest role after GWs in the momentum budget in the summer MLT region. QTDWs are radiated from regions with negative latitudinal gradients of the potential vorticity (PV) in the summer mesosphere, which is a necessary condition for BT/BC instabilities ([Sato et al. 2018](#), hereafter referred to as [SYM18](#)). This means that QTDWs are generated in situ by BT/BC instabilities in the summer mesosphere. Moreover, [SYM18](#) showed that these BT/BC instabilities are caused by the forcing of GWs originating in the troposphere.

Additionally, it was reported that QTDWs in the summer MLT region are amplified when an SSW occurs in the winter stratosphere. [France et al. \(2018\)](#) showed that BC instability is enhanced by strengthened westward jet in the summer stratosphere and mesosphere during an SSW. The westward forcing of the developed QTDWs weakens the residual-mean equatorward flow and causes a warm anomaly near the polar summer mesopause.

Moreover, it has been shown that secondary GW forcing in the mesosphere is also significant. Using the high-resolution KMCM, [Becker and Vadas \(2018\)](#) showed that the primary orographic GW forcing generates secondary GWs in the winter mesosphere. The secondary GWs can propagate into the lower thermosphere and break at high latitudes in the winter hemisphere. [Yasui et al. \(2018\)](#), hereafter referred to as [YSM18](#) also showed that westward-propagating GWs in the summer MLT region are likely generated in situ via shear instability. The shear instability is likely formed by primary GW forcing.

For these reasons, secondary GWs generated in the MLT region may contribute to the formation of warm anomalies in the summer MLT region, similar to QTDWs. In addition, the QTDW generation mechanisms during SSW events have not been examined in detail.

Therefore, the purpose of this study is to elucidate the roles of wave forcings due to RWs, Rossby-gravity waves (RGWs), and GWs generated in the middle atmosphere in IHC formation mechanisms associated with NH SSWs. This study is organized as follows. [Section 2](#) contains descriptions of the output data from the Ground-to-Topside Model of

Atmosphere and Ionosphere for Aeronomy (GAIA; [Jin et al. 2011](#)) and Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2; [Gelaro et al. 2017](#)), datasets used in this study. The analysis methods are also given. In [section 3](#), the zonal-mean temperature and wave forcing anomalies of each wave type and the causes of the wave forcing anomalies are examined. Additionally, the roles of these wave forcings are discussed. A summary is given in [section 4](#).

2. Data and analysis methods

a. Data description

1) GAIA SIMULATION DATA

GAIA is a whole atmosphere model that extends from the surface to the thermosphere/ionosphere ($z \sim 3000$ km), which is composed of a general circulation model (GCM), an ionospheric model, and an electrodynamical model ([Jin et al. 2011](#)). Outputs from the GCM part of GAIA ([Miyoshi and Fujiwara 2003](#)) are used in this study. The GCM model resolution is T42L150 corresponding to a horizontal grid of approximately 2.8° and a vertical grid of 0.2 scale height (~ 1.4 km in the middle atmosphere) for the altitude range from the surface to $z = \sim 600$ km (1.017×10^{-9} hPa). The time interval of the data is 1 h. The analyzed periods are 19 boreal winter (December–February) seasons of 1996–2015. This model is nudged using Japanese 25-year Reanalysis Project (JRA-25)/Japan Meteorological Agency (JMA) Climate Data Assimilation System (JCDAS) data ([Onogi et al. 2007](#)) with a relaxation time $\tau_{\text{nudge}} = 1$ day below approximately 30 km (12 hPa) in order to realistically simulate the quasi-biennial oscillation (QBO) in the equatorial lower stratosphere and planetary-scale RWs and tides originating from the troposphere. Moreover, solar ultraviolet (UV) and extreme UV (EUV) radiation processes are included using a parameterization by [Strobel \(1978\)](#) and using a EUV flux model for aeronomic calculation (EUVAC) ([Richards et al. 1994](#)), respectively. These parameterizations estimate the UV/EUV fluxes from daily values of the 10.7 cm solar radio flux (F10.7). The stratospheric O_3 mixing ratio is set to its climatological value at each grid point. The model also uses GW parameterizations by [McFarlane \(1987\)](#) and [Lindzen \(1981\)](#), respectively, for orographic and nonorographic GWs similar to [Garcia et al. \(2007\)](#). The parameterized nonorographic GWs are launched at the ~ 200 -hPa level. The phase speeds of the launched GWs are -30 to 30 m s^{-1} at an interval of 10 m s^{-1} . The GW source stress spectrum is specified as a Gaussian, where $\tau_b^* = 6.4 \times 10^{-4}$ hPa for (A21) of [Garcia et al. \(2007\)](#). Note that these GW parameterizations are different from those used in [KB10](#). Details regarding the model configuration are given in [SYM18](#).

2) MERRA-2 DATA

The MERRA-2 dataset is used to validate the GAIA data in the stratosphere and lower mesosphere in [section 3c](#). The MERRA-2 data are distributed on 42 vertical layers with an altitude range from the surface to 0.1 hPa, the horizontal grid has a resolution of 0.625° longitude $\times$ 0.5° latitude, and the time

interval is 3 h. The analyzed periods are the 28 boreal winter (December–February) seasons of 1990–2018.

3) AURA MLS OBSERVATION DATA

As the GAIA data are the outputs of numerical simulation, it is necessary to validate the reality. In this study, temperature and geopotential height data from the Earth Observation System *Aura* Microwave Limb Sounder (MLS), version 4.2, level 2 (Waters et al. 2006; Schwartz et al. 2008), are used for the validation of GAIA simulation. The analyzed height is from $z = 9.4$ km (261 hPa) to 97 km (1×10^{-3} hPa). The analyzed periods are the 14 boreal winter (December–February) seasons of 2004–18. The horizontal and vertical resolutions depend on the height, that is, respectively, 170 and 5 km at $z = 9.4$ and 220 km and 10–13 km at $z = 97$ km. It is known that the temperature data from *Aura* MLS have a cold bias of ~ 1 K in the upper troposphere and of ~ 10 K in the mesopause (Schwartz et al. 2008; Medvedeva et al. 2014).

b. Analysis methods

1) CLASSIFICATION OF ROSSBY WAVES AND ROSSBY–GRAVITY WAVES, GRAVITY WAVES, AND TIDES

In this study, RW/RGW, GWs, and tidal waves (TWs) were designated using the same method as in SYM18 and the contribution of each wave to IHC was examined. Respective wave disturbances are obtained from the original data by the following methods: first, migrating TWs were extracted using Fourier analysis as waves with zonal wavenumbers and periods (s, τ) = $(-1, 24\text{ h})$, $(-2, 12\text{ h})$, and $(-3, 8\text{ h})$. Second, the remaining components are divided into two: the waves with periods longer than 24 h were designated as RWs/RGWs, and those with wave periods shorter than or equal to 24 h were designated as “resolved” GWs. Note that the nonmigrating TWs are included in the resolved GWs. This is not too special since the nonmigrating TWs can be regarded as planetary-scale inertia–GWs as shown by Sakazaki et al. (2015). In this study, QTDWs are also extracted from the RWs/RGW by Fourier analysis and examined. The extracted QTDWs are the disturbances propagating westward with periods of 1.5–2.5 days and zonal wavenumbers of 2–4.

2) ANALYSIS OF WAVE ACTIVITY FLUX AND WAVE FORCING

The Eliassen–Palm (EP) flux in the transformed-Eulerian-mean (TEM) equation system in log-pressure coordinates was used to examine the momentum budget in the middle atmosphere (e.g., Andrews et al. 1987). The TEM equation of zonal momentum is written as

$$\bar{u}_t + \bar{v}^* \left[\frac{(\bar{u} \cos \phi)_\phi}{a \cos \phi} - f \right] + \bar{w}^* \bar{u}_z = \frac{1}{\rho_0 a \cos \phi} \nabla \cdot \mathbf{F} + \bar{X}, \quad (1)$$

where

$$\bar{v}^* = \bar{v} - \frac{1}{\rho_0} \left(\rho_0 \frac{\bar{v}'\theta'}{\bar{\theta}_z} \right)_z, \quad (2)$$

$$\bar{w}^* = \bar{w} + \frac{1}{a \cos \phi} \left(\cos \phi \frac{\bar{v}'\theta'}{\bar{\theta}_z} \right)_\phi, \quad (3)$$

$$\mathbf{F} = (0, F^{(\phi)}, F^{(z)}) \quad (4a)$$

$$= \rho_0 a \cos \phi \left(0, \bar{u}_z \frac{\bar{v}'\theta'}{\bar{\theta}_z} - \bar{v}'u', \left[f - \frac{(\bar{u} \cos \phi)_\phi}{a \cos \phi} \right] \frac{\bar{v}'\theta'}{\bar{\theta}_z} - \bar{w}'u' \right), \quad (4b)$$

$$\nabla \cdot \mathbf{F} = \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (F^{(\phi)} \cos \phi) + \frac{\partial F^{(z)}}{\partial z}. \quad (5)$$

Here, the overbar denotes the zonal mean, the prime denotes the anomaly from the zonal mean, ϕ is the latitude, $\bar{v}^*$ and $\bar{w}^*$ are the meridional and vertical components of the residual-mean flow, respectively, θ is the potential temperature, z is the log-pressure height, ρ_0 is the reference density, $\mathbf{F}$ is the mechanical forcing such as unspecified horizontal components of friction and/or subgrid-scale GW forcing, and $\mathbf{F}$ is the EP flux. The wave forcing due to the resolved waves is described as the divergence of the EP flux (EPFD).

3) EXTRACTION OF STRONG COLD EQUATORIAL STRATOSPHERE EVENTS

As described later (section 3a), a temperature anomaly response in the equatorial stratosphere appears earlier than the SSW central date as an initial IHC stage. The temperature anomaly that appears in the equatorial stratosphere extends to the midlatitudes of the summer hemisphere. Thus, this study focuses on events in which a strong cold anomaly appears in the equatorial stratosphere (hereafter referred to as a “cold equatorial stratosphere event”). IHC events were extracted as follows: 1) First, the temperature anomaly from the daily climatology, which is determined using GAIA (MERRA-2) data over 19 (28) years, is calculated. A 90-day running mean of the temperature anomaly is calculated and removed from the temperature anomaly to eliminate the possible modulation by the 11-yr solar cycle (i.e., cycle 22–24). This process is important since the 11-yr solar cycle has a significant influence on the thermospheric temperature anomaly. The 27-day cycle of solar variation may also affect the temperature variation. However, it is difficult to distinguish the temperature variations due to the 27-day solar cycle from those due to other mechanisms because the durations of the IHC events and associated warm anomalies in the SH MLT region are on a similar time scale (~ 20 – 25 days) (see Figs. 3 and 5). Thus, we did not exclude the 27-day period component from the temperature anomaly. 2) Next, the period when the temperature anomalies at both $(\phi, z) = (0^\circ\text{N}, 5\text{ hPa})$ and $(20^\circ\text{S}, 5\text{ hPa})$ are lower than twice their standard deviation ($\sigma_{0^\circ\text{N}} = 1.28\text{ K}$ and $\sigma_{20^\circ\text{S}} = 1.06\text{ K}$ in GAIA and $\sigma_{0^\circ\text{N}} = 1.58\text{ K}$ and $\sigma_{20^\circ\text{S}} = 1.28\text{ K}$ in MERRA-2, respectively) is extracted. The day with the coldest anomaly is defined as the central date (day = 0) of the cold equatorial stratosphere event as an IHC proxy. In total, 18 (14) cold equatorial stratosphere events were extracted from the 19 (28) years of GAIA (MERRA-2) data (Table 1). Note that GAIA uses climatological O_3 distribution for the shortwave heating

TABLE 1. Central dates and temperature anomalies δT (K) for $\phi = 0^\circ\text{N}$, 20°S , and $80^\circ\text{--}90^\circ\text{N}$ at 5 hPa during the cold equatorial stratosphere events in the GAIA (18 events) and MERRA-2 (14 events) datasets.

No.	GAIA				MERRA-2			
	Central date	δT (0°N)	δT (20°N)	δT ($80^\circ\text{--}90^\circ\text{N}$)	Central date	δT (0°N)	δT (20°N)	δT ($80^\circ\text{--}90^\circ\text{N}$)
1	17 Dec 1996	−5.1	−4.2	40.9	19 Dec 1992	−3.9	−2.8	0.7
2	8 Dec 1998	−4.6	−3.8	48.7	1 Jan 1995	−3.1	−3.6	16.0
3	24 Feb 1999	−2.5	−2.2	31.1	15 Dec 1998	−4.4	−4.0	48.8
4	1 Dec 2000	−3.8	−3.1	35.8	24 Feb 1999	−4.3	−4.0	47.2
5	27 Jan 2001	−4.1	−3.6	32.2	11 Dec 2000	−4.9	−4.2	54.0
6	24 Dec 2001	−3.4	−2.6	21.6	28 Jan 2001	−4.6	−3.6	18.3
7	28 Dec 2002	−3.9	−3.2	38.1	24 Dec 2001	−6.2	−4.4	56.1
8	24 Dec 2003	−3.2	−2.6	38.4	30 Dec 2002	−3.3	−3.5	45.5
9	2 Dec 2004	−2.7	−2.3	34.8	23 Dec 2003	−3.1	−2.9	47.6
10	23 Feb 2005	−3.5	−2.9	34.7	22 Jan 2009	−6.0	−4.9	56.7
11	13 Jan 2006	−2.8	−2.4	30.7	29 Jan 2010	−3.6	−3.1	29.0
12	25 Jan 2008	−2.9	−2.6	35.4	1 Jan 2012	−4.1	−3.4	29.5
13	24 Jan 2009	−2.9	−2.4	43.7	11 Jan 2013	−6.0	−5.3	41.5
14	31 Jan 2010	−3.5	−3.0	37.2	15 Feb 2018	−4.1	−2.4	23.3
15	12 Jan 2011	−3.5	−3.4	40.0				
16	15 Jan 2012	−4.5	−3.5	24.3				
17	30 Dec 2012	−6.3	−5.1	52.4				
18	7 Jan 2015	−2.5	−2.4	21.2				

calculation in the stratosphere. Thus, it is considered that the standard deviation of temperature anomaly in GAIA is smaller than that in MERRA-2. This is probably a reason why cold equatorial stratosphere events are more frequent in the GAIA simulation. Moreover, the “event climatology” is defined: the event climatology on day = n (n is an integer) is regarded as a composite of the climatologies on day = n for each event. In other words, the event climatology is equivalent to an epoch analysis of the climatology for cold equatorial stratosphere events. For example, the “event climatology” of day = 0 in GAIA is the average of the climatology values of 17 December, 8 December, 24 February, . . . , 30 December, and 7 January (see the left column of Table 1). Hereafter, the day = n is defined as the n th day from the central date of cold equatorial stratosphere events.

3. Zonal-mean field composites for IHC events

a. Response of temperature and wave forcing anomalies from the climatology in the middle atmosphere

Figure 1 shows latitude–height sections of composites of the event climatology and anomalies of zonal-mean temperature and zonal wind averaged over day = -4 to $+4$ for the cold equatorial stratosphere events using the GAIA data. Because the events mostly occurred in late December and January, the distribution of the event climatology of temperature and zonal wind has characteristics similar to the climatology for January (see Fig. 1 of SYM18), although the zonal wind near the summer mesopause in the event climatology is slightly weaker.

For the temperature anomaly, the cold anomaly around the equatorial stratosphere extends to SH midlatitudes. The temperature anomaly is largely positive in the Arctic stratosphere

and negative in the Arctic mesosphere. Additionally, a warm anomaly is observed in the equatorial mesosphere. Because most of the events extracted in this study are followed by SSWs, such a checkered pattern in the temperature anomaly is similar to the distribution of temperature anomalies when an SSW occurs.

At SH high latitudes, there are warm and cold anomalies at 100–115 and 70–80 km, respectively. The warm anomaly extends to and tilts downward at lower latitudes from $(\phi, z) = (80^\circ\text{S}, 110\text{ km})$ to $(0^\circ\text{N}, 75\text{ km})$. The anomaly maximum is $\sim +5\text{ K}$ near $(80^\circ\text{S}, 110\text{ km})$. At first glance, the distribution of the temperature anomaly maxima and minima is similar to that in the KB10 scenario. However, the warm anomaly in the SH MLT region in GAIA is located at higher altitudes than described by KB10. Another difference is that the cold anomaly in the equatorial stratosphere extends to SH midlatitudes in GAIA. Additionally, a westward wind anomaly is observed in the SH upper stratosphere and lower mesosphere (hereafter referred to as USLM) and its maximum is $\sim -10\text{ m s}^{-1}$ at $(10^\circ\text{S}, 55\text{ km})$. An eastward wind anomaly is located in the SH mesosphere above the westward anomaly. Maxima of the eastward anomaly are observed at two locations: at SH low latitudes near 80 km and at SH polar latitudes near 100 km.

Figure 2 shows latitude–height sections of the event climatology and anomaly of total wave forcing (i.e., all resolved wave and parameterized wave forcings), EPFD due to all resolved waves, resolved GWs, RWs/RGWs, migrating TWs, and QTDWs, and parameterized GW forcing [hereafter referred to as “parameterized GW flux divergence” (GW_pFD)] averaged over day = -4 to $+4$. The EPFD event climatology is similar to the January climatology shown in SYM18. For example, the EPFD event climatology due to all resolved waves is negative in the NH stratosphere and MLT region. This negative EPFD

FIG. 1. Latitude–height sections of composites of the (left) event climatology and (right) anomalies of the (top) zonal-mean temperature and (bottom) zonal winds averaged over day = -4 to $+4$ using the GAIA data. The contour intervals are (a) 10 K, (b) 2.5 K, (c) 10 m s^{-1} , and (d) 5 m s^{-1} , respectively. The dashed lines denote ± 1 , ± 2 , ± 3 , and ± 4 K for the temperature anomaly and ± 1 and ± 3 m s^{-1} for the wind anomaly. In (b) and (d), the color shading shows the regions with a 90% significance level.

is mainly due to RWs/RGWs and resolved GWs. The SH EPFD is largely positive below 110 km and negative above 110 km. This remarkable positive EPFD in the SH middle and upper mesosphere is mainly due to the resolved GWs and partly due to RWs/RGWs (particularly QTDWs) (Figs. 2e,m). The QTDWs give large negative wave forcing in the SH upper mesosphere and lower thermosphere. The GW_PFD (Fig. 2k) give as much wave forcing as the EPFD due to resolved GWs (Fig. 2e) in the SH mesosphere. The negative and positive maxima of GW_PFD are located lower than those of the EPFD due to resolved GWs.

The EPFD anomaly due to all resolved waves associated with the cold equatorial stratosphere events has five maxima in the middle atmosphere: (i) a negative maximum in the NH stratosphere ($\sim -20 \text{ m s}^{-1} \text{ day}^{-1}$), (ii) a pair of positive ($\sim +5 \text{ m s}^{-1} \text{ day}^{-1}$) and negative ($\sim -10 \text{ m s}^{-1} \text{ day}^{-1}$) maxima at NH high latitudes in the MLT region, and (iii) a pair of negative ($\sim -5 \text{ m s}^{-1} \text{ day}^{-1}$) and weakly positive [$\sim +2 \text{ m s}^{-1} \text{ day}^{-1}$ near (25°S , 70 km)] maxima in the SH MLT region. The negative EPFD anomaly (iii) in the SH lower thermosphere can cause a residual-mean southward flow anomaly. Thus, the region of the warm anomaly at SH high latitudes in the MLT region (Fig. 1b) is consistent with the negative

EPFD anomaly driving a downward flow anomaly at its southern end. The negative EPFD anomaly (iii) is caused by the resolved GWs (Fig. 2f) and QTDWs (Fig. 2n). The EPFD anomaly due to QTDWs is positive (negative) near (40°S , 70 km) [$(45^\circ\text{S}$, 90 km)]. This feature indicates that QTDWs generated in this region have significantly larger amplitudes during cold equatorial stratosphere events than usual and deposit more westward momentum in the SH upper mesosphere. This result is consistent with previous studies showing QTDW intensification during SSWs (e.g., McCormack et al. 2009). Moreover, the EPFD anomaly due to resolved GWs is mostly negative at all SH latitudes, except localized positive EPFD anomalies near (21°S , 70 km) and (57°S , 85 km). The altitude of the negative EPFD anomaly maximum due to the resolved GWs is higher than that of the negative EPFD anomaly due to QTDWs. The GW_PFD anomaly is largely positive in the NH mesosphere. The GW_PFD anomaly in the SH MLT region is weakly positive at (30°S , 75 km) and (60°S , 80 km) and negative near (65°S , 100 km) and this negative GW_PFD anomaly is smaller than the negative EPFD anomaly due to resolved GWs. This means that resolved GWs play an essential role in both hemispheres together with

FIG. 2. Latitude–height sections of the event climatology and anomalies of (a),(b) the total wave forcing (all resolved wave and parameterized wave forcings), divergence of the Eliassen–Palm flux (EPFD) due to (c),(d) all resolved waves, (e),(f) resolved gravity waves (GWs), (g),(h) Rossby waves (RWs)/Rossby–gravity waves (RGWs), (i),(j) migrating tidal waves (TWs) and (m),(n) quasi-two-day waves (QTDWs), and (k),(l) parameterized GW forcing (GW_{pFD}) from the GAIA simulation data. These panels are the composite results averaged over days = -4 to $+4$. The contour intervals are not uniform. The color shading shows the regions with a 90% significance level for the anomalies.

parameterized GWs for the IHC. As discussed in sections 3b and 3c, the weak GW_{pFD} anomaly in the SH mesosphere has an indirect impact on IHC. Note that the EPFD anomaly due to resolved GWs in the SH lower thermosphere is negative, but a low temperature anomaly is not observed in the equatorward region. In the tropical lower thermosphere, the EPFD anomaly due to TWs is positive which partly cancels the GW forcing. This is a reason why strong equatorial upwelling and associated significant low temperature are not formed in the equatorial lower thermosphere. It is possible that this TW forcing anomaly is modulated by zonal wind changes in the lower atmosphere (e.g., Jin et al. 2012). The detailed analysis is left for future work, however.

Next, the IHC time variations are examined. Figure 3 shows time–latitude sections of temperature anomalies at 42–47 km in the upper stratosphere, 70–80 km in the middle mesosphere, 85–95 km in the upper mesosphere, and 105–115 km in the lower thermosphere. First, at 42–47 km, the temperature anomaly is negative over a wide latitude range

from NH midlatitudes to the entire SH after day = ~ -10 . At the same time, a statistically significant warm anomaly is seen near 60°N and then extends to the North Pole. The time evolution of the warm anomaly in the NH is similar to that of SSWs. However, the cold anomaly almost simultaneously appears in the region extending to SH high latitudes when the warming starts in the NH polar region. Note that many cold equatorial stratosphere events occur 0 to +12 days before the major SSWs in the lower stratosphere as reported by Charlton and Polvani (2007) (not shown). At 70–80 km, the warm anomaly maximum appears near the equator on day = +2, while the cold anomaly in the NH polar region is largest at almost the same time as the warm anomaly at 42–47 km. At 105–115 km, a warm anomaly in the SH polar region is observed after day = -3 and reaches its maximum on day = +4. The warm anomaly maximum is observed at SH mid- and high latitudes at a later time near day = +9 at 85–95 km. Note that similar time evolution of the temperature anomalies is observed by the *Aura* MLS and in MERRA-2

FIG. 3. Time-latitude sections of the temperature anomalies at (a) $z = 42\text{--}47$ km in the upper stratosphere, (b) $z = 70\text{--}80$ km in the middle mesosphere, (c) $z = 85\text{--}95$ km in the upper mesosphere, and (d) $z = 105\text{--}115$ km in the lower thermosphere. The contour intervals are (a) 5, (b) 2, (c) 1.5, and (d) 1 K. The dashed lines denote -1 , -2 , -3 , and -4 K in (a) and $+1$ and $+3$ K in (b). The color shading shows the regions with a 90% significance level.

for the height region below 65 km where data are available (not shown).

The time variation of the wave forcing anomaly was also investigated. Figure 4 shows time-latitude sections of the EPFD anomalies due to all resolved waves, resolved GWs, RWs/RGWs, and QTDWs and the GW_pFD anomaly at 48, 90–100, and 105–115 km. A negative maximum is observed for the EPFD anomaly due to all resolved waves at NH high latitudes at 48 km near day = -3 . The time variation of the EPFD anomaly due to all resolved waves is mainly explained by that due to RWs/RGWs and is consistent with that of the temperature anomaly (Fig. 3a). However, the warm anomaly in the

NH polar region and the cold anomaly around the equator are maximized when the negative EPFD anomaly disappears. Because the radiative relaxation time (~ 7 days at 48 km; Mlynzak et al. 1999) is longer than the time period of the negative wave forcing anomaly, the temperature anomaly peaks can coincide with the end time when the negative EPFD anomaly is observed. Conversely, a positive GW_pFD anomaly ($\sim 1.8 \text{ m s}^{-1} \text{ day}^{-1}$) appears after day = 0 with a magnitude much smaller than the EPFD anomaly due to RWs/RGWs.

The EPFD anomaly due to all resolved waves at 105–115 km is largely negative at mid- and high latitudes after day = 0 in both hemispheres. The negative anomaly maximum in the SH appears ~ 3 days earlier (day = 0) than that in the NH (day = +3). This wave forcing anomaly is largely composed of EPFD anomaly due to the resolved GWs ($\sim 10 \text{ m s}^{-1} \text{ day}^{-1}$). While the negative EPFD anomaly due to the resolved GWs is seen at all SH latitudes from day = -3 to day = +1, it is confined to SH high latitudes after day = +1. The negative EPFD anomaly maximum due to the resolved GWs in the lower thermosphere appears before the warm anomaly maximum in the equatorial mesosphere (Fig. 3b). Conversely, the cold anomaly in the equatorial and SH stratosphere is maximized on day = 0 (Fig. 3a). Thus, the negative EPFD anomaly due to the resolved GWs observed before day = +1 may be affected by the cold anomaly in the equatorial and SH stratosphere. A positive EPFD anomaly due to the resolved GWs is seen near 40°S after day = +8, while negative EPFD anomaly due to RWs/RGWs is seen near 70°N near day = +8. A positive GW_pFD anomaly is seen at NH high latitudes near day = -2 . However, the contribution of RWs/RGWs and GW_pFD anomalies to all resolved wave forcing anomaly is small.

Next, a negative EPFD anomaly in the SH due to all resolved waves at 90–100 km is seen at midlatitudes ($\sim 10 \text{ m s}^{-1} \text{ day}^{-1}$). The negative anomaly is composed of EPFD anomaly due to RWs/RGWs, particularly QTDWs. The negative EPFD anomaly due to QTDWs has a peak ($\sim 15 \text{ m s}^{-1} \text{ day}^{-1}$) at 40°S on day = +7. For the resolved and parameterized GWs, there are few significant EPFD and GW_pFD anomalies in the SH. The EPFD and GW_pFD anomalies in the NH are positive, and their maxima are observed on day = -1 at 60°N . This feature is different from that in the lower thermosphere (i.e., 105–115 km).

Time variations of the warm anomaly in the SH lower thermosphere (Fig. 3d) correspond well to those of negative EPFD anomaly due to the resolved GWs at SH high latitudes in the lower thermosphere. Meanwhile, time variations of the warm anomaly in the SH upper mesosphere (Fig. 3c) correspond well to those of negative EPFD anomaly due to QTDWs at SH midlatitudes in the upper mesosphere. These results suggest that the warm anomalies in the SH MLT region are mainly caused by these different waves in their respective height regions. As shown in Figs. 3c and 3d, upper mesospheric warming occurs after lower thermospheric warming in the SH in association with cold equatorial stratosphere events. This is likely because different types of waves having different generation and propagation time scales are related to warming in these respective regions: GWs contribute to the warming in the lower thermosphere and QTDWs contribute to the warming in

FIG. 4. Time-latitude sections of EPFD anomalies due to (a),(d),(i) all resolved waves, (e),(j) GWs, (b),(f),(k) RWs/RGWs, (h),(m) QTDWs, and (c),(g),(l) GW_pFD anomaly for $z = 48, 90\text{--}100$, and $105\text{--}115$ km, respectively. The contour intervals are $12 \text{ m s}^{-1} \text{ day}^{-1}$ for (a) and (b), $0.5 \text{ m s}^{-1} \text{ day}^{-1}$ for (c), and $18 \text{ m s}^{-1} \text{ day}^{-1}$ for the other panels. The color shading shows the regions with a 90% significance level. The $0 \text{ m s}^{-1} \text{ day}^{-1}$ contour is not shown.

the upper mesosphere. Roughly speaking, GWs are expected to take ~ 1 day to propagate from the mesosphere to the lower thermosphere, while QTDWs take several days to develop and propagate into the upper mesosphere. A significantly long time lag of $\sim 5\text{--}10$ days between the NH SSW and subsequent SH MLT warming (e.g., Karlsson et al. 2009a) is likely explained by this mechanism in which not only GWs but also RWs/RGWs including QTDWs play crucial roles for IHC.

The time evolution of the temperature anomalies is also examined using *Aura* MLS data and MERRA-2 data. Note that the available height region for MERRA-2 data is below 65 km. Figure 5 shows latitude–height sections of a 9-day average of the temperature anomalies from GAIA (left panels), *Aura* MLS (middle panels), and MERRA-2 (right panels) at a time interval of 4 days. A checker pattern of the temperature anomaly in the NH stratosphere and mesosphere is seen for all data. The warm anomaly maximum that appeared in the SH MLT region in day = -4 to $+4$ moves downward and equatorward in GAIA. *Aura* MLS also exhibits a similar movement of the warm anomaly in the SH upper mesosphere after day = 0 to $+8$. In the MERRA-2 data, which assimilate *Aura* MLS observation, a similar evolution of the temperature anomaly is

observed. These results indicate that the temperature anomaly pattern and its time evolution related to the cold equatorial stratosphere events in the GAIA data are realistic.

b. Generation mechanism of resolved gravity waves related to interhemispheric coupling

In this section, the causes of the resolved GW forcing anomaly, which contributes to the warm anomaly in the SH lower thermosphere, are examined. As seen in Fig. 2l, the GW_pFD anomaly maxima near ($20^{\circ}\text{--}30^{\circ}\text{S}$, $68\text{--}75$ km) and (60°S , 80 km) are located slightly below the localized positive EPFD anomalies due to the resolved GWs near (21°S , 70 km) and (57°S , 85 km). SYM18 showed that GW_pFD forms a strong vertical shear of zonal winds in the summer mesosphere. The positive GW_pFD anomaly observed during cold equatorial stratosphere events can further accelerate the zonal winds eastward in the SH mesosphere, resulting in a strong eastward vertical shear of zonal winds. Figure 6 shows latitude–height sections of the event climatology and anomaly of vertical shear of zonal winds and occurrence frequency of Richardson number less than one ($Ri < 1$) averaged over day = -4 to $+4$. The vertical shear of zonal winds for the event climatology is largely

FIG. 5. Latitude–height sections of 9-day averaged zonal-mean temperature anomalies from (left) the GAIA, (center) the *Aura* MLS, and (right) the MERRA-2 data at a time interval of 4 days. The contour interval is 10 K. The dashed lines denote ± 1 , ± 2 , ± 3 , and ± 4 K. In the GAIA and the MERRA-2 data, the color shading shows the region with a 90% significant level.

positive in the region from (5°S , 55 km) to (75°S , 110 km). The maximum value is $\sim 5 \times 10^{-3} \text{ s}^{-1}$ at (45°S , 85 km). Enhanced strong vertical shear regions near (48°S , 80 km) and (20°S , 68 km) are considered due to the positive GW_{pFD} anomalies observed at these locations.

A large occurrence frequency of $\text{Ri} < 1$ for the event climatology is seen in the low and midlatitudes of the SH mesosphere and in all the displayed latitudes of the lower thermosphere below 115 km. The maximum is seen in the region from (20°S , 68 km) to (70°S , 105 km) and at low latitudes near 100 km.

During cold equatorial stratosphere events, the occurrence frequency anomaly of $\text{Ri} < 1$ is positive near (57°S , 80 km) and (20°S , 68 km) in the SH mesosphere. These regions agree well with those where largely positive vertical shear anomalies are observed, suggesting that the positive resolved GW forcing anomalies at (21°S , 70 km) and (57°S , 85 km) in the SH mesosphere are attributable to the enhanced shear instability, reflecting generation of the resolved GWs there. A few percent of occurrence frequency anomaly of $\text{Ri} < 1$ in the SH mesosphere may be small at a glance. However, the momentum that in situ

FIG. 6. Latitude–height sections of (a) the event climatology and (b) the anomaly of the vertical shear of the zonal-mean zonal winds in GAIA. Latitude–height sections of (c) the event climatology and (d) anomaly of occurrence frequency of Richardson number Ri lower than 1 ($Ri < 1$). These figures are the composite results averaged over days = -4 to $+4$. The contour intervals are (a) $1.5 \times 10^{-3} \text{ s}^{-1}$, (b) $0.25 \times 10^{-3} \text{ s}^{-1}$, (c) 3%, and (d) 0.5%. In (b) and (d), the color shading shows only the 90% significance level. The 0 s^{-1} and 0% contours are not shown in (b) and (d), respectively.

generated GWs deposit to the lower thermosphere can be large considering much smaller air density than at the generation level. Therefore, the occurrence frequency of $Ri < 1$ in the mesosphere is important even if its percentage anomaly is small.

Moreover, we examined the phase velocity of the resolved GWs radiated from the shear instability in the SH mesosphere. Figure 7 shows the event climatology and anomaly of the cospectrum of the Φ' and w' associated with resolved GWs weighted by density (i.e., vertical energy flux of the resolved GWs) at 20.9° and 57.2°S and averaged over day = -4 to $+4$ as a function of the zonal phase velocity C_{px} at each height, which were obtained using Fourier analysis. The solid black curves represent the zonal-mean zonal winds for the cold equatorial stratosphere events, and the dashed black curves represent the zonal-mean zonal winds for the event climatology (i.e., unperturbed state). At 20.9°S , the vertical energy flux for the event climatology is largely positive (i.e., upward) in the eastward ground-based phase velocity range in the height region of 40–120 km. Significant negative (i.e., downward) energy flux is seen for the phase velocities in the range

of $\sim -50 < C_{px} < 0 \text{ m s}^{-1}$ below 70 km. Above the downward energy flux region, an upward energy flux is observed. This feature shows that GWs propagating westward (eastward) relative to the mean winds are radiated upward (downward) from the height having a large occurrence frequency of shear instability for the event climatology.

Next, we focus on the range of the negative (i.e., westward) ground-based phase velocity for cold equatorial stratosphere events. For 20.9°S , upward (downward) vertical energy flux anomaly is observed over a broad range of phase velocity above (below) the height with high occurrence frequency of $Ri < 1$. The vertical energy flux of westward-propagating resolved GWs at 75 km is $\sim 1 \times 10^{-8} \text{ kg s}^{-1}$ for the event climatology, while its anomaly is $\sim 2 \times 10^{-9} \text{ kg s}^{-1}$ at the same altitude. The vertical energy flux increases by $\sim 20\%$. Similarly, the event climatology and anomaly of occurrence frequency of $Ri < 1$ is $\sim 6\%$ and $\sim 1.5\%$ at (20°S , 68 km). Thus, the occurrence frequency of $Ri < 1$ increases by $\sim 25\%$ and is consistent with the result of vertical energy flux of westward-propagating GWs. A similar structure is observed

FIG. 7. The event climatology and the anomaly of the cospectrum of Φ' and w' associated with the resolved GWs weighted by density (i.e., the vertical energy flux of the resolved GWs) at 20.9° and 57.2°S averaged over days = -4 to +4 as a function of the GW phase velocity C_{px} at the respective heights. The solid black curves represent the zonal-mean zonal winds for the cold equatorial stratosphere events, and the dashed black curves represent the zonal-mean zonal winds for the event climatology (i.e., unperturbed). The green lines denote the heights of the occurrence frequency anomaly maximum of $Ri < 1$.

at 57.2°S, even though the boundary height between the upward and downward energy flux anomalies is not as clear as the feature at 20.9°S.

Note that relatively large vertical energy flux anomaly of eastward-propagating GWs is also observed in the positive (i.e., eastward) ground-based phase velocity range. These could be due to modulations in the Brunt-Väisälä frequency squared N^2 or the existence of GWs having wavenumber vectors including meridional components. The structure is complex and is therefore left for future work.

The results obtained in this section show that the positive GWpFD anomaly in the SH mesosphere is responsible for the strong vertical shear anomaly of zonal winds. Additional westward-propagating GWs are likely radiated by the shear instability associated with the intensified vertical shear of zonal winds during cold equatorial stratosphere events according to theoretical studies (e.g., Bühler et al. 1999). These features are

fundamentally different from the KB10 scenario which describes IHC as being caused by modulation of forcing due to GWs originating only in the troposphere. Our results show that the shear instability generating westward-propagating GWs is more frequent in the middle atmosphere during cold equatorial stratosphere events and then in situ generated GWs cause significant negative wave forcing anomaly in the SH lower thermosphere.

Note that the structures of GWs resolved in the GAIA may be largely distorted due to relatively coarse horizontal model resolutions. However, the distribution of the momentum deposition transported by the model-resolved GWs was consistent with the results of higher-resolution model (see SYM18; YSM18). In addition, the shear instability is a characteristic of large-scale mean field, and hence the role of the GWs generated in situ in the middle atmosphere should be important regardless of the dependence of

GW structure on the model resolution from the viewpoint of the momentum budget.

c. QTDW enhancement during cold equatorial stratosphere events

In this section, we examine QTDW enhancement mechanisms during SSWs, focusing on their relationship with GW forcing. SYM18 showed that GW_pFD acts to strengthen the latitudinal gradient of zonal wind in the summer mesosphere. The zonal wind gradient leads to a negative latitudinal gradient of modified PV (MPV), which is a necessary condition for BT/BC instabilities. Here, MPV is defined as (Lait 1994)

$$\text{MPV} \equiv P \times \left(\frac{\theta}{\theta_0} \right)^{-9/2} = -g(f + \zeta) \frac{\partial \theta}{\partial p} \left(\frac{\theta}{\theta_0} \right)^{-9/2} \propto (f + \zeta) N^2, \quad (6)$$

where P , θ , θ_0 ($=420$ K), g , ζ , and p denote the Ertel's PV, potential temperature, reference potential temperature, gravitational acceleration, relative vorticity, and pressure, respectively. This study examined modulations of latitudinal gradient of MPV (MPV_ϕ) formed by GW_pFD during cold equatorial stratosphere events. In the quasigeostrophic equations, the quasigeostrophic PV (QGPV) tendency is written as follows:

$$\bar{q}_t = -\overline{v'q'}_y - \bar{X}_y, \quad (7a)$$

$$\overline{v'q'} = \rho_0^{-1} \nabla \cdot \mathbf{F}, \quad (7b)$$

where q is QGPV (Andrews et al. 1987). Thus, the contribution of GW_pFD to the QGPV tendency is obtained using

$$-\bar{X}_y = -\overline{\text{GW}_\text{pFD}_y} = -(a \cos \phi)^{-1} (\overline{\text{GW}_\text{pFD}} \cos \phi)_\phi. \quad (8)$$

A similar evaluation can be made for the primitive equation system using MPV instead of QGPV (see SYM18).

Figure 8 shows latitude–potential temperature sections of the event climatology and anomaly of $-\text{GW}_\text{pFD}_y$, MPV, and $\overline{\text{MPV}}_\phi$ averaged over day $= -4$ to $+4$ using the GAIA data. In the event climatology, two $-\overline{\text{GW}_\text{pFD}_y}$ maximum and minimum pairs are seen at SH low and midlatitudes in the mesosphere, corresponding to the GW_pFD maxima (Fig. 2k). The region of negative $-\overline{\text{GW}_\text{pFD}_y}$ near $(\phi, \theta) = [35^\circ\text{S}, 4500 \text{ K } (\sim 80 \text{ km})]$ corresponds to that of the negative MPV local maximum. Negative $\overline{\text{MPV}}_\phi$ is observed near $(48^\circ\text{S}, 4500 \text{ K})$, corresponding to the negative MPV local maximum. This region satisfies the necessary condition of BT/BC instability.

During cold equatorial stratosphere events, two pairs of positive and negative $-\text{GW}_\text{pFD}_y$ anomalies are observed near 30° and 60°S in the potential temperature region of 3500 – 7000 K (70 – 90 km), corresponding to positive GW_pFD anomalies. The MPV anomaly is negative in a similar region [i.e., near $(35^\circ\text{S}, 4000 \text{ K})$]. Thus, the negative local MPV maximum at SH midlatitudes in the mesosphere is strengthened during cold equatorial stratosphere events. The $\overline{\text{MPV}}_\phi$ anomaly is negative near $(40^\circ\text{S}, 4000$ – $5000 \text{ K})$. Therefore, negative $\overline{\text{MPV}}_\phi$ is also strengthened at SH midlatitudes in the mesosphere during cold equatorial stratosphere events.

The QTDW activity is generally expected to be higher during cold equatorial stratosphere events than in the climatology, likely owing to the enhanced probability of the BT/BC instability. However, there are four events occurred on 1 December 2000, 2 December 2004, 23 February 2005, and 24 January 2009, in which a negative QTDW forcing anomaly is seen in the middle mesosphere (not shown). Three of these occurred in early December or late February when the westward wind in the SH mesosphere was weak. The e -folding time of QTDW amplification depends on the background westward wind maximum, which is 5.2 (25.6) days for the maximum of -90 (-60) m s^{-1} (Salby and Callaghan 2001). SYM18 also demonstrated weak QTDW activity in early December and late February. Thus, the negative QTDW forcing in the middle mesosphere for these three cases is likely a result of the small QTDW growth rate in the seasonal westward wind in the SH mesosphere. In such cases, the contribution of QTDWs to the warming in the SH MLT region becomes small and the contribution of the resolved GWs becomes relatively large. The seasonal variation of the IHC time lag (i.e., $+2$ days in December and $+7$ days in January) indicated by Karlsson et al. (2009b) may be explained by this seasonal variation of the QTDW growth rate.

d. Effect of a cold anomaly of the equatorial stratosphere on interhemispheric coupling

As described above, the resolved GW and QTDW forcing anomalies observed in the SH MLT region are attributable to the strengthened shear and BT/BC instabilities, respectively. These instabilities are thought to be caused by strengthened vertical and meridional zonal wind shears formed by positive GW_pFD anomalies. In this section, causes of GW_pFD anomalies are discussed. Note that GW_pFD is a parameterization of GWs originating in the troposphere.

Figure 9 shows time–latitude sections of anomaly of occurrence frequency of $\text{Ri} < 1$, $\overline{\text{MPV}}_\phi$ and GW_pFD averaged over 65 – 70 km . The occurrence frequency anomaly of $\text{Ri} < 1$ is large near day $= -10$ and after day $= -2$ at $\sim 10^\circ$ – 20°S . The $\overline{\text{MPV}}_\phi$ anomaly is significantly negative at 40°S after day $= -3$. These features indicate that the shear instability condition is strengthened at $\sim 10^\circ$ – 20°S after day $= -2$ and BT/BC instability condition is strengthened at 40°S after day $= -3$. On the other hand, after day $= -2$, positive GW_pFD anomalies are seen at similar (lower) latitudes to the strengthened shear (BT/BC) instability regions. The positive anomaly at $\sim 10^\circ$ – 20°S is observed until day $= +6$. This means that the positive GW_pFD anomaly weakens eastward zonal winds and then the vertical and meridional shear of zonal winds is strengthened in the SH mesosphere. These results suggest that the positive GW_pFD anomaly at least partly causes the shear and BT/BC instability.

Next, we examine how the positive GW_pFD anomaly forms in the SH mesosphere. Figures 10a–c show latitude–height sections of the event climatology and anomaly of GW_pFD weighted by density and anomaly of zonal-mean zonal winds averaged over day $= -4$ to $+4$. In the SH, eastward and westward GW_pFD for the event climatology are observed above and below 50 km , respectively. The eastward GW_pFD maxima are seen at $(12^\circ\text{S}, 60 \text{ km})$ and $(70^\circ\text{S}, 65 \text{ km})$.

FIG. 8. Latitude–potential temperature sections of the event climatology and anomalies of (a),(b) $-\overline{\text{GW}_p\text{FD}_y}$, (c),(d) $\overline{\text{MPV}}$, and (e),(f) $\overline{\text{MPV}_\phi}$ averaged over days = -4 to $+4$ in the GAIA data. The contour intervals are (a) $4 \times 10^{-6} \text{ s}^{-1} \text{ day}^{-1}$, (b) $1 \times 10^{-6} \text{ s}^{-1} \text{ day}^{-1}$, (c) 5 PVU ($1 \text{ PVU} = 10^{-6} \text{ K kg}^{-1} \text{ m}^2 \text{ s}^{-1}$), (d) 0.4 PVU, (e) 15 PVU rad^{-1} , and (f) 4 PVU rad^{-1} , respectively. The $0 \text{ s}^{-1} \text{ day}^{-1}$, 0 PVU, and 0 PVU rad^{-1} contours are not shown in (b), (d), and (f), respectively. The green curves in (f) denote the contour of 0 PVU rad^{-1} for the event climatology. The color shading shows the regions with a 90% significance level for the anomalies.

Conversely, negative and positive GW_pFD anomalies are seen in and above the region of $(\phi, z) = (10^\circ\text{--}20^\circ\text{S}, 30\text{--}60 \text{ km})$, respectively. Thus, the positive GW_pFD peak shifts upward during cold equatorial stratosphere events. According to the KB10 scenario, the positive GW_pFD anomaly appears in the eastward wind anomaly region in the SH mesosphere. However, the positive GW_pFD anomaly is seen near or below the height of the zero zonal wind anomaly in the GAIA data. This feature implies

that a different mechanism from that in the KB10 scenario causes the positive GW_pFD anomaly. The GW parameterization used in this study assumes only upward propagation like many common parameterizations. Thus, this positive mesospheric GW_pFD anomaly is associated with a negative stratospheric GW_pFD below, which is likely filtered out by the negative zonal wind anomaly in the SH USLM. A positive GW_pFD anomaly is also observed at 60–85 km near 70°S . The positive GW_pFD anomaly

FIG. 9. Time–latitude sections of (a) the anomaly of occurrence frequency of $Ri < 1$ averaged over $z = 65\text{--}70$ km, (b) anomaly of MPV_ϕ averaged over $\theta = 3680\text{--}4460$ K ($z \sim 65\text{--}70$ km), and (c) anomaly of GW_pFD averaged over $z = 65\text{--}70$ km. The contour intervals are (a) 0.6%, (b) 3 PVU rad^{-1} , and (c) $6 \times 10^{-10} \text{ kg m}^{-2} \text{ s}^{-2}$. The 0%, 0 PVU rad^{-1} , and $0 \text{ kg m}^{-2} \text{ s}^{-2}$ contours are not shown. The color shading shows the regions with a 90% significance level. The green contours denote the 2% contour of the event climatology in (a) and the 0 PVU rad^{-1} contour of the event climatology in (b).

at SH high latitudes in the mesosphere may occur at least partly due to a similar mechanism acting at SH low latitudes.

These results indicate that GW_pFD in the SH mesosphere is modulated by zonal winds in the SH USLM. In this study, a new mechanism for the appearance of the positive GW_pFD anomaly in the SH mesosphere is proposed, as shown schematically in Fig. 10d: (i) A westward wind anomaly is formed in the SH USLM via thermal wind balance with the cold equatorial stratosphere. (ii) The westward wind anomaly in the SH USLM provides a critical level for the parameterized westward GWs, which leads to a negative GW_pFD anomaly. (iii) Because the parameterized westward GWs no longer reach into the SH mesosphere, a positive GW_pFD anomaly forms in the SH mesosphere. For this new mechanism, during cold equatorial stratosphere events, the zonal wind anomaly in the SH USLM may be essential for IHC because it causes in situ GW and QTDW generation in the SH mesosphere.

e. Causes of the anomalous cold equatorial stratosphere and Rossby wave generation in the winter stratosphere

In this section, causes of the anomalous cold equatorial stratosphere are examined. In the equatorial stratosphere,

middle-atmosphere Hadley circulation is characterized by strong flows (Dunkerton 1989). Semeniuk and Shepherd (2001a,b) demonstrated that middle-atmosphere Hadley circulation is strengthened by wave forcing in the winter stratosphere. The enhanced middle-atmosphere Hadley circulation causes a cold (warm) anomaly at low (mid-)latitudes in the summer (winter) stratosphere via adiabatic cooling (heating) because of its upward (downward) flows.

To examine the influence of middle-atmosphere Hadley circulation on IHC, differences between strong and weak cold equatorial stratosphere events, as defined below, were analyzed. During the cold equatorial stratosphere events defined in section 2b(3) (hereafter referred to as “strong cold events”), most of the Arctic temperature anomalies at $80^\circ\text{--}90^\circ\text{N}$ and 5 hPa are more than 15 K. Accordingly, weak cold equatorial stratosphere events (hereafter referred to as “weak cold events”) are defined as follows. 1) Days when a warm anomaly exceeds 15 K at 5 hPa in the Arctic region ($80^\circ\text{--}90^\circ\text{N}$) in December–February are regarded as candidates for weak cold event periods. These candidates include strong cold event periods. The day with the highest Arctic warm anomaly at 5 hPa during each event period is a candidate for the central date. 2) Next, if the next event period starts within 14 days after the end of the previous event period, this event is excluded from the weak cold event candidates as a series of the previous event. This procedure reduces the influence of previous cold events and accurately estimates the response time of the SH MLT region to cold events. 3) The rest of the candidates for the central dates, except for the strong cold events (Table 1), are defined as central days of weak cold events. Based on this procedure, 9 (23) events are extracted in GAIA (MERRA-2) for the weak cold events (Table 2).

Figure 11 shows latitude–height sections of temperature anomalies averaged over day = -4 to $+4$ and EPFD anomalies averaged over day = -8 to 0 during strong and weak cold equatorial stratosphere events in GAIA and MERRA-2. While the cold anomaly in the region from near the equator to 20°S in the upper stratosphere is weaker during weak cold events than during strong cold events, as expected, the temperature anomaly in the Arctic stratosphere is largely positive during both strong and weak cold events (Figs. 11a,c). For the weak cold events in GAIA, a checkerboard pattern of the temperature anomaly is observed poleward of 20°N in the NH USLM, similar to usual SS events. Conversely, the pattern in the strong cold events extends to 30°S . The equatorward extension of the negative EPFD anomaly in the NH USLM is also small in the weak cold events. Warm anomalies in the SH MLT region are weak for the weak cold events. This is likely related to the absence of a negative EPFD anomaly in this region.

The difference in the temperature and EPFD anomaly patterns between strong and weak cold events is also observed in MERRA-2, except in the equatorial stratosphere. In MERRA-2, the cold anomaly has local minima not only at $(\phi, z) = (25^\circ\text{N}, 35 \text{ km})$ but also at $(0^\circ\text{N}, 45 \text{ km})$ even during weak cold events.

Figure 12 shows latitude–height sections of differences in the negative EPFD anomalies and the angular momentum

FIG. 10. Latitude–height sections of (a) the event climatology and (b) anomaly of GW_pFD weighted by the density, and (c) anomaly of the zonal-mean zonal wind averaged over days = -4 to $+4$. (d) Schematic illustration of a new inferred mechanism for the positive GW_pFD anomaly in the SH mesosphere as a function of the zonal-mean zonal wind or GW phase velocity and height. The solid black curve indicates the zonal-mean zonal wind in the event climatology. The colored curve indicates the zonal-mean zonal wind anomaly during cold equatorial stratosphere events. The wavy curves denote upward-propagating GWs. The dashed wavy curves denote GWs that can no longer propagate upward. The dashed line denotes the zonal-mean zonal wind minimum. δ represents the anomaly from the event climatology. The contour intervals in (a) and (b) are not uniform. The contour interval in (c) is 5 m s^{-1} . The dashed contours denote ± 1 and $\pm 3 \text{ m s}^{-1}$. The $0 \text{ kg m}^{-2} \text{ s}^{-2}$ contour is not shown in (b). In (b) and (c), the color shading shows the regions with a 90% significance level. The green curves in (b) denote the 0 m s^{-1} contour for the zonal-mean zonal wind anomaly.

$M = a \cos \phi (u + a \Omega \cos \phi)$ and its latitudinal gradient between strong and weak cold events averaged over day = -8 to 0 in GAIA and MERRA-2, respectively. Here, a is Earth’s radius and Ω is Earth’s angular velocity. The differences in the EPFD anomalies are largely negative in the NH USLM. The difference in the negative EPFD anomaly extends from 87° to 10°N in the NH above 35 km and has a maximum at $(55^\circ\text{N}, 55 \text{ km})$. In other words, significant negative wave forcing occurs at NH low and midlatitudes, as well as at NH high

latitudes, in the USLM during strong cold events. The latitudinal gradient of the angular momentum is small equatorward of 20°N particularly near $40\text{--}50 \text{ km}$. These results indicate that, if strong westward wave forcing occurs at NH low latitudes, middle-atmosphere Hadley circulation is effectively intensified. As a result, strong northward flow over the equator should form. Note that it has also been pointed out by Zülicke and Becker (2017) that the cross-equator flow at the stratopause called “easterly nose” is enhanced by

TABLE 2. Central dates and temperature anomalies δT for $\phi = 0^\circ\text{N}$, 20°S , and $80^\circ\text{--}90^\circ\text{N}$ at 5 hPa during the weak cold equatorial stratosphere events in the GAIA (9 events) and MERRA-2 (23 events) datasets.

No.	GAIA				MERRA-2			
	Central date	δT (0°N)	δT (20°N)	δT ($80^\circ\text{--}90^\circ\text{N}$)	Central date	δT (0°N)	δT (20°N)	δT ($80^\circ\text{--}90^\circ\text{N}$)
1	6 Dec 1997	−0.7	−0.3	24.3	10 Jan 1991	−2.3	−1.6	16.3
2	3 Jan 1998	−0.7	−0.7	20.4	27 Jan 1991	−2.6	−2.6	53.2
3	27 Feb 1998	−2.4	−1.8	22.8	11 Jan 1992	−1.6	−1.9	42.5
4	8 Feb 2000	−1.8	−1.7	42.7	20 Feb 1993	−1.0	−1.7	31.9
5	16 Dec 2006	−0.5	−0.6	20.1	30 Dec 1993	−2.3	−1.7	28.7
6	4 Jan 2007	−1.6	−1.7	41.6	28 Jan 1995	−1.8	−1.8	36.8
7	1 Feb 2007	−1.2	−1.0	27.8	18 Feb 1996	−1.1	−1.3	28.1
8	3 Feb 2011	−1.3	−1.1	23.5	8 Dec 1997	−2.6	−1.5	21.6
9	12 Feb 2012	−1.0	−0.6	20.4	10 Feb 1998	−2.2	−2.3	34.0
10					26 Jan 2000	−1.9	−1.4	17.8
11					17 Feb 2002	−1.9	−1.7	31.2
12					22 Feb 2005	−1.9	−1.2	17.5
13					3 Jan 2006	0.1	−0.1	25.0
14					1 Jan 2007	−2.0	−1.9	33.3
15					24 Jan 2008	−2.6	−2.8	45.6
16					7 Dec 2009	−3.2	−2.1	23.8
17					1 Feb 2011	−1.0	−1.6	31.2
18					8 Feb 2014	−2.4	−2.1	23.8
19					31 Dec 2014	−0.9	−1.4	18.0
20					8 Feb 2016	−2.7	−2.2	32.7
21					28 Jan 2017	−2.1	−1.7	37.4
22					27 Feb 2017	−0.9	−1.2	23.3
23					1 Feb 2018	−1.2	−0.5	16.8

FIG. 11. (a)–(d) Latitude–height sections of the temperature anomaly averaged over days = -4 to $+4$ and EPFD anomalies averaged over day = -8 to 0 in GAIA, with (a) and (b) showing strong cold equatorial stratosphere events and (c) and (d) showing weak cold equatorial stratosphere events. (e)–(h) As in (a)–(d), but for MERRA-2. The contour interval is 2.5 K for the temperature anomaly. The contour intervals of the EPFD anomaly are not uniform. The dashed contours denote $\pm 1, \pm 2, \pm 3,$ and ± 4 K in (a), (c), and (e) and $\pm 1, \pm 1.5, \pm 2, \pm 3,$ and ± 4 K in (g). The color shading shows the regions with a 90% significance level.

westward-propagating GW breaking in the tropics during SSWs. Additionally, in this study, we show that it is also enhanced by RW forcing at NH low latitudes in the USLM.

Next, we examine waves responsible for the negative EPFD anomaly at NH low latitudes in the USLM. Figure 13 shows the ratio of the zonal wavenumber–frequency amplitude spectra of Φ' during strong and weak cold events to those of the event climatology at $(30^\circ\text{N}, 50\text{ km})$ for each dataset, where $\overline{\text{MPV}}_\phi$ is negative as described later. The amplitude spectral densities are enhanced over a broad range of wave periods $\tau = \sim 1$ – 5 days and zonal wavenumbers $s \leq -4$ during strong cold

events. A broad peak is observed near $(s, \tau) = (-6, 2\text{ days})$ for the strong cold events in both datasets. This feature is not obvious in MERRA-2 and is weak in GAIA for the weak cold events.

Figure 14 shows latitude–height sections of the EP flux and EPFD associated with westward-traveling RWs with periods of 1–5 days and zonal wavenumbers of -4 to -10 , the ratio of the EPFD due to westward-traveling RWs to the EPFD due to all resolved waves, and the occurrence frequency anomaly of $\overline{\text{MPV}}_\phi < 0$ during the strong cold events in GAIA and MERRA-2. A pair of positive and negative EPFDs is aligned latitudinally

FIG. 12. (a),(d) Latitude–height sections of differences in the negative EPFD anomalies between the strong and weak cold events averaged over days = -8 to 0 in GAIA and MERRA-2. (b),(c) and (e),(f) Latitude–height sections of the angular momentum and its latitudinal gradient, respectively, for the strong cold events in each dataset. The contour intervals are $1.5 \times 10^8 \text{ m}^2 \text{ s}^{-1}$ for the angular momentum and $1 \times 10^9 \text{ m}^2 \text{ s}^{-1} \text{ rad}^{-1}$ for the latitudinal gradient of the angular momentum. The contour intervals in (a) and (d) are not uniform. In (a) and (d), the color shading shows the regions with a 90% significance level.

Ratio of amplitude spectra of geopotential Φ' to climatology at (30N, 50km)

FIG. 13. Ratio of the zonal wavenumber–frequency amplitude spectra of geopotential disturbances from the zonal mean during strong and weak cold equatorial stratosphere events to the event climatology at $(\phi, z) = (30^\circ\text{N}, 50\text{ km})$ in the (top) GAIA and (bottom) MERRA-2 datasets. The spectra are during the (left) strong and (right) weak cold events. The black curves denote the zonal phase velocity C_{px} (m s^{-1}). Positive and negative zonal wavenumber s indicate eastward and westward propagation, respectively.

at latitudes lower than 50°N in the NH USLM. The negative EPFD ratio at NH low latitudes reaches $\sim 5\%$ (20%) at 20°N and $\sim 20\%$ (40%) at 10°N in GAIA (MERRA-2). These features strongly suggest that a significant part of all resolved wave forcing at NH low latitudes is due to secondary RWs in the USLM. Additionally, positive occurrence frequency anomalies of $\overline{\text{MPV}}_\phi < 0$ are seen at $(\phi, \theta) = [30^\circ\text{--}45^\circ\text{N} (20^\circ\text{--}40^\circ\text{N}), 1300\text{--}2800\text{ K} (1100\text{--}2600\text{ K})]$ for GAIA (MERRA-2). Because the EP fluxes point horizontally in these regions, the secondary RWs are likely generated by BT instability. The negative $\overline{\text{MPV}}_\phi$ is attributable to a largely elongated comma-shaped polar vortex (not shown). Therefore, the primary stationary RW, which breaks in the NH upper stratosphere, and the westward-traveling secondary RWs play complementary roles in strengthening the middle-atmosphere Hadley circulation.

4. Summary

In this study, characteristics of the zonal-mean fields for IHC using GAIA data were analyzed with a focus on cold equatorial stratosphere events. The cold equatorial stratosphere events often appear before SSW central dates. Figure 15 shows schematic illustrations of newly proposed IHC mechanisms based on the results obtained in this study. After the checked

temperature anomaly pattern forms in the NH stratosphere and mesosphere (day = +0), a warm anomaly appears at SH high latitudes in the lower thermosphere on day = +4. Next, a warm anomaly in the SH upper mesosphere appears on average +5 days later (day = +9). These two warm anomalies occur in succession as if the anomaly slowly moves downward and slightly equatorward from the lower thermosphere to the upper mesosphere.

First, the wave forcing anomaly in the SH was examined. Two significant negative EPFD anomalies in the MLT region are observed, which is consistent with the presence of two warm anomalies. The two negative EPFD anomalies are due mainly to the resolved GWs in the lower thermosphere and QTDWs in the upper mesosphere, respectively. Below these two negative anomalies, positive EPFD anomalies, which are respectively attributable to the resolved GWs and QTDWs, are observed in the mesosphere. This demonstrates the enhanced in situ generation of GWs and QTDWs during cold equatorial stratosphere events.

Second, the causes of positive anomalies of resolved GW and QTDW forcings in the SH mesosphere were investigated. The region of positive anomaly of resolved GW (QTDW) forcing accords well with that of enhanced vertical (meridional) shear of zonal winds and enhanced shear (BT/BC)

FIG. 14. Latitude–height sections of EP flux and EPFD associated with westward-traveling RWs with periods of $\tau = 1$ –5 days and zonal wavenumbers of $s = -4$ to -10 obtained using the (a) GAIA and (d) MERRA-2 datasets. Ratio of the EPFD due to westward-traveling RWs with $\tau = 1$ –5 days and $s = -4$ to -10 to the EPFD due to all resolved waves in the (b) GAIA and (e) MERRA-2 datasets. Latitude–potential temperature sections of the occurrence frequency anomaly of $\text{MPV}_\phi < 0$ for the strong cold equatorial stratosphere events in the (c) GAIA and (f) MERRA-2 datasets. These are the results averaged over days $= -8$ to 0 . Note that (b) and (e) only show the regions where the EPFD associated with all resolved waves is lower than $-1 \text{ m s}^{-1} \text{ day}^{-1}$. The contour intervals are 10% for (b) and (d) and 5% for (c) and (f). The contour intervals in (a) and (d) are not uniform.

instability. Therefore, it is thought that stronger resolved GWs and QTDWs are generated by enhanced shear and BT/BC instabilities in the mesosphere, respectively. Moreover, it is shown that the enhancements of vertical and meridional shear of zonal winds are due to the positive GW_PFD anomaly in the mesosphere, which is likely caused by negative GW_PFD modulated by the westward wind anomaly in the stratosphere. In this study, we examined the contribution of secondary GWs which are generated by shear instability in the SH mesosphere to the IHC. However, it is also important to consider the contribution of secondary GWs generated by other mechanisms such as those directly emitted from the momentum deposit by primary GWs (e.g., Becker and Vadas 2018) to the IHC. To examine the respective contribution to the IHC by GWs generated through different mechanisms in the middle atmosphere is left for the issue of future study.

Third, the causes of the cold equatorial stratosphere (Fig. 15a), which is thought to induce the westward wind anomaly in the SH USLM via thermal wind balance, were examined using the GAIA and MERRA-2 datasets. A largely negative RW forcing anomaly is observed in the NH upper stratosphere during strong cold equatorial stratosphere events. The negative RW forcing anomaly extends to NH low latitudes in the upper stratosphere. The low latitude part of the negative RW forcing anomaly is mainly caused by westward-traveling RWs generated in association with the primary stationary RWs breaking in the NH upper stratosphere. The middle-atmosphere Hadley circulation can be efficiently strengthened by this negative wave forcing anomaly at low latitudes. Therefore, it is thought that strong cold anomaly near the

equator and SH low latitudes in the upper stratosphere is generated by the upward flow anomaly of the middle-atmosphere Hadley circulation.

Finally, differences in the IHC scenario from the previous study are discussed. First, the mechanism proposed by Smith et al. (2020) cannot explain the intraseasonal variation in the time lag of the temperature anomaly response in the SH MLT region to SSWs which is evident in observations Karlsson et al. (2009b). The present study showed that the temperature anomaly in the SH MLT region is formed by the forcings of two different kinds of waves, that is GWs and QTDWs. The QTDWs have characteristic seasonal variation in which the amplitudes are large during the period from the solstice to the end of January (see SYM18). It is likely that such seasonal dependence causes intraseasonal variation in the lag of the temperature anomaly in the SH MLT region. In addition, Smith et al. (2020) indicated low correlation between wave forcing anomalies in the SH MLT region and those at high latitudes in the NH stratosphere. However, this low correlation may be explained by the following: 1) The wave forcing at low latitudes, which they did not evaluate, in addition to that at high latitudes in the NH stratosphere is important for the appearance of the cold equatorial stratosphere events (section 3e). 2) The difference between the central date of SSWs and that of the cold equatorial stratosphere events largely depends on the case in a range of 0 to +12 days (section 3a). In other words, the date of the strong wave forcing at NH low latitudes and that at NH high latitudes are different, depending on the case. 3) The GWs and QTDWs are generated by instabilities originating from the strong wave forcing at NH low latitudes. The time required for the GWs and QTDWs to

FIG. 16. Comparison of IHC mechanisms causing the GW forcing in the SH MLT region between (right) this study (red) and (left) KB10 (blue). The horizontal axes denote the zonal-mean zonal wind or GW zonal phase velocity. The solid curves and solid wavy lines denote the zonal-mean zonal wind and the upward-propagating GWs, respectively. The event climatology, the SSW case according to KB10, and the cold equatorial stratosphere events in the present study are represented in black, blue, and red, respectively. δ represents the anomaly from the event climatology.

the SH middle atmosphere as shown by the present study, there are at least the above-mentioned three reasons why Smith et al. (2020) did not detect a high correlation between the wave forcing in the SH MLT region and that at high latitudes in the NH stratosphere.

In KB10, the axisymmetric model was used (i.e., no resolved wave forcing), and the effect of GW forcing expressed by a parameterization on the IHC was examined. In contrast, in our study, it is revealed that not only GWs generated in the troposphere but also GWs and QTDWs generated in the SH mesosphere play a crucial role in the IHC. These GWs and QTDWs are generated in the SH mesosphere by the shear and BT/BC instabilities which are caused by the enhanced middle-atmosphere Hadley circulation. This feature shows that the temperature anomaly in the SH MLT is caused by the response of temperature anomaly in the NH stratosphere through the equatorial stratosphere, rather than through the equatorial mesosphere showed by KB10. Figure 16 shows a comparison of IHC mechanisms causing GW forcing in the MLT region between the present study and KB10. In the KB10 scenario, the GW forcing anomaly is negative in the upper mesosphere because the eastward wind anomaly in the mesosphere modulates the propagation of GWs originating in the troposphere. Conversely, the present study shows that a westward wind anomaly is observed in the USLM. GW_{pFD} can be modulated by this westward wind anomaly. In the mesosphere, a positive GW_{pFD} anomaly

appears and leads to strong vertical shear of zonal winds. Therefore, the shear instability is enhanced in the mesosphere, which leads to additional enhanced secondary GW generation. It is possible that the secondary GWs cause a negative wave forcing anomaly in the lower thermosphere. The mechanisms proposed in this study include the in situ generation of GWs and QTDWs in the SH mesosphere, as essential players in IHC.

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Data availability statement. The GAIA data are available from NICT Science Cloud System, Japan (https://gaia-web.nict.go.jp/index_e.html). The MERRA-2 data are available from Goddard Earth Sciences Data and Information Services Center.

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